

Age

Composition and origin of Nuevo Laredo Trend eucrites

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Abstract—New bulk-rock analyses are reported for the eucritic meteorites ALHA81001, Lakangaon and Nuevo Laredo. ALHA81001 resembles Ibitira, as both have far lower concentrations of Na than any other eucrite. Lakangaon and Nuevo Laredo are the two most “ferroan” (low-Mg/Fe) of all eucrites. Monomict eucrites have previously been divided into three geochemical classes: a Main “Group,” and the Nuevo Laredo and Stannern “trends.” The Nuevo Laredo Trend, which is characterized by limited variations of incompatible element contents despite considerable variations in Mg/(Mg + Fe) ratio and V content, clearly formed out of melt(s) undergoing fractional crystallization, probably as partial cumulates. The less distinctive Main Group eucrites are often assumed to be primary basalts, geochemically but one step (equilibrium partial melting) removed from the bulk composition of the eucrite parent asteroid. This partial melting model seems more appropriate for the Stannern Trend, which is characterized by moderate-high Mg/(Mg + Fe) despite moderate-high incompatible element contents. Our new data and a review of literature data suggest that most of the so-called Main Group eucrites actually belong to the Nuevo Laredo Trend; while few, if any, belong to the Stannern Trend, or any other partial melting trend. The Nuevo Laredo Trend may be part of a series of fractional crystallization products that began with orthopyroxenite cumulates (diogenites). The bulk MgO/(MgO + FeO) ratio of the eucrite parent asteroid is generally estimated at ~0.67 by assuming that typical eucrites are primary basalts. But if typical eucrites are diogenite-comagmatic fractional crystallization products, the bulk MgO/(MgO + FeO) of the parent asteroid must be considerably higher, probably about 0.81.

1. INTRODUCTION

THE EUCRITES, WHICH constitute the most common type of meteoritic basalt, are probably closely related to the diogenites, which are meteoritic orthopyroxenites, and the howardites, which are regolith breccias composed mainly of eucrites and diogenites (CLAYTON *et al.*, 1986; see also reviews by DODD, 1981, and WASSON, 1985). The eucrite parent body is probably an asteroid, possibly Vesta (DRAKE, 1979). The eucrites are the oldest known basalts in the solar system. Five reported Sm-Nd ages range from 4.41 ± 0.02 to 4.60 ± 0.04 Ga (BVSP, 1981). The lower end of this range presumably reflects the extended subsolidus annealing that is evident among all monomict eucrites except Pasamonte (TAKEDA *et al.*, 1983), for which the Sm-Nd age is 4.58 ± 0.12 Ga. A Rb-Sr age determined for a similarly unannealed clast in the Yamato-75011 eucrite is 4.56 ± 0.11 Ga (BANSAL *et al.*, 1985). A minority of eucrites, characterized by coarse, equant textures, are interpreted as cumulates. However, more typical eucrites have subophitic (basaltic) textures, indicative of relatively rapid cooling—the “noncumulate” eucrites (BVSP, 1981, pp. 218–219).

A longstanding debate concerns whether typical noncumulate eucrites represent ideal “primary” partial melts, *i.e.*, melts with compositions unmodified by differentiation (*e.g.*, crystal fractionation) following segregation from their source regions. STOLPER (1975, 1977) argued that most eucrites formed as essentially primary basalts, mainly because his experiments showed that nearly all noncumulate eucrites have bulk compositions that cluster close to a plagioclase-olivine-pigeonite peritectic. However, the petrogenetic signif-

icance of this clustering is not entirely clear (WARREN, 1985; DELANEY, 1986). An older model (MASON, 1962, 1967; MCCARTHY *et al.*, 1973) proposed that eucrites represent residual liquids from fractional crystallization of more mafic melt(s) which had earlier produced the diogenites as cumulate pyroxenites. Recently, TAKEDA *et al.* (1984), WARREN (1985), DELANEY (1986) and HEWINS (1986), all have emphasized (see also BVSP (1981), p. 595) that many processes, including partial melting, fractional crystallization, and magma mixing, may have been involved in the production of eucrites.

Previous work (STOLPER, 1977; REID *et al.*, 1979; BVSP, 1981, p. 229ff.) has suggested that the noncumulate eucrites may be classified into three types, based on a plot of whole-rock data for Mg/(Mg + Fe) (hereafter abbreviated as *mg*) versus an incompatible element, such as Ti. According to BVSP (1981), the “Main Group” of eucrites are those with *mg* ~ 0.38–0.41 and Ti ~ 3–4 mg/g. The “Stannern Trend” comprises four eucrites with *mg* similar to the Main Group, but with higher Ti, ranging up to 5.7 mg/g. BVSP (1981) assigned only three eucrites (Nuevo Laredo, Lakangaon, and Pasamonte) to the “Nuevo Laredo Trend,” defined as extending from the Main Group composition towards Nuevo Laredo, which has Ti = 5.7 mg/g and *mg* = 0.32. Figure 1 shows this same plot, *mg* vs. Ti, based on a more extensive data base (see Appendix) than used by BVSP (1981). The three groups are not as well resolved in Fig. 1 as in Fig. 1.2.8.7 of BVSP (1981), mainly because Stannern’s *mg* value is lower here, so the intermediate members of Stannern Trend no longer fall along a line between Stannern and the center of the Main Group. BVSP (1981) evidently used a single analysis (MCCARTHY *et al.*, 1973)

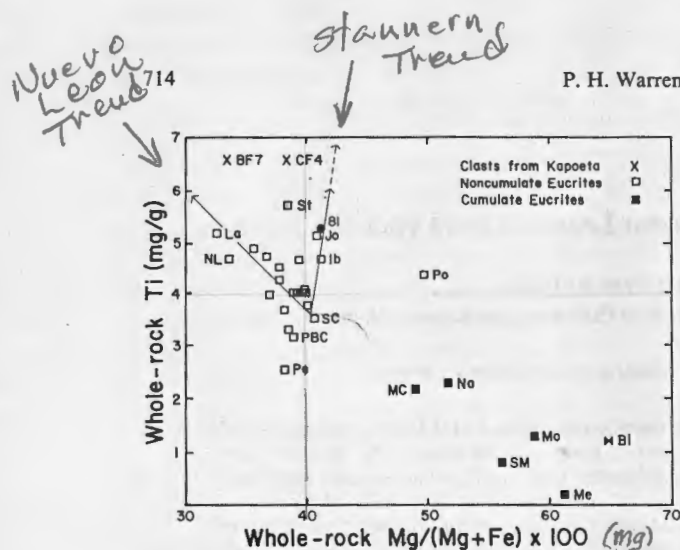


FIG. 1. The *mg* ratio vs. Ti for monomict eucrites, showing Nuevo Laredo and Stannern Trends (negative and positive slopes, respectively) after BVSP (1981). Also shown are two exceptionally "evolved" clasts from the Kapoeta howardite (SMITH, 1982). Binda (Bi) is shown with unusual symbol because it is sometimes classified as a howardite. See Appendix regarding meteorite abbreviations and sources of data.

for the *mg* of Stannern. The *mg* value shown in Fig. 1 is based averaging the MCCARTHY *et al.* (1973) *mg* datum, 0.411, with four others, which range from 0.363

to 0.380 (DUKE, 1963; DUKE and SILVER, 1967; JEFFOME, 1970; MITTFELDLT, 1979). In any event, note that the Stannern Trend and the Nuevo Laredo Trend are supposed to converge (or diverge) at the center of the Main Group. Thus, the Main Group is presumably related to one or both of the Trends. In this paper we present new data for three eucrites, including two key members of the Nuevo Laredo Trend. We attempt to resolve whether the Main Group is more closely related to the Stannern Trend or to the Nuevo Laredo Trend, and we examine the implications of these trends for the abundance of ideal "primary" basalts among the eucrites.

2. DATA, NEW AND OLD

Table I shows our new bulk-rock data for Nuevo Laredo, Lakangaon, and Antarctic eucrite ALHA81001. These data were obtained mainly by instrumental neutron activation analysis (INAA), using essentially the same procedure described by KALLEMEYN and WARREN (1983). Data for Al, Si, Mg, and Ti were obtained by electron microprobe analysis of fused beads (using typically 30 mg of aliquot from the powdered INAA sample), which also provided checks for Ca and Fe. Data for Ni, Re, Os, Ir and Au were obtained by radiochemical NAA (RNAA), using a procedure modified from SUNDBERG and BOYNTON (1977). The two masses shown for each meteorite in Table I pertain to separate powders produced

Table I. Concentrations of 37 elements in Lakangaon, Nuevo Laredo and ALHA81001.

	Lakangaon			Nuevo Laredo			ALHA81001,16		
	A	B	A&B avg.	A	B	A&B avg.	A	B	A&B avg.
Mass, mg	174.3	173.9		307.3	325.7		214.7	173.4	
Na, mg/g	3.72	3.79	3.76	3.76	3.77	3.77	1.88	1.96	1.82
Mg, mg/g	34.4	35.2	34.8	34	33	33.5	48	47	47.5
Al, mg/g	63	63	63	63	66	64.5	61	61	61
Si, mg/g	232	230	231	231	232	231.5	232	231	231.5
K, mg/g	0.42	0.48	0.45	0.40	0.43	0.415	0.23	0.34	0.29
Ca, mg/g	72	75	73.5	74	73	73.5	65	73	69
Sc, µg/g	32.5	32.3	32.4	33.3	33.3	33.3	30.2	30.2	30.2
Ti, mg/g	4.9	5.6	5.3	4.9	5.0	5.0	5.9	5.8	5.9
V, µg/g	62	50	56	64	58	61	64	71	67
Cr, mg/g	1.98	1.89	1.93	1.93	1.92	1.93	3.08	3.09	3.08
Mn, mg/g	4.20	4.23	4.22	4.50	4.54	4.52	4.35	4.35	4.35
Fe, mg/g	159	160	159.5	153	151	152	148	150	149
Co, µg/g	4.9	5.2	5.1	2.17	2.12	2.15	7.9	7.9	7.9
Ni, µg/g	1.3	—	1.3	2.9	—	2.9	<31	<26	<29
Ga, µg/g	1.8	3.1	2.5±7	1.0±2	1.6±3	1.2±3	1.2±2	1.3±3	1.2
Sr, µg/g	57±20	77±14	70	80±16	97±39	83±15	74±19	64±23	70±15
Zr, µg/g	<140	45±14	45±14	71±30	<70	70±30	110±20	58±25	90±30
Ce, µg/g	<0.46	<0.10	<0.10	<0.32	<0.32	<0.32	<0.35	<0.44	<0.35
Ba, µg/g	49±6	40±7	44	44±8	33±8	39±7	45±9	57±12	49±8
La, µg/g	3.80	3.33	3.57	3.79	3.87	3.83	3.6	5.0	4.3
Co, µg/g	9.9	9.2	9.5	10.0	9.8	9.9	10.3	13.5	11.9
Nd, µg/g	6.6	7.2	7.0	6.5	6.9	6.7	7.3	8.8	8.0
Sm, µg/g	2.11	2.27	2.19	2.33	2.31	2.32	2.33	3.15	2.74
Eu, µg/g	0.80	0.66	0.73	0.71	0.71	0.71	0.68	0.78	0.72
Tb, µg/g	0.57	0.53	0.55	0.54	0.54	0.54	0.56	0.74	0.65
Dy, µg/g	4.2±6	2.8	3.4	3.3±8	4.4±1.0	3.7±7	3.3±9	4.2±8	3.8±8
Ho, µg/g	—	0.84	0.84	—	—	—	—	—	—
Yb, µg/g	2.44	2.04	2.24	2.41	2.39	2.40	2.40	3.07	2.73
Lu, µg/g	0.36	0.33	0.35	0.35	0.34	0.35	0.35	0.43	0.39
Hf, µg/g	1.60	1.55	1.53	1.60	1.61	1.61	2.04	2.04	2.04
Ta, µg/g	0.19	0.21	0.20	0.178	0.183	0.181	0.24	0.26	0.25
Re, pg/g	2.3	—	2.3	6.8	—	6.8	—	—	—
Os, pg/g	8.4	—	8.4	80	—	80	—	—	—
Ir, ng/g	0.14	—	0.14	0.083	—	0.083	<6	<7	<7
Au, ng/g	4.6	—	4.6	1.5	—	1.5	<0.5	—	<0.5
Th, ng/g	330	360	350	430	440	440	560	560	560
U, ng/g	150±28	110±20	120±20	140±40	—	140±40	150±40	—	150±40

Except as noted, uncertainties are within limits listed, by element, in Warren *et al.* (1986).

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from different chips, albeit for each meteorite the chips were delivered to us in a single package, and thus presumably were derived from a single portion of the stone (however, the Nuevo Laredo "B" sample consisted partly of selvages of a black glassy material (shock-melted veins?, fusion crust?), whereas the "A" sample appeared to be free of this material). Following INAA, small aliquots (20–30 mg) of the INAA powders were used to produce the fused beads, and the remainders were used for RNAA.

Nuevo Laredo had previously been analyzed for many of the same elements, albeit never before simultaneously (utilizing a single sample) for trace elements and major elements (DUKE, 1963; SCHMITT *et al.*, 1963; MORGAN and LOVERING, 1964, 1965; HASKIN *et al.*, 1966; DUKE and SILVER, 1967; PAPANASTASSIOU and WASSERBURG, 1969; GAST *et al.*, 1970; TERA *et al.*, 1970). Our data are generally in good agreement with these older analyses, except for Sc, for which the three oldest measurements (DUKE, 1963; SCHMITT *et al.*, 1963; HASKIN *et al.*, 1966) are far higher (40–43 $\mu\text{g/g}$) than our two replicates (both 33.3 $\mu\text{g/g}$). An analysis of intermediate vintage (SCHMITT *et al.*, 1972) gave an intermediate result (38 $\mu\text{g/g}$). These Sc discrepancies might be due to sampling problems. Although Nuevo Laredo is, on the whole, a relatively fine-grained eucrite (DUKE and SILVER, 1967; TAKEDA *et al.*, 1983), it is a breccia, and Fig. 1.2.8.1 of BVSP (1981) indicates that its texture ranges all the way from vitric-vesicular to subophitic. However, considering the uniquely high Sc content that the older analyses imply for Nuevo Laredo (no other eucrite contains more than about 33 $\mu\text{g/g}$), we suspect that the older Sc analyses are simply inaccurate.

Lakangaon had been analyzed previously only for 10 elements: Mg, Al, Si, K, Ca, Ti, Mn and Fe (MCCARTHY *et al.*, 1974); Na and Cr (MASON *et al.*, 1979). Our data agree well with these earlier analyses, except we find lower K, Cr and Fe, by factors of 0.70, 0.80 and 0.90, respectively. In the case of Fe, this disparity may be largely due to sampling problems, because our microprobe analyses of pyroxenes, obtained from the same portion of Lakangaon that supplied our INAA sample, are likewise less ferroan than analyses reported by MASON *et al.* (1979) (Fig. 2).

DELANEY *et al.* (1984c) report that both Lakangaon and Nuevo Laredo have uncommonly high pigeonite/orthopyroxene ratios, but otherwise typical eucritic modes; *e.g.*, plagioclase contents of 44.9 and 43.5 vol%, respectively. Figure 3 shows our microprobe data for plagioclase compositions. Mean plagioclase An ratios are 82.2 ± 4.0 (one sigma) in Nuevo Laredo (59 analyses), and 84.1 ± 3.6 in Lakangaon (48 analyses). Similar plagioclase compositions have previously been reported for Nuevo Laredo by DUKE and SILVER (1967) and TAKEDA *et al.* (1983), and for Lakangaon by MASON *et al.* (1979).

Siderophile elements such as Au, Ir, Ni, Os and Re are useful as tracers of core formation, and (assuming a certain composition for the core) as indicators of core size, in the

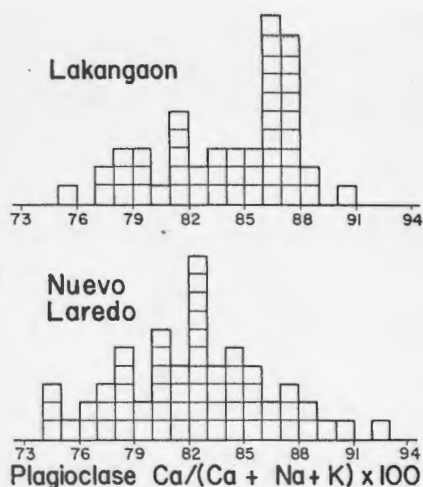


FIG. 3. An ratios of Lakangaon and Nuevo Laredo plagioclase.

eucrite parent asteroid (MORGAN *et al.*, 1978). The literature contains few data for these elements, particularly Os and Re, in eucrites; and those few data tend to scatter widely. For example, analyses of Ir in Juvinas range from 0.028 ng/g (MORGAN *et al.*, 1978) to 27 ng/g (WÄNKE *et al.*, 1972); two analyses of Os in Moore County range from 3 pg/g (MORGAN *et al.*, 1978) to 400 pg/g (MORGAN, 1965). Our new data for Os and Re in Lakangaon and Nuevo Laredo are low compared to most literature data, but similar to analyses of Ibitira, Juvinas, Moore County, and Serra de Magé (Os data for Juvinas and Moore County only), reported by MORGAN *et al.* (1978). By analogy with lunar rocks, eucrite siderophile element contents can be expected to scatter widely, as these elements tend to concentrate into rare, unevenly distributed metal grains. Nevertheless, many of the lunar data reported for these elements prior to *circa* 1973 appear to be erroneous due to laboratory contamination (WARREN *et al.*, 1986). Evidently many of the older data reported for eucrites also suffer from this problem.

Ours are the first bulk-rock analyses for ALHA81001, except for a broad-beam analysis by MASON (1983) and a modal recombination analysis by DELANEY *et al.* (1984b). MASON (1983) described ALHA81001 as an anomalous eucrite, based on its "quenched" texture (Fig. 4), which is reminiscent of rapidly cooled terrestrial mafic-ultramafic lavas (CAMERON and NISBET, 1982); and its unusually low Na content. MASON (1983) and DELANEY *et al.* (1984b) reported Na_2O contents of 0.2 and 0.18 wt%, respectively. Our two analyses (Table 1) average 0.245 wt% Na_2O . Among non-Antarctic "noncumulate" eucrites only Ibitira, which contains 0.19 wt% (WÄNKE *et al.*, 1974; HIGUCHI and MORGAN, 1975), has an Na_2O content much below 0.34 wt%. An average noncumulate eucrite has about 0.50 wt% Na_2O (DODD, 1981, p. 243). For many elements, our analyses appear to suffer from sampling problems. For example, REE in sample "B" are systematically 1.25 to 1.40 times higher than REE in "A." Potassium and to a lesser extent Na seem to be correlated, in this respect, with REE. Although our two analyses agree well with one another for major elements, they indicate a higher content of mafic silicates than the analyses of MASON (1983) and DELANEY *et al.* (1984b). Relative to the average of the data of MASON (1983) and DELANEY *et al.* (1984b), our Mg data are higher by a factor of 1.17, and our Fe data are higher by a factor of 1.10.

As discussed by DELANEY *et al.* (1984b), ALHA81001 is essentially unbrecciated, and might be either (a) the quenched equivalent of pristine (endogenously produced) basaltic melt, which DELANEY *et al.* suggest would have to be unusually Al-

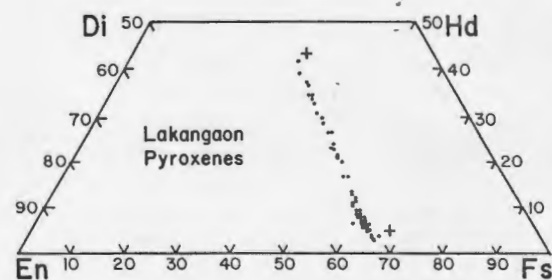


FIG. 2. Lakangaon pyroxene Mg-Fe-Ca proportions. Range for pyroxenes in Nuevo Laredo (58 points) is practically identical to range shown for Lakangaon (89 points). Two "+" symbols show pyroxene compositions reported for Lakangaon by MASON *et al.*, (1979).

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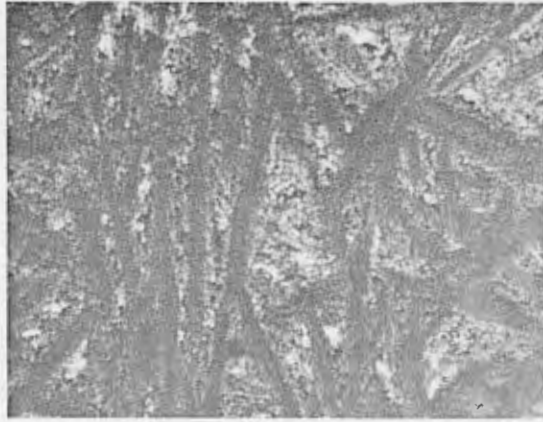


FIG. 4. Transmitted light photomicrograph of ALHA81001. View is 1.5 mm long. The parallelism of the "soda straw" (DELANEY *et al.*, 1984b) pyroxene phenocrysts in this area is unique in the thin section; phenocrysts are also uncommonly abundant here. Otherwise, however, this view is typical.

rich and Na-poor; or (b) an impact melt, which DELANEY *et al.* suggest might have been produced from an unusually Fe-rich and Na-poor cumulate eucrite. Model (a) seems more likely—the upper limits we have determined for Ni and Au (Table 1) indicate little or no meteoritic contamination. However, our analyses do not indicate a particularly Al-rich composition. Although MASON (1983) and DELANEY *et al.* (1984b) reported identical Al_2O_3 contents of 14 wt%, our analyses (Table 1) both indicate 11.5 wt%. Apparently the Al_2O_3 content of ALHA81001 is not significantly different from that of an average noncumulate eucrite, about 12.6 wt% (DODD, 1981). Regarding model (b), ALHA81001 has far higher incompatible element contents than any recognized cumulate eucrite. For example, its Sm content of 2.74 $\mu\text{g/g}$ compares with an average of 0.43 $\mu\text{g/g}$ (range 0.16–0.89 $\mu\text{g/g}$) for cumulate eucrites (Fig. 5). Thus, if ALHA81001 is an impact melt (which seems unlikely, anyway) it is apparently not predominantly of cumulate-eucrite derivation.

The Appendix cites 184 literature analyses of eucrites that have been used to construct Figs. 1 and 5–8. Bulk analyses of polymict Antarctic eucrites (SMITH and SCHMITT, 1981) are not included in the variation diagrams, because mixing can easily scramble igneous compositional relationships.

3. INCOMPATIBLE ELEMENTS vs. $\text{Mg}/(\text{Mg} + \text{Fe})$

Incompatible elements, which ideally are totally excluded from all solids, are excellent tracers of crystal-melt fractionation. During fractional crystallization, the concentration of an incompatible element, such as the "typical" rare earth element Sm, increases in the melt in direct proportion to $1/f$, where f is the fraction of the original system that is molten. The same relationship holds during equilibrium (batch) partial melting, where as f increases the concentration of an incompatible element decreases. In short,

$$c_E = C_E/f \quad (1)$$

where c_E is the melt concentration of an incompatible element, and C_E is the melt concentration when $f = 1$; either at the onset of fractional crystallization, or, in the case of partial melting, at the point where (in theory) melting becomes total.

Another excellent tracer is the mg ratio, which, in basaltic, low- $f(\text{O}_2)$ systems, is always moderately lower in the melt than in coexisting mafic silicates. This relationship is governed by an exchange reaction distribution coefficient, K_D , defined (ROEDER and EMSLIE, 1970) as

$$K_D = ([\text{FeO}]_{\text{xtl}}/[\text{FeO}]_{\text{liq}})/([\text{MgO}]_{\text{xtl}}/[\text{MgO}]_{\text{liq}}) \quad (2)$$

where the square brackets denote molar concentrations, and the subscripts are xtl = crystal and liq = liquid. Equation (1) implies

$$mg_{\text{xtl}} = 1/(1 + K_D/mg_{\text{liq}} - K_D) \quad (3)$$

and

$$mg_{\text{liq}} = K_D/(K_D + 1/mg_{\text{xtl}} - 1). \quad (4)$$

The only important mafic silicates found among the eucrite-diogenite-howardite clan are pigeonite, orthopyroxene, and olivine. Many literature data (reviewed by WARREN, 1986), indicate that at low pressure K_D

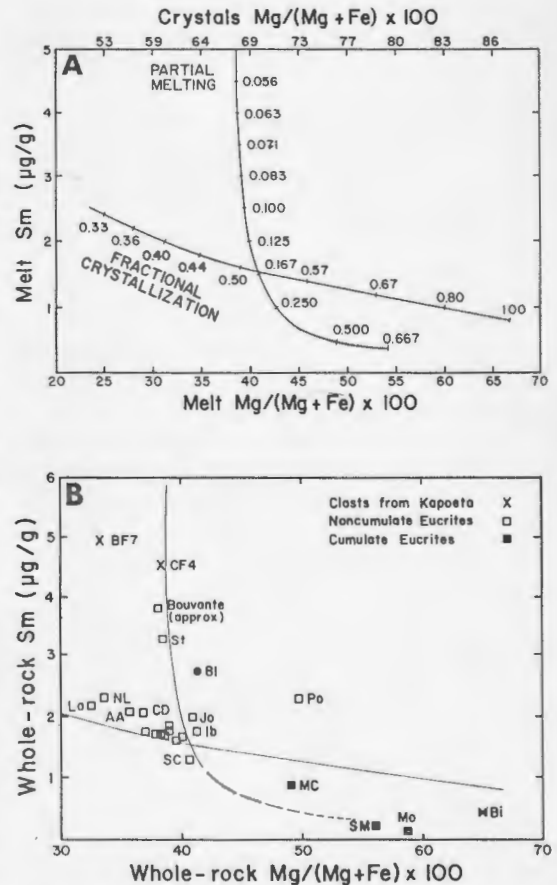


FIG. 5. (a) Model results: melt mg ratio vs. Sm concentration, during partial melting and fractional crystallization. Scale at top of diagram indicates mg ratios of coexisting mafic silicates. Numbers beside points indicate f , the fraction of the initial "system" currently molten (in the case of partial melting the system is the bulk source region; in the case of fractional crystallization the system is the initial melt, at the onset of crystallization). (b) Bulk-rock mg ratio vs. Sm for monomict eucrites; analogous to Fig. 1.

tracer is the *mg* ratio, which, in systems, is always moderately lower than that of coexisting mafic silicates. This result is derived by an exchange reaction distribution coefficient, K_D , defined (ROEDER and EMSLIE,

$$K_D = \frac{[FeO]_{xtl}}{[MgO]_{xtl}} / \frac{[FeO]_{liq}}{[MgO]_{liq}} \quad (2)$$

where $[FeO]_{xtl}$ and $[MgO]_{xtl}$ denote molar concentrations in the crystal and $[FeO]_{liq}$ and $[MgO]_{liq}$ denote molar concentrations in the liquid.

$$a = \frac{[MgO + FeO]_{xtl}}{[MgO + FeO]_{liq}} \quad (3)$$

$$f = \frac{K_D / (K_D + 1/mg_{xtl} - 1)}{1 + K_D/mg_{liq} - K_D} \quad (4)$$

where mg_{xtl} and mg_{liq} are the *mg* ratios of mafic silicates found among the various mineral assemblages. Many literature data (reviewed by GROVE, 1978) indicate that at low pressure K_D

is $0.30 \pm$ about 0.02 for olivine coexisting with basaltic melt. STOLPER (1977) found that K_D is 0.30 for pigeonite coexisting with eucritic melt. GROVE (1978) and LONGHI (1980) found that K_D is also 0.30 for pigeonite coexisting with lunar basaltic melts. For orthopyroxene, literature data (WARREN, 1986) indicate that K_D is probably in the range 0.25–0.32.

In order to apply Eqns. (2)–(4) to the evolution of the melt *mg* ratio during partial melting, we must constrain the ratio *a*, defined (WARREN, 1986) as

$$a = \frac{[MgO + FeO]_{xtl}}{[MgO + FeO]_{liq}} \quad (5)$$

i.e., *a* equals the mole fraction of (MgO + FeO) in the crystals divided by the same mole fraction in the melt. Literature data (JAQUES and GREEN, 1980; MYSEN and KUSHIRO, 1977) indicate that during partial melting of a peridotitic source $a = 2.0 \pm$ about 0.4. Little or no significant variation in *a* occurs with increasing *f* (WARREN, 1986). Starting from Eqns. (3) and (4), WARREN (1986) derived equations interrelating mg_{sys} (the *mg* ratio of the source region or "system" undergoing melting), mg_{liq} , mg_{xtl} , *a*, *f*, and K_D (assuming a single K_D is representative for the combined unmelted mafic silicates):

$$mg_{sys} = \frac{f \cdot mg_{liq} + (1-f) \cdot a / (1 + (K_D/mg_{liq} - K_D))}{f + (1-f) \cdot a} \quad (6)$$

and

$$mg_{sys} = \frac{f \cdot K_D / (K_D + 1/mg_{xtl} - 1) + (1-f) \cdot mg_{xtl} \cdot a}{f + (1-f) \cdot a} \quad (7)$$

Equation (6) can be solved for mg_{liq} , but the solution is long and cumbersome. In practice we use a simple computer program to solve for mg_{liq} directly from Eqn. (6) by successive approximation (testing different values of mg_{liq} until the assumed value of mg_{sys} is precisely matched). Thus, Eqn. (6) can be used to model evolution of the melt *mg* ratio as *f* increases during partial melting of a eucrite source region.

Unlike the melt concentration of an incompatible element, behavior of the melt *mg* ratio as a function of *f* is sensitive to whether crystal-melt fractionation occurs *via* partial melting or *via* fractional crystallization. Figure 5a shows these contrasting scenarios, based in the case of partial melting on Eqn. (6), and in the case of fractional crystallization on a finite-difference model, *à la* LONGHI (1977). For the partial melting model, the assumptions are that the source region mg ratio (mg_{sys}) = 0.670, its content of Sm = 250 ng/g (i.e., about the same as the Sm content of the nonmetallic, nonsulfide fraction of an ordinary chondrite (MASON, 1979)), *a* = 2.0, and K_D = 0.30 (based on a presumption that the main residual mafic phases are olivine and low-Ca pyroxene). For the sake of simplicity, Sm is treated as an ideal incompatible element ($D = 0$). The value of mg_{sys} was arbitrarily chosen to

"fit" the *mg* ratios of Stannern and (with less emphasis) Bouvante and the highest-*mg* "noncumulate" eucrites (Ibitira and Jonzac).

For the fractional crystallization model, the initial composition is again somewhat arbitrary (the main purpose of the model is to illustrate the general nature of a fractional crystallization sequence, not to provide a precise "fit" to the eucrite data array). As discussed by WARREN (1985), mass-balance considerations suggest that most eucrites formed as later products of the same melts that had earlier produced diogenites. We assumed the initial composition for the fractional crystallization model to be the 1:1 average of two components: (a) the relatively high-*mg* diogenite Tatahouine (MCCARTHY *et al.*, 1972), albeit with the 10 mg/g of reduced Fe (7.9 mg/g as Fe-metal and 2.2 mg/g as FeS) that this meteorite contains (MASON, 1963) subtracted out of the bulk composition; and (b) the relatively high-*mg* eucrite, Ibitira (WÄNKE *et al.*, 1974; HIGUCHI and MORGAN, 1975; PALME and RAMMENSEE, 1981), albeit with 1.5 mg/g of Na and 0.12 mg/g of K added. Thus, the assumed initial composition, in wt%, is Na₂O 0.2, MgO 18, Al₂O₃ 6.6, SiO₂ 52, K₂O 0.15, CaO 5.8, TiO₂ 0.4, Cr₂O₃ 0.6, MnO 0.5, FeO 16.

Being fine-grained and vesicular (STEELE and SMITH, 1976), Ibitira is particularly likely to have a "primary" composition, little modified by fractional crystallization. However, Ibitira is also unusually Na- and K-poor (see STOLPER, 1977), hence the adjustments to Na and K. Although STEELE and SMITH (1976) interpret Ibitira as a surface lava or shallow intrusive, it might conceivably be an impact melt breccia. However, Ibitira has extremely low contents of trace siderophile elements: e.g., Ir = 3.9 pg/g, Re = 2.0 pg/g (MORGAN *et al.*, 1978); for comparison, these same authors reported for Juvinas Ir = 28 pg/g, Re = 9.7 pg/g. Assuming the Moon (RYDER and NORMAN, 1980) is representative of small, atmosphereless bodies, siderophile elements probably tend to be enriched in impact melts. In any case, a relatively high-*mg* initial composition, or else one that is relatively Al-rich, appears necessary to have the melt arrive at plagioclase saturation with an appropriate (Ibitira-like, Jonzac-like) *mg*. Although the composition ("Y") proposed as a fractional crystallization progenitor by IKEDA and TAKEDA (1985) has an *mg* ratio of 0.704, its Al₂O₃ content is only 4.1 wt%. Finite-difference modeling of olivine, and later pyroxene fractionation from this initial melt indicates that by the time plagioclase saturation (Al₂O₃ = 12 wt%, at *f* = 0.31) is reached the melt *mg* has dropped to about 0.14—much lower than the *mg* of any eucrite.

In our own model, the initial melt composition has an *mg* ratio of 0.667. The initial Sm content is assumed to be 0.80 μg/g, which is a compromise between the Sm content derived as the simple average of the two "components" (0.85 μg/g), and the Sm content suggested by assuming chondritic proportionality between Sm and several other refractory lithophile elements (Al, Ca and Ti) that should be nearly incompatible with the silicates (olivine and low-Ca pyroxene) that

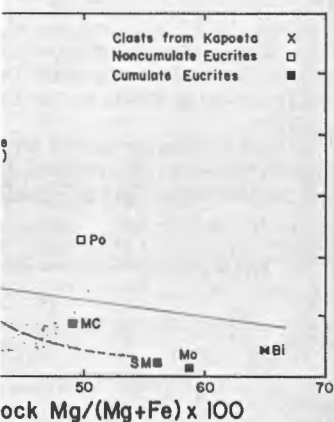
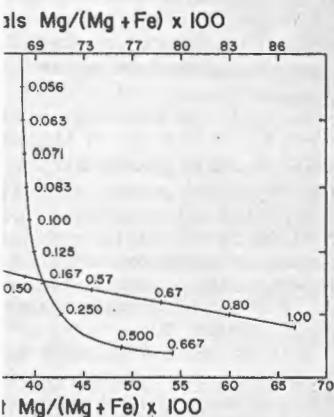


Figure 5a: melt *mg* ratio vs. Sm concentration, and fractional crystallization. Scale at top is *mg* ratios of coexisting mafic silicates. *f* indicates the fraction of the initial melt (in the case of partial melting the source region; in the case of fractional crystallization the initial melt, at the onset of crystallization). Scale at bottom is Sm for monomict eucrites.

might have constituted the residual solids left behind when this melt formed in the interior of the asteroid (about 0.70 $\mu\text{g/g}$). In the model, this composition evolves as first pure orthopyroxene, and later (after f falls below 0.5) pigeonite plus plagioclase fractionally crystallize, in proportions controlled by STOLPER's (1977) phase diagram. The model assumes that $K_D = 0.30$ for orthopyroxene as well as pigeonite. Again, for the sake of simplicity, Sm is treated as an ideal incompatible element ($D = 0$). The reader is referred to LONGHI (1977) regarding the mechanics of constructing this finite-difference type of fractional crystallization model.

Note that for any realistic value of f (i.e., $f < 0.5$) the slope of the partial melting trend on Fig. 5a is far steeper than the slope of the fractional crystallization trend. This same tendency is apparent from similar calculations by MAALØE (1982) for evolution of melt mg ratio as a result of olivine fractionation during fractional crystallization (his Fig. 10) and partial melting (his Fig. 2). The actual eucrite compositions are shown in Fig. 5b. Bouvante and Stannern are clearly associated with the partial melting trend, and Lakangaon and Nuevo Laredo are clearly associated with the fractional crystallization trend. Further, the 10 or so Main Group eucrites appear to distribute mainly along the fractional crystallization trend. A few members of the Main Group might be high- f members of the Stannern Trend. But the wide range in mg among the full Main Group would imply partial melting over a wide range of f , whereas the Group's narrow range in Sm would require a nearly constant f . If Pomozdino, "a monomict breccia, the breccia fragments consisting of a subophitic intergrowth of plagioclase and pigeonite" (MASON *et al.*, 1979)—analyzed by KVASHA and DYAKONOVA (1972) and KOLOSOV (1976), represents a primary equilibrium partial melt, then none of the other eucrites, all of which have lower mg than Pomozdino, could possibly be derived by partial melting of the same source composition.

Other elements, particularly rare earths, could substitute for Sm with similar results. Although Ti and K are also essentially incompatible with the relevant major minerals, the data (e.g., Fig. 1) scatter too much to be amenable to this type of analysis. Scatter for Hf (Fig. 6) is not as bad. The trends are similar for mg vs. Ba, and mg vs. Zr, albeit the utility of these combinations is limited (in terms of assessing the relative importance of fractional crystallization vs. partial melting) by scarcity of data: Only 11 noncumulate eucrites have been analyzed for Ba, and only 10 have been analyzed for Zr—whereas 17 have been analyzed for Hf, and 18 for Sm. Nevertheless, STOLPER (1977) noted that mg vs. Zr trends are consistent with Cachari, Juvinas, Pasamonte, Lakangaon, and Nuevo Laredo, all being produced by "crystallization differentiation" of a liquid similar to Sioux County. Note that two "evolved" clasts from Kapoeta (SMITH, 1982) appear to have affinities mainly with the Stannern Trend.

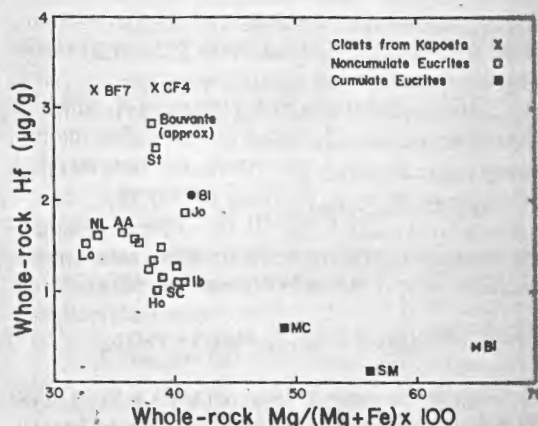


FIG. 6. Bulk-rock mg ratio vs. Hf for monomict eucrites; analogous to Fig. 5b.

4. INCOMPATIBLE ELEMENTS vs. VANADIUM

In terms of oxygen fugacities, eucrites "have affinities that closely parallel those of lunar basalts, characterized by metallic iron with crystallization $f(\text{O}_2)$ below that of the IW buffer curve" (BVSP, 1981, p. 378). In lunar basaltic systems, and hence by implication in eucritic systems, crystal/melt distribution coefficients for V (D_V) are 1.3 for olivine (RINGWOOD, 1970), ~ 3.1 for low-Ca pyroxene (2.8 according to RINGWOOD, 1970; 3.4 according to LINDSTROM, 1976), and ~ 38 for chromite (RINGWOOD, 1970). Evolution of the melt V content, c_V , during equilibrium partial melting can be calculated using the following equation from SHAW (1970):

$$c_V = c_0(D_0 + f(1 - P)) \quad (8)$$

where c_0 is the initial V content of the source region, D_0 is the bulk (weighted mean) distribution coefficient for the initial solids in the source region, and the term P is also a (weighted mean) distribution coefficient, but in this case weighted according to the fraction of each solid phase that enters the melt. SHAW's (1970) model is often simplified into a "modal melting" model, in which P is assumed to be equal to D_0 . For the case of the eucrites, assuming that the initial source region consists of about 1 wt% chromite, 7 wt% plagioclase, plus olivine and low-Ca pyroxene in a roughly 2:1 ratio, and that the phases entering the melt are 1% chromite, 42 wt% plagioclase and 57 wt% low-Ca pyroxene (STOLPER, 1977), the simplified model gives precisely the same results as the more rigorous model; i.e., $D_0 = P = 2.1$. Figure 7a shows the evolution of the melt V content vs. Sm content, based on Eqn. (8), during equilibrium partial melting. As for Fig. 5, for the sake of simplicity Sm is assumed to be perfectly incompatible (i.e., D_0 and P both = 0), and the initial source region Sm content is assumed to be 250 ng/g. The initial source region V content, constrained to give a reasonably good "fit" to the V contents of Stannern

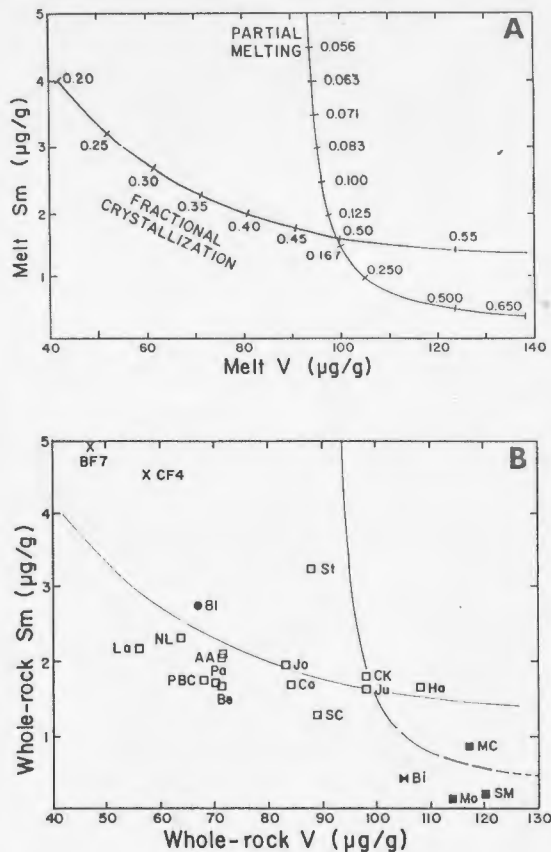


FIG. 7. (a) Model results: melt V concentration vs. Sm concentration, during partial melting and fractional crystallization; analogous to Fig. 5a. (b) Bulk-rock V vs. Sm for monomict eucrites; analogous to Fig. 5b.

and the more V-rich noncumulate eucrites (Fig. 7b), is assumed to be 190 μg/g.

Figure 7a also shows a contrasting melt evolution trend based on the fractional crystallization model used for Fig. 5, with V fractionation base on D_V -pyroxene = 3.1. The initial melt V content, constrained to have the melt V content = 100 μg/g at $f = 0.5$, is assumed = 429 μg/g. Possible further fractionation of V by chromite removal is ignored in this model. An average noncumulate eucrite contains about 0.2 ± 0.2 vol% (i.e., 0.3 ± 0.3 wt%) chromite (DELANEY *et al.*, 1984c); an upper limit of 0.7 wt% can be inferred from the average Cr₂O₃ content of noncumulate eucrites, 0.34 wt% (DODD, 1981). These chromites hold an average of about 5400 ± 900 μg/g V (BUNCH and KEIL, 1971). Thus, of the total V contained in an average noncumulate eucrite (about 81 μg/g—Fig. 7b), probably only about 20%, and almost certainly less than 50%, resides in chromite. An average cumulate eucrite contains about 0.51 ± 0.27 vol% (0.85 ± 0.45 wt%) chromite (DELANEY *et al.*, 1984c); which holds about 5000 ± 500 μg/g V (BUNCH and KEIL, 1971; LOVERING, 1975). Thus, of the total V contained in an average cumulate eucrite (about 114 μg/g—Fig. 7b), probably

only about 37% resides in chromite. Also, the Cr contents of Lakangaon and Nuevo Laredo (both 1.9 mg/g) (Table 1) are virtually identical to those of high-mg noncumulate eucrites such as Ibitira, Juvinas, and Sioux County (2.1–2.3 mg/g). Thus, we assume that the chromites in eucrites are products of trapped interstitial liquid, and that pyroxene and plagioclase were the only cumulus phases affecting the fractionation of a large body of melt. If the fractional crystallization model were altered by adding 0.3–0.9 wt% chromite, with $D_V = 38$, into the cumulus matter, the effect would be to slightly increase the bulk-cumulus D_V , roughly from 1.8 to 2.0; which, in terms of Fig. 7, would slightly improve the “fit” of the fractional crystallization model to the array of noncumulate eucrites (other than Stannern).

The mean V content of chondritic silicates is only roughly 100 μg/g (MASON, 1979). Note that the high initial V content that must be invoked to model the eucrites, whether they are modeled as partial melts or as fractional crystallization residues, suggests that one or more of the distribution coefficients cited above may be too high. For example, even if D for olivine were 0, and olivine were the sole residual phase, primary partial melt from a source with $V = 100$ μg/g would have $V < 429$ μg/g for any $f > 0.233$. Likewise, moderate- f primary partial melt of a source with $V = 100$ μg/g could not have $V = 100$ μg/g unless D_0 and P of Eqn. (8) were ~ 1 . The effect of a lower pyroxene D on Fig. 7a would be to steepen the negative slopes of both trends; a lower olivine D would steepen the partial melting trend only. In any case, provided that D for pyroxene is significantly > 1 , the basic relationship between the slopes would not change.

The two trends in Fig. 7 are closely analogous to the two trends in Fig. 5. The partial melting model predicts (for any reasonable f) only limited variation in V as Sm increases. In contrast, the fractional crystallization model predicts abruptly decreasing V as Sm increases. With the glaring exception of Stannern, the eucrites (Fig. 7b) are distributed along a trend of widely-varying V despite only modest variation in Sm, which suggests that most are more closely related to the Nuevo Laredo (fractional crystallization) Trend than to the Stannern (partial melting?) Trend.

5. DISCUSSION

Vanadium vs. Mg/(Mg + Fe)

The eucrites exhibit a correlation between mg and V (Fig. 8). The slope of this trend is steeper (by about a factor of three) than predicted by the simple fractional crystallization model (however, if cumulus matter is a significant component of Nuevo Laredo Trend “non-cumulate” eucrites, as discussed at length below, this discrepancy tends to disappear). On the other hand, the partial melting model predicts that as long as f remains in the range 0.05–0.25, the melt mg ratio does

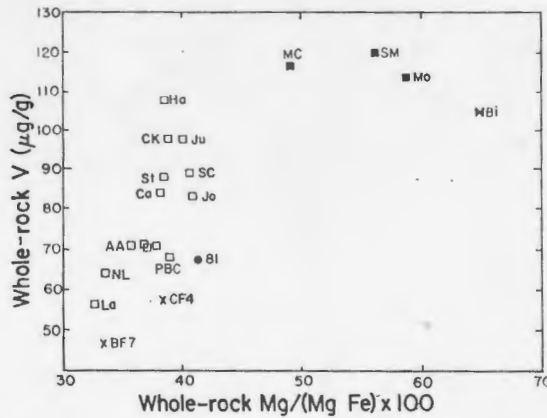


FIG. 8. Bulk-rock mg vs. V for monomict eucrites; symbols, data base as for Fig. 1.

not change by more than a total of 3.8 mole percent, and the melt V content does not change by more a factor of 1.10. Variation of f over more than this range can be ruled out on several grounds. First, a 5-fold variation in f would imply a 5-fold variation in Sm , which is not observed (Figs. 5b and 7b). Second, increasing f much beyond 0.17 would be inconsistent with the observation that all noncumulate eucrites are, as melts, saturated (or at least nearly saturated) with plagioclase (STOLPER, 1977). The only way that such a high-degree partial melt could be plagioclase-saturated would be if the initial source region was far more plagioclase-rich than any model for the eucrite parent body's mantle composition (CONSOLMAGNO and DRAKE, 1977; HERTOGEN *et al.*, 1977; MORGAN *et al.*, 1978; DREIBUS and WÄNKE, 1980; JONES, 1984; IKEDA and TAKEDA, 1985).

Origin of the Nuevo Laredo Trend

In the fractional crystallization model, the decrease in f , as gauged by the decrease in mg , is surprisingly small as the melt evolves from the point of initial plagioclase saturation (at the high- mg end of the Main Group, *i.e.*, a composition similar to Ibitira or Sioux County) to a Lakangaon- or Nuevo Laredo-like composition. Hereafter we refer to the average composition of Ibitira and Sioux County as "ISC" and the average composition of Lakangaon and Nuevo Laredo as "LNL". The mere 9 mol% decrease in mg from ISC to LNL implies only about 20% crystallization of the ISC-like melt (*i.e.*, f drops from about 0.5 to 0.4; Fig. 5). In contrast, the 40% decrease in V from ISC to LNL suggests about 40 wt% crystallization of the ISC-like melt (*i.e.*, f drops from 0.5 to 0.3; see Fig. 7). The roughly 48% increase in Sm from ISC to LNL implies about 32% crystallization of the ISC-like melt (Eqn. (1)). STOLPER (1977), based primarily on bulk-rock Zr data (but also on observations of "crystallinities of the products of melting experiments on the Sioux County eucrite in which the $Fe/(Fe + Mg)$ ratios and Ti concentrations of the experimental liquids are similar to

those of other eucrites"), similarly estimated that "30-40% crystallization of liquids similar to Sioux County" could have produced residual liquids similar to LNL.

Even though Lakangaon and Nuevo Laredo are considered "noncumulate" eucrites, they (and, for that matter, most other eucrites of the Nuevo Laredo Trend) might consist in part of "cumulus" material. By "cumulus" we mean material formed not by quenching of the parent melt, but by equilibration with a large (*i.e.*, not merely interstitial) mass of the parent melt at a particular stage in the compositional evolution of that parent melt (the compositional evolution of the parent melt being a by-product of the isolation of "cumulus" matter as the parent intrusion crystallizes). Thus, crystals may be mainly of "cumulus" origin and still be suspended in a magmatic mush of crystals + melt near the margin (generally the bottom) of an intrusion (IRVINE, 1980). The distinction of cumulus matter from "trapped liquid" is notoriously difficult on the basis of petrographic criteria alone. Consider the 300-m thick Palisades Sill (WALKER, 1969) as an analog. Except for a single 5-m thick layer near its base (the famous olivine-bearing layer), the rocks of this diabase intrusion have subophitic textures (WALKER, 1969) that give no hint of any cumulus origin; a meteorite with such a composition and texture would be assumed to be basalt. Yet, the steady fractionation of incompatible elements in vertical sections through the sill (WALKER, 1969; SHIRLEY, 1986) can only mean that these rocks are in fact typically about 50-60 wt% "cumulus," in the sense just described. Assume, for example, that the average composition of Lakangaon and Nuevo Laredo (LNL) represents 65 wt% trapped liquid, 21 wt% "cumulus" pyroxene, and 14 wt% "cumulus" plagioclase. The implied composition of the LNL parent melt, using the distribution coefficients cited above, has $V = 46 \mu\text{g/g}$ (vs. $60 \mu\text{g/g}$ in LNL), $Sm = 3.46 \mu\text{g/g}$ (vs. $2.25 \mu\text{g/g}$ in LNL), and $mg = 0.21$ (vs. 0.33 in LNL). If this melt formed by fractional crystallization of an ISC-like melt, the implied extent of crystallization is about 56 wt% based on V , about 56 wt% based on Sm , and about 40 wt% based on mg —resulting in better agreement than the alternative approach (previous paragraph) of assuming that LNL represents a quenched melt.

Even with corrections for possible cumulus matter in the LNL composition, the extent of crystallization implied for fractional crystallization to produce a LNL-like melt out of an ISC-like melt is still far less than 100%. The compositional course of fractional crystallization is essentially independent of whether the cumulates are pure cumulus matter or partly trapped liquid—the effect of "trapping" liquid is merely to reduce the quantity of the melt; its composition is not affected. However, "trapping" of liquid would cause the physical amount of melt remaining to be less than the f used in the modeling, and thus would help explain the observed scarcity of samples formed at low f . Any intrusion tends to "congeal" as the physical scale of

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its melt fraction diminishes; and the slight density contrast between cumulate eucrites and their parent melts would not be conducive (particularly in the low- g environment of an asteroid) to continued fractional crystallization down to very low f (WARREN, 1985).

Relationship of Nuevo Laredo Trend eucrites to diogenites and cumulate eucrites

The most primitive (high mg , high V , low Sm , etc.) members of the Nuevo Laredo Trend are ISC-like eucrites near the intersection of the Nuevo Laredo and Stannern trends. But the same fractional crystallization sequence that produced these eucrites might have earlier produced a series of more mafic cumulates—possibly, according to a model originally proposed by MASON (1962), the diogenites. WARREN (1985) adduced a mass balance constraint that suggests that most eucrites are indeed products of residual melts from the intrusion(s) that earlier produced the diogenites. Among observed falls, diogenites are almost as numerous as eucrites, and among the howardites (which probably provide a better “sample” of the eucrite parent body than do the observed falls), diogenitic and eucritic components are approximately equal in abundance. Assuming that the diogenitic orthopyroxenites formed as cumulates from melts saturated with pyroxene but not with plagioclase, these same melts must have later become plagioclase-saturated, *i.e.*, they must have later become eucrite-like. In fact, unless these melts began with compositions implausibly far from plagioclase saturation, they probably ultimately yielded at least nearly as much eucrite-like basalt as diogenitic orthopyroxenite. But eucritic components are not much more abundant than diogenitic components, so mass balance suggests that most eucrites are basaltic residual materials from the same magma(s) that produced the diogenites.

The high density contrast between the diogenites (which are nearly pure pyroxene) and a eucrite-like melt would promote efficient fractional crystallization. However, with the onset of plagioclase saturation, the density contrast between the melt and its aggregate liquid (= potential cumulus) crystals would suddenly diminish, by a factor of roughly 2.8 (densities calculated after BOTTINGA and WEILL (1970), assuming cumulus proportions are 42 wt% plagioclase and 58 wt% pyroxene (STOLPER, 1977)), which would in turn diminish the efficiency of fractional crystallization—and enhance the chances that the melt, or some large portion of it, would congeal into a basaltic, “noncumulate” rock. Considering this density effect, reinforcing the normal tendency for any intrusion to congeal as the physical scale of its melt fraction diminishes, it is not surprising that the Nuevo Laredo Trend of noncumulate eucrites extends only from $Sm \sim 1.5 \mu\text{g/g}$ to $Sm \sim 2.3 \mu\text{g/g}$ (Fig. 5). During low-pressure fractional crystallization of melts derived from sources with chondritic ratios of Al, Ca and Sm (all of which are refractory lithophile elements), plagioclase saturation

begins when the melt Sm content reaches $\sim 1.5 \mu\text{g/g}$ (WARREN, 1983).

The bulk $MgO/(MgO + FeO)$ ratio of the eucrite parent body is generally estimated at ~ 0.67 (MORGAN *et al.*, 1978; HERTOGEN *et al.*, 1977; CONSOLMAGNO and DRAKE, 1977), which is appropriate to account for the Stannern Trend eucrites as primary, moderate- f partial melts. However, DREIBUS and WÄNKE (1980) propose a composition with $MgO/(MgO + FeO) = 0.791$. A similarly high bulk-asteroid $MgO/(MgO + FeO)$ would be necessary to account for the diogenites and Nuevo Laredo Trend eucrites as a single fractional crystallization series, because even after producing the diogenites the melt mg ratio must still be as high as that of an “ISC-like” eucrite (*i.e.*, ~ 0.40). The model applied above assumed that the mg of the initial melt was 0.667. Assuming that this melt was produced as a partial melt of the asteroid’s mantle, at a moderately high f (say $0.3 < f < 0.6$), application of Eqn. (6) suggests that the source region’s $MgO/(MgO + FeO)$ must have been in the range 0.78–0.83. An advantage of a high $MgO/(MgO + FeO)$ for the bulk asteroid is that the pallasites, essentially biminerallitic olivine-metal meteorites with olivines as magnesian as $FO_{89.2}$ (BUSECK and GOLDSTEIN, 1969), might then be modeled as pieces of the same asteroid. Oxygen isotopic data (CLAYTON, 1977) suggest a genetic link between eucrites, diogenites and pallasites. Unless K_D for olivine is only 0.26—the extreme fringe of the range of K_D for olivine from experimental studies (WARREN, 1986), even a total melt of an $mg=0.67$ mantle would not crystallize FO_{89} olivine; with $K_D = 0.30$, the first olivine crystallized by such a melt would be $FO_{87.1}$. Assuming that eucrites, diogenites, and pallasites are from one asteroid, the pallasites may have formed when deep mantle layers, “depleted” of all major phases other than olivine by extensive partial melting, mixed with molten metal from the core of the asteroid (SCOTT, 1977). If the bulk-silicates mg of such an asteroid were 0.80, Eqn. (7), with $K_D = 0.30$, implies that the mg of these residual olivines would reach $FO_{89.2}$ at $f = 0.67$ (and $mg_{liq} = 0.71$).

If the parent melts of the Nuevo Laredo Trend indeed evolved out of Main Group (ISC-like) melts by fractional crystallization, then a series of magnesian cumulate eucrites must have formed in the process. However, the trend among “noncumulate” eucrites on Fig. 8 does not extrapolate to the area of the cumulate eucrites. Although the V contents of individual cumulates are sensitive to their plagioclase/pyroxene/chromite ratios, any bulk-rock mg analysis, being governed almost exclusively by pyroxene, should be reasonably representative. STOLPER (1977) noted that the melts that produced four of the five cumulate eucrites evidently had mg ratios in the range 0.23–0.29—significantly lower than the mg of any noncumulate eucrite (Fig. 5); the exception is Binda, which apparently formed from a melt with $mg \sim 0.37$. Considering that the complementary low- mg parent melts of typical cumulate eucrites are “missing” from our museum col-

lections, it is not entirely surprising that the magnesian cumulates that (hypothetically) formed as ISC-like melt(s) fractionally crystallized into the Nuevo Laredo Trend are also (except for Binda) "missing." At least, they are "missing" in monomict-eucrite form. Actually, both types of "missing" material might be abundantly available, albeit in the difficult-to-study form of small clasts among the howardites and/or the polymict eucrites (SMITH, 1982; DELANEY *et al.*, 1984c; IKEDA and TAKEDA, 1985; TAKEDA and MORI, 1985; TREIMAN and DRAKE, 1985). Unfortunately, except for SMITH (1982), none of these studies report an *mg* ratio or V content of a clast sample for which incompatible trace elements have also been determined.

6. CONCLUSIONS

The Stannern Trend eucrites (comprising at least Stannern and Bouvante) may have formed as primary partial melts. However, as originally suggested by STOLPER (1977), the Nuevo Laredo Trend eucrites (comprising at least Nuevo Laredo and Lakangaon) clearly are not primary, having formed out of residual liquids from fractional crystallization. Limited variations of incompatible element contents among the remaining "Main Group" of eucrites, despite considerable variations in *mg* ratio and V content, suggest that most Main Group eucrites belong to the Nuevo Laredo Trend; and few, if any, belong to a partial melting trend. In fact, if the high-*mg* eucrite Pomozdino (KVASHA and DYAKONOVA, 1972; KOLOSOV, 1976) represents a primary equilibrium partial melt, then no other eucrite could possibly be a primary partial melt derived from the same source composition.

Detailed comparison between a fractional crystallization model and the actual Nuevo Laredo Trend suggests that Lakangaon and Nuevo Laredo may be partly "cumulus" in origin, and not simply quenched melts. This model would imply that the parent melts of these meteorites had considerably lower *mg* ratios than the meteorites themselves. By occasionally forming rocks with far higher proportions of cumulus matter, such melts might have also produced cumulate eucrites such as Moama, Moore County, and Serra de Magé, all of which clearly (STOLPER, 1977) formed out of liquids with *mg* ratios far lower than any known eucrite.

The Nuevo Laredo Trend eucrites may be only part of a series of fractional crystallization products that began with diogenites (MASON, 1962). If so, the bulk MgO/(MgO + FeO) ratio of the eucrite parent asteroid is probably higher (~0.81) than generally estimated (~0.67). A similarly high MgO/(MgO + FeO) ratio is also consistent with suggestions (CLAYTON, 1977; WASSON, 1985; CLAYTON *et al.*, 1986) that pallasites and eucrites are derived from a single asteroid.

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- Millbillille Ma79
Moama (Mo) Ha78, Lo75, Mi79
Moore County (MC) Du63, Je70, Mc73, Mo78, Ni67, Ph70, Sc72, Te70, Ur53
Nagaria (Na) Ma79, Sc72
Nuevo Laredo (NL) Bi78, Du63, Du67, Ga70, Ha66, Mo65, Sc63, Sc72, Te70, Wa86
Padvarninkai Ko76, Ma79
Palo Blanco Creek (PBC) Di85, Ma79
Pasamonte (Pa) Bi78, Ch76, Du63, Du67, Ga70, Ha66, Je70, Mc73, Mi79, Ni67, Sc63, Sc72, Sh86, Te70, Un77, Ur53, VM69, Wä77
Peramiho (Pe) Sc72, Ur53
Pomozdino (Po) Ko76, Kv72
Serra de Magé (SM) Bi78, Je70, La72, Ma79, Mc73, Mo78, MS79, Pa78, Ph70, Sc72, Ur53
Sioux County (SC) Bi78, Ch76, Du63, Du67, Je70, La72, Mc73, Mi79, Ni67, Pa78, Sc72, Te70, VM69
Stannern (St) Bi78, Ch76, Du63, Du67, Ga70, Ha66, Je70, La72, Mc73, Mi79, Ni67, Pa81, Ph70, Sc64, Sc72, Sh86, Te70
Vetluga Kh82

* In parentheses: Abbreviations used as labels in the Figs

* Key to Appendix references:

- Bi78 = BIRCK J. L. and ALLÈGRE C. J. (1978) *Earth Planet. Sci. Lett.* **39**, 37-51.
Ch46 = CHIRVINSKY P. N. and SOKOLOVA A. I. (1946) *Meteoritika* **3**, 37-44.
Ch76 = CHOU C.-L. et al. (1976) *Proc. Lunar Planet. Sci. Conf. 7th*, 3501-3518.
De84 = DELANEY et al. (1984b).
Di85 = DICKINSON T. et al. (1985) *Chem. Erde* **44**, 245-257.
Du63 = DUKE (1963) (Data for Ti, which seem systematically low, are ignored.)
Du67 = DUKE and SILVER (1967).
Du78 = DUKE M. B. (1978) *Meteoritics* **13**, 443-448.
Ga70 = GAST et al. (1970).
Go79 = GOODING J. L. et al. (1979) *Lunar Planet. Sci. X*, 446-448.
Ha66 = HASKIN et al. (1966).
Ha78 = HAMET J. et al. (1978) *Proc. Lunar Planet. Sci. Conf. 9th*, 1115-1136.
Je70 = JEROME (1970).
Je71 = JEROME D. Y. et al. (1971) *Comptes Rend. Acad. Sci. Paris* **273**, 1657-1659.
Kh82 = KHARITONOVA V. YA. and BARSUKOVA L. D. (1982) *Meteoritika* **48**, 41-44.
Ko76 = KOLOSOV (1976) (Data for Hf, which seem systematically low, are ignored.)
Kv72 = KVASHA and DYAKONOVA (1972).
La72 = LAUL J. C. et al. (1972) *Geochim. Cosmochim. Acta* **36**, 329-345.
Lo75 = LOVERING J. F. (1975) *Meteoritics* **10**, 101-114.
Ma79 = MASON et al. (1979).
Ma83 = MASON (1983).
Ma86 = MASON B. and JAROSEWICH E. (1986) *Meteoritics* **21**, (in press).
Mc73 = MCCARTHY et al. (1973).
Mc74 = MCCARTHY et al. (1974).
Mi79 = MITTFELDELT D. W. (1979) *Geochim. Cosmochim. Acta* **43**, 1917-1935.
Mo65 = MORGAN and LOVERING (1965); also MORGAN and LOVERING (1964).
Mo73 = MORGAN J. W. and LOVERING J. F. (1973).
Mo78 = MORGAN et al. (1978).

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St77 = STOLP
Sy70 = SYME
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Te70 = TERA
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APPENDIX: REFERENCES FOR FIGURES 1, 5, 6, 7 AND 8

Meteorite (Abbrev.)*	References*
Adalia	Ma79
Aioun el Atrouss (AA)	Du78
ALHA81001 (81)	De84, Ma83, Wa86
Bereba (Be)	Bi78, Ch76, Je70, La72, Mc73, Mi79, Mo73, Pa78, Sc72, Ur53, Wä77
Binda (Bi)	Ch76, Je70, Mc73, Mi79, Ni67, Pa78, Sc72, Ur53
Bouvante	Pa81*
Cachari (Ca)	Je70, Mc73, Pa78, Sc72, Sc81, Sh86, Ur53
Camel Donga (CD)	Ma86, Pa86
Chervony Kut (CK)	Ch46, Go79, Je70, Pa78, Sc72
Emmaville	Ma79, Mo73, St77
Haraiya (Ha)	Je70, Ma79, Mc73, Ni67, Sc72
Ibitira (Ib)	Bi78, Mo78, Pa81, Wä74
Jonzac (Jo)	Bi78, Je70, La72, Ma79, Ph70, St77, Te70, Ur53, Wä77
Juvinas (Ju)	Du63, Du67, Ga70, Ha66, Je70, Je71, La72, Mc73, Mo73, Mo78, Na80, Ni67, Pa81, Ph70, Sc64, Sc72, Sh86, Te70, Ur53, Wä72
Kirbyville	Ma79
Lakangaon (La)	Ma79, Mc74, Wa86
Medanitos (Me)	Sc72, Sy70

- MS79 = MA M.-S. and SCHMITT R. A. (1979) *Meteoritics* **14**, 81-89.
- Na80 = NAKAMURA N. and MASUDA A. (1980) In *Proc. 5th Symp. Antarctic Meteorites* (ed. T. NAGATA), pp. 159-167. National Inst. Polar Res., Tokyo.
- Ni67 = NICHIPORUK W. *et al.* (1967) *Geochim. Cosmochim. Acta* **31**, 1911-1930. (Data for Fe, which seem systematically low, are ignored.)
- Pa78 = PALME H. *et al.* (1978) *Proc. Lunar Planet. Sci. Conf. 9th*, 25-57.
- Pa81 = PALME and RAMMENSEE (1981).
- Pa86 = PALME H. *et al.* (1986) *Meteoritics* **21** (in press).
- Ph70 = PHILPOTTS J. A. and SCHNETZLER J. C. (1970) *Proc. Apollo 11 Lunar Sci. Conf.*, 1471-1486; also SCHNETZLER J. C. and PHILPOTTS J. A. (1969) In *Meteorite Research* (ed. P. M. MILLMAN), pp. 206-216. Reidel.
- Sc63 = SCHMITT *et al.* (1963).
- Sc64 = SCHMITT R. A. *et al.* (1964) *Geochim. Cosmochim. Acta* **28**, 67-86.
- Sc72 = SCHMITT R. A. *et al.* (1972) *Meteoritics* **7**, 131-214.
- Sc81 = SCHMITT R. A. *et al.* (1981) private communication.
- Sh86 = SHIMIZU H. and MASUDA A. (1986) *Geochim. Cosmochim. Acta* **50**, 2453-2460.
- St77 = STOLPER E. (1977).
- Sy70 = SYMES R. F. and HUTCHISON R. (1970) *Mineral. Mag.* **37**, 721-723.
- Te70 = TERA F. *et al.* (1970) *Proc. Apollo 11 Lunar Sci. Conf.*, 1637-1656.
- Un77 = UNRUH D. M. *et al.* (1977) *Earth Planet. Sci. Lett.* **37**, 1-12.
- Ur53 = UREY H. C. and CRAIG H. (1953) *Geochim. Cosmochim. Acta* **4**, 36-82.
- VM69 = VON MICHAELIS H. *et al.* (1969) *Earth Planet. Sci. Lett.* **5**, 387-394.
- Wä72 = WÄNKE H. *et al.* (1972) *Proc. Lunar Sci. Conf. 3rd*, 1251-1268.
- Wä74 = WÄNKE H. *et al.* (1974) *Proc. Lunar Sci. Conf. 5th*, 1307-1335.
- Wä77 = WÄNKE H. *et al.* (1977) *Proc. Lunar Sci. Conf. 8th*, 2191-2213.
- Wa86 = Table 1 of this work.

* In the absence of any bulk-rock Mg datum for Bouvante, *mg* is estimated to be ~0.38 based on pyroxene compositions reported by DELANEY *et al.* (1984a).

The figures show data averaged (straightforward linear means) from all sources, giving equal "weight" to each analysis of a separate sample. Thus, each of the two Wa86 analyses of separate chips from Lakangaon was assigned a weight equal to that of the single Mc74 Lakangaon analysis. A few data of dubious accuracy were excluded from the averages; all instances where these exclusions might have significant impact on the figures are noted above (see Du63, Ko76 and Ni67). The recognized "cumulate" eucrites (STOLPER, 1977; MASON *et al.*, 1979) comprise Binda, Medanitos, Moama, Moore County, Nagaria, and Serra de Magé. A photocopy of the complete table of data, including a few additional references to analyses for limited sets of elements (*e.g.*, exclusively Rb and Sr), is available from the senior author.