Origin of Fragmental and Regolith Meteorite Breccias—Evidence from the Kendleton L Chondrite Breccia

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The Kendleton L chondrite breccia is unusual in that it has most of the characteristics of ordinary chondrite breccias rich in solar-wind gases, but there is no evidence that it contains, or ever lost, these gases. The dark fragmental matrix in Kendleton is largely composed of type 4 material and contains at least four kinds of light- and dark-colored clasts with diverse origins. These clasts are composed of L3, L5, shock-blackened and melt rock material and were probably all derived from normal L chondrite material. The shock-blackened clasts were produced in post-metamorphic impacts, whereas the melt rock formed before the end of metamorphism. We also found a unique tridymite-rich inclusion that has an oxygen isotopic composition consistent with an origin in an H chondrite asteroid. However, the absence of both tridymite-rich inclusions in H chondrites and H chondrite clasts in L chondrites suggests a different origin for the tridymite-rich inclusion. The presence of martensite in some shock-blackened clasts shows that Kendleton, like regolith breccias, was not metamorphosed after compaction. If solar-flare irradiated grains in gas-rich meteorite breccias were irradiated on asteroid surfaces by an early active (T-Tauri) sun, as *Caffee et al.* (1987) argue, then some of these grains were stored inside parent bodies for up to 10^9 years before being mixed with larger volumes of material to form gas-rich breccias. Most of the material in gas-rich chondrite breccias may never have been near the surface of their parent bodies as the mixing could have occurred, for example, during break-up and reassembly of asteroids. Thus meteoritic and lunar regolith breccias may have different origins. Kendleton and other gas-poor, light-dark structured breccias probably formed in an identical fashion to gas-rich breccias, except that the gas-poor breccias did not receive a small admixture of irradiated grains.

INTRODUCTION

Regolith breccias are fragmental rocks that contain implanted solar-wind gases and are widely believed to have formed by lithification of regolith material that once resided at the surface of a body (*Taylor and Wilkening,* 1982). Almost all chondrites that have dark matrices and light- and dark-colored clasts also contain solar-wind gases and are believed to have formed from regolith material on asteroids (*Goswami et al.,* 1984). But a few such light-dark structured chondrites are known that do not contain solar-wind gases (*Keil,* 1982; *Rubin,* 1982). None of these chondrites has been studied in detail. Did such chondrites lose solar-wind gases from heating, or were the gases never acquired? How is it possible for two meteorite breccias to appear identical petrographically, yet differ so much in the abundance of irradiated grains in their hosts?

To elucidate the origin of regolith and fragmental breccias we chose to study the Kendleton L chondrite, which fell in 1939 and contains many prominent clasts. Limited published data (*Taylor and Heymann,* 1969) and our preliminary inspection suggested that Kendleton might be a light-dark structured chondrite that lacks detectable solar-wind gases. Our petrographic and noble gas analyses confirm this and show that Kendleton contains a wide variety of clasts. L3, L5, shock-blackened and melt rock clasts, and a unique tridymite-rich clast were identified. We suggest that Kendleton and some other fragmental breccias lacking solar-wind gases may be cogenetic with gas-rich breccias, except that the former did not receive a small admixture of solar-irradiated grains. We speculate on the origin of fragmental and regolith breccias in the light of the recent studies of *Caffee et al.* (1983, 1986, 1987). We infer from these studies that meteoritic and lunar regolith breccias may have different origins.

METHODS AND MATERIALS

Fifty-two specimens of Kendleton with a total weight of 9.7 kg were inspected at the Monnig Meteorite Collection at Texas Christian University, Fort Worth. Nine were chosen for study because hand-specimen examination revealed the presence of a wide variety of clasts. Polished thin sections, which are now in the collection at the University of New Mexico, were studied microscopically in transmitted and reflected light. Electron microprobe analyses were made using an ARL EMX-SM operated at 15 keV and about 20 nA sample current. We corrected for differential matrix effects using the method of *Bence and Albee* (1968); minerals of well-known compositions were used as standards. Precision was such that 20 analyses of a homogeneous olivine (Fa 17) have a standard deviation of 0.4% Fa.

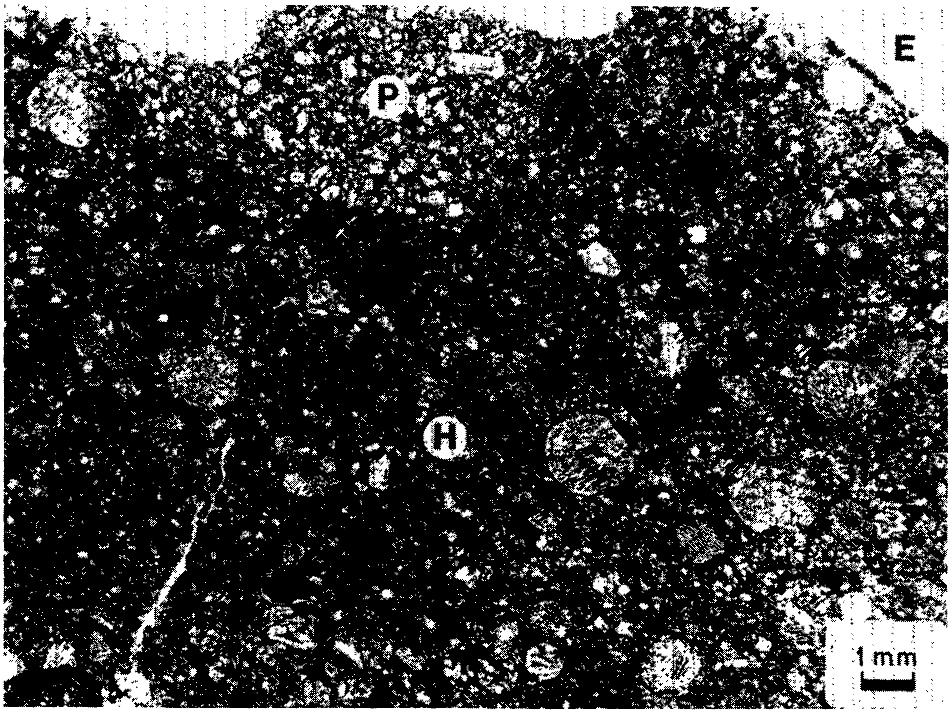

Fig. 1. Transmitted-light photomicrograph of the Kendleton fragmental breccia (section UNM 711) showing chondritic host (H) and a light-colored, porphyritic, melt rock clast (P); E = epoxy. This meteorite resembles chondritic regolith breccias in having a dark host that contains abundant fragmental material and a few prominent chondrules and encloses light-colored clasts.

Oxygen isotopic compositions of host and clast material from Kendleton were analyzed using standard techniques (*Clayton et al.*, 1976). Noble gas concentrations and isotopic ratios were analyzed using techniques of *Weber et al.* (1982).

RESULTS

Chondritic Host

About 80% of the Kendleton breccia consists of dark chondritic host; discernible clasts greater than 1 mm in size make up the remainder. Some of the chondrules are well defined (Fig. 1) and chondrules and chondrule fragments have mesostases that are microgranoblastic to feathery, as in type 4 chondrites.

Compositions of olivines and low-Ca pyroxenes in the host were measured in nine thin sections (Table 1). All but one have compositional characteristics of type 4 chondrites: percent mean deviations (PMD) of wt % FeO in olivine in eight of the sections are 0.9-3.7. In the other section, UNM 710, which contains a type 3 clast, the host has the compositional heterogeneity of type 3: PMD of FeO in olivine is 10.5. Low-Ca pyroxenes are much more heterogeneous than olivines, implying that most of the host material is low type 4. Histograms

TABLE 1. Mean compositions of olivine and low-Ca pyroxene in nine sections of the chondritic host of Kendleton.

Section	Olivine				Low-Ca Pyroxene			
No.	Fa (mol %)		PMD*	No.	Fs (mol %)		PMD*	No.
	Mean	σ		Anal.	Mean	σ		Anal.
649	23.3	1.50	3.7	24	18.7	3.1	12.9	23
650	22.8	0.63	1.5	38	18.3	5.0	19.2	23
709	22.3	0.32	1.1	26	18.3	2.1	8.3	14
710	22.6	4.7	10.5	15	15.1	7.0	41	13
711	22.9	0.34	1.0	28	17.0	3.5	15.4	20
712	21.9	0.27	0.9	53	17.7	3.6	15.1	36
713	22.9	0.32	1.0	36	19.4	3.8	13.2	22
714	22.8	0.79	2.5	19	16.2	3.1	12.8	11
715	22.8	0.50	1.4	24	19.6	5.2	19.9	19
Mean	22.6			263	18.0			181

*PMD, percent mean deviation of FeO analyses in wt %.

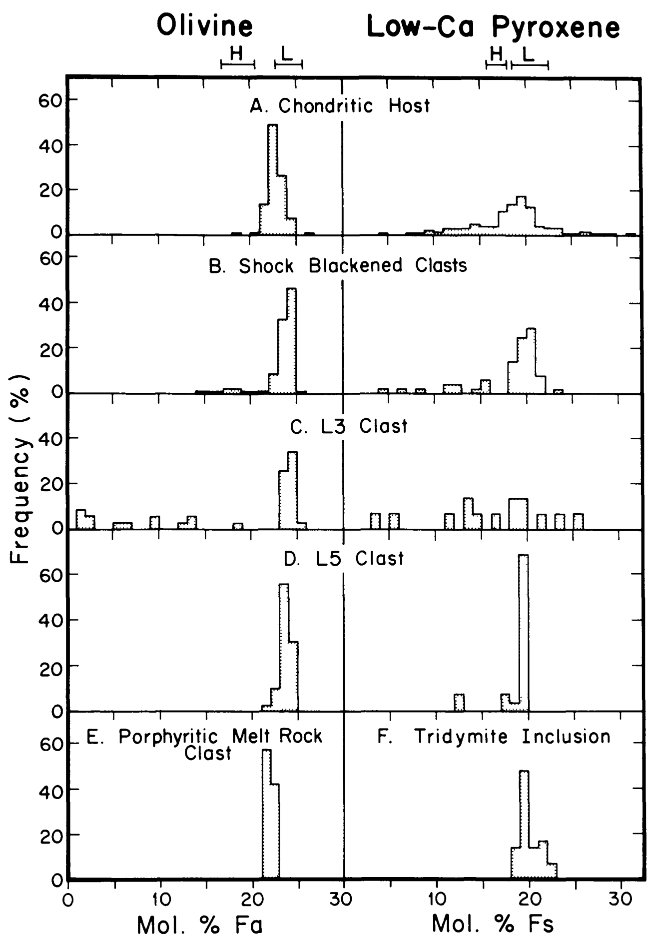

Fig. 2. Histograms of electron probe analyses of randomly chosen olivine and low-Ca pyroxene grains in host (A), and clasts and other objects (B-F) in the Kendleton fragmental breccia. Means and standard deviations are given in Tables 1 and 2. Kendleton host has the compositional characteristics of an L4 chondrite, but it was not metamorphosed to type 4 levels after lithification. Compositional ranges for equilibrated H and L chondrites are from *Gomes and Keil* (1980).

of all the analyses in nine sections are shown in Fig. 2A. Most of the host silicates have iron concentrations that lie in or close to the range of equilibrated L group chondrites (*Gomes and Keil,* 1980); mean concentrations of Fa 22.6 and Fs 18.0 lie at the low end of the L range. Calcium concentrations in host olivines (0.06 ± 0.02 wt % CaO) are marginally higher than those for most equilibrated chondrites (0.01-0.05 wt % CaO; *Busche,* 1975).

Olivines in the host show undulose extinction, implying shock facies b or c and shock pressures of 5-24 GPa (*Dodd and Jarosewich,* 1979). This estimate is consistent with the classification of shock class C by *Taylor and Heymann* (1969), who used X-ray diffraction techniques. Etched sections of host show that kamacite is polycrystalline with occasional taenite blebs; taenite contains partially or completely resorbed cloudy taenite (clear taenite) or plessite. This implies a metallographic shock class of III in the scheme of *Taylor and Heymann* (1969). Pockets of shock melt were not observed in the host, although a few shock veins are visible, especially around some shock-blackened clasts.

Noble gas concentrations were determined on a 0.31 g sample of the Kendleton host. The results are shown in Table 2 with the data of *Taylor and Heymann* (1969) for an uncharacterized sample of Kendleton. The host clearly lacks solar wind gases. The rather high concentrations of trapped planetary gases suggest a type 3 classification for the Kendleton host. The concentrations of trapped ^{36}Ar, for example, are appropriate to type 3.8 ordinary chondrites, according to *Sears et al.* (1980). The cosmic ray exposure age is 2.7 Myr using the ^{21}Ne production rate and shielding correction given by *Nishiizumi et al.* (1980). This value is low for L chondrites; about 95% have higher exposure ages (*Crabb and Schultz,* 1981). The concentration of ^{4}He is somewhat depleted, consistent with the shock level of b or c for Kendleton's host, and the decrease in L chondrites of ^{4}He and ^{40}Ar concentrations with increasing shock level (*Dodd and Jarosewich,* 1979). Data from *Taylor and Heymann* (1969) are broadly similar, although the lower concentration of ^{36}Ar is more consistent with the assignment of petrologic type 4 based on olivine heterogeneity.

An analysis of the oxygen isotopic composition of a host sample yielded values for δ^{18}O of 4.64‰ and δ^{17}O of 3.45‰ relative to standard mean ocean water. These values are entirely consistent with the L group classification derived from FeO concentrations of silicates.

Shock-Blackened Clasts

Five black objects ranging from 3 to 15 mm in diameter were studied in detail; many similar but smaller objects are distributed throughout the host. Although some of these objects appear at first sight to have formed in situ, we argue below that they are clasts of shocked rock. Similarities with black ordinary chondrites (*Heymann,* 1967) indicate that the blackening was produced by shock. Textures vary enormously, as in other heavily shocked chondritic material. A few regions show a faint chondritic texture, partly obscured by troilite veining of silicate fractures. However, much of the black

TABLE 2. Noble gas concentrations (10^{-8}cc STP/g) in Kendleton: (a) host, (b) uncharacterized sample.

Source	^{3}He	^{4}He	^{20}Ne	^{21}Ne	^{22}Ne	^{36}Ar	^{38}Ar	^{40}Ar	^{84}Kr	^{132}Xe
(a) This work	3.25	840	0.84	0.77	0.87	6.8	1.34	5300	0.10	0.14
(b) *Taylor and Heymann* (1969)	3.3	792	1.7	0.92	0.96	3.2	0.68	3380	*	*

*Not determined.

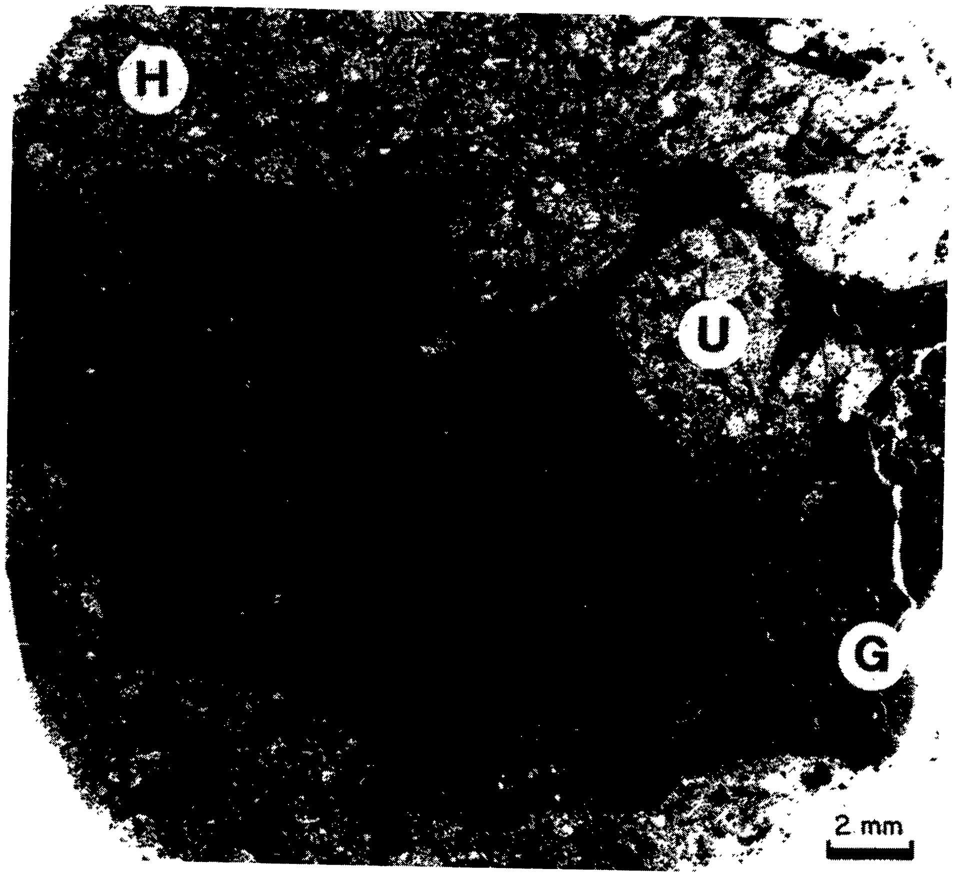

Fig. 3. Transmitted light photomicrograph of the Kendleton fragmental breccia (section UNM 710) showing a large shock-blackened clast, which contains glassy, clast-laden, shock melt (G), light-colored relict chondritic material, and an unequilibrated L3 clast (W), embedded in chondritic host (H). Although its upper horizontal edge is ill defined, this shock-blackened object is a clast of a heavily shocked, preexisting breccia and did not form in situ from the Kendleton host.

material is completely opaque in thin section because of the intrusion of extensive veins of troilite into olivine and low-Ca pyroxene grains (Fig. 3). These veins are generally much less than a micron in width and only a few microns apart. Some regions up to 1 cm across resemble shock melt veins with relict silicates 5-50 μm in size contained in a glassy looking matrix, which contains many submicron metal grains and silicate phenocrysts. Transparent olivine crystals show undulatory to mosaic extinction patterns, indicating diverse shock intensities. Metal and troilite are largely irregular in shape. Composite grains are rare and the dendritic or igneous metal-troilite textures are rare, unlike nearly all other shock-blackened chondritic material that we have observed. However, some regions in Fig. 3 do contain troilite-rimmed globules of metallic

TABLE 3. Mean compositions of olivines and low-Ca pyroxenes in clasts and inclusions in Kendleton.

Object	Section No.	Olivine Fa (mol %) Mean	Olivine Fa (mol %) σ	Olivine PMD*	Olivine No. Anal.	Low-Ca Pyroxene Fs (mol %) Mean	Low-Ca Pyroxene Fs (mol %) σ	Low-Ca Pyroxene PMD*	Low-Ca Pyroxene No. Anal.
(a) Shock-blackened clasts	649	23.7	1.42	2.6	25	18.7	3.0	12.5	9
	650	23.8	1.35	2.9	23	20.1	-	-	3
	710	22.7	3.0	9.6	38	15.8	5.7	30	18
	712	22.2	0.20	0.6	11	19.4	0.40	1.6	11
	715	24.2	0.4	1.3	20	20.1	0.96	3.9	8
	Mean	23.3	2.0		117	18.1	4.0		49
(b) L3 clast	710	17.9	8.9	43	35	15.9	6.2	30	14
(c) L5 clast	714	23.5	0.66	1.9	59	19.2	2.6	6.7	27
(d) Melt rock clast	711	21.9	0.27	0.9	45	+			
(e) Tridymite-rich inclusion	709	-	-	-	-	20.0	1.03	4.3	29

*PMD, percent mean deviation of FeO analyses in wt %.
+Only high-Ca pyroxenes found with Wo 3.0-12.4.

Fe,Ni that show a martensitic texture on etching. Clast boundaries are well defined in many places, but where the objects contain transparent chondritic material or the host contains shock veins, the boundaries are ill defined (Fig. 3).

Electron microprobe analyses of the larger relict olivine and low-Ca pyroxene crystals are given in Table 3. Olivines in three of the five clasts are distinctly richer in FeO than host olivines (Fig. 2). All but one of the objects have mean fayalite concentrations in the L group range; the exception is only marginally lower. Olivine and pyroxene data suggest that the precursor material was largely type 4 for clasts in two of the sections (649 and 710), but possibly a higher type for two others (712 and 715). The black clast in section 710 also contains an L3 clast (see below), suggesting that the precursor material for the shock-blackened clasts was a breccia or regolith containing type 3 and type 4 or 5 material.

The oxygen isotopic composition of one shock-blackened clast is $\delta^{18}O$ = 4.88‰, $\delta^{17}O$ = 3.70‰. These values are marginally higher than those of the host and most L chondrites. However, the displacement from the terrestrial fractionation line, $\delta^{17}O - 0.52\ \delta^{18}O$, is 1.16‰, a value which is typical for equilibrated L chondrites (mean is 1.11 ± 0.08‰). These data and the silicate compositions (Table 3) show that the shock-blackened clasts were derived from L group chondritic materials.

L3 Chondritic Clast

A light-colored clast or region in the shock-blackened clast shown in Fig. 3 was found to be much more heterogeneous than other host or clast material (Table 3 and Fig. 2C). Olivine heterogeneity suggests a type 3.5 classification (*Sears et al.,* 1982). A single barred olivine chondrule ranges in composition from Fa 1 to Fa 20. The presence of chondrules with mesostases of slightly turbid glass is consistent with a type 3 classification for this clast.

L5 Chondritic Clast

The largest light-colored clast that was observed is 2.5 cm in length and our petrologic studies indicate that it is an L5 chondritic clast. Chondrules are less well defined than those of the host, and olivines and low-Ca pyroxenes (Fig. 2D) are more homogeneous than those in the host. Mean olivine compositions in the clast, Fa 23.5 ± 0.7, are close to those of the neighboring host, Fa 22.8 ± 0.8 (Tables 3 and 1). Metallic Fe,Ni grains on etching show a range of structures like those observed in the host. Residual patches of cloudy taenite show that the L5 material was originally slowly cooled; plessite and polycrystalline kamacite indicate that it was subsequently shock heated. Silicates in the clast indicate a shock facies of b or c according to the scheme of *Dodd and Jarosewich* (1979).

Porphyritic Melt Rock Clast

One well-defined, light-colored clast, 8 mm in length, was found to have a porphyritic texture with little metallic Fe,Ni (Fig. 1). It contains euhedral to subhedral olivine phenocrysts up to 1mm in length, embedded in a matrix of smaller subhedral to anhedral olivine crystals, poorly resolvable pyroxene crystals and turbid brown glass. Microprobe analysis shows that the large and small olivines are homogeneous and have similar compositions of Fa 21.9 ± 0.3 and 0.054 ± 0.026 wt % CaO (Table 3). This fayalite concentration is slightly lower than that of the neighboring host (Fa 22.9 ± 0.3) and is matched by only one host region (section UNM 712; Table 1). Ten pyroxenes were found to have a mean composition of Fs 19 ± 2, Wo 5 ± 3. Shock features in the clast are similar but not identical to those of the host: olivine has undulatory extinction, small grains contain partly resorbed, cloudy taenite

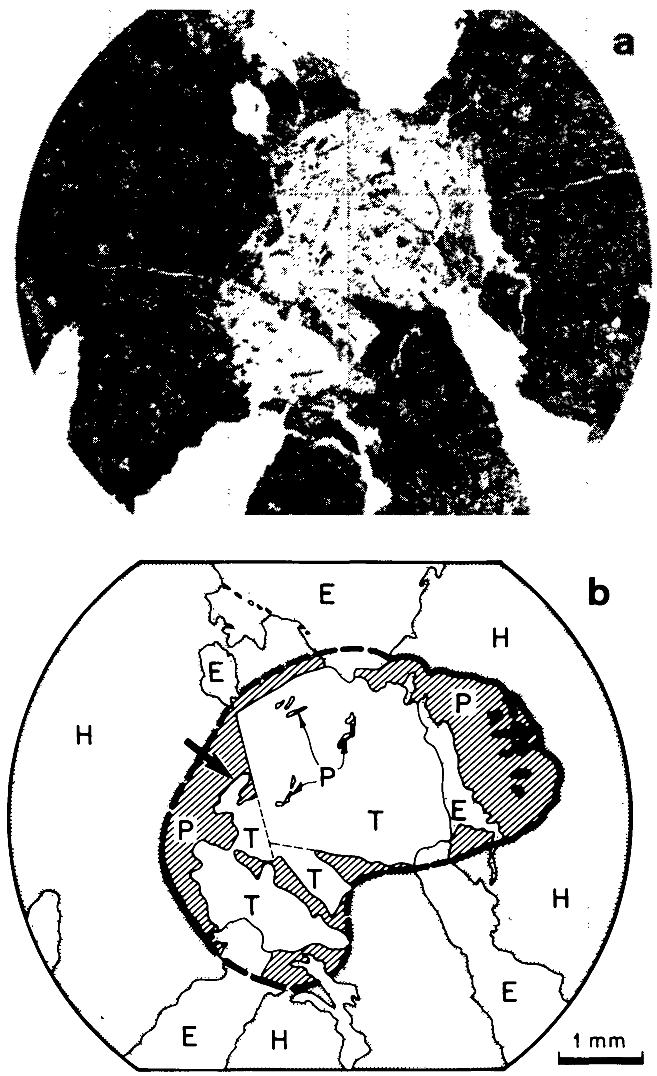

Fig. 4. Tridymite-rich inclusion in the Kendleton breccia (section UNM 709); **(a)** transmitted light photomicrograph, **(b)** sketch showing euhedral and subhedral tridymite (T) enclosed by low-Ca pyroxene (P), host (H) and epoxy (E). Tridymite occurs in four distinctly bounded segments, which are euhedral or subhedral in places; large arrow points to small tridymite crystal with terminating faces. Tridymite is surrounded by an ill defined rim of low-Ca pyroxene, which does not appear to have formed by metamorphic reaction of silica and olivine, unlike other silica-pyroxene occurrences in types 4-6 chondrites.

and tiny taenite grains in kamacite. Lack of plessite and polycrystalline kamacite is more likely to reflect the low abundance of metal rather than a different shock history for host and clast.

This clast is too large and irregular to be a porphyritic chondrule and is very likely to be an impact melt rock clast lacking relict material. In this respect it resembles four melt rock clasts in Dimmitt (*Rubin et al.,* 1983a) and some in other chondrites (*Keil,* 1982).

Tridymite-rich Inclusion

A single tridymite-rich inclusion about 5 mm in diameter was found in the Kendleton host. We describe it as an inclusion, not a clast, because there is no textural evidence that it was broken from a preexisting rock, although due to the large size of the tridymite this is likely. Tridymite occurs in four segments, which are euhedral and subhedral in places, and is largely enclosed by an ill-defined zone of low-Ca clinopyroxene (Fig. 4). Troilite grains are present in the clinopyroxene rim, and traces of troilite, chromite, metallic Fe,Ni and K-feldspar (?) were found within the tridymite. The tridymite is heavily fractured and largely isotropic, due to shock. The optical and electron probe identification of tridymite was confirmed by X-ray powder diffraction (G. Lumpkin, private communication, 1984). The rim of low-Ca pyroxene and the small stringers within the tridymite have identical compositions, Fs 20.0 ± 1.0, Wo 1 (Table 3, Fig. 2F).

Oxygen isotopic analysis of the tridymite (exclusive of visible impurities of low-Ca pyroxene and other phases) gave 5.71‰ for $\delta^{18}O$ and 3.70‰ for $\delta^{17}O$ (relative to SMOW). The displacement from the terrestrial fractionation line, $\delta^{17}O - 0.52\ \delta^{18}O$, is 0.73‰. This value is much less than that for L chondrites (1.11 ± 0.08‰) but entirely appropriate for equilibrated H chondrites (0.75 ± 0.08‰).

DISCUSSION

Tridymite-rich Inclusion

The tridymite-rich inclusion in Kendleton is poorly defined (Fig. 4). We do not know whether the outer edge of the low-Ca pyroxene zone marks the edge of the object in which the tridymite crystal formed. This object might include some of the material around the pyroxenes or it is possible that the pyroxene-tridymite inclusion is a fragment of a much larger object.

Silica-pyroxene aggregates greater than a few millimeters in size have been described in several ordinary type 5-6 chondrites: Farmington (*Binns,* 1967), Nadiabondi (*Christophe Michel-Lévy and Curien,* 1965), ALHA76003 (*Olsen et al.,* 1981), Knyahinya and Lissa (*Brandstätter and Kurat,* 1985). In addition silica has been reported in a large silica-pyroxene-plagioclase intergrowth in Vishnupur (*Wlotzka et al.,* 1983) and many chondrules and millimeter and submillimeter sized inclusions (*Fredriksson and Wlotzka,* 1985; *Brigham et al.,* 1986, and references therein). In all cases where the silica polymorph was identified, it was found to be cristobalite, except in

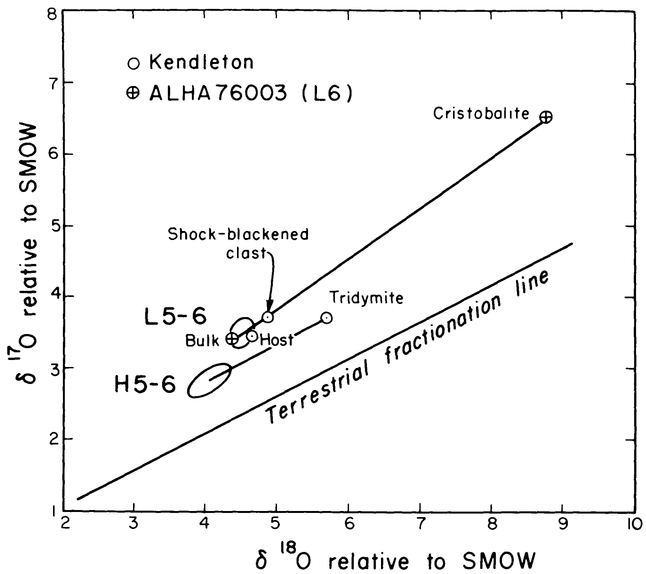

Fig. 5. Oxygen isotopic composition of the host, a shock-blackened clast and tridymite from a tridymite-rich inclusion in the Kendleton breccia; also shown are data from *Olsen et al.* (1981) for Allan Hills A76003 (L6) and its cristobalite inclusion. The host and shock-blackened clast of Kendleton have O isotopic compositions typical of equilibrated L chondrites. The Kendleton tridymite, like the cristobalite in ALHA76003, does not plot on a fractionation line through the L5-6 composition. The composition of the tridymite is consistent with equilibration with H5-6 chondrites at 900°C, however, other evidence suggests that it may have come from an H-like asteroid and not the H chondrite asteroid.

Vishnupur, where it is tridymite (*Wlotzka et al.,* 1983). The occurrence of tridymite in Kendleton is also unusual in that most of the surrounding pyroxene is not a reaction rim between silica and olivine as in Nadiabondi, Farmington, ALHA76003, Knyahinya and Lissa, because the tridymite is euhedral in places. However, the relative homogeneity of the pyroxene associated with the Kendleton tridymite (Fig. 2F, Table 3) may reflect metamorphism in an L4-6 environment.

Conditions favoring silica formation in chondrites have been discussed at length by *Brigham et al.* (1986). They argue that silica-low-Ca-pyroxene objects could not have formed by planetary igneous processes because of the low concentrations of Ca and Al in these objects. Thus, unless the Kendleton inclusion is a very unrepresentative clast of a feldspar-bearing rock, it is unlikely that it formed by planetary igneous fractionation.

Oxygen isotopic compositions of the silica occurrences in Kendleton and ALHA76003 (*Olsen et al.,* 1981) are shown in Fig. 5; other occurrences have not been analyzed. In both cases, the silica could not have formed by any kind of differentiation process that operated on bulk L chondrites. Olsen et al. conclude that their cristobalite must be a foreign grain that accreted into the L chondrite parent body. The Kendleton tridymite has an oxygen isotopic composition that is quite different from that of the ALHA76003 cristobalite, but as noted above, it is compatible with an H chondrite source.

The ^{18}O fractionation between quartz and pyroxene at 900°C has been experimentally determined to be 1.5‰ (*Matthews et al.,* 1983). Hence pyroxene in equilibrium with the Kendleton tridymite at 900°C would have $\delta^{18}O = 4.21$, $\delta^{17}O = 2.92$; for associated olivine, values would be a little lower. These values agree well with the means for equilibrated H chondrites: $\delta^{18}O = 4.13$, $\delta^{17}O = 2.90$. From the study of *Olsen et al.* (1981), we know that large cristobalite crystals can retain an H group oxygen isotopic composition during metamorphism in an L4-6 environment while the Fe/Mg ratio in associated low-Ca pyroxene changes from an H to an L value.

Four centimeter-sized, igneous-textured clasts with oxygen isotopic compositions close to those of equilibrated H chondrites have also been found in L4-6 chondrites. Two similar olivine-rich clasts in the L6 chondrites, Yamato 75097 and 793241, which are believed to be paired, have mean oxygen isotopic compositions of $\delta^{18}O = 4.54$, $\delta^{17}O = 3.05$ and $\delta^{18}O = 3.91$, $\delta^{17}O = 2.46$, respectively (*Clayton et al.,* 1984; *Prinz et al.,* 1984; *Mayeda et al.,* 1987). The occurrence of a relict chondrule in the Yamato 793241 clast implies that both clasts were derived from a chondritic source (*Prinz et al.,* 1984). The third clast, which was found in Bovedy (L4), has $\delta^{18}O = 4.3$ and $\delta^{17}O = 2.6$, and is thought to have formed from a chondritic source by impact melting (*Rubin et al.,* 1981). The fourth clast was found in the Barwell (L6) chondrite (*Hutchison et al.,* 1986) and has $\delta^{18}O = 4.14$ and $\delta^{17}O = 2.73$. All clasts except that in Bovedy, which has heterogeneous silicates, have olivines with compositions appropriate to equilibrated L chondrites. Besides chondrules and inclusions in L3 chondrites (*Gooding et al.,* 1983) and inclusions of carbonaceous chondrites (*Wilkening,* 1977), only one other inclusion in L chondrites has been found with an anomalous oxygen isotopic composition. A 7-mm-long igneous inclusion in Roosevelt County 014 (L5) has an oxygen isotopic composition close to those of EL chondrites (*Scott et al.,* 1986). Chondritic inclusions with H-like oxygen isotopic compositions have not been found in L chondrites.

Some of the four igneous clasts with H-like oxygen isotopic compositions may have been derived from H chondrite or H-like projectiles that melted on impact with the L chondrite parent body, but the Kendleton tridymite inclusion is so different in mineralogy and texture from chondritic impact melts that it is unlikely to have formed in this way. The absence of reports of tridymite inclusions in H chondrites and H chondritic clasts in L chondrites suggests that the Kendleton tridymite inclusion was not derived directly from an H chondrite projectile. L chondritic clasts have not been found in H and LL chondrites, although the presence of an H chondritic clast in an LL chondrite, St. Mesmin, and an LL clast in an H chondrite, Dimmitt (see references in *Clayton et al.,* 1984), suggests that H and LL asteroids may have experienced limited exchange of material. Because there are iron meteorites that formed on asteroids with oxygen isotopic compositions indistinguishable from those of H, L, and LL chondritic asteroids (*Clayton et al.,* 1983), it is possible that the Kendleton tridymite inclusion and some of the four igneous clasts with H-like oxygen are samples of additional asteroids with oxygen isotopic compositions like those of ordinary chondrites. However, it is also possible that the Kendleton inclusion was added to the L body during accretion and was never part of another body.

Shock-Blackened Clasts

We suggest that the shock-blackened objects in Kendleton, like similar objects in Dimmitt (DT7; *Rubin et al.,* 1983a) and Nulles (clast 5; *Williams et al.,* 1985), are clasts from preexisting rocks. However, *Fodor and Keil* (1976) conclude that similar objects in the Plainview regolith breccia, which they call "pseudo-fragments," formed in situ. They note that silicates in these objects are compositionally identical with those of the host, and veins or protrusions of shocked material appear to extend into the host. We have examined three of these objects in Plainview (PV10 in University of New Mexico [UNM] section number 281, PV12 in UNM 274, and one in UNM 282) and conclude that at least two of them are clasts. (Clast PV10 is too close to the edge of the section to make any judgement).

We have compared the Plainview and Kendleton shock-blackened clasts with shocked regions in several heavily shocked chondrites, such as Walters, Orvinio, and Rose City, and conclude that the clasts differ from the regions in two ways. The clasts appear to be largely equant, not vein-like, and textural and mineralogical features in shock-melted parts are not aligned parallel to the edge of the shock-blackened areas. We have observed a large equant pocket of shock melt that contains clasts in the Bluff (L5) chondrite (section L6986 from the Naturhistorisches Museum, Vienna). However, it clearly formed in situ as the edges of olivines adjacent to the melt are shock granulated, and there are extensive protrusions of melt in the surrounding host. For three of the five Kendleton objects, our conclusion that they did not form in situ is strengthened by small but significant compositional differences between clast and host olivines (Fig. 2).

We recognize that the boundaries of shock-blackened objects in Kendleton and Plainview are indistinct in many places, but we attribute this to the host-like chondritic regions within the shock-blackened rock (Fig. 3), which camouflage the clast-host boundary. In addition, in both chondrites there are small shock veins in the host material, which intersect clasts and were formed in situ. These features convey the false impression that the shocked clasts formed in situ.

Thermal and Shock History of Kendleton's Ingredients

To help understand when the host material and clasts were mixed and lithified, we have summarized some of their metamorphic and shock features in Table 4. Host material is largely type 4 with some type 3 material. Since host olivine and low-Ca pyroxene have concentrations of iron that are consistently low for L chondrites, the host material must be derived from only a part of the total L4 reservoir.

The shock-blackened clasts did not experience any planet-wide metamorphism and associated slow cooling because martensite, which is characteristic of rapid cooling (*Heymann,* 1967; *Smith and Goldstein,* 1977), is present. We do not know

TABLE 4. Comparison of metamorphic and shock features of host and clasts in the Kendleton breccia.

Feature	Host Material	Clasts			
		Shock bl.	L5	Melt-rock	SiO_2-rich
Metamorphic history					
Petrologic type	4 (3)	4 (3,5)	5	3?	4
Cloudy taenite*	yes	no	yes	yes	-
Shock history					
Silicate shock level	b or c	e or f (b/c)	b or c	b or c	b, c or d
Metallic Fe,Ni†	ct, pk, pl	ct, pk, ma	ct, pk, pl	ct	-

*Cloudy taenite is a submicroscopic mixture of tetrataenite and martensite (*Reuter et al.,* 1986).
†Shock-heating characteristics of metal: ct, clear taenite (i.e., resorbed cloudy taenite); ma, martensite; pk, polycrystalline kamacite; pl, plessite.

whether the L5 clast was mixed with host material after or during late-stage metamorphism, as metallographic cooling rates could not be obtained. The melt rock clast contains cloudy taenite, which is characteristic of slowly cooled irons and chondrites (*Reuter et al.,* 1986), so that the impact melting must have occurred before metamorphism had ended. Turbid glass and heterogeneous pyroxenes in the melt rock clast indicate that this clast was metamorphosed less than type 4 chondrites. Olivine in the clast may have been homogenized during metamorphism, but homogenization during crystallization cannot be excluded. The tridymite-rich inclusion probably experienced type 4 levels of metamorphism in view of the abundance and homogeneity of the clinopyroxene.

Thus the strongest constraint on the time of lithification of the Kendleton breccia is provided by martensite in the shock-blackened clasts. Studies of similar metal in Ramsdorf and Orvinio (*Smith and Goldstein,* 1977) suggest that cooling rates below 700°C were higher than 10^{-4}°C sec^{-1} or 10^{3}°C yr^{-1}. Thus, Kendleton, like chondritic regolith breccias, must have been lithified after planet-wide metamorphism had ended, because metamorphic cooling rates were 10^{-6} to 10^{-3}°C yr^{-1}.

All of the material in Kendleton has been shocked to low or high levels. Thus it is possible that the low shock to b or c levels occurred after lithification of the breccia. The only feature that might require lower levels of shock after lithification is the martensite in parts of the shock-blackened clasts. We do not know the relative rates at which cloudy taenite and martensite are converted into plessite. If martensite decomposes faster, this would suggest that the host was shocked prior to lithification. It is possible that a single event produced the shock-blackened material and shocked the host material, but multiple events cannot be excluded. A much earlier impact formed the porphyritic melt rock.

Comparison with Regolith Breccias

Noble gas analyses of Kendleton (Table 2) show no evidence for the presence of solar-wind gases, nor does it seem likely that solar-wind gases were lost. Shock reheating of Kendleton is responsible for its low ^{4}He and ^{40}Ar retention ages of 2.6 and 3.6 Gyr (*Taylor and Heymann,* 1969). However, heating was relatively mild as resorption of cloudy taenite, which is incomplete in Kendleton host, takes only a few seconds at 700°C in the atmospherically heated rims of iron meteorites (*Buchwald,* 1975, Figs. 45 and 53). The high concentration of trapped ^{36}Ar also suggests that solar wind gases were not removed by shock heating. Finally, because *Binns* (1968) found martensite with 14-18% Ni and kamacite with 7-8% Ni throughout the host and clasts of a gas-rich, regolith breccia, Ghubara, we infer that shock heating much more severe than that experienced by the Kendleton host is insufficient to remove solar-wind gases.

Except for the absence of solar-wind gases and associated irradiation features, Kendleton has nearly all features found in ordinary chondrite regolith breccias (*Keil,* 1982). It has a dark matrix that encloses a wide variety of light- and dark-colored clasts. The chondritic clasts in Kendleton range from type 3 to type 5; some clastic material was formed by impacts during metamorphism, other material was shocked after metamorphism, and the rock contains a unique tridymite-rich inclusion that probably comes from another body. The only petrographic feature that is somewhat unusual for regolith breccias is the high abundance of host, about 80%. In the six regolith breccias for which quantitative estimates have been made, matrix abundances are 40-70% (*Rubin,* 1982; *Rubin et al.,* 1983a; *Williams et al.,* 1985). However, the LL3 regolith breccia, Ngawi, appears to have a matrix abundance that is comparable to that of Kendleton.

Origins of Fragmental and Regolith Breccias

Kendleton has the diversity of clasts expected in regolith breccias but lacks the irradiation features associated with them. Other possible examples of such chondritic breccias include L'Aigle (*Keil,* 1982), Castalia (*Taylor and Heymann,* 1969; *Van Schmus,* 1969), and Waconda (*Weber et al.,* 1983; *Rubin,* 1982). Presumably material in these fragmental breccias, unlike that in regolith breccias, was not close to the asteroid surface for any significant period. However, recent studies of gas-rich meteorites suggest that conventional ideas on the origin of meteoritic regolith breccias may need revision.

Caffee et al. (1983, 1986, 1987) find very large enrichments of cosmogenic Ne in solar-flare irradiated grains in Murchison and Kapoeta, which are absent in nonirradiated grains. They argue that the most likely source is solar cosmic rays from a more active, early sun. Since Kapoeta and other regolith breccias contain some clasts with ages of 3.7 Gyr or less (*Rajan et al.*, 1979; *Schultz and Signer*, 1977; *Keil et al.*, 1980), this requires that irradiated grains were stored inside the parent body for as much as 10^9 years before mixing with other fragmental material.

In order for solar-wind irradiated grains to be introduced into meteorite breccias in this way, these grains must have been loosely consolidated inside the parent body, so they could have been readily mixed with fragmental material over the next 10^9 years. Since many ordinary chondrites are fragmental breccias that were lithified after metamorphism (*Scott et al.*, 1985; *Rubin et al.*, 1983b), there must have been much mixing of weakly lithified or unconsolidated chondritic material after metamorphism as a result of cratering, spallation, and disruption and reassembly of parent bodies. The results of *Caffee et al.* (1983, 1986) do not exclude the possibility that some early irradiated grains were mixed with recently irradiated grains in a regolith as their analyses were made on sets of 6-11 grains. However, it is also possible that most of the material in a meteorite containing solar-flare irradiated grains was never present near the surface of an asteroid. Then the proportion of irradiated grains in a meteorite may have no simple bearing on the proportion in an asteroidal regolith, and the term "regolith breccia" may be inappropriate for some meteorites that contain solar-flare irradiated grains.

One possible test of this scenario would be a search for chemical differences between irradiated and unirradiated grains in the hosts of meteorite regolith breccias. Because most irradiated and unirradiated grains are postulated to have different origins, chemical differences should be detectable in favorable cases. *Wilkening et al.* (1971) analyzed 110 pyroxenes in Kapoeta and did not observe any significant compositional differences between irradiated and unirradiated grains. However, both groups of pyroxenes have a wide compositional range in Kapoeta, so that further searches would be worthwhile.

If gas-rich meteorites did not form in surface regoliths, it is possible that Kendleton and some other gas-poor, fragmental breccias formed in the same way as gas-rich breccias except that Kendleton et al. failed to acquire an admixture of early irradiated grains. If some gas-rich meteorites were lithified in asteroidal regolith, however, all of the constituents of Kendleton et al. may also have resided in a regolith, but for too short a period for surface grains to have acquired detectable solar-wind gases.

Acknowledgments. We thank G. J. Taylor for discussions, G. Lumpkin for technical assistance, and A. E. Rubin for a helpful review. This work was partly supported by NASA Grant NAG-9-30 (Keil) and NSF Grant EAR 8316812 (Clayton).

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