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## A petrologic and isotopic study of lodranites: Evidence for early formation as partial melt residues from heterogeneous precursors

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**Abstract**—We have conducted petrologic, chemical, and isotopic studies of lodranites in an attempt to constrain their genesis. Lodran, Gibson, Y-791491, Y-791493, Y-74357, Y-8002, Y-75274, MAC 88177, LEW 88280, EET 84302, FRO 90011, and QUE 93148 are classified as lodranites. Lodranites and acapulcoites are indistinguishable on the basis of oxygen isotopic compositions but are distinct in average grain sizes of their mafic silicates, with lodranites being significantly coarser-grained. Lodranites exhibit a diverse range of petrologic and mineralogic features: they range widely in mafic silicate compositions (Fa<sub>3–13</sub>), plagioclase (0–11.4 vol%), Fe,Ni metal (0.5–20 vol%), and troilite (0.2–5.3 vol%) contents; and shock levels (S1–S4). They appear to have experienced high peak temperatures and rapid cooling in the temperature range recorded by metallographic cooling rates (i.e., 700–350°C). The only dated lodranite, Gibson, cooled to Ar closure temperatures at  $4.49 \pm 0.01$  Ga.

Lodranites formed from chemically and isotopically heterogeneous precursors in which the mineral and oxygen isotopic compositions were correlated. Heating of their parent body to temperatures between ~1050–1200°C resulted in formation of Fe,Ni-FeS and basaltic partial melts. Depletions of troilite and/or plagioclase in most lodranites testify to the removal of some of these partial melts, although melt migration was complex. Lodranites appear to have experienced a complex cooling history of slow cooling at high temperatures, followed by rapid cooling at intermediate temperatures, possibly related to breakup of the parent body. Lodranites were liberated from their parent body during 1–3 impact events, with most having cosmic ray exposure ages of 5.5–7 Ma. The acapulcoites are samples from the same parent body but were heated to lower temperatures and, thus, experienced lower degrees of partial melting. Copyright © 1997 Elsevier Science Ltd

### 1. INTRODUCTION

In recent years, considerable progress has been made through the study of meteorites in deciphering the igneous histories of differentiated asteroids (e.g., Taylor et al., 1993). We have studied acapulcoites (McCoy et al., 1996a) and here present the results of our studies of the closely related lodranites. Ever since the work of Bild and Wasson (1976), it has been recognized that Lodran is probably the residue of a partially molten region of an asteroid. With the recent recovery of a diverse suite of lodranites and acapulcoites (mostly from Antarctica), our understanding of the partial melting history of the acapulcoite-lodranite parent body has increased dramatically. We find that in contrast to acapulcoites, lodranites appear to have experienced significantly higher degrees of partial melting, including silicate partial melting, as well as melt migration. Thus, these rocks provide insights into the physical and chemical processes that took place during partial melting and differentiation in relatively small asteroidal parent bodies. The present paper addresses the properties and genesis of lodranites. A third paper (McCoy

et al., 1997) will address details of partial melting and melt migration on the acapulcoite-lodranite parent body.

Several criteria can be used to distinguish acapulcoites from lodranites, and authors have differed on the application of these criteria to establish these groupings. These criteria were discussed in detail by McCoy et al. (1996a) and are only briefly reiterated here. Both acapulcoites and lodranites have mafic silicates with compositions between those of enstatite and ordinary chondrites and oxygen isotopic compositions which cluster around  $\delta^{18}\text{O} \sim 3.5\text{‰}$  and  $\delta^{17}\text{O} \sim 0.75\text{‰}$ . Note that both the mineral and oxygen isotopic compositions of these groups are heterogeneous and, thus, are not useful for discrimination between acapulcoites and lodranites. A fundamental distinction between these two groups is that lodranites are coarse-grained whereas acapulcoites are fine-grained (McCoy et al., 1992a, 1993). As we show later, we suggest that this difference in grain size stems from a basic difference in their petrogenesis, and we use grain size for distinction between the groups. Some, but not all, lodranites have high contents of metallic Fe,Ni and, thus, have been called stony-irons by some previous workers (e.g., Prior, 1916). Based on our criteria, Lodran, Gibson, Yamato 791491, Yamato 791493, Yamato 74357, Yamato 8002, Yamato 75274, MAC 88177, LEW 88280, EET 84302, FRO 90011, and QUE 93148 are lodranites. We recognize that other workers (e.g., Zipfel and Palme, 1993) have utilized

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different criteria, particularly noting that lodranites are depleted in elements enriched in early partial melts (e.g., Al, Na, LREE). By this definition, some plagioclase-rich lodranites (e.g., Gibson, Y-8002, EET 84302) would be classified as acapulcoites. While we include these in our lodranite group, we recognize that these meteorites are, in many ways, transitional between the two groups, a topic we return to later in this paper.

## 2. SAMPLES AND ANALYTICAL TECHNIQUES

We studied ten of the twelve known lodranites (except Yamato 75274 and QUE 93148; the latter was only recently identified as a lodranite by Mason, 1995), including at least one sample from each of the paired samples. Pairing has been suggested for Yamato 8002 with Yamato 75274 (Yanai et al., 1984) and is evident for Yamato 791491 with Yamato 791493 from the mineralogical data of Nagahara and Ozawa (1986) and Hiroi and Takeda (1991), consistent with our data. The following thin sections were studied: Gibson, UH 192; Lodran, USNM 481-4; Yamato 791491, 61-2; Yamato 791493, 91-2; Yamato 74357, 62-3; Yamato 8002, 51-3; MAC 88177, 26; LEW 88280, 16; EET 84302, 12; and FRO 90011, 1 (UH = University of Hawaii; USNM = United States National Museum; Yamato = National Institute of Polar Research; MAC, LEW, and EET = Meteorite Working Group; FRO = European Meteorite Group).

Several of the meteorites we studied have only recently been recovered and/or recognized as lodranites. Gibson was found as a complete stone of 67.1 grams in October 1991. It was recovered by a prospector surveying a sandy area 150 m from a highway ~15 km north of Esperance, Western Australia (~33°41'S, 121°48'E). The main mass is in the possession of David New. FRO 90011 was recovered in the 1990–1991 Antarctic field season and originally classified by Folco (1992) as an acapulcoite, but is actually a lodranite (McCoy et al., 1993). EET 84302 was classified as a unique achondrite (Score and Lindstrom, 1990) but is a lodranite (McCoy et al., 1993).

Polished thin sections were studied in transmitted and reflected light. Mineral compositions were measured on Cameca Camebax and Cameca SX-50 electron microprobes at the University of Hawaii and corrections were made using a manufacturer supplied PAP ZAF program. Natural and synthetic minerals of well-known compositions were used as standards. The maximum dimensions of twenty-five mafic silicate grains in a single section were measured to determine the average grain sizes. Of these, twenty-three were randomly selected, and we also sought the largest and smallest grains in each meteorite. Inclusion of these grains did not significantly affect the average grain size. Olivines poikilitically enclosed in pyroxenes are not included in these measurements, since they could not have grown by the same process as other grains.

Modal analyses were conducted using optical point counting techniques in reflected light at a magnification of 200–400 $\times$ . At this magnification, plagioclase, Fe,Ni metal, troilite, and weathering products can be confidently discriminated from mafic silicates. Point counting of 500–1000 points per section covered the entire thin section. Uncertainty ( $\pm 2\sigma$ ) in the modal abundance for a phase which comprises 5–10% of the rock is typically  $\pm 0.2$ –0.3% and for phases which comprise 0.5–1%, typically  $\pm 0.7$ –1%. As we show later, errors in determining the abundances within individual lodranites are small compared to differences in mineral abundances between lodranites. Unfortunately, meteorite slices on many thin sections of Antarctic lodranites are very small (e.g., Y-8002, 51-3 has an area of ~8 mm<sup>2</sup>). While it is possible that these small samples may not be representative of a larger volume of rock, we show later that this is unlikely given the extensive studies of the 1 kg Lodran meteorite, the uniform enrichments of metal and depletions of plagioclase found by all workers, and correspondence of bulk chemistry calculated from modal data and mineral chemistry with that directly measured.

Polished thin sections were etched with a mixture of 1% nitric acid and alcohol for 20–30 sec to reveal the kamacite-taenite structure for

the purpose of determining metallographic cooling rates. We did not etch Gibson, which is highly weathered, and MAC 88177, which is highly shocked, since these meteorites would not yield reliable metallographic cooling rates. Taenite was found in Lodran, LEW 88280 and Yamato 791491, but not in Yamato 8002, Yamato 74357, EET 84302, and Yamato 791493. We used the method of Wood (1967) and the revised cooling rate curves of Willis and Goldstein (1981). Taenite grains typically have rims of strongly zoned taenite 1–20  $\mu$ m in thickness, surrounding regions of plessite. Due to their small sizes and low central Ni contents, Ni was measured on etched taenite grains. This may introduce small, albeit insignificant errors considering the uncertainties in the absolute metallographic cooling rates. Measurements were made in the approximate centers of semi-equant taenite grains, the majority of which are rimmed by kamacite. Point selection is based on identification of the minimum Ni content, using 1 sec counting times on an LIF crystal centered on the Ni K $\alpha$  peak. These taenite grains have an average P content of <0.04 wt% and, thus, P content should not add uncertainties to the rates.

Measurements of oxygen isotopic compositions were made on Gibson, Yamato 8002, Yamato 75274, Yamato 74357, EET 84302, Lodran LEW 88280, and MAC 88177 (see Table 2), using techniques described by Clayton and Mayeda (1963, 1983). Weathering contaminants are abundant in some lodranites and, therefore, samples of Yamato 74357, Yamato 8002, LEW 88280, and EET 84302 were acid-washed for 2–3 min with 6M HCl at 70°C prior to isotopic analysis. Data reported by Clayton et al. (1984, 1992) on lodranites were preliminary and have been adjusted slightly. Oxygen isotopic compositions of FRO 90011 were collected using a laser fluorination technique (Franchi et al., 1992).

A sample of Gibson (44 mg) was irradiated in the Los Alamos Omega reactor (irradiation constant  $J = 0.03368 \pm 0.00026$ ). Step-wise temperature extractions were made in a high vacuum RF furnace equipped with a thermocouple, and the argon isotopes released were measured on a mass spectrometer. Corrections were made for extraction blanks, radioactive decay, and reactor-produced interferences. Errors reported for individual ages include uncertainties in these corrections plus analytical uncertainty in measuring the <sup>40</sup>Ar/<sup>39</sup>Ar ratios in the samples and the hornblende neutron flux monitors. Errors in individual ages do not include an additional uncertainty of about  $\pm 1\%$  due to the accuracy of determining the flux and the absolute age of the hornblende.

Noble gases were analyzed by mass spectrometry in three different chips of Gibson using techniques described by Graf et al. (1990a). Argon was separated from Kr and Xe by freezing the latter two gases in a charcoal trap at –90 to –100°C. In two samples, the gases were extracted in a single temperature step at 1800°C, and in the other (73.9 mg), we tried to remove possibly loosely bound atmospheric noble gases in a first step at about 300°C. This was not successful, however, since the 300°C step contains less than 1% of either noble gas, including Kr and Xe, which clearly are predominantly atmospheric; the data of this step are, therefore, not given.

## 3. RESULTS

Lodranites have been the subject of numerous publications, including broader reviews (e.g., Nagahara, 1992; Takeda et al., 1994; Mittlefehldt et al., 1996). Here, we focus on those properties of lodranites which help unravel their origins, and reference will be made to previous work as appropriate. We will also contrast the properties of lodranites with those of acapulcoites (McCoy et al., 1996a). Relevant data on lodranites are summarized in Tables 1 and 2.

### 3.1. Petrography and Mineral Compositions

#### 3.1.1. Texture

Lodranites are coarse-grained, equigranular rocks and consist mostly of low-Ca pyroxene and olivine, minor amounts of chromite, troilite, and chromian diopside, vari-

TABLE 1. Modes and mineral compositions of lodranites.

	Olivine		Low-Ca Pyroxene			High-Ca Pyroxene		Plagioclase		Fe,Ni (vol.%)	FeS (vol%)	Weathering (vol.%)
	Fa	Rev. Zone?†	Fs	Rev. Zoning? FeO	CaO	Fs	Wo	vol. %	An			
Gibson	3.1 (1)	N (1)	5.8 (1)	Y (1)	W (1)	4.3	45.1 (1)	6.0 (1)	18.1 (1)	0.2	0.5	19.4 (1)
{ Yamato 8002	3.5 (2)	N (2)	3.7 (2)	N (2)	Y (2)	1.1	45.3 (11)	10.3 (2)	30.9 (2)	6.3	0.3	2.8 (2)
{ Yamato 75274	3.9 (3)	N (3)	4.0 (3)	N (3)	Y (3)	1.4	44.3 (11)	3.4 (11)	28 (11)	19.6	0.2	- (11)
{ Yamato 74357	7.9 (4)	N (4)	13.8 (4)	Y (4,9)	Y (9)	6.3	42.8 (11)	Tr. <sup>2</sup> (1,11)	15 (4)	9.6	2.4	- (11)
EET 84302	8.4 (1)	W (1)	8.3 (1)	N (1)	Y (1)	3	42 (13)	11.4 (1)	23 (13)	13.1	0.1	2.8 (1)
FRO 90011	9.4 (12)	Y (1)	12.6 (12)	Y (1)	Y (1)	5.2	44.9 (1)	Tr. (1)	12.3 (12)	7.7	3.0	4.7 (1)
{ Yamato 791491	~12 (4)	Y (4)	11.8 (4)	N (4)	Y (4)	5.2	43.9 (4)	0.25 (4)	18.2 (4)	4.2	5.3	4.5 (1)
{ Yamato 791493	11.6 (5)	Y (5)	12.2 (5)	W (10)	Y (10)	5.0	44.5 (10)	1.6 (11)	18.3 (10)	9.0	3.0	- (11)
Lodran	12.6 (6)	Y (6)	13.8 (6)	N (6)	W (1)	6.4	43 (6)	0 (6)	-	20	2.7	- (6)
MAC 88177	13.3 (7)	N (7,8)	12.2 (10)	N (7,8)	Y (1)	6	42 (7,8)	0 (7)	-	0.5	1.9	- (8)
LEW 88280	12.9 (1)	N (1)	11.5 (1)	N (1)	W (1)	4.8	45.0 (1)	0 (1)	-	10.3	2.5	5.5 (1)

† Reverse Zoning: Y - Yes, N - No (Homogeneous), W - Weak.

<sup>2</sup> Previous analyses report no plagioclase. We identified a 2.2 mm long plagioclase (PTS ,62-3).

( Possibly paired meteorites.

References: (1) This Work (2) Nagahara, 1992 (3) Mori et al., 1984 (4) Hiroi and Takeda, 1991 (5) Yanai and Kojima, 1982 (6) Bild and Wasson, 1976 (7) Takeda et al., 1994 (8) Prinz et al., 1991 (9) Miyamoto and Takeda, 1991 (10) Nagahara and Ozawa, 1986 (11) Yanai et al., 1984 (12) Folco, 1992 (13) Mason, 1986.

able amounts of Fe,Ni metal, plagioclase, and hydrated iron oxides of terrestrial origin, and traces of phosphates and schreibersite. Numerous 120° triple junctions between grains are indicative of extensive recrystallization. Average mafic silicate grain sizes for most lodranites are ~500–600 μm (Table 2). Yamato 8002 has a larger average grain size (700 μm) but only thirteen grains were present in the section examined and, thus, this average has a large uncertainty. EET 84302 is finer-grained (340 μm) than most lodranites but significantly coarser-grained than most acapulcoites (150–200 μm; McCoy et al., 1996a). Figure 1 illustrates the textures of Acapulco and Lodran at the same scale, demonstrating the coarser-grained nature of the latter.

### 3.1.2. Shock effects

None of the lodranites are brecciated, and many are unshocked (shock stage S1 of Stöffler et al., 1991), such as Lodran, EET 84302, FRO 90011, Yamato 791491, Yamato 791493, and LEW 88280. Gibson is also unshocked (S1

but exhibits numerous semi-parallel veins of hydrated iron oxides of terrestrial origin cutting silicates. There is no evidence to suggest that these veins are related to shock. One olivine grain in Yamato 8002 exhibits undulatory extinction. However, the small number of grains present in our thin section (S1-3) prevents definitive classification, and we, therefore, consider Yamato 8002 to be unshocked to very weakly shocked (S1–S2). Yamato 74357 is weakly shocked (S3), exhibiting undulatory extinction of olivines. It also has μm-sized troilite and, rarely, metal grains decorating planar fractures in mafic silicates. MAC 88177 has olivines with multiple, intersecting sets of planar fractures and strong mosaicism. Shock veins of whole-rock melt, including finely-dispersed opaques, crosscut the specimen. Large (>100 μm) troilite grains are polycrystalline and Fe,Ni metal-troilite composite grains are finely intergrown, indicative of rapid post-shock cooling. These features suggest that the rock is moderately shocked (S4) and that the shock event took place after recrystallization and partial melting. Some of these features in Yamato 74357 and MAC 88177 were

TABLE 2. Petrologic, chemical and isotopic data for lodranites.

	Grain Size (μm)	Two-Px (°C)†	Shock Stage	δ <sup>18</sup> O (‰)	δ <sup>17</sup> O (‰)	Δ <sup>17</sup> O (‰)	Cooling Rate		Cosmic-Ray Exposure Age* (Ma)
							°C/Myr	Method	
Gibson	540	1020	S1	3.57	0.53	-1.33 (1)			6.0-6.4 (1)
{ Yamato 8002	700	1030	S1-2	3.85	0.51	-1.49 (1)			
{ Yamato 75274	-	1080		2.52	0.23	-1.08 (1)			
{ Yamato 74357	580	1100	S3	3.44	0.48	-1.31 (1)	1.5x10 <sup>6</sup>	Diffusion	(7) 17.5-20.5 (8)
EET 84302	340	1150	S1	3.31	0.53	-1.19 (1)			
FRO 90011	540	1020	S1	3.73	0.98	-0.96 (4)			5.9-6.4 (6)
{ Yamato 791491	560	1060	S1	-	-	-	~10 <sup>3</sup>	MCR**	(1) 6.1-6.6 (6)
{ Yamato 791493	570	1030	S1	-	-	-	~10 <sup>3</sup>	Chromite Size	(5)
Lodran	580	1070	S1	3.41	0.92	-0.85 (3)	~10 <sup>5</sup>	MCR**	(1) 4.0-4.3 (6)
MAC 88177	620	1140	S4	3.52	0.60	-1.23 (1)			7.2-7.5 (6)
LEW 88280	610	1020	S1	3.40	0.78	-0.99 (2)	~10 <sup>3</sup>	MCR**	(1) 4.0-4.3 (6)

† Two-pyroxene closure temperatures calculated from pyroxene compositions given in Table 1.

\* Reference indicates source of data used in calculation.

\*\* Metallographic cooling rate method.

( Possibly paired meteorites.

References: (1) This Work (2) Clayton et al., 1992 (3) Clayton et al., 1984 (4) Franchi et al., 1992 (5) Nagahara and Ozawa, 1986 (6) Eugster and Weigel, 1993 (7) Miyamoto and Takeda, 1991 (8) Takaoka et al., 1993

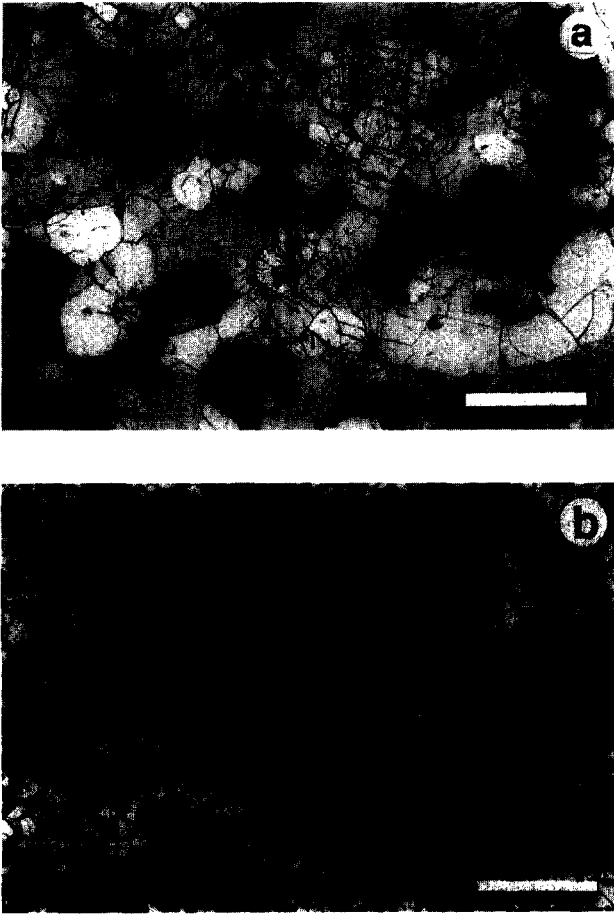


Fig. 1. Plane polarized, transmitted light photomicrographs of Lodran (a) and Acapulco (b) to the same scale (bar = 1 mm). Although acapulcoites and lodranites have similar mineralogies and mineral and oxygen isotopic compositions, they exhibit markedly different grain sizes. Lodranites experienced silicate partial melting and became coarse-grained, whereas silicates in acapulcoites did not melt and, hence, acapulcoites are finer-grained.

noted by Takeda et al. (1994), although not specifically attributed to shock.

### 3.1.3. Compositions of mafic silicates

Lodranites exhibit a broad range of mafic silicate compositions (Tables 1, 3). Olivine ranges from  $Fa_{3.1}$  (Gibson) to  $Fa_{13.3}$  (MAC 88177). Individual olivine grains in a number of lodranites exhibit reverse zoning (FeO decreasing from core to rim), consistent with reduction. Reverse zoning is prominent in lodranites of intermediate olivine composition ( $Fa_{8.4-12.6}$ ). FRO 90011, Yamato 791491, Yamato 791493, and Lodran have strong reverse zoning, with cores of grains enriched in FeO by up to 1 wt% relative to the rims. Weak reverse zoning is observed in EET 84302, with cores higher by  $\sim 0.5$  wt% relative to rims. No zoning is observed in Gibson, Yamato 8002, Yamato 75274, Yamato 74357, MAC 88177, and LEW 88280.

Pyroxene Fs (molar Fe/(molar Fe + Mg + Ca)) approximately correlates with Fa of co-existing olivines. Pyroxenes also show a broad range of compositions ( $Fs_{3.7-13.8}$ ) and

reverse Fs zoning. In general, lodranites with higher Fs contents in low-Ca pyroxene than Fa contents in olivine (e.g., Gibson,  $Fa_{3.1}$ ,  $Fs_{5.8}$ ) also have inverse Fs zoning in the pyroxene. This is the case for Gibson, Yamato 74357, FRO 90011, and Yamato 791493. However, Bild and Wasson (1976) and Papike et al. (1995) found no reverse zoning in low-Ca pyroxenes of Lodran ( $Fa_{12.6}$ ,  $Fs_{13.8}$ ), whereas Takeda et al. (1994) note both exsolution and zonation in pyroxenes, although the zoning is slight. In many other lodranites (e.g., Yamato 8002, Yamato 75274, EET 84302, Yamato 791491), olivine and low-Ca pyroxene have nearly equal Fa and Fs contents. The two lodranites with the highest Fa (MAC 88177, LEW 88280) have Fa 1.1–1.4 mol% higher than Fs. Ratios of Fa/Fs vary from 0.53 (Gibson) to 1.12 (LEW 88280). All lodranites exhibit reverse CaO zoning in orthopyroxene. In addition, many lodranites show very fine scale ( $\sim 1-5 \mu\text{m}$ ) lamellae of clinopyroxene in orthopyroxene, and vice versa (e.g., Yamato 791491, Nagahara and Ozawa, 1986; MAC 88177, Takeda et al., 1994).

High-Ca pyroxene is a volumetrically minor phase ( $< 3$  vol%; Takeda et al., 1994) and, thus, compositional data are frequently from a single grain (Table 1). All contain significant amounts of  $\text{Cr}_2\text{O}_3$  (0.74–1.74 wt%) (Table 3) and, hence, are properly referred to as chromian diopsides. Both our studies and published analyses (e.g., Nagahara and Ozawa, 1986) reveal that high-Ca pyroxenes are also reversely zoned in Fs.

### 3.1.4. Equilibration temperatures

The coexistence of low-Ca pyroxene and chromian diopside in lodranites allows estimation of two-pyroxene equilibration temperatures. The use of a two-pyroxene geothermometer in zoned pyroxenes which may have experienced continuous cooling (e.g., Nagahara, 1992) could introduce errors. It is important to note that these are closure temperatures and, thus, reflect minimum peak temperatures for these rocks. We, therefore, use these data only to constrain these minimum peak temperatures. We have used the transfer equations of Kretz (1982) to calculate two-pyroxene equilibration temperatures utilizing the average compositions of low-Ca and high-Ca pyroxene as determined with the electron microprobe. Uncertainties resulting from pyroxene heterogeneities are estimated to be  $\pm 50^\circ\text{C}$ , using  $\pm 1\sigma$  compositional variabilities in the average compositions. Errors resulting from derivation of the Kretz (1982) equations are  $\pm 60^\circ\text{C}$ , and the effect of  $\text{Cr}_2\text{O}_3$  on the equilibration temperatures is unknown. We also recognize that a variety of other two-pyroxene thermometer formulations are available (e.g., Lindsley and Frost, 1992; Frost and Lindsley, 1992), but we have not applied these because these authors also did not explicitly consider the possible influence of  $\text{Cr}_2\text{O}_3$  on equilibration temperatures.

We find that two-pyroxene equilibration temperatures range from  $1020^\circ$  to  $1150^\circ\text{C}$ , with the highest for EET 84302 ( $1150^\circ\text{C}$ ) and MAC 88177 ( $1140^\circ\text{C}$ ) (Table 2). We did not observe chromian diopside in EET 84302 (MWG PTS ,12) but use the analysis by Mason (1986) of chromian diopside given in Table 1. The equilibration temperatures of the paired meteorites Yamato 8002 and Yamato 75274 differ by

TABLE 3. Compositions of silicate minerals from lodranites analyzed in this work.

	Olivine			Orthopyroxene			Clinopyroxene			Plagioclase
	Gibson	BET 84302	LEW 88280	Gibson	BET 84302	LEW 88280	Gibson	FRO 90011	LEW 88280	Gibson
SiO <sub>2</sub>	42.0 <i>0.37</i>	40.4 <i>0.25</i>	39.8 <i>0.19</i>	57.9 <i>0.38</i>	57.2 <i>0.28</i>	55.6 <i>0.72</i>	53.6 <i>0.06</i>	53.9 <i>0.39</i>	52.1 <i>0.39</i>	64.3
TiO <sub>2</sub>	b.d.	n.d.	b.d.	0.16 <i>0.05</i>	n.d.	0.12 <i>0.03</i>	0.41 <i>0.02</i>	n.d.	0.31	n.d.
Al <sub>2</sub> O <sub>3</sub>	b.d.	n.d.	b.d.	0.42 <i>0.10</i>	n.d.	0.46 <i>0.04</i>	1.00 <i>0.09</i>	n.d.	1.07 <i>0.13</i>	22.6
Cr <sub>2</sub> O <sub>3</sub>	0.07 <i>0.03</i>	n.d.	b.d.	0.47 <i>0.05</i>	n.d.	0.50 <i>0.12</i>	1.74 <i>0.12</i>	n.d.	1.60	n.d.
FeO	3.10 <i>0.11</i>	8.17 <i>0.11</i>	12.2 <i>0.20</i>	3.86 <i>0.90</i>	5.67 <i>0.13</i>	7.88 <i>0.19</i>	2.62 <i>0.20</i>	3.31 <i>0.11</i>	2.98	0.28
MnO	0.39 <i>0.03</i>	0.62 <i>0.03</i>	0.48 <i>0.03</i>	0.51 <i>0.06</i>	0.48 <i>0.07</i>	0.54 <i>0.05</i>	0.30 <i>0.03</i>	0.37	0.34	n.d.
MgO	54.8 <i>0.21</i>	50.2 <i>0.31</i>	46.3 <i>0.24</i>	34.4 <i>0.75</i>	34.5 <i>0.13</i>	32.9 <i>0.41</i>	17.4 <i>0.41</i>	17.7	17.7	b.d.
CaO	b.d.	b.d.	b.d.	1.09 <i>0.20</i>	0.96 <i>0.12</i>	1.32 <i>0.36</i>	21.5 <i>0.60</i>	22.1	22.0	3.87 <i>0.13</i>
Na <sub>2</sub> O	n.d.	n.d.	n.d.	b.d.	n.d.	b.d.	0.65 <i>0.02</i>	n.d.	0.71	9.27 <i>0.12</i>
K <sub>2</sub> O	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0.66 <i>0.06</i>
Total	100.36	99.39	98.78	98.81	99.06	99.32	99.22	97.38 <sup>1</sup>	98.81	100.98
N	20	10	10	10	10	10	3	1	1	10
Fa	3.1	8.4	12.9	--	--	--	--	--	--	--
Fs	--	--	--	5.8	8.3	11.5	4.3	5.2	4.8	--
Wo	--	--	--	2.1	1.8	2.5	45.1	44.9	45.0	--
An	--	--	--	--	--	--	--	--	--	18.1
Or	--	--	--	--	--	--	--	--	--	3.7

N= Number of analyses; b.d. = below detection; n.d. = not determined  
<sup>1</sup>Low totals reflect incomplete analysis of all oxides.  
 Italicized figures are 1σ of compositional variability; not given for averages of < 3 analyses.

~50°C, consistent with our estimated uncertainties owing to zoning. Although lodranites exhibit a range in temperatures, no obvious correlation exists between these and other parameters (e.g., mafic silicate compositions, plagioclase contents).

3.1.5. Plagioclase composition, abundance, and morphology

Plagioclase plays a particularly important role in understanding the genesis of lodranites, in large part because of the variability it exhibits in composition, abundance, and morphology between different meteorites. Compositions range from An<sub>12.3-30.9</sub> (Table 1; Fig. 2), but we find no

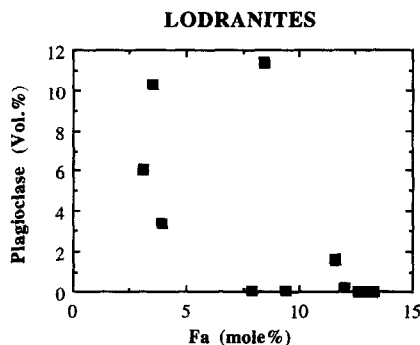


Fig. 2. Modal plagioclase contents of lodranites (vol%) vs. Fa (mol%) of olivines. Except for EET 84302, lodranites with low Fa tend to have higher plagioclase abundances, whereas those with high Fa have low plagioclase abundances.

correlation with mafic silicate compositions or plagioclase abundances, although Nagahara (1992) suggested that such a correlation existed, based on limited data for four lodranites (Yamato 8002, Yamato 75274, Yamato 791493, Yamato 74357). Lodranites also vary widely in plagioclase contents (0–11.4 vol%, Table 1), with a tendency for those with high contents of Fa in olivine (Fa<sub>11.6-13.3</sub>) to have very low plagioclase contents and those with lowest olivine Fa contents (Gibson and Yamato 8002) to have high plagioclase contents (6.0–10.3 vol%; Fig. 2). The latter are only slightly less than or equal to those typically found in ordinary chondrites (9.6–10.3 wt%; McSween et al., 1991). EET 84302 does not follow this trend and has 11.4 vol% plagioclase, despite having intermediate mafic silicate compositions (Fa<sub>8.4</sub>, Fs<sub>8.3</sub>).

Bulk chemical analyses are available for some, but not all, lodranites. Our modal estimates of plagioclase content agree well with plagioclase contents calculated from bulk chemical analyses of lodranites. Fukuoka et al. (1978) reported that Lodran is highly depleted in Al, Ca, Na, and K, consistent with the absence of plagioclase in the mode. Zipfel and Palme (1993) confirmed the highly plagioclase-depleted nature of MAC 88177 (0% modal plagioclase) and FRO 90011 (trace). Furthermore, Mittlefehldt et al. (1996) also confirmed the highly plagioclase-depleted nature of LEW 88280 (0%) and MAC 88177 (0%), while demonstrating that EET 84302 contains near-chondritic abundances of Na and Ca, consistent with our modal analysis of 11.4 vol% plagioclase. The good agreement in plagioclase content determined by modal analyses and calculated from bulk chemical analyses suggests that differences in plagioclase modal



Fig. 3. Photomicrograph in transmitted light with crossed polars of Yamato 8002. Plagioclase (pl; striated phase) is interstitial to, and sometimes surrounds, mafic silicates (si) and Fe,Ni metal (m), suggesting that it crystallized from an interstitial, basaltic melt. Scale bar = 250  $\mu\text{m}$ . The dark area to the lower right is epoxy.

abundances between lodranites are real and do not simply reflect nonrepresentative sampling of heterogeneous rocks. In particular, numerous studies of different samples of the 1 kg Lodran mass have consistently shown the lack of plagioclase. Clearly, this rock does not contain plagioclase-rich areas comparable to EET 84302.

Plagioclase often occurs as mm-sized interstitial grains which surround multiple mafic silicate and opaque mineral grains (Fig. 3). The best example of this type is in Yamato 8002, although a single 2.2 mm long plagioclase crystal observed by us in Yamato 74357 (PTS ,62-3) has the same texture. Plagioclase grains in Gibson and EET 84302 often surround single mafic silicate grains and pinch between multiple mafic silicates. In Yamato 791493, which contains only 1.6 vol% plagioclase, mm-long plagioclase grains pinch between mafic silicates, with multiple grains in contact and surrounding mafic silicates.

Lodran, MAC 88177, and LEW 88280 are free of plagioclase but contain an Al-rich phase first noted in Lodran by Bild and Wasson (1976) that is close to stoichiometric  $(\text{K},\text{Na})\text{AlSi}_3\text{O}_{12}$ . Prinz et al. (1991) report the occurrence of a  $\text{SiO}_2$ -rich feldspathic glass of unknown origin along grain boundaries in MAC 88177. Prinz et al. (1978) argue that the phase in Lodran was trapped melt, whereas Bild and Tallant (1984) concluded that it is a crystalline mineral, based on Raman spectroscopy. The origin of this phase thus remains enigmatic.

### 3.1.6. Opaque mineral occurrences and abundances

Metallic Fe,Ni and troilite are present in all lodranites and occur predominantly as relatively large (typically 0.5–1 mm maximum dimension; 0.1–2.6 mm range) interstitial grains. Metallic Fe,Ni and, less frequently, Fe,Ni metal-troilite intergrowths or pure troilite also occur as volumetrically insignificant, 10–20  $\mu\text{m}$ -sized blebs in the centers (but not the rims) of mafic silicate grains, primarily orthopyroxene. Similar occurrences have been observed in acapulcoites (e.g.,

McCoy et al., 1996a). A third occurrence is in association with the enigmatic Al-rich material in Lodran, which Prinz et al. (1978) noted to contain some Fe-rich phase. We observed that the Fe-rich material consists of irregular,  $<5 \mu\text{m}$ -sized grains of Fe-rich metal and troilite, intergrown with a darker (in reflected light), silicate material of uncertain composition and structure (Fig. 4). Due to its fine grain size and intimate intergrowths, we were unable to obtain reliable quantitative analyses of this material. TEM studies could elucidate these compositions and contribute to our understanding of their formation, but such analyses were outside the scope of this study. This material occurs mostly along grain boundaries and less frequently as diffuse areas at the edges of, and extending into, mafic silicate grains. In addition to the  $<5 \mu\text{m}$ -sized grains of Fe,Ni metal and troilite, distinct veins of Fe,Ni metal and troilite  $\leq 2 \mu\text{m}$  wide and  $\leq 160 \mu\text{m}$  in length also occur, with each phase forming segments of these veins. A similar occurrence is noted in the paired meteorites Yamato 791491 and 791493 and in FRO 90011, where  $\mu\text{m}$ -sized irregular blebs of troilite are observed at grain boundaries. This material is also intergrown with an unknown silicate phase which, while not darker in reflected light, does have low birefringence in transmitted light and crossed polars. The origin of these troilite-silicate intergrowths is uncertain, but it may be related to reduction of mafic silicates as the result of interaction with a S-rich fluid. Papike et al. (1995) previously suggested that mafic silicate zoning in Lodran was due to interaction with a S-rich fluid, rather than a carbon-rich material, as in the ureilites. We note that the four lodranites (Lodran, Yamato 791491, Yamato 791493, FRO 90011), in which we observed these intergrowths, also show the most dramatic inverse FeO zoning in olivine.

Lodranites contain variable abundances of Fe,Ni metal (0.5–20 vol%; Table 1). This appears not to be an artifact due to terrestrial weathering, since many of the meteorites in Table 1 (except Gibson) are relatively fresh. No correla-



Fig. 4. Reflected light photomicrograph of Lodran. A  $\mu\text{m}$ -sized intergrowth of troilite and a  $\text{SiO}_2$ -rich material occurs at the edges of large metal (m) and troilite (tr) grains and along silicate (si) grain boundaries. Scale bar = 50  $\mu\text{m}$ .

tion is observed between metal abundances and mafic silicate compositions.

Lodranites also contain variable abundances of troilite (0.1–5.3 vol%). The apparent depletion of troilite (0.5%) in Gibson is undoubtedly an artifact due to its highly weathered nature (19.4% terrestrial weathering products), whereas the enrichment of troilite in Yamato 791493 (5.3%) may be the result of a bias due to extensive plucking of the silicates in the thin section studied. Other lodranites exhibit only modest effects of terrestrial weathering. Some lodranites (e.g., the paired Yamato 8002 and Yamato 75274) appear to be quite depleted in troilite relative to likely chondritic precursors, while others may have troilite contents only slightly less or comparable to ordinary chondrites (3.6–7.2 wt%; Keil, 1962). This range in troilite abundances is supported by measurements of S and Se in lodranites by Dreibus et al. (1995). These authors found 2.00–2.56 wt% S and 8.85–9.76 ppm Se in ordinary chondrites. In comparison, the lodranites FRO 90011 (2.42% S, 5.85 ppm Se) and MAC 88177 (1.49% S, 6.14 ppm Se) were inferred to be only slightly depleted in troilite, consistent with modal analyses of FRO 90011 (3.0 vol% troilite) and MAC 88177 (1.9 vol% troilite). Gibson is severely depleted in S and Se, which Dreibus et al. (1995) also attributed to its extensive terrestrial weathering. Mittlefehldt et al. (1996) used the Se/Co ratio for LEW 88280, MAC 88177, EET 84302, and FRO 90011 to examine the removal of troilite. These authors found that LEW 88280, MAC 88177, and FRO 90011 have approximately the same Se/Co ratio as acapulcoites, consistent with moderate to high modal troilite abundances. However, EET 84302 has a much lower Se/Co ratio, consistent with its 0.1 vol% modal troilite content. As in the case of plagioclase, bulk chemical analyses for S and Se are broadly

consistent with modal data and suggest that the wide range of troilite abundances in lodranites is not due to poor sampling but is real.

### 3.1.7. Cooling rates

Previous work on lodranites has generated a wide range of cooling rates. Cooling rates on several lodranites by a variety of methods have yielded rapid cooling rates at moderate temperatures. Bild and Wasson (1976) used the metallographic method of Wood (1967) to derive a rate of  $\sim 10^4$ °C/Myr for Lodran in the range of 600–400°C. Miyamoto and Takeda (1991) determined a rate of  $1.5 \times 10^6$ °C/Myr for Yamato 74357, based on modeling of Fe-Mg and Ca zoning in orthopyroxene in the range of 1000–600°C. The presence of CaO zoning of comparable magnitude in other lodranites also suggests rapid cooling. The presence of high-Ca pyroxene lamellae in low-Ca pyroxene and vice versa might provide further constraints on the thermal history of lodranites, although we are unaware of any such studies (Takeda et al., 1994). Nagahara and Ozawa (1986) derived a rate for Yamato 791493 of  $\sim 10^3$ °C/Myr between 800–600°C, based on spinel grain size and spinel-olivine equilibration temperatures.

Slower cooling rates have been inferred at lower temperatures. Bild and Wasson (1976) calculated a low temperature cooling rate of  $\sim 10$ –30°C/Myr for Lodran, using the method of Short and Goldstein (1967) which involves the measurement of maximum Ni content at the edge of taenite grains. Prinz et al. (1978) determined a rate based on size and composition of schreibersite lamellae of 10°C/Myr using the technique of Hewins and Goldstein (1977). J. I. Goldstein (pers. commun.) suggested that the metallographic

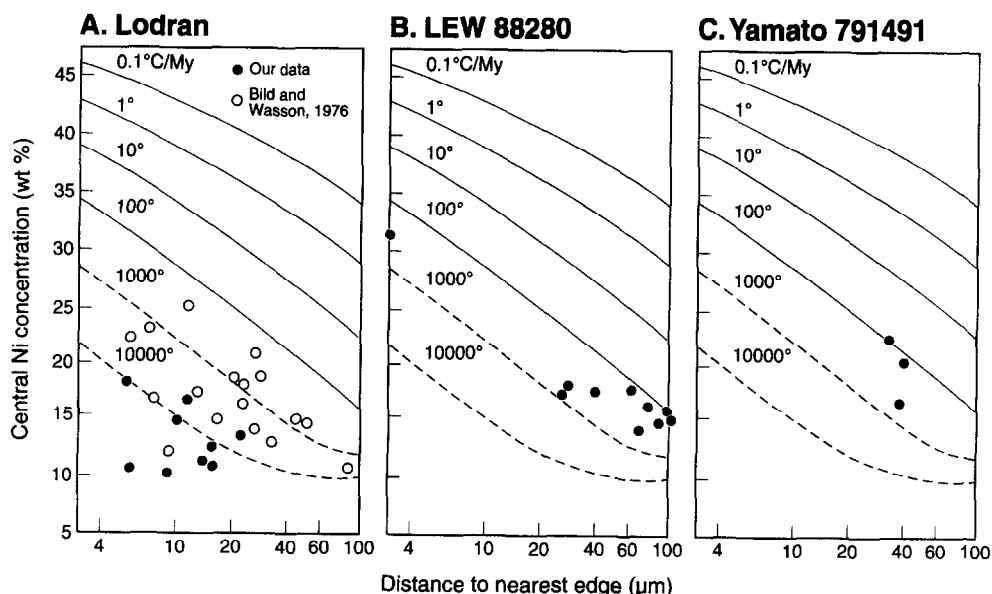


Fig. 5. Plots of central Ni concentration vs. distance to nearest edge for taenite grains in lodranites. Curves for cooling at 0.1–100°C/Myr from Willis and Goldstein (1981). Curves for 1000–10000°C/Myr (dashed lines) are approximate. Our data suggest a rate of  $\sim 10^4$ – $10^3$ °C/Myr for Lodran, whereas those of Bild and Wasson (1976) suggest  $\sim 10^4$ °C/Myr. Our data suggest that LEW 88280 cooled at a rate of  $\sim 10^3$ °C/Myr and Yamato 791491 at  $\sim 10^3$ °C/Myr.

techniques used by Bild and Wasson (1976) and Prinz et al. (1978) may have been inappropriate for use in the lodranites, which contain high-Ni metal. Thus, evidence for slow cooling at low temperature is, at best, tenuous.

We measured metallographic cooling rates for those lodranites that contain taenite (Lodran, LEW 88280, Yamato 791491; Fig. 5). Our data for Lodran scatter but suggest a rate of  $\sim 10^4$ – $10^5$ °C/Myr. This is somewhat faster than that found by Bild and Wasson (1976) ( $\sim 10^4$ °C/Myr) using the same technique, although both groups find fast rates for the temperature interval of  $\sim 700$ – $400$ °C, to which metallographic methods apply. We also find fast rates ( $\sim 10^3$ °C/Myr) for LEW 88280 and Yamato 791491, although we detected only three taenite grains (35–40  $\mu\text{m}$  diameter) in the latter, resulting in a greater uncertainty in the cooling rate estimate.

### 3.2. Oxygen Isotopic Compositions

Oxygen isotopic compositions of lodranites are similar to those of acapulcoites but differ from those of IAB and IIICD irons (the most abundant groups of silicate-bearing iron meteorites; Choi et al., 1995), winonaites, and ureilites (Table 2; Fig. 6). Clayton et al. (1992) discussed possible relationships of acapulcoites and lodranites with other meteorite groups (e.g., Kakangari, CR chondrites, a group of carbonaceous chondrites; Kallemeyn et al., 1994) but considered similarities in oxygen isotopic compositions fortuitous. Thus, acapulcoites and lodranites are from a different parent body than the other meteorite groups. Acapulcoites and lodranites cannot be distinguished from one another on the basis of oxygen isotopes but can be distinguished based on texture and composition (McCoy et al., 1996a). Clayton et al. (1992) also noted that acapulcoites and lodranites, as a group, exhibit variations beyond what could be attributed to mass-dependent fractionation and analytical uncertainty.

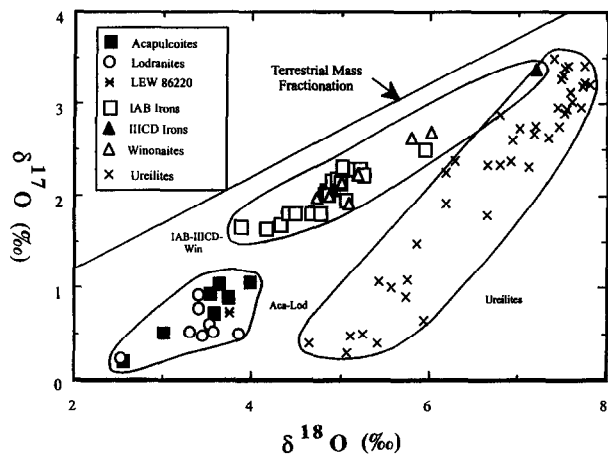


Fig. 6. Three-isotope oxygen plot for primitive achondrites. Acapulcoites, lodranites, and the related, unique meteorite LEW 86220 are readily distinguished from IAB irons, IIICD irons, and winonaites, as well as ureilites, but acapulcoites and lodranites cannot be distinguished from one another on the basis of oxygen isotopic composition. Data from Clayton and Mayeda (1988, 1996), Clayton et al. (1983, 1984, 1992), and Franchi et al. (1992).

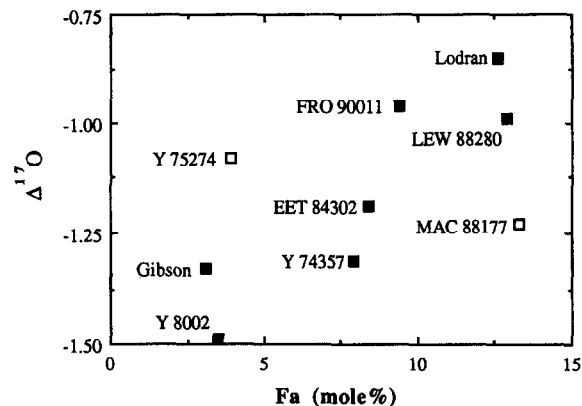


Fig. 7. Fa in olivine and  $\Delta^{17}\text{O}$  for lodranites are weakly positively correlated. Non-Antarctic and acid-washed Antarctic lodranites are represented by solid squares; nonacid washed Antarctic lodranites (Yamato 75274, MAC 88177) are represented by open squares and have higher uncertainties. This correlation is of nebular origin and suggests that lodranites formed from a chemically and isotopically heterogeneous precursor.

Nine analyzed lodranites have an average  $\Delta^{17}\text{O} = -1.16 \pm 0.21\text{‰}$ . In contrast, equilibrated H chondrite falls have an average  $\Delta^{17}\text{O} = 0.73 \pm 0.09\text{‰}$  ( $N = 22$ ) (Clayton et al., 1991) and, thus, lodranites exhibit an isotopic variation more than twice that of H chondrites. Note that  $\pm$  values refer to standard deviations of the analyses, not standard error of the means.

Oxygen isotopic compositions of lodranites are weakly positively ( $r^2 = 0.378$ ) correlated with Fa contents of olivine (Fig. 7). However, the Antarctic lodranites Yamato 75274 and MAC 88177 were analyzed without acid-washing, which at the time was not generally applied as a means of removing terrestrial weathering, and new samples for acid washing and subsequent analyses were, unfortunately, unavailable. Clayton and Mayeda (1988) showed that acid-washing, which removes terrestrial contamination, causes small ( $\sim 0.1\text{‰}$ ) but significant, positive or negative shifts in  $\Delta^{17}\text{O}$ . While such shifts do not affect our conclusions regarding the classification of these meteorites as lodranites, they obscure the subtle correlations examined here: When these two meteorites are excluded, the correlation coefficient  $r^2 = 0.795$  and, thus,  $\Delta^{17}\text{O}$  and Fa are then positively correlated at the 95% confidence level. Positive correlation between these parameters is also observed for the ureilites: For the non-Antarctic ureilites Dingo Pup Donga and Dyalpur and the acid-washed Antarctic ureilites ALH 84136, Yamato 74659, Yamato 791538, and LEW 85440, which have Fa concentrations between  $\text{Fa}_{2.5-16}$  and  $\Delta^{17}\text{O}$  of  $-1.29$  to  $-2.42$  (oxygen isotopic data from Clayton and Mayeda, 1988), we calculate a comparable  $r^2 = 0.764$ .

Although we argue that acapulcoites and lodranites are related and are from the same parent body (e.g., McCoy et al., 1996a), acapulcoites do not show a clear relationship between  $\Delta^{17}\text{O}$  and Fa, as do the lodranites: Although the low-Fa acapulcoites ALH A81187 and ALH 84190 also have the most negative  $\Delta^{17}\text{O}$ , the overall Fa distribution of acapulcoites is bimodal and no clear correlation exists with oxygen isotopic composition (Note that the lodranites were



Table 4. Argon isotopic data for Gibson. From left to right, columns are extraction temperature (°C), <sup>39</sup>Ar concentration (10<sup>-9</sup> cm<sup>3</sup>STP/g), age in Ga, K/Ca ratios, and corrected <sup>40</sup>Ar/<sup>39</sup>Ar, <sup>38</sup>Ar/<sup>39</sup>Ar, <sup>37</sup>Ar/<sup>39</sup>Ar, and <sup>36</sup>Ar/<sup>39</sup>Ar ratios. The K/Ca and <sup>36</sup>Ar/<sup>39</sup>Ar ratios have been multiplied by 100. Uncertainties were derived from those associated with spectrometer measurements and decay, blank, and reactor corrections. They do not include the uncertainty in irradiation constant, J = 0.03368 ± 0.00026.

Temp	<sup>39</sup> Ar	AGE Ga	K/Ca x100	<sup>40</sup> / <sup>39</sup>	<sup>38</sup> / <sup>39</sup>	<sup>37</sup> / <sup>39</sup>	<sup>36</sup> / <sup>39</sup> x100
400	0.80	5.468 ± 0.028	4.66 ± 0.10	585.4 ± 10.4	61.72 ± 1.11	11.33 ± 0.23	196.8 ± 6.0
500	0.71	3.702 ± 0.029	3.96 ± 0.09	201.4 ± 4.1	34.78 ± 0.74	13.31 ± 0.30	24.9 ± 4.1
650	2.16	4.159 ± 0.014	6.99 ± 0.10	268.1 ± 2.6	15.27 ± 0.15	7.54 ± 0.11	13.5 ± 1.5
750	5.39	4.481 ± 0.009	17.15 ± 0.20	326.3 ± 1.9	3.16 ± 0.02	3.08 ± 0.04	2.3 ± 0.5
825	10.20	4.499 ± 0.003	20.29 ± 0.21	329.7 ± 0.7	1.42 ± 0.00	2.60 ± 0.03	1.1 ± 0.2
875	9.76	4.502 ± 0.003	18.98 ± 0.19	330.3 ± 0.7	1.50 ± 0.00	2.78 ± 0.03	1.6 ± 0.3
925	6.62	4.482 ± 0.004	15.83 ± 0.17	326.4 ± 0.9	2.04 ± 0.01	3.33 ± 0.03	2.2 ± 0.4
1000	4.13	4.453 ± 0.007	6.46 ± 0.07	320.7 ± 1.6	4.06 ± 0.02	8.17 ± 0.09	5.0 ± 0.7
1100	3.82	4.473 ± 0.006	1.75 ± 0.02	324.7 ± 1.2	2.49 ± 0.01	30.20 ± 0.32	6.7 ± 0.8
1200	2.14	4.498 ± 0.023	0.13 ± 0.00	329.5 ± 5.0	7.56 ± 0.12	394.23 ± 7.20	44.7 ± 2.6
1300	0.21	5.443 ± 0.128	0.05 ± 0.00	577.0 ± 47.	21.90 ± 1.82	1064.6 ± 86.9	380 ± 39.
1450	0.15	5.439 ± 0.150	0.09 ± 0.01	575.5 ± 54.	21.13 ± 2.15	610.9 ± 58.4	477 ± 58.
1600	0.03	8.031 ± 0.533	0.18 ± 0.06	2516 ± 814	40.59 ± 13.7	298.5 ± 96.9	1388 ± 510

subject to the same pre-analysis acid wash as the acapulcoites). We selected Δ<sup>17</sup>O as the measure of oxygen isotopic composition, since it is unaffected by mass-dependent fractionation. Such fractionation could have occurred in the lodranites as a result of plagioclase loss. Plagioclase is enriched in <sup>18</sup>O relative to mafic silicates (Clayton, 1993) and, thus, plagioclase removal could move the residues to lower δ<sup>18</sup>O values along mass fractionation lines. We do not, however, believe that this process determined the isotopic compositions of lodranites, since all acapulcoites and some lodranites (e.g., Gibson, Y-8002) contain chondritic abundances of plagioclase but still differ significantly in δ<sup>18</sup>O.

### 3.3. <sup>39</sup>Ar-<sup>40</sup>Ar Chronology

Argon isotopic data for stepwise temperature extractions of an irradiated sample of Gibson are given in Table 4 and calculated <sup>39</sup>Ar-<sup>40</sup>Ar ages and K/Ca ratios as a function of fractional release of <sup>39</sup>Ar are shown in Fig. 8. If we discount

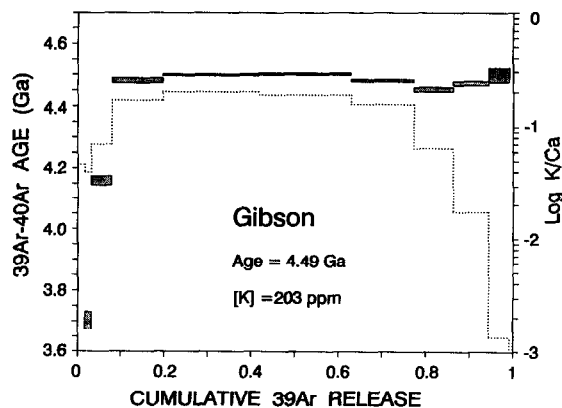


Fig. 8. <sup>39</sup>Ar-<sup>40</sup>Ar age (rectangles) and K/Ca ratio (dotted lines) as a function of cumulative release of <sup>39</sup>Ar for stepwise extractions of Gibson. Individual age uncertainties are indicated by the width of the rectangle. Potassium concentration and the temperature releasing 50% of the total <sup>39</sup>Ar are also indicated. Gibson has a plateau age of 4.49 Ga.

the first three extractions that indicate recent Ar diffusive loss and the three last extractions, then seven extractions (750–1200°C) releasing 90% of the <sup>39</sup>Ar define a <sup>39</sup>Ar-<sup>40</sup>Ar age of 4.484 ± 0.018 Ga (where the error is 1σ of deviations from the mean). The measured K content for our sample of Gibson (203 ppm) is less than that for chondrites, indicating that our sample contained less feldspar than that present in the thin sections studied (Table 1).

The Ar release spectrum for Gibson shows some age variations that deserve closer examination. Most of the <sup>39</sup>Ar release is associated with a phase having a K/Ca ratio of 0.15–0.2 (Table 4; Fig. 8), which is probably feldspar. The first ~8% of the <sup>39</sup>Ar release show a somewhat lower ratio and younger ages. These may represent effects of terrestrial weathering. Both the decrease in the K/Ca ratio (Fig. 8) and changes in the rate of release of Ar as a function of extraction temperature suggest that a different phase degasses <sup>39</sup>Ar at temperatures of ~1000°C and above. We ascribe the slight decrease in age at ~80–85% <sup>39</sup>Ar release to weathering effects of grain surfaces of this second phase, a phenomenon we have observed in other weathered meteorites. Because of the coarse grain size of Gibson (Table 2) and the relatively large fraction of the total <sup>39</sup>Ar released from this higher temperature mineral phase, we doubt that the slight decrease in age at ~1000°C is caused by <sup>39</sup>Ar recoil redistribution in the irradiation. However, we cannot totally rule out this explanation. Because of the uncertainty in interpretation of the Ar data from the 1000°C extraction, a preferred <sup>39</sup>Ar-<sup>40</sup>Ar degassing age of Gibson would be 4.493 ± 0.011 Ga, which considers only the 825–925°C and 1100–1200°C extractions releasing 71% of the total <sup>39</sup>Ar.

The 1300–1450°C extractions suggest substantially older ages of ~5.4 Ga associated with a phase with quite low K/Ca. It is unlikely that these older ages can be explained by uncertainties in <sup>40</sup>Ar blank corrections, which are only ~5% for these extractions. (The large blank correction for the 1600°C extraction, which released much less <sup>40</sup>Ar, does make its age very uncertain). Reactor-generated corrections are even less important. We note that these high temperature

TABLE 5. Noble gases in Gibson.

Mass (mg)	<sup>3</sup> He	<sup>4</sup> He	<sup>20</sup> Ne	<sup>20</sup> Ne/ <sup>22</sup> Ne	<sup>22</sup> Ne/ <sup>21</sup> Ne	<sup>36</sup> Ar	<sup>36</sup> Ar/ <sup>38</sup> Ar	<sup>40</sup> Ar	<sup>84</sup> Kr	<sup>132</sup> Xe	Cosmogenic Fraction		
											<sup>21</sup> Ne	<sup>38</sup> Ar	<sup>22</sup> Ne/ <sup>21</sup> Ne
66.7	8.31 <i>0.40</i>	793 <i>35</i>	2.34 <i>0.09</i>	1.348 <i>0.007</i>	1.313 <i>0.007</i>	8.24 <i>0.33</i>	5.14 <i>0.03</i>	3190 <i>130</i>	0.307 <i>0.020</i>	0.040 <i>0.003</i>	1.32	0.067	1.211
144.1	8.66 <i>0.40</i>	1090 <i>45</i>	1.74 <i>0.07</i>	1.010 <i>0.005</i>	1.253 <i>0.006</i>	2.21 <i>0.09</i>	4.39 <i>0.02</i>	1830 <i>70</i>	0.097 <i>0.008</i>	0.024 <i>0.002</i>	1.37	0.099	1.211
73.9	8.62 <i>0.40</i>	945 <i>40</i>	2.17 <i>0.08</i>	1.225 <i>0.005</i>	1.290 <i>0.006</i>	4.95 <i>0.20</i>	5.08 <i>0.03</i>	3470 <i>140</i>	0.238 <i>0.018</i>	0.042 <i>0.003</i>	1.37	0.054	1.211

Noble gas concentrations in 10<sup>-8</sup> cm<sup>3</sup>STP/g. 1σ errors are given in italics.

extractions also release the overwhelming amount of <sup>36</sup>Ar (Table 4), most of which is trapped rather than cosmogenic gas (see next section). (However, the concentration of <sup>36</sup>Ar in our sample is approximately an order of magnitude less than that in the unirradiated samples reported below). Thus, it is possible that these older ages reflect the presence of excess radiogenic <sup>40</sup>Ar, which was produced prior to the time of strong metamorphism, but was not totally lost from this retentive phase because of significant burial in the parent body.

The <sup>39</sup>Ar-<sup>40</sup>Ar age uncertainty quoted above does not include the uncertainty in the irradiation constant. If we include this additional uncertainty, the absolute degassing age becomes 4.49 ± 0.04 Ga. This is the first age determination of a lodranite, although Torigoye et al. (1993) reported that whole-rock Rb-Sr systematics of the lodranite Yamato 8002 are consistent with a 4.50 Ga age, assuming the same initial isotopic composition as for a CAI in Allende. They did not, however, determine the age of the meteorite. We can also compare the age of Gibson with <sup>39</sup>Ar-<sup>40</sup>Ar ages of two acapulcoites (McCoy et al., 1996a). Because Gibson and the acapulcoite Monument Draw were included in the same irradiation, their <sup>39</sup>Ar-<sup>40</sup>Ar ages can be compared using uncertainties that do not include those in the irradiation constant. The age of Monument Draw is precisely defined at 4.517 ± 0.006 Ga (1σ) and Gibson at 4.493 ± 0.011 Ga. Thus, these ages agree within 2σ, but not 1σ, and, thus, it is not impossible that the Ar degassing time for Gibson was actually ~0.02 Ga later than that for Monument Draw. Such an age difference, if real, may reflect the somewhat higher metamorphic heating of the lodranites compared to the acapulcoites and may suggest a longer cooling time to Ar closure temperatures for the former.

### 3.4. Noble Gases

#### 3.4.1. Cosmic ray exposure age

The concentrations of cosmogenic <sup>21</sup>Ne and the shielding parameters (<sup>22</sup>Ne/<sup>21</sup>Ne)<sub>cos</sub> are very similar in all splits of the Gibson lodranite (Table 5). The rather high (<sup>22</sup>Ne/<sup>21</sup>Ne)<sub>cos</sub> ratios indicate a low shielding of a few cm at most and presumably a small preatmospheric size of Gibson. With the nuclide production model of Graf et al. (1990b), we estimate a shielding-corrected <sup>21</sup>Ne exposure age of 6.0–6.4 Ma. We assume a <sup>21</sup>Ne production rate typical for L chondrites, which

is appropriate in view of the Mg concentration (~15.9 wt%) estimated from modes and mineral compositions. We also make the reasonable assumption that the preatmospheric radius of Gibson (which was found as a whole stone of only 67.1 g) was less than 30 cm.

Eugster and Weigel (1993) report noble gas data for five lodranites. The <sup>21</sup>Ne<sub>cos</sub> and also the (<sup>22</sup>Ne/<sup>21</sup>Ne)<sub>cos</sub> values in Yamato 791491, MAC 88177, and FRO 90011 are similar to those in Gibson and, thus, their nominal exposure ages are also similar (Yamato 791491, 6.1–6.6 Ma; MAC 88177, 7.2–7.5 Ma; FRO 90011, 5.9–6.4 Ma; all calculated with the Graf et al. (1990b) model and the same assumptions as stated above for Gibson). For Lodran and LEW 88280 (Eugster and Weigel, 1993), we deduce nominal shielding-corrected <sup>21</sup>Ne exposure ages of 4.0–4.3 Ma. The differences between these ages and those of the former four lodranites seem to be significant, although errors are on the order of 20%. Takaoka et al. (1993) report a <sup>21</sup>Ne exposure age of 11.7 ± 1.5 Ma for the lodranite Yamato 74357. However, with our method, we calculate a considerably higher <sup>21</sup>Ne age of 17.5–21.5 Ma, whereby the upper limits reflect the assumption that the preatmospheric radius of Yamato 74357 did not exceed 70 cm. This discrepancy is due to the fact that Yamato 74357 has a low (<sup>22</sup>Ne/<sup>21</sup>Ne)<sub>cos</sub> ratio of only ~1.075, which indicates rather high shielding. As is shown by Graf et al. (1990b), the shielding correction of Eugster (1988) used by Takaoka et al. (1993) overestimates <sup>21</sup>Ne production rates at (<sup>22</sup>Ne/<sup>21</sup>Ne)<sub>cos</sub> < 1.08. In any case, Yamato 74357 has a higher cosmic ray exposure age than any of the other lodranites discussed here. Thus, at least two, and possibly three, ejection events are needed to account for the exposure ages of all lodranites studied. It is interesting to note that four of seven lodranites have exposure ages in the same ~5.5–7 Ma range as all acapulcoites (e.g., McCoy et al., 1996a). This is consistent with a single ejection event on a common source body for most acapulcoites and lodranites for which cosmic ray exposure ages have been measured.

#### 3.4.2. Trapped noble gases

Lodranites and acapulcoites usually contain relatively large amounts of trapped noble gases, similar to those in type 3 or 4 chondrites (Palme et al., 1981; Eugster and Weigel, 1993; Takaoka et al., 1993; McCoy et al., 1996a). The isotopic compositions of Kr and Xe are often identical

to those of component  $Q$  found in carbonaceous chondrites (Wieler et al., 1992). In all three analyses of Gibson, however, Kr and Xe have atmospheric composition (data not given). The only exception is  $^{129}\text{Xe}$ . The ratio of  $^{129}\text{Xe}/^{132}\text{Xe}$  ranges between 1.12–1.27, indicating the presence of radiogenic  $^{129}\text{Xe}$ . Otherwise, Kr and Xe in Gibson are obviously mostly atmospheric contamination, presumably trapped in the abundant weathering products. A generous upper limit for primordial  $^{132}\text{Xe}_0$  is  $0.005 \times 10^{-8} \text{ cm}^3\text{STP/g}$ , which is roughly  $20\times$  less than the trapped Xe in, for example, Monument Draw and is also less than what is usually found in type 5 or 6 chondrites. Unfortunately, the concentration of trapped  $^{36}\text{Ar}$  cannot be reliably determined because of the possible contamination with atmospheric Ar and because the concentration of radiogenic  $^{40}\text{Ar}$  is difficult to estimate reliably. Since, at least in chondrites, a large fraction of the trapped Xe survives dissolution of the meteorite by HF/HCl, it seems unlikely that the primordial trapped Xe in Gibson could have been lost nearly completely during terrestrial weathering.

#### 4. DISCUSSION

##### 4.1. Properties of the Precursor Chondritic Material

We have previously presented evidence that the acapulcoites and, by inference, the lodranites, formed by thermal metamorphism and partial melting from chondritic precursor materials on the same parent body (McCoy et al., 1992a,b, 1996a). In the case of lodranites, all evidence for the original texture of this chondritic precursor has been erased by heating and partial melting. However, we suggest that the mineral and oxygen isotopic compositions of lodranites were largely inherited from the chondritic precursor. We further suggest that some properties of lodranites resulted from nebular processes. This is because similar ranges in mineral and isotopic compositions as observed in lodranites have been noted in other groups of meteorites and attributed to a nebular origin. For example, ordinary H, L, and LL chondrites (Clayton et al., 1991) and ureilites (Clayton and Mayeda, 1988) show a correlation between  $\Delta^{17}\text{O}$  and mafic silicate compositions, not unlike what is observed in lodranites.

##### 4.2. Heating and Melting

Lodranites have experienced extensive modification by heating (as is evident from the equigranular textures and abundant  $120^\circ$  triple junctions) and partial melting. This has previously been suggested by Nagahara (1992), Takeda et al. (1994), and Mittlefehldt et al. (1996).

###### 4.2.1. Evidence for Fe,Ni-FeS partial melting

Lodranites have experienced Fe,Ni-FeS cotectic melting. The first partial melt in a chondritic system forms between 950 and 980°C, depending on composition, and consists of ~85 wt% troilite (FeS) and ~15 wt% Fe,Ni metal (Kullerud, 1963; Kubaschewski, 1982). Thus, removal of a small amount of cotectic melt has a dramatic effect on the troilite abundance of the residue. It appears that Fe,Ni-FeS cotectic melting has occurred in the lodranites at 950–1050°C. Some

unweathered lodranites are highly depleted in modal troilite (e.g., Y-8002, Y-75274, EET 84302) relative to likely chondritic precursors (e.g., 3.6–7.2 wt% FeS in ordinary chondrites; Keil, 1962), reflecting nearly complete removal of the Fe,Ni-FeS cotectic melt. Many other lodranites have modal troilite abundances and bulk S contents at chondritic or only slightly below chondritic levels. Evidence for Fe,Ni-FeS cotectic melting and melt migration in the form of irregular patches and occasional veins of metal and troilite is found in some lodranites (e.g., Lodran, Y-791491, Y-791493, FRO 90011). McCoy et al. (1996a) argued that acapulcoites had experienced melting at the Fe,Ni-FeS cotectic, but not silicate partial melting. In the acapulcoites, the Fe,Ni-FeS cotectic melt is trapped in veins which range from  $\mu\text{m}$ - to cm-sized. In contrast, such veining is not common in lodranites.

###### 4.2.2. Silicate partial melting

It is also apparent that lodranites have experienced silicate partial melting. The first melt in a chondritic system is basaltic (Morse, 1980). Silicate partial melting of an olivine-rich rock in the system Fo-An-SiO<sub>2</sub> occurs at the peritectic point. This melt contains ~55% plagioclase (Morse, 1980). We recognize that plagioclase in lodranites is albitic and the phase boundaries would shift in a Fo-Ab-SiO<sub>2</sub> phase diagram, which is currently unstudied. Examination of the Ab-SiO<sub>2</sub> binary phase diagram (Tuttle and Bowen, 1958) suggests that the boundaries would shift so that the initial melt would contain an even greater abundance of plagioclase. Thus, removal of this basaltic partial melt results in a significant depletion of plagioclase in the residual source rocks. In fact, most lodranites contain significantly less plagioclase than likely chondritic parental rocks (9.6–10.3 wt% in ordinary chondrites; McSween et al., 1991). Yamato 74357, FRO 90011, Yamato 791491, Yamato 791493, Lodran, MAC 88177, and LEW 88280 contain 0–1.6 vol% plagioclase. We interpret these lower than chondritic plagioclase contents as an indication of the removal of basaltic partial melts from lodranites.

Some lodranites, however, contain abundant plagioclase. Plagioclase abundances in Gibson (6.0 vol%), Yamato 8002 (10.3), and EET 84302 (11.4) are all equal to or only slightly less than plagioclase abundances in ordinary chondrites. Evidence for silicate partial melting is present in these rocks as well. In Yamato 8002, single plagioclase crystals surround multiple mafic silicate grains. Interstitial textures for plagioclase grains are also observed in Gibson and EET 84302. In addition, clinopyroxene is always associated with plagioclase in EET 84302 (Field et al., 1993; Mittlefehldt et al., 1996). This association is typical of that expected in a basaltic partial melt and suggests to us that basaltic partial melting did occur, but that those melts did not migrate from the rock, but were trapped in situ, or could have migrated into the area sampled by these meteorites (see McCoy et al., 1997).

###### 4.2.3. Formation of minor phases

Redistribution of minor phases may also result from silicate partial melting. Takeda et al. (1994) noted the existence

of a chromite-rich region in EET 84302. We did not observe this chromite-rich lithology in our section (,12). Records from other workers (e.g., M. M. Lindstrom, pers. commun.) indicate that only a small portion, perhaps an area 2 mm on a side, of the 59.6 g stone contains this chromite-rich lithology. Most of EET 84302 resembles normal, metal-bearing lodranites. Takeda et al. (1994) argued that this lithology formed by deposition of chromite in a small area as a result of movement of large amounts of Cr-rich silicate partial melts. They cite the absence of olivine and enrichment of pyroxene in the vicinity of the chromite-rich lithology as evidence of reaction between  $\text{Cr}_2\text{O}_3$  in the melt and olivine to form chromite and pyroxene. If correct, this is another consequence of silicate partial melting in the lodranites.

#### 4.2.4. Unexplained features

Some features of lodranites are likely due, at least in part, to the igneous processing they experienced. Yet the exact mechanisms which controlled these features are uncertain. First among these is the composition of plagioclase. Lodranites exhibit a broad range of plagioclase compositions ( $\text{An}_{12.3-30.9}$ ). This range is much wider than that exhibited by the acapulcoites ( $\text{An}_{\sim 14-18}$ ), which are the likely direct precursors to the lodranites. It is possible to argue that the variability in the lodranites was inherited from the precursor chondrite, and we have yet to sample this full range of variability in the acapulcoites. This seems unlikely to us. More likely, the partial melting experienced by the lodranites altered the plagioclase compositions from those of the precursors, probably by fractionation prior to melt removal or crystallization of other Ca-bearing phases (e.g., pyroxenes). However, plagioclase compositions show no correlation with other parameters (plagioclase abundances, mafic silicate composition), so these processes did not operate in any systematic manner.

A second unexplained feature relates to the relative abundances of Fe,Ni metal, FeS, and plagioclase. While it is clear that the abundances of these minerals were altered in response to removal of the Fe,Ni-FeS and silicate melts, how this removal affected individual lodranites is less clear. Lodranites exhibit a wide range of Fe,Ni metal abundances (0.5–19.6 vol%, excluding the highly-weathered Gibson). Unrepresentative sampling is not likely to be responsible for this wide range, given that the 1 kg Lodran has never been shown to contain metal-poor regions despite extensive study of this meteorite. This range is considerably greater than that for acapulcoites (14–23 wt%), the likely direct precursors of the lodranites. It also appears unlikely that the disparity in these two ranges reflects poor-sampling of the acapulcoites, since no metal-poor regions within acapulcoites have been noted in this study or in previous work. Rather, some lodranites (e.g., Lodran) must have gained metal during igneous processing, whereas other lodranites (e.g., MAC 88177, 0.5 vol% Fe,Ni metal) lost much of their original metal. We would note that it appears that oxidation of FeO from silicates was not responsible for the increase in metal abundances, since the Fe,Ni metal-troilite irregular patches and veins are volumetrically insignificant.

The relative abundances of troilite and plagioclase in indi-

vidual lodranites are similarly puzzling. Lodranites tend to be depleted in these phases relative to likely precursor chondritic materials. However, some lodranites (EET 84302, Yamato 8002, Gibson, Yamato 75274) retain moderate to high abundances of plagioclase but are highly depleted in FeS. In contrast, other lodranites (MAC 88177, Yamato 74357, LEW 88280, Lodran, FRO 90011, Yamato 791491, Yamato 791493) are highly depleted in plagioclase but contain moderate to abundant FeS. Thus, some lodranites appear to have lost only the Fe,Ni-FeS partial melt, whereas others appear to have lost primarily the silicate partial melt. We suggest that once melt migration begins, the coexistence of a dense, immiscible Fe,Ni-FeS melt in a silicate melt will lead to highly heterogeneous and complex assemblages, since it is not to be expected that these melts would be removed and deposited with equal ease. (A more complete discussion of melt migration in the acapulcoite-lodranite parent body is given by McCoy et al., 1997). Some authors (Nagahara, 1992; Mittlefehldt et al., 1996) believe that one or more of these melts may have been introduced into the depleted lodranites. Nagahara (1992) used this mechanism to explain the textures of plagioclase in Yamato 8002, whereas Mittlefehldt et al. (1996) argued for injection of an Fe,Ni-FeS melt to explain enriched Se/Co ratios in MAC 88177 and LEW 88280 relative to most acapulcoites.

#### 4.2.5. Peak temperatures

The peak temperatures experienced by lodranites can be estimated an estimation of the degree of partial melting. Taylor et al. (1993) used the melting of peridotite (McKenzie and Bickel, 1988) to model the degree of partial melting as a function of temperature for chondritic systems. Basaltic partial melts begin forming at  $\sim 1150^\circ\text{C}$  in a chondritic system. Evidence for silicate partial melting in all lodranites suggests this as the minimum temperature to which lodranites were heated. The maximum temperature is estimated from the degree of partial melting necessary for complete melting and removal of plagioclase. Melting and removal of  $\sim 10$  vol% plagioclase which is a typical content for chondrites, would require  $\sim 15$ – $20\%$  silicate partial melting (depending on the composition of the plagioclase) and peak temperatures of the order of  $1200^\circ\text{C}$ . The estimate of  $1200^\circ\text{C}$  is relatively insensitive to the amount of melting in the range of  $15$ – $20\%$  (McKenzie and Bickel, 1988). Lodranites which experienced lower degrees of partial melting (e.g., the plagioclase-rich Gibson) might have experienced a temperature closer to  $1150^\circ\text{C}$ . Thus, a relatively small range of peak temperatures can result in a relatively wide range in the degree of silicate partial melting.

It is interesting to note that these temperatures are considerably higher than the two-pyroxene geothermometer temperatures. However, the latter are closure temperatures and a period of relatively slow cooling might have occurred between the peak temperature and pyroxene closure. Further, the two-pyroxene geothermometer temperatures show no correlation with degree of partial melting. We suggest that an additional factor may have influenced the degree of partial melting. It is known that FeO contents of mafic silicates strongly influence melting temperatures. Lodranites with low

FeO contents exhibit relatively low degrees of silicate partial melting (as measured by plagioclase abundances, Fig. 2), whereas high-FeO lodranites exhibit relatively extensive silicate partial melting. We suggest that the wide range of FeO contents of these meteorites may also have strongly influenced the degree of silicate partial melting. There are, however, exceptions to this rule (e.g., EET 84302), and we suggest that these exceptions were caused by real differences in peak temperatures. We also note that we cannot unequivocally eliminate the possibility that increased plagioclase abundances in low-FeO lodranites reflect in situ crystallization of partial melts, although we could offer no explanation of why this crystallization would occur only in low-FeO lodranites.

### 4.3. Chronology and Thermal History

We are interested in two aspects of the chronology of the lodranites: their age of formation and their cooling history. The 4.49 Ga  $^{39}\text{Ar}$ - $^{40}\text{Ar}$  age of Gibson is the only measured age of any lodranite. Considering the prolonged thermal history of the lodranites and its effect on the  $^{39}\text{Ar}$ - $^{40}\text{Ar}$  chronometer, this age clearly does not represent the time of formation of the lodranite parent body. That time is probably best given by the 4.556 Ga Pb-Pb age for Acapulco (Göpel et al., 1992). The  $^{39}\text{Ar}$ - $^{40}\text{Ar}$  age for Gibson, along with similar ages for acapulcoites (McCoy et al., 1996a), thus place a lower limit on the time of significant cooling after the metamorphic heating experienced by the parent body. We note that the 4.49 Ga  $^{39}\text{Ar}$ - $^{40}\text{Ar}$  age for Gibson lies within the measured range of  $^{39}\text{Ar}$ - $^{40}\text{Ar}$  ages of unshocked ordinary chondrites of 4.44–4.52 Ga ages that also presumably date the time of Ar closure at the end of parent body metamorphism (Turner et al., 1978).

The cooling history of lodranites appears to be complex. Several lines of evidence suggest that lodranites experienced a period of slow cooling at high temperatures. Two-pyroxene equilibration temperatures of lodranites are 100–200°C below peak temperatures inferred from the degree of partial melting. As two-pyroxene equilibration temperatures reflect the cessation of diffusion, relatively slow (1–100°C/Myr) cooling must have occurred between the peak temperatures and the establishment of the two-pyroxene equilibration temperature. A second argument for slow cooling comes from the ~65 Ma difference between the likely formation age of the lodranite parent body and the ~4.49 Ga  $^{39}\text{Ar}$ - $^{40}\text{Ar}$  age. After this slow cooling, a period of very rapid cooling occurred between perhaps 1000°C and 400°C, as indicated by cooling rates of  $\sim 10^3$ – $10^6$ °C/Myr by a variety of methods (Table 2). This suggests that the lodranite parent body was at least partially disrupted by a large impact while at a temperature of  $\geq 900$ °C. The cooling rates in the middle temperature range suggest fragmentation into pieces which range in radius from approximately 200 m to 6 km (using Eqn. 12 of Haack et al., 1990). McCoy et al. (1996a) argued for breakup and reassembly of this parent body based on a more complete dataset for acapulcoites.

### 4.4. Acapulcoites and Lodranites are from the Same Parent Body

All available data suggest that acapulcoites and lodranites almost certainly originated on a common parent body. They

have the same mineralogy (opx + ol + Fe,Ni metal  $\pm$  FeS  $\pm$  plag) but differ in their mineral abundances. Lodranites have heterogeneous and correlated mafic silicate and oxygen isotopic compositions which overlap those of acapulcoites, although acapulcoites do not show a strong correlation between oxygen isotopic and mafic silicate compositions. Both groups are enriched in trapped noble gases. Complex thermal histories suggest breakup and reassembly of the acapulcoite-lodranite parent body. In addition, four acapulcoites and five lodranites share a common cosmic ray exposure age (5.5–7 Ma). Further, EET 84302 is intermediate in many of its properties (e.g., mafic silicate grain size, plagioclase abundances) between the two groups (McCoy et al., 1993; Field et al., 1993; Mittlefehldt et al., 1996).

There are, however, important differences between the two rock types, and these can be attributed to different degrees of partial melting. We suggest that acapulcoites were heated to ~950–1050°C and experienced Fe,Ni-FeS cotectic melting (McCoy et al., 1996a). We note, however, that others argue for higher peak temperatures for at least some acapulcoites, without melt migration (Zipfel et al., 1995). A full discussion of these opposing viewpoints is given in McCoy et al. (1996a). Melt migration distances in acapulcoites were small ( $\mu\text{m}$  to cm), and chemical fractionation did not occur (McCoy et al., 1996a). In contrast, lodranites appear to have been heated to ~1100–1250°C, resulting in the formation of both Fe,Ni-FeS cotectic and basaltic partial melts. Melts migrated from the source rocks, leaving residues (e.g., lodranites) which tend to be depleted in FeS and plagioclase. Another possible consequence of this difference in peak temperature may be the slightly younger  $^{39}\text{Ar}$ - $^{40}\text{Ar}$  age for Gibson compared to several acapulcoites. Thus, we conclude that acapulcoites and lodranites appear to represent the products of differing degrees of heating and partial melting of common, isotopically and chemically heterogeneous precursors.

The differences in grain sizes between the acapulcoites and lodranites could also be related to the presence or absence of a silicate partial melt. Grain growth in these rocks probably occurs by Ostwald ripening, with the diffusion rate through the intergrain medium as the controlling factor in determining the grain sizes. Diffusion rates in silicate melts (summarized by Taylor, 1992) are up to six orders of magnitude faster than those in crystalline rocks. Thus, the lodranites, which once may have contained silicate partial melts, became much coarser-grained than the acapulcoites, which lacked silicate partial melts. Because of the speed of grain growth in the presence of silicate melt, Taylor (1993) argued that the silicate melts must have migrated from their source regions in times of the order of  $10^3$  years. Otherwise, lodranites would be even coarser grained.

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