

Origin of four- and five-phase ultramafic xenoliths from Sonora, Mexico

JAMES T. GUTMANN

Department of Earth and Environmental Sciences, Wesleyan University, Middletown, Connecticut 06457, U.S.A.

ABSTRACT

Ultramafic nodules occur in hawaiite tuff breccias of the Pinacate volcanic field, Sonora, Mexico. These include igneous xenoliths yielding high pyroxene solvus temperatures and metamorphic and allotriomorphic granular rocks yielding lower apparent temperatures. In spinel lherzolite facies assemblages, temperature estimates based on two corrected pyroxene solvus methods agree well with one another and indicate temperatures of about 930–1030°C. These estimates are 6–44° higher than most of those based on Al content of orthopyroxene coexisting with spinel and olivine, possibly owing to bulk compositional differences between these rocks and the systems upon which the Al-in-Opx thermometers are based. Temperatures calculated from uncorrected compositions of Pinacate aluminous diopsides correlate strongly with tetrahedral Al contents and can be seriously in error.

Pinacate magmas may have passed upward through many kilometers of plagioclase-spinel-bearing ultramafic rocks before reaching the very thin section of continental crust in this region of high heat flow. Calculations based on the reaction of $Ol + Pl = Opx + Cpx + Spl$ indicate that several nodules containing these five phases last equilibrated at about 9 kbar some 8–12 km or more below the base of the crust. This is near the upper limit of pressure for stability of this five-phase assemblage in material of pyrolite composition. The corrected pyroxene solvus thermometers indicate that xenoliths with spinel lherzolite facies assemblages record higher temperatures than do the plagioclase-bearing, nonigneous nodules.

The igneous xenoliths are inhomogeneous, but the relatively high temperatures (about 1100–1200°C) of these samples are real. These rocks may have been heated prior to eruption; alternatively, they were still-hot igneous masses when entrained in Pinacate magmas although their pyroxenes differ from those typical of erupted Pinacate lavas, in which orthopyroxene is unstable at low pressure.

INTRODUCTION

The Pinacate volcanic field, northwestern Sonora, Mexico, lies in Basin and Range terrane immediately east of the northern end of the Gulf of California. Its Pleistocene (Lynch, 1981) lava flows and numerous cinder cones are dominantly alkali basalts and hawaiites (Donnelly, 1974), many of which contain megacrysts of augite and labradorite together with smaller, rapidly grown phenocrysts (Gutmann, 1977).

Continental crust beneath the Pinacate evidently is only 20 km or less in thickness (Smith, 1978, Fig. 6-2). Heat flow in the region is high. The nearest published heat-flow measurements are from Yuma to the northwest (1.92 HFU, Sass et al., 1971) and Ajo to the northeast (2.4 HFU, Roy et al., 1968). Lachenbruch and Sass (1978) presented a set of geotherms for a typical Basin and Range surface heat flow of 2.1 HFU (1.6 reduced HFU); depending on the thermal model chosen for heat transfer through the crust, the temperature at a depth of 20 km (base of the crust) will be in the range 530–610°C, and the 800°C isotherm

will be intersected before a depth of 34 km is reached ($P = 10$ kbar assuming crustal thickness of 20 km).

The phase diagram for pyrolite presented by Saxena and Eriksson (1983) indicates that, at temperatures of 800°C and higher, the stable mineral assemblage at pressures less than 9.7 kbar will be the five-phase assemblage $Ol + Opx + Cpx + Pl + Spl$, with plagioclase disappearing at higher pressure or lower temperature. Accordingly, owing to the exceptionally thin crust and probable steep geotherms in this area, Pinacate magmas rising through the upper mantle may reasonably be expected to pass through ultramafic rocks first with spinel lherzolite facies assemblages ($Ol + Opx + Cpx + Spl$) and then with plagioclase-spinel (five-phase) assemblages during their ascent. Rocks with bulk compositions other than that of the pyrolite investigated by Saxena and Eriksson (1983) can contain plagioclase at greater depth than indicated above, especially if enriched in the constituents of albite (Emslie and Lindsley, 1969).

Gabbroic nodules occur at many of the cinder cones

Table 1. Pinacate xenoliths: modes and microprobe analyses of clinopyroxene

	P-29	P-28	P-33	P-125A	P-125C	P-5	80-11	80-9	P-125B	P-31	P-61	P-25	P-100F	P-32
Modes														
Name	ol web	web	web	web	ol cpx	lherz	ol web	wehrl	ol cpx	wehrl	lherz	ol web	lherz	wehrl
Texture	porphc	porphc	eq mo	al gr	al gr	eq mo	porphc	eq mo	porphc	eq mo	ig	ig	ig	ig
Cpx%	68	81	14	70	65	10	23	8	62	43	50	58	26	25
Opx%	13	19	86	26	4	7	40	2	5	3	7	18	6	2
Ol%	15			<1	27	79	36	76	31	53	39	22	65	66
Spl%	2			1	2	4	1	14	2	1				6
Pl%	2	<1	<1	2	2						4	2	3	1
Clinopyroxene														
SiO ₂	49.47	50.51	51.00	49.98	49.37	51.18	51.53	51.88	51.71	51.46	50.33	50.67	50.03	48.73
Al ₂ O ₃	6.04	6.01	6.05	5.48	6.37	5.55	5.37	4.55	6.01	5.31	5.58	5.11	5.91	7.28
TiO ₂	0.89	0.79	0.96	0.71	1.31	0.95	0.65	0.68	0.59	0.57	1.15	0.96	0.92	1.85
Cr ₂ O ₃	0.98	0.44	0.62	0.49	0.39	0.84	0.74	1.22	0.59	0.76	0.63	0.63	0.83	0.67
FeO*	4.11	5.19	4.63	4.86	5.96	3.62	4.39	3.61	4.38	4.57	7.52	6.85	7.16	7.70
MnO	0.12	0.15	0.16	0.16	0.15	0.12	0.10	0.12	0.11	0.14	0.17	0.17	0.19	0.20
MgO	15.22	14.90	15.26	15.10	14.08	15.59	15.87	16.30	15.69	16.14	16.21	16.43	15.87	14.46
NiO	0.01	0.03	0.07	0.05	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0.09	n.d.	n.d.	n.d.
CaO	21.69	21.77	21.93	21.92	21.38	21.69	21.27	20.83	21.07	20.80	18.52	18.63	18.36	19.17
Na ₂ O	0.90	0.75	0.75	0.75	0.81	1.05	0.82	1.10	0.64	0.65	0.72	0.56	0.62	0.93
Total	99.43	100.54	101.43	99.50	99.82	100.59	100.74	100.29	100.79	100.40	100.92	100.01	99.89	100.99
Cations on the basis of six oxygens														
Si	1.816	1.840	1.839	1.838	1.819	1.852	1.864	1.880	1.869	1.868	1.831	1.856	1.838	1.780
Al(IV)	0.184	0.160	0.161	0.162	0.181	0.148	0.136	0.120	0.131	0.132	0.169	0.144	0.162	0.220
Al(VI)	0.078	0.098	0.096	0.076	0.096	0.089	0.093	0.074	0.125	0.095	0.070	0.077	0.094	0.094
Ti	0.025	0.022	0.026	0.020	0.036	0.026	0.018	0.019	0.016	0.016	0.031	0.026	0.025	0.051
Cr	0.028	0.013	0.018	0.014	0.011	0.024	0.021	0.035	0.017	0.022	0.018	0.018	0.024	0.019
Fe ²⁺	0.057	0.114	0.104	0.085	0.140	0.067	0.100	0.071	0.131	0.117	0.178	0.183	0.192	0.182
Fe ³⁺ †	0.069	0.044	0.036	0.064	0.044	0.043	0.033	0.039	0.001	0.022	0.051	0.027	0.028	0.053
Mn	0.004	0.005	0.005	0.005	0.005	0.004	0.003	0.004	0.003	0.004	0.005	0.005	0.006	0.006
Mg	0.833	0.809	0.820	0.828	0.773	0.841	0.856	0.880	0.845	0.873	0.879	0.897	0.869	0.787
Ni	0.000	0.001	0.002	0.001						0.003				
Ca	0.853	0.850	0.847	0.864	0.844	0.841	0.824	0.809	0.816	0.809	0.722	0.731	0.723	0.750
Na	0.064	0.053	0.052	0.053	0.058	0.074	0.058	0.077	0.045	0.046	0.051	0.040	0.044	0.066
Total	4.011	4.009	4.006	4.010	4.007	4.009	4.006	4.008	3.999	4.004	4.008	4.004	4.005	4.008

Note: Abbreviations: ol web, olivine websterite; web, websterite; ol cpx, olivine clinopyroxenite; lherz, lherzolite; wehrl, wehrlite; porphc, porphyroclastic; eq mo, equigranular mosaic; al gr, allotropic granular; ig, igneous; n.d., not determined.

* Total Fe as FeO.

† Fe³⁺ by charge balance.

and maar craters in the Pinacate field, and ultramafic nodules are present at a few of these. Tuff breccias surrounding maars in the northwestern part of the volcanic field contain rare ultramafic nodules with metamorphic textures. Both spinel lherzolite facies assemblages and five-phase assemblages are present, as are Ol + Opx + Cpx + Pl ultramafic rocks with igneous textures. Most abundant among the ultramafic nodules are clinopyroxenites, although websterite and wehrlite also are well represented and lherzolite and dunite occur sparingly. Harzburgite evidently is absent. In contrast to some nodule localities in Arizona, only one among more than 100 ultramafic nodules examined in hand specimen was composite (wehrlite with a narrow selvage of olivine clinopyroxenite). This nodule is inhomogeneous in phase composition and will not be described further here. Although kaersutite megacrysts and rare kaersutite gabbro nodules occur in the Pinacate, hydrous phases are absent from all ultramafic nodules examined.

The magmas that contained the nodules described here

are hawaiites with Mg values [= 100 Mg/(Mg + Fe²⁺)] of 44 and 51 (Gutmann, in prep.).

PROCEDURE AND DESCRIPTION OF XENOLITHS

Mineral analyses were carried out on an automated MAC electron probe with Krisel Control operating system at the Geophysical Laboratory, Carnegie Institution. Accelerating potential was 15 kV, beam current 0.10 μ A, and beam diameter 1–2 μ m (10 μ m for plagioclase). Oxide values were calculated using alpha corrections following Bence and Albee (1968) and Albee and Ray (1970). Three to eight grains of each mineral phase in each rock were analyzed (four grains in most instances). The outer 1 cm of nodules was avoided in the probe work when possible, and porphyroclasts were avoided in nearly all instances. The analyzed spots are within *cores* of crystals: as emphasized by Ozawa (1983), it is difficult to obtain true rim compositions, and those of cores are more meaningful for purposes of comparison with composition of coexisting phases. No systematic study of rim compositions was

Table 2. Pinacate xenoliths: microprobe analyses of orthopyroxene and olivine

	P-29	P-28	P-33	P-125A	P-125C	P-5	80-11	80-9	P-125B	P-31	P-61	P-25	P-100F	P-32
Orthopyroxene														
SiO ₂	54.06	54.23	54.22	52.94	53.07	55.01	55.12	55.74	54.64	54.81	53.59	53.93	54.42	52.46
Al ₂ O ₃	3.63	4.33	4.17	4.26	4.11	3.46	3.59	2.57	4.35	3.92	3.60	3.56	3.17	4.70
TiO ₂	0.18	0.21	0.23	0.17	0.29	0.25	0.16	0.19	0.16	0.18	0.47	0.43	0.25	0.63
Cr ₂ O ₃	0.34	0.24	0.32	0.29	0.19	0.38	0.39	0.45	0.30	0.41	0.35	0.39	0.55	0.32
FeO*	9.29	11.38	10.46	11.64	12.91	8.29	9.43	8.05	9.27	9.35	12.71	11.82	11.50	14.09
MnO	0.22	0.27	0.19	0.29	0.26	0.19	0.20	0.18	0.20	0.20	0.23	0.22	0.22	0.28
MgO	31.91	28.96	30.59	29.73	28.42	32.41	31.86	33.00	30.86	31.15	28.65	28.63	28.39	27.18
NiO	0.05	0.05	0.07	0.05	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0.04	n.d.	n.d.	n.d.
CaO	0.62	0.81	0.77	0.84	0.71	0.79	0.88	0.95	1.06	1.16	1.80	1.89	2.14	1.48
Na ₂ O	0.03	0.03	0.04	0.04	0.03	0.06	0.05	0.07	0.04	0.05	0.11	0.08	0.09	0.10
Total	100.33	100.51	101.06	100.25	99.99	100.84	101.68	101.20	100.88	101.23	101.55	100.95	100.73	101.24
Cations on the basis of six oxygens														
Si	1.884	1.908	1.888	1.868	1.889	1.900	1.897	1.917	1.897	1.897	1.881	1.899	1.920	1.859
Al(IV)	0.116	0.092	0.112	0.132	0.111	0.100	0.103	0.083	0.103	0.103	0.119	0.101	0.080	0.141
Al(VI)	0.033	0.087	0.059	0.045	0.062	0.041	0.043	0.021	0.075	0.057	0.030	0.047	0.052	0.056
Ti	0.005	0.006	0.006	0.005	0.008	0.006	0.004	0.005	0.004	0.005	0.012	0.011	0.007	0.017
Cr	0.009	0.007	0.009	0.008	0.005	0.010	0.011	0.012	0.008	0.011	0.010	0.011	0.015	0.009
Fe ²⁺	0.221	0.335	0.278	0.289	0.361	0.210	0.238	0.198	0.259	0.250	0.327	0.329	0.336	0.380
Fe ^{3+***}	0.050	0.000	0.027	0.055	0.023	0.030	0.033	0.034	0.010	0.021	0.046	0.019	0.004	0.037
Mn	0.006	0.008	0.006	0.009	0.008	0.006	0.006	0.005	0.006	0.006	0.007	0.007	0.007	0.008
Mg	1.657	1.518	1.587	1.563	1.508	1.669	1.635	1.691	1.597	1.607	1.499	1.503	1.493	1.436
Ni	0.001	0.001	0.002	0.001							0.001			
Ca	0.023	0.031	0.029	0.032	0.027	0.029	0.032	0.035	0.039	0.043	0.068	0.071	0.081	0.056
Na	0.002	0.002	0.003	0.003	0.002	0.004	0.003	0.005	0.003	0.003	0.007	0.005	0.006	0.007
Total	4.007	3.995	4.006	4.010	4.004	4.005	4.005	4.006	4.001	4.003	4.007	4.003	4.001	4.006
Olivine														
SiO ₂	39.11			38.84	38.27	39.67	39.44	40.13	39.50	39.58	39.45	38.45	38.23	38.20
TiO ₂	0.02			0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.02	0.02	0.02	0.03
Al ₂ O ₃	0.06			0.02	0.02	0.00	0.02	0.02	0.04	0.04	0.04	0.04	0.05	0.05
Cr ₂ O ₃	0.02			0.01	0.02	0.01	0.03	0.02	0.00	0.02	0.03	0.01	0.03	0.02
FeO*	14.89			17.39	20.52	13.36	15.11	12.81	14.91	15.14	20.38	19.41	19.29	22.72
MnO	0.21			0.23	0.29	0.20	0.20	0.21	0.21	0.21	0.24	0.24	0.24	0.32
MgO	46.32			42.89	41.48	47.25	46.44	47.54	45.91	45.71	40.26	42.41	40.37	40.03
NiO	0.20			0.25	0.14	0.32	0.20	0.23	0.17	0.20	0.25	0.23	0.21	0.20
CaO	0.07			0.11	0.07	0.09	0.09	0.09	0.12	0.11	0.19	0.17	0.18	0.15
Total	100.90			99.74	100.81	100.90	101.53	101.05	100.86	101.01	100.86	100.98	98.62	101.72
Cations on the basis of four oxygens														
Si	0.976			0.991	0.981	0.983	0.979	0.988	0.985	0.986	1.007	0.979	0.996	0.981
Ti	0.000			0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.001
Al	0.001			0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.001	0.000	0.001	0.002
Cr	0.000			0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.001	0.000	0.000	0.000
Fe	0.311			0.371	0.439	0.276	0.313	0.264	0.310	0.315	0.435	0.413	0.420	0.488
Mn	0.003			0.004	0.006	0.003	0.003	0.004	0.004	0.004	0.005	0.004	0.004	0.007
Mg	1.723			1.632	1.586	1.745	1.717	1.745	1.707	1.697	1.532	1.610	1.569	1.532
Ni	0.003			0.004	0.002	0.005	0.003	0.004	0.002	0.003	0.005	0.004	0.003	0.004
Ca	0.001			0.002	0.001	0.002	0.002	0.002	0.002	0.002	0.005	0.004	0.004	0.004
Total	3.108			3.004	3.015	3.014	3.017	3.007	3.010	3.007	2.991	3.014	2.997	3.019

Note: n.d.: not determined.

* Total Fe as FeO.

** Fe³⁺ by charge balance.

made, and zonation reflecting changes late in the history of these rocks may be present. Indeed, this was observed in the igneous-textured samples. However, homogeneity of crystal cores in the nonigneous rocks (see below) speaks close approach to equilibrium at some time in their history. The analyses are presented in Tables 1–3 together with approximate modal compositions and textures of the samples.

Following the textural classification of Pike and Schwarzman (1977), the 14 xenoliths described in this report include 4 with porphyroclastic textures (although

their matrix grains show slight strain), 4 with equigranular mosaic textures, 4 with igneous textures and 2 with allotriomorphic granular textures of uncertain origin. Some of the samples have traces of pyrometamorphic texture represented by tachylitic or tan glass along a few grain boundaries and indicative of incipient melting, perhaps during transport in the magmas. The two allotriomorphic granular rocks are medium and fine grained, contain minor tachylitic(?) material along grain boundaries, and show traces of zoning at the margins of a few clinopyroxene or plagioclase grains although probe data indicate that cli-

Table 3. Pinacate xenoliths: microprobe analyses of spinel and plagioclase

	P-29	P-28	P-33	P-125A	P-125C	P-5	80-11	80-9	P-125B	P-31	P-61	P-25	P-100F	P-32
Spinel														
SiO ₂	0.07			0.10	0.09	0.08	0.12	0.09	0.09	0.11				0.12
TiO ₂	0.33			0.22	0.64	0.44	0.32	0.71	0.21	0.39				4.07
Al ₂ O ₃	50.79			55.82	51.79	47.02	49.63	35.49	55.84	47.58				37.35
Cr ₂ O ₃	13.04			7.64	7.72	18.32	15.11	30.27	10.10	17.51				12.71
FeO*	17.38			16.98	23.43	16.25	18.48	18.61	15.50	18.13				33.79
MnO	0.15			0.14	0.18	0.20	0.16	0.22	0.15	0.17				0.25
MgO	18.48			18.57	16.25	17.95	17.41	15.77	19.00	17.13				12.54
NiO	n.d.			n.d.	0.25	0.28	0.21	0.14	0.19	0.17				0.24
ZnO	0.16			0.24	n.d.	0.13	n.d.	0.08	n.d.	n.d.				0.16
Total	100.40			99.71	100.35	100.67	101.44	101.38	101.08	101.19				101.23
Cations on the basis of three metals and four oxygens														
Si	0.002			0.003	0.002	0.002	0.003	0.003	0.002	0.003				0.003
Ti	0.007			0.004	0.013	0.009	0.006	0.015	0.004	0.008				0.088
Al	1.592			1.727	1.637	1.496	1.560	1.186	1.708	1.512				1.261
Cr	0.274			0.159	0.164	0.391	0.319	0.679	0.207	0.373				0.288
Fe ²⁺	0.269			0.273	0.356	0.276	0.310	0.341	0.264	0.315				0.541
Fe ³⁺	0.117			0.100	0.169	0.091	0.103	0.100	0.072	0.094				0.269
Mn	0.003			0.003	0.004	0.005	0.004	0.005	0.003	0.004				0.006
Mg	0.733			0.727	0.649	0.722	0.692	0.666	0.735	0.688				0.535
Ni					0.005	0.006	0.005	0.003	0.004	0.004				0.006
Zn	0.003			0.005		0.003		0.002						0.003
Plagioclase														
SiO ₂	54.91	54.43	53.49	51.19	53.39						54.71	53.91	53.33	54.59
Al ₂ O ₃	27.45	28.85	29.01	29.41	28.46						28.35	28.20	29.39	28.62
TiO ₂	0.03	0.03	0.02	0.01	0.03						0.07	0.07	0.07	0.09
FeO*	0.11	0.29	0.22	0.18	0.18						0.32	0.31	0.30	0.30
MgO	0.05	0.07	0.07	0.07	0.05						0.10	0.10	0.11	0.07
CaO	11.30	11.68	12.09	13.13	11.99						11.33	11.65	12.56	11.08
Na ₂ O	4.78	4.51	4.61	4.00	4.41						5.11	4.77	4.12	5.04
K ₂ O	0.57	0.68	0.23	0.23	0.22						0.38	0.29	0.20	0.45
Total	99.20	100.54	99.74	98.22	98.73						100.41**	99.30	100.08	100.24
Cations on the basis of eight oxygens														
Si	2.500	2.451	2.428	2.369	2.446						2.465	2.457	2.414	2.462
Al	1.473	1.531	1.552	1.604	1.536						1.504	1.514	1.567	1.521
Ti	0.000	0.000	0.000	0.000	0.000						0.001	0.001	0.001	0.002
Fe	0.004	0.010	0.008	0.006	0.006						0.012	0.010	0.010	0.010
Mg	0.002	0.004	0.004	0.004	0.002						0.006	0.006	0.006	0.004
Ca	0.550	0.563	0.587	0.650	0.587						0.546	0.568	0.608	0.535
Na	0.421	0.392	0.405	0.358	0.391						0.445	0.420	0.361	0.440
K	0.032	0.038	0.013	0.012	0.012						0.021	0.016	0.010	0.025
Total	4.982	4.989	4.997	5.003	4.980						5.001**	4.992	4.977	4.999

Note: n.d.: not determined.

* Total Fe as FeO.

** Totals include 0.04 wt% MnO, 0.001 Mn.

nopyroxene grain interiors are homogeneous in major-element composition through each rock.

A quantitative test of homogeneity is afforded by the homogeneity index of Boyd and Finger (1975), which relates compositional variation to that which would be expected from counting statistics in a perfectly homogeneous sample population; homogeneity indices greater than three for major elements indicate real inhomogeneity. The probe data show that the four igneous rocks are somewhat inhomogeneous: homogeneity indices are greater than three for one or more major elements in at least one mineral phase in all of them. In contrast, analyses of grain interiors in the metamorphic and allotropic granular rocks indicate that all are apparently homogeneous or nearly so except for plagioclase and spinel in P-125A and slight

inhomogeneity (for one element only) in olivine of P-29 and plagioclase of P-28, P-29, and P-33.

Clinopyroxenes of the four igneous rocks are Fe-poor augites. Based on the composition of their olivines (Fo₇₆₋₈₀) and Ti-rich clinopyroxenes (0.92–1.85 wt% TiO₂), the igneous rocks belong to the Al-augite group of Wilshire and Shervais (1975) and group II of Frey and Prinz (1978), although Cr₂O₃ content of the Pinacate clinopyroxenes is high (0.63–0.83 wt%).

The ten rocks with metamorphic and allotropic granular textures (hereafter referred to as "nonigneous" although the two with allotropic granular textures could be igneous) range widely in modal composition. Their clinopyroxenes all are diopsides except for metamorphic sample P-31, which carries endiopsidite. The five

Table 4. Pinacate xenoliths: estimates of temperature and pressure

Sample no.	Wo in Opx T (°C)	Opx in Cpx, corrected T (°C)	Opx in Cpx, uncor- rected T (°C)	Al in Opx T (°C)	P at T (kbar at °C)
Plagioclase-bearing, nonigneous xenoliths					
P-29	870	920	517	924	9.0 at 870°C
P-28		933	638		
P-33		944	655		
P-125A		916	559	974	9.4 at 916°C
P-125C	860	917	611	1020	8.3 at 860°C
Spinel lherzolite facies xenoliths					
P-5	930	927	642	918	
80-11	950	988	826	944	
80-9	980	991	885	905	
P-125B		1014	878	988	
P-31	1020	1033	941	995	
Igneous xenoliths					
P-61		1120	1126		
P-25	1180	1126	1131		
P-100F	> 1200	1116	1122		
P-32		1049	986	1264	8.6 at 1049°C

Note: Wo in Opx by method of Lindsley (1983); Opx in Cpx, corrected, by method of Wells (1977); Opx in Cpx, uncorrected, from solvus equation of Kretz (1982) as described in text; Al in Opx by method of Sachtleben and Seck (1981); P from five-phase reaction evaluated following Powell (1978). See text for details.

rocks with spinel lherzolite facies assemblages, and five-phase metamorphic sample P-29, might be assigned to the Cr-diopside group of Wilshire and Shervais or group I of Frey and Prinz; however, their diopsides are relatively titaniferous (0.57–0.95 wt% TiO₂) and their ferromagnesian phases are somewhat less magnesian (e.g., Fo_{84–87}) than many group I rocks. These six samples may represent group I rocks that were metasomatized at some time in their history.

The remaining four nonigneous rocks, all of which are plagioclase-bearing pyroxenites, differ in several ways from the six rocks described above. The group I-related samples described above tend to be rich in olivine (average 48%) whereas the four remaining pyroxenites contain little or no olivine except for one sample with 27%. Ferromagnesian phases in the group I-related samples are more magnesian than in these pyroxenites (Fo_{84–87} vs. Fo_{78–82}; En_{86–88} vs. En_{80–84}; 100(Mg/Mg + Fe_{tot}) of diopside = 86–89 vs. 81–85). Orthopyroxene, clinopyroxene, and spinel of these four pyroxenites tend to be poorer in Cr than their counterparts in the group I-related samples, and their pyroxenes even richer in Ti (0.71–1.31 wt% TiO₂). These four plagioclase-bearing rocks might best be assigned to group II on the basis of their relatively Fe- and Ti-rich compositions although they contain moderately chromian diopsides (0.39–0.62 wt% Cr₂O₃).

Thus, ultramafic rocks of the Pinacate volcanic field investigated to date range from metamorphic rocks very similar to (e.g., samples P-5, 80-9), but generally more Fe- and Ti-rich than, those of the Cr-diopside group

through metamorphic and allotriomorphic granular rocks that are even more rich in Fe and Ti. The latter are plagioclase-bearing and pyroxene-rich and may represent metagneous or greatly metasomatized rocks. In addition, the suite includes igneous rocks of the Al-augite group bearing relatively chromian augites.

GEOOTHERMOMETRY

Geothermometry of spinel-bearing ultramafic xenoliths is fraught with pitfalls. Apart from the problem of internal disequilibrium, which seems usually to be present at some scale, available geothermometers face substantial difficulties both analytical (e.g., estimation of Fe³⁺ from probe data) and theoretical (e.g., nonideal solution behavior). Even in rocks where grain interiors are reasonably homogeneous in composition across a thin section, and using geothermometers that minimize the sorts of difficulties noted above or seem to deal with them successfully, the meaning of the resulting apparent temperatures remains unclear. Equilibration of the phases involved may have been blocked kinetically at some stage and, in any event, geotherms must have been perturbed at least locally by the very processes of alkali basaltic volcanism that brought the nodules to the surface. Nevertheless, consistency of phase composition and results of different thermometers, if obtained, is a regularity worthy of note and affords information on relative temperatures of last equilibration of grain interiors within a suite of nodules and, within broad limits, on absolute temperatures.

Pyroxene solvus geothermometry is among the best calibrated of currently available, applicable methods. Clinopyroxenes of Pinacate xenoliths contain at least 10% of nonquadrilateral components and so cannot properly be treated by the method of Lindsley (1983). However, many of the orthopyroxenes have suitable compositions, and temperatures were estimated from those orthopyroxenes wherein Wo + En + Fs constitute at least 90.0% of the orthopyroxene components as defined by Lindsley (1983). Following this method, cations were calculated on the basis of six oxygens, Fe³⁺ was calculated from charge-balance considerations (Papike et al., 1974), and metals were then recomputed to a basis of six oxygens. After reduction of Fe²⁺ and Mg by the amount needed for non-quadrilateral components, Wo, En, and Fs were plotted on an enlargement of the quadrilateral (Lindsley, 1983) displaying relationships at 10 kbar (choice of pressure is justified below). Temperatures were recorded to the nearest 10° although such small differences in temperature may not be significant in this graphical projection technique and, in any case, uncertainty of the method is not expected to be better than 75–90°. This uncertainty includes accuracy of ±50° combined with additional uncertainty of 25–40° owing to amounts of nonquadrilateral components in these pyroxenes (Lindsley and Andersen, 1983). The results are presented in Table 4.

Two of the three nonigneous rocks containing the five-phase assemblage Ol + Pl + Cpx + Opx + Spl can be treated by this method and yield temperatures of 860 and

870°C. Four of the five spinel lherzolite facies assemblages can be so treated and give temperatures in the range 930–1020°C. Orthopyroxenes of only two of the four igneous rocks are low enough in nonquadrilateral components, and these indicate temperatures of 1180 and >1200°C.

The pyroxene solvus method of Wells (1977) is among the most widely employed geothermometric methods for basaltic xenoliths. This method reproduces relationships in the CMAS system at 10–15 kbar well (Sen, 1985) although its application to natural samples can present problems (e.g., Harley, 1984). Temperatures were estimated for the Pinacate xenoliths by this method using cation contents calculated on the basis of six oxygens and charge-balance considerations as described above. The results are presented in Table 4.

Temperatures estimated by Wells's method range from 916 to 944°C in the plagioclase-bearing, nonigneous rocks and are 50° and 57° higher than those indicated by the method of Lindsley in two samples. In the spinel lherzolite facies xenoliths, the temperatures determined by the method of Wells range from 927 to 1033°C and are 3° lower and 11°, 13°, and 38° higher than those indicated by the method of Lindsley. Orthopyroxene of sample P-125B contains a few tenths of 1% too many nonquadrilateral components for proper application of the Ca-in-Opx method but, if this stricture is disregarded for this sample, it indicates a temperature only 14° lower than the method of Wells. In the four igneous rocks, the temperatures from the method of Wells are 1049–1126°C and are 54° and at least 84° lower than those indicated by the method of Lindsley. These results support the idea (Lindsley, 1983) that assumptions involved in Wells's method lead to overestimates of temperature for metamorphic pyroxenes. They also support Harley's (1984) suggestion that Wells's thermometer may *underestimate* the temperature of higher temperature (>1100°C) xenoliths. In the temperature range 930–1030°C, however, the very close agreement between these methods is encouraging. This agreement may also be somewhat fortuitous in that the small effect of pressure on the solvus cannot be treated by Wells's method, which rests on experimental data obtained at various pressures both above and below 10 kbar.

Application of another pyroxene solvus formulation leads to erroneously low temperatures but underscores the great importance of correcting for nonquadrilateral components in highly aluminous clinopyroxenes. The equation of Kretz (1982) for the diopside limb of the solvus takes account of Fe content explicitly and of Al content implicitly by virtue of its calibration against natural, mildly aluminous clinopyroxenes. Lindsley (1983) found that this method is successful over a wide range of temperatures. However, the clinopyroxenes in Pinacate xenoliths contain 4.55–7.28 wt% Al₂O₃, substantially more than those by which the thermometer of Kretz (1982) was calibrated (1.54–3.22 wt% Al₂O₃). Apparent temperatures were calculated for the Pinacate xenoliths from the pyroxene solvus equation of Kretz (1982) using cation contents calculated as described above for the method of

Lindsley (1983). The resulting apparent temperatures (Table 4) range from 517 to 941°C in the ten nonigneous rocks and from 986 to 1131°C in the four igneous rocks. However, these temperatures in the nonigneous rocks correlate negatively and strongly ($r = -0.91$) with Al(IV) in their diopsides.

The surface of the clinopyroxene side of the solvus generally decreases in temperature with decrease in relative abundance of ferrosilite. It was suspected that the wide temperature range (424°) calculated for the ten nonigneous rocks using the equation of Kretz partly reflected erroneous calculated values of Fe³⁺ due to imprecision in silica determination and that the correlation with Al(IV) is an artifact of that imprecision. Accordingly, pyroxene cation contents were recalculated assuming $Fe^{3+} = 0.3 Fe_{tot}$, the average abundance of Fe³⁺ among the eight diopsides, Cr-rich diopsides, and endiopsides from ultramafic xenoliths tabulated by Deer et al. (1978). The temperatures were then recalculated. The resulting temperature range among the nonigneous rocks is reduced somewhat (to 334°), but the strong negative correlation with Al(IV) is unchanged ($r = -0.91$). Even if these pyroxenes are recalculated assuming that *all* Fe is Fe²⁺, although the resulting apparent temperatures are higher (>717°C) and the temperature range is further reduced (to 253°), the strong negative correlation with Al(IV) remains ($r = -0.87$). Of course, in the lherzolite facies of the CMAS system, Al₂O₃ content of diopside *increases* with temperature rather than the reverse (e.g., Herzberg, 1978, Fig. 6).

Mysen (1976) plotted the diopside limb of the solvus for three natural lherzolites, correcting for nonquadrilateral components by subtracting Al(IV) from Ca prior to calculating Ca/(Ca + Mg). In the correction method of Lindsley (1983), the three nonquadrilateral components subtracted from "Ca" of Ca-rich pyroxene each contain one Al(IV) per Ca. It seems clear from the correlations with Al(IV) noted in the preceding paragraph that temperatures inferred from uncorrected compositions of aluminous diopsides in the present study are *very* greatly depressed owing to nonquadrilateral components. Whereas the four igneous rocks truly record much higher temperatures than the ten nonigneous rocks, the latter apparently represent conditions within a thermal range of about 117° (as judged from Wells's method) to 160° (as inferred by the method of Lindsley). The much wider apparent range indicated by uncorrected clinopyroxene compositions is largely spurious. The elegant and otherwise successful solvus formulation of Kretz (1982) evidently should be employed only with caution for estimating temperature in such highly aluminous clinopyroxenes as these.

Results of this study suggest influence of Ti-Al(IV) coupled substitution on the clinopyroxene limb of the solvus. Ti in diopsides of the ten nonigneous rocks correlates negatively ($r = -0.75$) with apparent temperature calculated from the solvus equation of Kretz (1982) assuming $Fe^{3+} = 0.3 Fe_{tot}$. As would be expected, Ti in these pyroxenes correlates positively ($r = 0.76$) with Al(IV) and

presumably is charge-compensated by 2Al(IV) , e.g., as $\text{CaTiAl}_2\text{O}_6$. The association of Ti with lowered orthopyroxene solubility is supported by positive correlation ($r = 0.67$) between Ti in diopside and the *difference* in temperature between that apparent from the solvus equation of Kretz (1982) and the (higher) temperature inferred by the Ca-in-Opx method of Lindsley (1983).

Other types of geothermometers likely to be successful in these rocks are few. Those based on Mg-Fe²⁺ partitioning depend critically on determination of Fe³⁺, which can only be estimated approximately from electron-probe data on pyroxenes. However, Fujii (1976) presented an equation relating temperature to the Al content of enstatite coexisting with spinel and forsterite, and Sachtleben and Seck (1981) calibrated this relationship empirically using natural spinel peridotites. Apparent temperatures of the Ol + Sp + Opx-bearing Pinacate nodules calculated by the geothermometer of Sachtleben and Seck (1981) are given in Table 4.

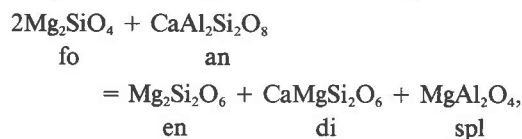
In at least three of the four samples that also contain plagioclase, these Al-in-Opx temperatures are considerably higher than those indicated by the pyroxene solvus methods, evidently reflecting the influence of plagioclase on the equilibria. In the plagioclase-free rocks, temperatures inferred following Sachtleben and Seck (1981) range from 6° to 75° lower than those judged from the Ca-in-Opx method of Lindsley (1983) and from 9° to 86° lower than those indicated by the method of Wells (1977). Excepting one sample (80-9), the Al-in-Opx temperatures for the plagioclase-free rocks are within 44° of those indicated by the pyroxene solvus methods.

Tendency of the Al-in-Opx temperatures from spinel lherzolite facies rocks to be less than the pyroxene solvus temperatures could reflect differences in bulk composition between the xenoliths studied by Sachtleben and Seck (1981) and those of the present study. Tracy (1980) also obtained low apparent temperatures for spinel peridotites when adjusted Al_2O_3 content of orthopyroxene was interpreted via the isopleths of Danckwerth and Newton (1978) and the result was compared to pyroxene solvus temperatures. The isopleths of Danckwerth and Newton are based on the (Fe-free) system $\text{MgO-Al}_2\text{O}_3\text{-SiO}_2$ (MAS), and Tracy (1980) noted Danckwerth and Newton's (1978) suggestion that the presence of Fe may result in anomalously low estimated temperatures. Minerals of the present study are more rich in Fe than those studied by Sachtleben and Seck (1981). To test this hypothesis concerning the importance of Fe, apparent temperatures were determined for the plagioclase-free rocks via the improved Al-in-Opx formulation of Gasparik and Newton (1984), which also is based on relationships in the MAS system. Pressure of 10 kbar was assumed in the calculations and the computation was iterated once (c.f. Gasparik and Newton, 1984). Again excepting sample 80-9, the resulting temperatures are even further (79°–134°) below those indicated by pyroxene solvus methods, tending to support the importance of Fe in this geothermometric approach. The caveat expressed by Gasparik and Newton (1984) that use of their method

should be limited to natural systems closely approaching the MAS system should be heeded.

GEOBAROMETRY

Reliable methods of estimating pressure in spinel lherzolite facies assemblages are not currently available. However, Powell (1978) has evaluated ΔG_1^0 for the reaction



which can be employed as a geobarometer if all five phases are present in close approach to equilibrium and if temperature is known. The formulation of activities used here differs somewhat from those used by Powell (1978) and Arculus and Wills (1980) in plutonic blocks from the Lesser Antilles. Specifically, $a_{\text{MgAl}_2\text{O}_4}$ was calculated here from the temperature-activity-composition relationship of Sack (1982) as reduced by Gasparik and Newton (1984) to consider the components MgAl_2O_4 , FeCr_2O_4 , and Fe_3O_4 . Furthermore, activities of diopside in clinopyroxene and of enstatite in orthopyroxene were formulated to take account of the aluminous nature of the pyroxenes in Pinacate rocks. Activities other than in spinel were expressed as

$$a_{\text{Mg}_2\text{Si}_2\text{O}_6} = (X_{\text{Mg}}^{\text{M1}})(X_{\text{Mg}}^{\text{M2}})(X_{\text{Si}}^{\text{T}})^2$$

$$a_{\text{CaMgSi}_2\text{O}_6} = (X_{\text{Ca}}^{\text{M2}})(X_{\text{Mg}}^{\text{M1}})(X_{\text{Si}}^{\text{T}})^2$$

$$a_{\text{Mg}_2\text{SiO}_4} = (X_{\text{Mg}}^{\text{O1}})^2$$

and

$$a_{\text{CaAl}_2\text{Si}_2\text{O}_8} = (X_{\text{Ca}}^{\text{Pl}}).$$

Expressions used for the equilibrium constant and relationship between K , pressure, and temperature are as in Powell (1978) and Arculus and Wills (1980).

Four of the xenoliths contain these five phases. Pressures calculated for their compositions do not depend strongly on assumed temperatures (about 0.1-kbar increase per 10° increase). Thus, assuming ideal behavior of the silicate solid solutions, the calculated pressures should be reasonably good estimates of the pressures indicated by this univariant reaction. The results are presented in Table 4 and imply pressures ranging from 8.3 to 9.4 kbar. This justifies choice of the 10-kbar quadrilateral in application of the method of Lindsley (1983). The inferred P - T conditions are within the stability field of the five-phase assemblage in pyrolite as calculated by Saxena and Eriksson (1983, Fig. 4). They are near the upper limit of that field, beyond which plagioclase would disappear with increase in pressure, but these rocks plainly do not have the composition of that pyrolite.

DISCUSSION

Spinel lherzolite is abundant in many suites of ultramafic nodules from nephelinitic and basanitic host lavas. Host rocks in the Pinacate are hawaiites, however; and, as in hawaiitic host lavas from Hawaii (White, 1966),

lherzolite is rare. Indeed, among Pinacate ultramafic nodules found to date, no more than 12% are peridotites (>40% olivine) and harzburgite appears to be absent.

The five Pinacate xenoliths with spinel lherzolite facies assemblages tend to be more Fe-rich and titaniferous than typical Cr-diopside or group I nodules in basaltic rocks from the southwestern United States. For example, 100Mg/(Mg + Fe_{tot}) of diopside in such rocks is about 89.8 (Wilshire and Shervais, 1975) to 89.6 (average of 16 group I clinopyroxenes from Frey and Prinz, 1978) whereas Pinacate diopsides average 87.4 and none are more magnesian than 88.9. TiO₂ in these Pinacate diopsides averages 0.69 wt% versus 0.56 and 0.64 wt% average TiO₂ content in diopsides of the two studies just cited.

Although hydrous phases are absent from the ultramafic xenoliths and composite nodules are extremely rare, the five nodules with spinel lherzolite facies assemblages nevertheless could have been metasomatically enriched in Fe and Ti at some time. Similarly, the five plagioclase-bearing nonigneous rocks, all of which are pyroxene-rich and all but one of which (P-29) are best assigned to the Al-augite group, could be either metaigneous rocks or rocks metasomatically enriched in Fe, Ti, Na, and Al from nearby intrusions (Wilshire, 1984). Their plagioclases are more sodic (An₅₅₋₆₅) than those typical of alpine-type ultramafic tectonites (An₆₅₋₉₀). Further compositional studies of these samples are in progress.

Grain interiors of the five-phase, nonigneous rocks record lower temperatures than those of the spinel lherzolite facies assemblages judging either by the method of Lindsley (1983) (860–870°C versus 930–1020°C) or by that of Wells (1977) (916–920°C versus 927–1033°C). These results permit the interpretation that, as suggested by the phase assemblages, the five-phase nonigneous rocks last recrystallized at shallower depth and lower temperature than the spinel lherzolite facies assemblages. This is not required by the phase assemblages owing to the possible effect of variation in bulk composition. However, if the higher pyroxene solvus temperatures of the plagioclase-free rocks represent either heating by nearby intrusions or blocking phenomena, it is peculiar that these effects influenced all of the plagioclase-free rocks and none of the five-phase assemblages. If the apparent temperatures and phase assemblages reflect relative depth of last grain interior equilibration and relative depth of the xenoliths at the time of entrapment in magma, it follows that ascending Pinacate hawaiites first incorporated fragments of spinel lherzolite facies rocks and then of rocks from a zone containing five-phase metaigneous, igneous, and/or metasomatized materials.

All these rocks may have undergone some thermal or chemical changes since grain interiors acquired their compositions. In the five-phase rocks, these compositions are appropriate for equilibration at depths of 28–32 km (8.3–9.4 kbar) below the volcanic field. Such depths are some 8–12 km or more below the base of the crust in this region.

The four igneous rocks yield relatively high temperatures judging by every geothermometer employed. These

rocks either were heated by Pinacate magmas or represent young igneous masses that still were hot when fragments of them were entrained. The former explanation would seem to be corroborated by presence in three of these xenoliths of blebs of tan glass (partial melt?) at grain boundaries. Such tan glass is absent from the nonigneous rocks, however, and it is not clear why samples with igneous texture should be singled out for heating and partial re-equilibration. But neither is it evident that these rocks were still-hot, cognate plutonic inclusions: all contain orthopyroxene whereas orthopyroxene is unstable in Pinacate magmas at low pressure and occurs in them only in xenoliths or as rare, corroded crystals in a few units, generally together with xenolithic debris. Furthermore, clinopyroxene phenocrysts and megacrysts typical of Pinacate magmas are much less rich in Cr than are the clinopyroxenes of these xenoliths (Gutmann, in prep.). Magma genetically related to these xenoliths may nevertheless occur at depth, where orthopyroxene could be stabilized by high pressure.

One of these igneous nodules contains the assemblage Ol + Opx + Cpx + Pl + Spl and is devoid of tan glass. The pressure calculated for this sample (8.6 kbar, Table 4) indicates a depth within the range inferred for the five-phase nonigneous rocks. This suggests that still-hot igneous masses may reside at that depth, where they would have heated the surrounding rocks. Thus, in conclusion, it seems especially appropriate to emphasize that the *P-T* conditions calculated in this study should *not* be taken to represent a Pinacate geotherm. The extent to which these samples reflect equilibration to the locally perturbed and presumably dynamic geotherm associated with this basaltic volcanic system cannot at the present be known.

ACKNOWLEDGMENTS

I thank F. R. Boyd and H. S. Yoder for their encouragement and support of this work. Some early petrographic studies were carried out by Paul B. Gold at Wesleyan. Robert J. Tracy facilitated use of the probe at Yale for some preliminary analyses. Electron-probe facilities were provided by the Geophysical Laboratory, Carnegie Institution, and C. G. Hadidiacos and D. George assisted greatly in their use. The manuscript benefitted from reviews by Joe Boyd, Don Lindsley, Doug Smith, and Joop Varekamp.

REFERENCES

- Albee, A.L., and Ray, L. (1970) Correction factors for electron probe microanalysis of silicates, oxides, carbonates, phosphates, and sulfates. *Analytical Chemistry*, 42, 1408–1414.
- Arculus, R.J., and Wills, K.J.A. (1980) The petrology of plutonic blocks and inclusions from the Lesser Antilles island arc. *Journal of Petrology*, 21, 743–799.
- Bence, A.E., and Albee, A.L. (1968) Empirical correction factors for the electron microanalysis of silicates and oxides. *Journal of Geology*, 76, 382–403.
- Boyd, F.R., and Finger, L.W. (1975) Homogeneity of minerals in mantle rocks from Lesotho. *Carnegie Institution of Washington Year Book* 74, 519–525.
- Danckwerth, P.A., and Newton, R.C. (1978) Experimental determination of the spinel peridotite to garnet peridotite reaction in the system MgO-Al₂O₃-SiO₂ in the range 900°–1100°C and

- Al₂O₃ isopleths of enstatite in the spinel field. *Contributions to Mineralogy and Petrology*, 66, 189–201.
- Deer, W.A., Howie, R.A., and Zussman, J. (1978) *Rock-forming minerals*, volume 2A, single-chain silicates. Wiley, New York.
- Donnelly, M.F. (1974) *Geology of the Sierra del Pinacate Volcanic Field*, northern Sonora, Mexico, and southern Arizona, U.S.A. Ph.D. thesis, Stanford University, Stanford, California.
- Emslie, R.F., and Lindsley, D.H. (1969) Experiments bearing on the origin of anorthositic intrusions. *Carnegie Institution of Washington Year Book* 67, 108–112.
- Frey, F.A., and Prinz, Martin. (1978) Ultramafic inclusions from San Carlos, Arizona: Petrologic and geochemical data bearing on their petrogenesis. *Earth and Planetary Science Letters*, 38, 129–176.
- Fujii, Toshitsugu. (1976) Solubility of Al₂O₃ in enstatite coexisting with forsterite and spinel. *Carnegie Institution of Washington Year Book* 75, 566–571.
- Gasparik, T., and Newton, R.C. (1984) The reversed alumina contents of orthopyroxene in equilibrium with spinel and forsterite in the system MgO-Al₂O₃-SiO₂. *Contributions to Mineralogy and Petrology*, 85, 186–196.
- Gutmann, J.T. (1977) Textures and genesis of phenocrysts and megacrysts in basaltic lavas from the Pinacate volcanic field. *American Journal of Science*, 277, 833–861.
- Harley, S.L. (1984) Comparison of the garnet-orthopyroxene geobarometer with recent experimental studies, and applications to natural assemblages. *Journal of Petrology*, 25, 697–712.
- Herzberg, C.T. (1978) Pyroxene geothermometry and geobarometry: Experimental and thermodynamic evaluation of some subsolidus phase relations involving pyroxenes in the system CaO-MgO-Al₂O₃-SiO₂. *Geochimica et Cosmochimica Acta*, 42, 945–957.
- Kretz, Ralph. (1982) Transfer and exchange equilibria in a portion of the pyroxene quadrilateral as deduced from natural and experimental data. *Geochimica et Cosmochimica Acta*, 46, 411–421.
- Lachenbruch, A.H., and Sass, J.H. (1978) Models of an extending lithosphere and heat flow in the Basin and Range province. In R.B. Smith and G.P. Eaton, Eds. *Cenozoic tectonics and regional geophysics of the Western Cordillera*: Geological Society of America Memoir 152, 209–250.
- Lindsley, D.H. (1983) Pyroxene thermometry. *American Mineralogist*, 68, 477–493.
- Lindsley, D.H., and Andersen, D.J. (1983) A two-pyroxene thermometer. *Proceedings of the Thirteenth Lunar and Planetary Science Conference*, Part 2. *Journal of Geophysical Research*, 88, Supplement, A887–A906.
- Lynch, D.J. (1981) A model for volcanism in the Pinacate volcanic field of northwestern Sonora, Mexico. *Geological Society of America Abstracts with Programs*, 13, 93.
- Mysen, B.O. (1976) Experimental determination of some geochemical parameters relating to conditions of equilibration of peridotite in the upper mantle. *American Mineralogist*, 61, 677–683.
- Ozawa, K. (1983) Evaluation of olivine-spinel geothermometry as an indicator of thermal history for peridotites. *Contributions to Mineralogy and Petrology*, 82, 52–65.
- Papike, J.J., Cameron, K.L., and Baldwin, K. (1974) Amphiboles and pyroxenes: Characterization of OTHER than quadrilateral components and estimates of ferric iron from microprobe data. *Geological Society of America Abstracts with Programs*, 6, 1053–1054.
- Pike, J.E.N., and Schwarzman, E.C. (1977) Classification of textures in ultramafic xenoliths. *Journal of Geology*, 85, 49–61.
- Powell, Marjorie. (1978) Crystallisation conditions of low-pressure cumulate nodules from the Lesser Antilles island arc. *Earth and Planetary Science Letters*, 39, 162–172.
- Roy, R.F., Decker, E.R., Blackwell, D.D., and Birch F. (1968) Heat flow in the United States. *Journal of Geophysical Research*, 73, 5207–5221.
- Sachtleben, Th., and Seck, H.A. (1981) Chemical control of Al-solubility in orthopyroxene and its implications on pyroxene geothermometry. *Contributions to Mineralogy and Petrology*, 78, 157–165.
- Sack, R.O. (1982) Spinel as petrogenetic indicators: Activity-composition relations at low pressures. *Contributions to Mineralogy and Petrology*, 79, 169–186.
- Sass, J.H., Lachenbruch, A.H., Munroe, R.J., Greene, G.W., and Moses, T.H., Jr. (1971) Heat flow in the western United States. *Journal of Geophysical Research*, 76, 6376–6413.
- Saxena, S.K., and Eriksson, G. (1983) Theoretical computation of mineral assemblages in pyrolite and lherzolite. *Journal of Petrology*, 24, 538–555.
- Sen, Gautam. (1985) Experimental determination of pyroxene compositions in the system CaO-MgO-Al₂O₃-SiO₂ at 900–1200°C and 10–15 kbar using PbO and H₂O fluxes. *American Mineralogist*, 70, 678–695.
- Smith, R.B. (1978) Seismicity, crustal structure, and intraplate tectonics of the interior of the western Cordillera. In R.B. Smith and G.P. Eaton, Eds. *Cenozoic tectonics and regional geophysics of the Western Cordillera*. Geological Society of America Memoir 152, 111–144.
- Tracy, R.J. (1980) Petrology and genetic significance of an ultramafic xenolith suite from Tahiti. *Earth and Planetary Science Letters*, 48, 80–96.
- Wells, P.R.A. (1977) Pyroxene thermometry in simple and complex systems. *Contributions to Mineralogy and Petrology*, 62, 129–139.
- White, R.W. (1966) Ultramafic inclusions in basaltic rocks from Hawaii. *Contributions to Mineralogy and Petrology*, 12, 245–314.
- Wilshire, H.G. (1984) Mantle metasomatism: The REE story. *Geology*, 12, 395–398.
- Wilshire, H.G., and Shervais, J.W. (1975) Al-augite and Cr-diopside ultramafic xenoliths in basaltic rocks from western United States. In L.H. Ahrens et al., Eds. *Physics and Chemistry of the Earth*, 9, 257–272. Pergamon, New York.

MANUSCRIPT RECEIVED AUGUST 22, 1985

MANUSCRIPT ACCEPTED MAY 16, 1986