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The Antarctic achondrite, Grove Mountains 021663: An olivine-rich winonaite

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Abstract–The Grove Mountains (GRV) 021663 meteorite was collected from the Grove Mountains region of Antarctica. The meteorite is composed primarily of olivine ($Fa_{5.4}$), orthopyroxene ($Fs_{4.7}Wo_{3.0}$), chromian diopside ($En_{53.6}Fs_{2.4}Wo_{44}$), troilite, kamacite, and plagioclase ($Ab_{74.5}Or_4An_{21.5}$). Minor phases include schreibersite and K-feldspar. The meteorite is highly weathered (W3) and weakly shocked (S2). We determine a whole rock oxygen isotopic composition of $\delta^{18}O = 7.50‰$, $\delta^{17}O = 3.52‰$. Comparisons of these data with other primitive achondrites have resulted in the reclassification of this meteorite as a member of the winonaite group. The occurrences of troilite, metal, and schreibersite in GRV 021663 indicate that these minerals were once completely molten. Euhedral inclusions of pyroxene within plagioclase further suggest that these may have crystallized from a silicate melt, while the depletion of plagioclase, metal, and troilite indicates that GRV 021663 could represent a residuum following partial melting on its parent asteroid. Trace element distributions in silicate minerals do not, however, confirm this scenario. As with other winonaite meteorites, the formation of GRV 021663 probably relates to brecciation and mixing of heterogeneous lithologies, followed by varying degrees of thermal metamorphism on the parent body asteroid. Peak metamorphic conditions may have resulted in localized partial melting of metal and silicate mineralogies, but our data are not conclusive.

INTRODUCTION

The primitive achondrite meteorites, winonaites, are characterized by having chondritic mineralogies and chemical compositions, but achondritic, recrystallized, textures. For the most part, they are fine-to-medium grained, equigranular rocks; some samples, however, contain what appear to be relic chondrules (e.g., Pontlyfni, Mt. Morris [Wisconsin] and Northwest Africa (NWA) 1463; Benedix et al. 1998, 2003). Their oxidation states are intermediate between those of enstatite and H chondrites, and FeNi-FeS veins that constitute the first partial melts of a chondritic precursor material are common (see Figs. 1a and 2a). Meteorites of this group are primarily composed of orthopyroxene, olivine, diopside, plagioclase, troilite, Fe-Ni alloy, and minor amounts of chromite and phosphate. In addition, they may contain characteristic accessory minerals, such as alabandite, daubreélite, schreibersite (Benedix et al. 1998). Amphibole is reported in at least one sample (Hammadah al Hamra [HaH] 193; Floss et al. 2007).

There are several attributes that collectively distinguish the winonaites from other achondrite meteorite groups, such as (1) highly reduced mineral assemblage(s), (2) distinct oxygen isotopic compositions, (3) the abundance and distribution of metal(s) and troilite, and (4) characteristic accessory mineralogies (Benedix et al. 1998). Winonaites may derive from a

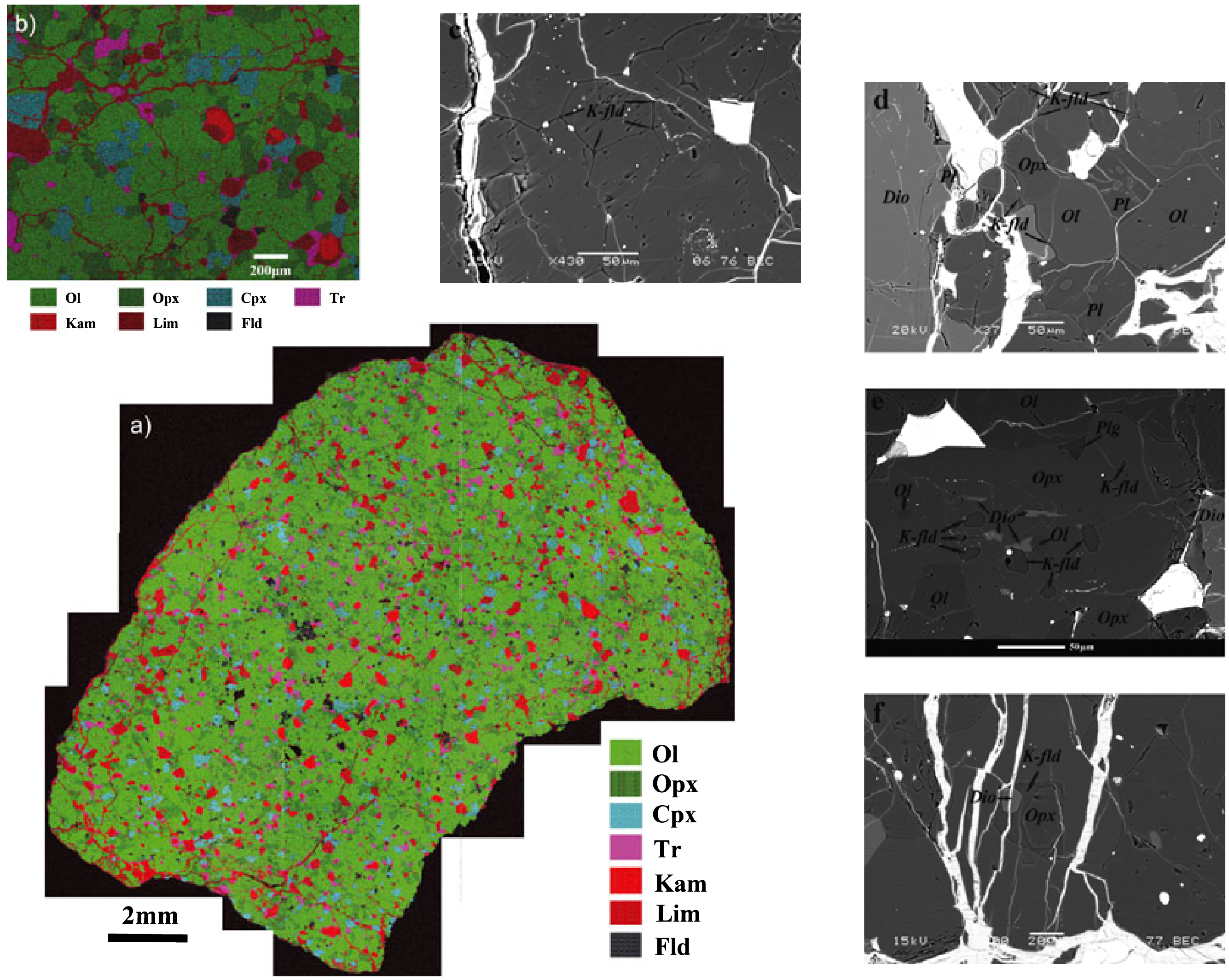

Fig. 1. Elemental composite and backscattered electron images of GRV 021663. a) Composite image of thin section GRV 021663-2, combining 14 individual X-ray element maps obtained for the elements: Ca, Mg, Fe, and S. b) One of these combined single area element map images that highlights the textural relationships and abundances of the various minerals present. c) Clearly evident are 120° triple junctions among the silicate minerals in GRV 021663. The fractures between silicate grains are filled with K-feldspar. Scale bar represents 50 μm. d) Interstitial, plagioclase crystals poikilitically enclose mafic silicate grains, two of which are euhedral. The fractures between silicate grains are filled with K-feldspar. e) Silicate melt and olivine inclusions in orthopyroxene consist principally of K-feldspar, with small clinopyroxene daughter crystals along the inclusion walls. Olivine boundaries abutting their pyroxene host are rounded but the inclusions are not suggesting they may have formed from a melt. f) Inclusions in orthopyroxene consist of clinopyroxene and orthopyroxene daughter crystals, set in a matrix of K-feldspar. Both the chromian diopside and orthopyroxene daughter crystals are euhedral. Opx = orthopyroxene, Ol = olivine, Dio = chromian diopside (clinopyroxene), Pl = plagioclase, K-fld = K-feldspar.

common parent body with the IAB and IIICD irons (Bild 1977; Benedix et al. 2000 and references therein), but differ in that they have experienced extensive metamorphism and partial melting of troilite and Fe-Ni metal. For the silicate minerals, Benedix et al. (1998) argued, on the basis of textural observations of clinopyroxene- and plagioclase-rich areas in the Pontlyfni meteorite, and coarse-grained, olivine-rich lithologies in both Winona and Mt. Morris, that partial melting also affected these silicate assemblages to some degree. More recent studies of Winona and several of the IAB iron silicate inclusions, thought to relate to the winonaites, provide further support for melting at peak metamorphic conditions associated with the possible collisional fragmentation and reassembly on the parent body of these meteorites (Benedix et al. 2005). However, this is in contrast to the work of Kimura et al. (1992), who suggested that the silicate fractions of winonaite

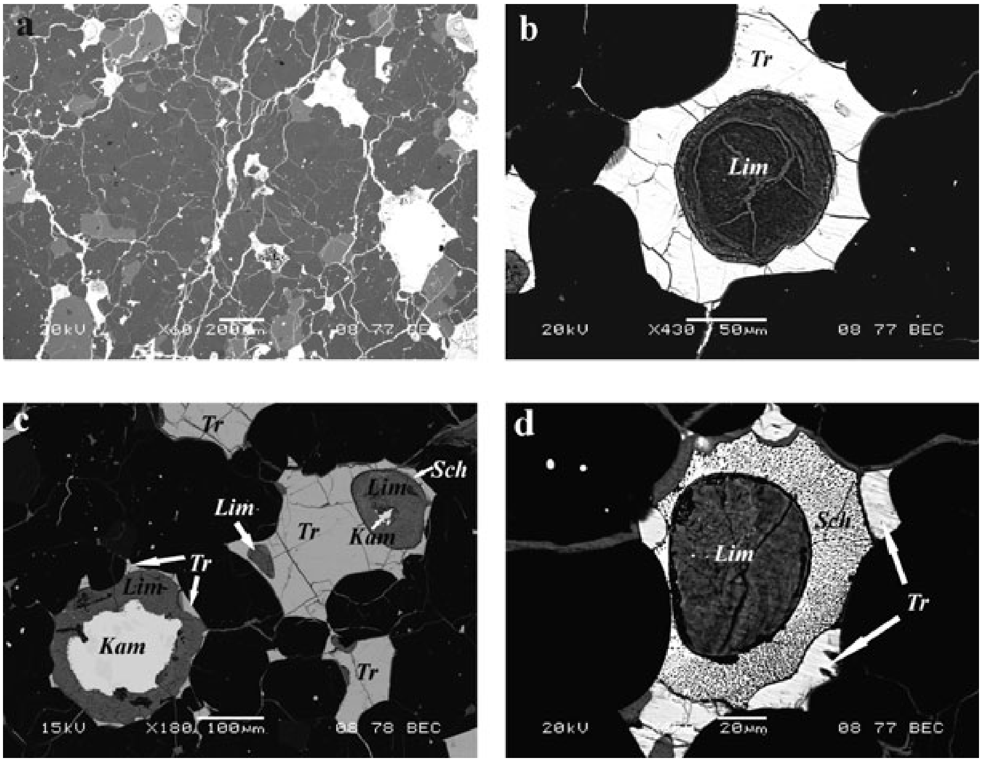

Fig. 2. Backscattered electron images of occurrences of opaque minerals in GRV 021663. a) Extensive troilite and limonite veining. b) A limonite grain is completely enclosed by troilite. c) Kamacite and limonite grains are similarly enclosed by troilite and schreibersite. d) A limonite grain is almost fully enclosed by schreibersite. Tr = troilite, Lim = limonite, Kam = kamacite, Sch = schreibersite.

meteorites did not show any textural and mineralogical evidence for melting. Moreover, the distributions of trace elements within the winonaites lack any clear evidence suggestive of silicate partial melting (Floss et al. 2008). The question of whether or not silicate minerals experienced some degree of partial melting is still, therefore, open to question.

To date, only 20 winonaites have been identified, most of which have been recovered from hot and cold desert environments. The meteorite, Grove Mountains (GRV) 021663, was recovered by the Chinese Antarctic Research Expedition in 2003 from No. 4 moraine in the Grove Mountains region of Antarctica (Lat. 72°46′25″ S, Long. 75°20′23″ E). The meteorite weights 2.19 g and measures 14 × 14 × 18 mm in size; it is completely covered by a black fusion crust. While GRV 021663 has been classified as an acapulcoite (Weisberg et al. 2009), data presented in this paper support the reclassification of this meteorite as a member of the winonaites. Accordingly, in this paper, we present a detailed study of the petrography, mineral chemistry, and whole rock oxygen isotopic composition of GRV 021663. In doing so, we provide further evidence for inclusion of this meteorite within the winonaite primitive achondrite group, and conclusions concerning its formation, origin, and parentage.

EXPERIMENTAL

Sample Description and Preparation

A polished thin section of GRV 021663 of standard thickness (30 μm) was prepared for mineralogical and petrographic investigation. This section, labeled GRV 021663-2, measures approximately 10 mm × 16 mm. For whole rock, oxygen isotopic analysis, a chip of approximately 100 mg was sawn from the meteorite, utilizing a low-speed diamond saw. The fusion crust was removed from the sample, and procedures for the analysis of a weathered meteorite (after Clayton et al. 1991) were subsequently followed.

Experimental Procedure

The petrography and identification of minerals present in the GRV 021663 sample were carried out on an Olympus optical microscope and a JEOL JSM-6460LV scanning electron microscope, coupled to an EDAX Genesis energy-dispersive X-ray spectrometer. Shock effects and the degree of weathering were also observed at this time. The mineral distribution in GRV 021663 is homogeneous at all scales. Determinations of modal mineralogy were completed with the aid of image analysis facilities within Adobe Photoshop. Modal abundances were measured in four discrete rectangular areas within the GRV 021663-2 thin section (each approximately 1.6 × 2 mm).

Electron probe microanalysis (EPMA) was performed utilizing a Shimadzu EPMA-1600 instrument in wavelength-dispersion mode, at the State Key Laboratory of Ore Deposit Geochemistry, Institute of Geochemistry, Chinese Academy of Sciences. Operating conditions were: accelerating voltage of 25 kV, beam diameter of approximately 10 μm. Conventional ZAF corrections were applied to all mineral analyses. A variety of well-characterized minerals and synthetic phases were used as standards: olivine (for Mg, Ni, and Fe in olivine and pyroxene), plagioclase (Al, Si, and Ca in olivine and pyroxene; Na, K, Al, Ca, and Si in feldspar), pyrope garnet (Ti, Cr, and Mn in olivine and pyroxene; Mg, Fe, and Ti in feldspar), Fe metal (Fe in metal and sulfide), Fe_3P (P in metal), Ni metal (Ni), marcasite (S in metal and troilite), silicon (Si), Cr metal (Cr), Co metal (Co), and Cu metal (Cu).

Trace element analysis of silicate minerals was carried out using a laser-ablation microprobe coupled to an inductively couple plasma–mass spectrometry (ICP-MS) instrument at the Northwest University, Xi'an, China. The laser-ablation system is a GeolLas 200M, equipped with a 193 nm ArF excimer laser and a homogenizing imaging optical system. The ICP-MS is an Agilent 7500a. The laser beam diameter used for ablation was approximately 44 μm. Data were normalized to known Si contents of the analyzed minerals and were reduced using GLITTER software. NIST 612 was used as an external standard. Analytical details and results for NIST SRM (610, 612, 614) standards have been reported in Gao et al. (2002).

Oxygen isotope measurements at LSU OASIC were conducted on a Finnigan MAT253 instrument. All measurements were run above certain thresholds of gas pressure (~20–25 mbar) and are based on an extrapolation of the V-SMOW measurements, assuming ideal linear mass-spectrometric performance (i.e., a single reference approach). The $\delta^{17}O$ value was initially calibrated against UWG-2, assuming a value for $\delta^{18}O = +5.8‰$ (Valley et al. 1995) and $\delta^{17}O = 0.520 \times \delta^{18}O = 3.016‰$. Errors for $\delta^{18}O$ and $\Delta^{17}O$ in each independent laser fluorination run were: ~0.1‰ (1σ) and ~0.05‰ (1σ), respectively.

Mineralogical and Petrographic Features

The meteorite, GRV 021663, consists predominantly of olivine, orthopyroxene, chromian diopside, troilite, and metallic Fe-Ni; minor phases include schreibersite ($(Fe,Ni)_3P$), plagioclase, and K-feldspar. Chromite, phosphate, amphibole, and taenite were not identified. In comparison with other winonaites, GRV 021663 contains more olivine and less orthopyroxene, plagioclase, metal, and troilite (see Table 1; Figs. 1a–c). In common with other winonaites, GRV 021663 exhibits a range of petrographic features that reflect the varying degrees of thermal metamorphism it has experienced (e.g., equigranular textures and 120° triple junctions between grains; Fig. 1c). The grain size variation for olivine, orthopyroxene, and chromian diopside is in the range of 50–250 μm (ave. 150 μm).

Olivine is the most abundant mineral in GRV 021663; it has a modal abundance of approximately 56 vol%. With the exception of a few locally distributed grains within orthopyroxene (small rounded inclusions ~10 μm in size), olivine grains are sub- to anhedral. Orthopyroxene and chromian diopside are texturally similar to the olivine and impart an overall hypidiomorphic texture to the rock. Modal abundances for orthopyroxene and chromium diopside are ~17 and ~8 vol%, respectively. The majority of olivine grains display undulatory extinction under cross-polarized light and, in accordance with the shock classification scheme for ordinary chondrites (after Stöffler et al. 1991), the shock stage of GRV 021663 is S2. Plagioclase feldspar and K-feldspar are both present as minor constituents (modal abundance of ~1.4 vol%). Plagioclase is the more abundant, occurring as sub- to anhedral grains of up to approximately 150 μm; it is mainly distributed in the interstices between the mafic silicates or as small inclusions within them (see Figs. 1a, 1b, and 1d). More rarely, plagioclase poikilitically encloses the mafic silicate grains (Fig. 1d), and is perhaps indicative of interstitial basaltic partial melt (McCoy 1994). K-feldspar occurs along the edges of plagioclase grains or infilling fractures between other minerals (Figs. 1c and 1d). K-rich feldspar is also present as inclusions of variable shape within mafic minerals (Figs. 1d and 1e). Importantly, some euhedral crystals of pyroxene occur as inclusions within feldspar (see Figs. 1d and 1f), implying also their crystallization from a silicate melt phase.

Opaque minerals, including metal, troilite, schreibersite, and limonite, account for approximately 17 vol% of GRV 021663 meteorite. Textural observations coupled with compositional data demonstrate that limonite is the main alteration/weathering product of kamacite metal because essentially no sulfur has been detected by electron microprobe analysis. Limonite and veins of troilite occur throughout the entire sample, forming an interconnecting network (Fig. 2a). The degree of weathering is characterized as W3 according to Wlotzka's (1993) criteria.

Troilite occurs in GRV 021663 dominantly as irregular, interstitial grains, but also in lower abundances

Table 1. Modal mineralogy (vol%) for GRV 021663, winonaites, and A–L clan meteorites.

Meteorite	Opx	Ol	Pl	Cpx	Amp	Php	Opq	Source
GRV 021663	17	56	1.4[a]	8	0	0	17	This work
NWA 1463(W)	46.4	8.6	6.5	7.7	0	0.7	30.6	Floss et al. (2008)
Pontlyfni(W)	29.9	20.8	10.5	3.6	0	Trace	35.1	Floss et al. (2008)
Winona(W)	53.1	14.6	11.4	0.6	0	0.4	20.0	Floss et al. (2008)
Mt. Morris(W)	47.5	16.2	6.8	2.9	0	0.1	26.5	Floss et al. (2008)
Tierra Blanca(W)	39.3	16.5	7.7	3.0	0	Trace	33.6	Floss et al. (2008)
HaH 193(W)	48.0	9.7	10.9	0.7	4.1	0.1	26.7	Floss et al. (2008)
Y-74025(W)	41.4	14.9	16	6.1	0	0	21.0	Yugami et al. (1998)
Y-75305(W)	36.5	5.8	8.6	0.4	0	0.01	48.2	Yugami et al. (1998)
Y-8005(W)	28.4	11.8	13.5	6.2	0	0	39.9	Yugami et al. (1998)
Y-74063(A)	46.2	24.3	7.2	6.2	0	Trace	16.2	Kimura et al. (1992)
ALH 77081(A)	43.6	25.3	13.4	3.4	0	1.1	13.2	Kimura et al. (1992)
Acapulco(A)	41.1	25.2	14.8	2.5	0	1.21	15.2	Yugami et al. (1998)
ALH 77081(A)	38.7	27.9	14.2	7.3	0	0.42	11.4	Yugami et al. (1998)
ALH 78230(A)	37.7	28.4	13.6	5.3	0	0.52	14.5	Yugami et al. (1998)
Y-8307(A)	32.4	25.9	13.5	6.7	0	0.90	20.7	Yugami et al. (1998)
Dho 290(A)	23.0	30.0	13.0	11.0	0	1.2	21.7	Patzer et al. (2004)
GRA 98028(A)	15.0	30.5	11.5	9.5	0	1.1	30.4	Patzer et al. (2004)
NWA 725(A)	11.0	29.0	11.0	9.0	0	1.0	37.7	Patzer et al. (2004)
NWA 1058(A)	14.0	29.0	10.0	6.0	0	0.5	40.2	Patzer et al. (2004)
TIL 99002(A)	28.0	38.0	12.0	4.0	0	1.4	16.4	Patzer et al. (2004)
EET 84302(A/L)	38.6	17.7	15.2	0.5	0	0.04	28.0	Yugami et al. (1998)
Gibson(L)	41.9	27.3	3.3	6.3	0	0.09	20.9	Yugami et al. (1998)
MAC 88177(L)	44.1	48.7	0	1.9	0	0.03	5.2	Yugami et al. (1998)
Y-74357(L)	9.7	71.9	0	4.7	0	0.09	13.7	Yugami et al. (1998)
Y-75274(L)	50.5	20.3	2.3	0.4	0	0	26.5	Yugami et al. (1998)
Y-791491(L)	32.0	57.4	0.2	0.2	0	0.16	10.1	Yugami et al. (1998)
Y-8002(L)	43.7	31.7	10.0	1.7	0	0.01	12.8	Yugami et al. (1998)

Notes: Mineral abbreviations follow those of Whitney and Evans (2010) except for Php = phosphate minerals. Opq = opaque minerals include: Fe-Ni metal, troilite, chromite, schreibersite and weathering products.

[a] Trace K-feldspar. W = winonaite, A = acapulcoite, L = lodranite.

as inclusions within mafic silicates, and as veins. Iron-nickel metal alloys in this meteorite are all low Ni, kamacite, and dominant grains are spherical or ovoid; nearly 60% of this material has been weathered to limonite. A limited quantity of metal grains occurs as inclusions within mafic silicates, mostly olivine. A significant textural characteristic found in this meteorite is the common occurrence of metal grains and their weathering products enclosed by troilite and/or schreibersite (Figs. 2b–d). These textures suggest that metal, troilite, and schreibersite may recrystallize from Fe-Ni-S-P eutectic melt (Rubin et al. 1999). Rubin (2007) has shown, for example, that schreibersite may form by the melting of P-bearing metal.

Neither chromite nor phosphate has been observed in this meteorite.

Mineral Compositions: Major Elements

Detailed electron microprobe analysis data for the mineral phases in GRV 021663 are reported in Tables 2 and 3.

The fayalite content of olivine varies between 5.3 and 5.6 mole% (ave. = 5.4 mole%); chemical zoning in olivine has not been observed. In orthopyroxene, the ferrosilite content ranges from 4.3 to 5.2 mole% between grains, while the wollastonite component does not exceed 3.0 ± 0.2 mole%. Chromium content for the orthopyroxene is high at 0.85–0.96 wt% Cr_2O_3. The composition of clinopyroxene is also fairly homogeneous in this meteorite; average values of En, Fs, and Wo are 53.6, 2.4, and 44.0 mole%, respectively. Cr_2O_3 contents of 1.44–1.57 wt% help classify this pyroxene as chromian diopside. Two-pyroxene thermometry indicates equilibrium temperatures of 1102 ± 30 °C (utilizing the QUILF95 program; Andersen et al. 1993).

From the analysis of five feldspar grains, we determine plagioclase compositions as (ave. of three grains: $Ab_{74.5}Or_4An_{21.5}$) and K-feldspar as (ave. of two grains $Ab_{60.3}Or_{32.4}An_{7.2}$). The compositional homogeneity of K-feldspar probably reflects more effective homogenization due to the small size of grains.

Metal present in GRV 021663 is Ni-poor kamacite: Fe content averages 96.4 wt%; Ni and Co are 3.01 and

Table 2. EPMA analyses of silicate mineral chemical compositions (in wt%) for GRV 021663.

	Olivine		Orthopyroxene		Chromian diopside			Plagioclase			K-feldspar	
	Av. (n = 10)	Range	Av. (n = 10)	Range	Av. (n = 8)	Range		1	2	3	1	2
SiO_2	41.72	41.01–42.62	57.59	56.15–58.52	56.85	56.46–57.25		62.33	63.90	63.94	70.44	70.02
TiO_2	0.03	0–0.04	n.d.	n.d.	0.77	0.66–0.88		0.07	0.05	0.07	0.57	0.64
Al_2O_3	0.01	0–0.03	0.41	0.34–0.48	0.93	0.84–1.01		22.19	21.84	21.97	17.78	17.40
Cr_2O_3	0.47	0.40–0.53	0.90	0.85–0.96	1.50	1.44–1.57		n.d.	n.d.	n.d.	n.d.	n.d.
FeO	5.18	5.01–5.33	3.15	2.88–3.55	1.48	1.28–1.68		0.95	0.23	0.22	0.78	1.00
NiO	0.01	0–0.04	n.d.	n.d.	0.02	0–0.04		n.d.	n.d.	n.d.	n.d.	n.d.
MnO	0.47	0.41–0.52	0.47	0.40–0.49	0.31	0.28–0.33		n.d.	n.d.	n.d.	n.d.	n.d.
MgO	50.74	50.31–51.03	34.37	33.79–34.74	18.12	17.78–18.37		0.06	0.03	0.05	0.89	0.96
CaO	0.03	0.02–0.04	1.59	1.31–1.69	20.69	20.32–21.06		4.49	4.14	4.16	0.83	0.88
Na_2O	n.d.	n.d.	n.d.	n.d.	0.61	0.51–0.65		8.25	8.25	8.08	3.74	4.18
K_2O	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.		0.64	0.68	0.69	3.27	3.18
Total	98.65	97.53–99.84	98.49	97.73–99.18	101.27	100.59–102.06		98.98	99.11	99.17	98.30	98.26
Fa/Fs	5.4	5.3–5.6	4.7	4.3–5.2	2.4	2.1–2.8	Ab	74.0	75.1	74.6	58.9	61.8
Wo			3.0	2.9–3.2	44.0	43.5–44.5	Or	3.8	4.1	4.2	33.9	31.0
							An	22.3	20.9	21.2	7.3	7.2

Note: n.d. = not determined.

Table 3. EPMA analyses of metal and troilite compositions (wt%) for GRV 021663.

	Si	Cr	S	P	Co	Fe	Cu	Ni	Total
kam1	b.d.	0.07	b.d.	0.77	0.29	97.8	b.d.	2.90	101.8
kam2	0.01	0.04	0.01	0.72	0.25	97.6	0.07	2.90	101.6
kam3	b.d.	b.d.	0.01	0.85	0.33	97.1	b.d.	3.13	101.4
kam4	b.d.	0.04	b.d.	0.56	0.25	97.5	b.d.	2.94	101.3
kam5	0.01	0.05	0.02	0.56	0.35	96.8	0.05	3.14	101
kam6	0.01	b.d.	0.03	0.81	0.18	97.0	b.d.	2.95	100.9
kam7	b.d.	b.d.	0.01	0.89	0.23	96.5	0.05	3.00	100.6
kam8	b.d.	0.04	0.03	0.85	0.20	96.4	0.07	3.04	100.6
kam9	b.d.	0.05	0.02	0.73	0.23	96.4	0.07	3.04	100.5
kam10	b.d.	0.05	0.04	0.69	0.22	96.4	b.d.	2.95	100.3
kam11	0.01	0.05	0.03	0.66	0.25	96.1	b.d.	3.04	100.1
kam12	b.d.	0.05	b.d.	0.73	0.26	95.5	b.d.	3.07	99.6
kam13	0.02	0.06	0.02	0.83	0.23	95.3	0.08	3.05	99.6
kam14	0.02	0.06	0.02	0.46	0.21	95.9	b.d.	2.87	99.5
kam15	b.d.	0.05	0.03	0.58	0.32	95.0	b.d.	3.16	99.1
kam16	b.d.	0.05	0.03	1.42	0.21	94.4	b.d.	2.90	99.0
Tr1	b.d.	1.29	34.1	n.d.	0.08	63.2	n.d.	b.d.	98.6
Tr2	0.03	1.18	34.8	n.d.	0.06	62.8	n.d.	b.d.	98.9
Tr3	0.01	1.07	34.9	n.d.	0.09	63.3	n.d.	b.d.	99.4
Tr4	0.01	1.80	33.9	n.d.	0.06	62.2	n.d.	b.d.	98.0
Tr5	b.d.	1.12	33.5	n.d.	0.09	62.5	n.d.	b.d.	97.1
Tr6	b.d.	1.10	34.7	n.d.	0.08	62.5	n.d.	b.d.	98.5

Note: kam = kamacite; Tr = troilite; b.d. = below detection limit; n.d. = not determined.

0.25 wt%, respectively. Phosphorus content in the kamacite is 0.46–1.42 wt% (ave. 0.76). The composition of troilite is homogeneous, with the exception of Cr, which ranges from 1.07 to 1.8 wt%.

Oxygen Isotopic Composition

Whole-rock oxygen isotopic compositions for GRV 021663 are plotted on a three-isotope diagram (Fig. 3). Multiple analyses of samples of GRV 021663 by laser fluorination give $\delta^{17}O = +3.54$, $+3.46$, and $+3.54$, $\delta^{18}O = +7.57$, $+7.40$, and $+7.53$, and $\delta^{17}O = -0.389$, -0.385, and -0.380 (all in ‰). Average values for $\delta^{18}O$ and $\delta^{17}O$ lie below the terrestrial fractionation line at 7.50‰ and 3.52‰, respectively. The GRV 021663 oxygen isotopic composition ($\delta^{17}O$ values) of GRV 021661 is similar to that of other winonaites, and the silicate inclusions in IAB and IIICD irons, and different

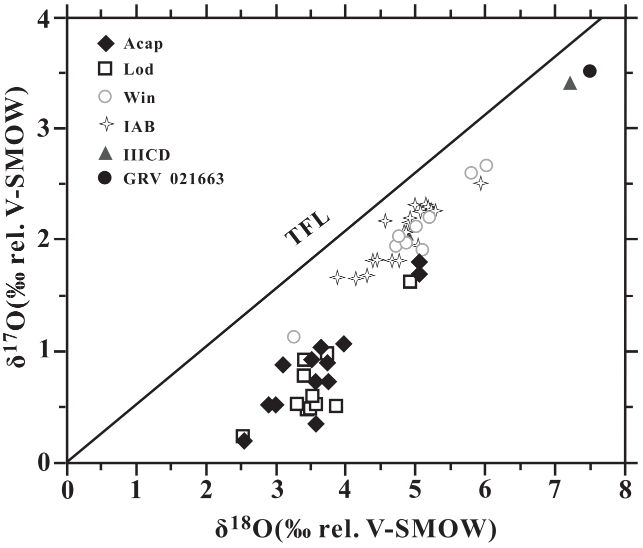

Fig. 3. Oxygen three-isotope diagram for GRV 021663 whole rock data. TFL is the terrestrial fractionation line. The oxygen isotopic compositions of acapulcoites (Acap), lodranites (Lod), winonaites (Win), and silicate inclusions in iron IAB and IIICD type meteorites are used for comparison with our data for GRV 021663. Other meteorite data derive from the literature: Mayeda and Clayton (1989), Clayton and Mayeda (1996), Grossman (2000), Russell et al. (2005), and Irving et al. (2007).

from that of the acapulcoites and lodranites (cf. Clayton and Mayeda 1996).

Trace and Minor Elements in Silicate Mineral Phases

While textural evidence in several winonaites suggests that these meteorites have experienced some degree of partial melting in their origin, trace element distributions in these lithologies do not show clear evidence for such melting (Floss et al. 2008). In an attempt to further constrain whether partial melting of silicate phases occurred in GRV 021663, and by analogy the winonaites, the concentrations of rare earth element (REE) and other minor and trace elements have been determined in the various silicate minerals present. The abundances of minor and trace elements present in olivine, orthopyroxene, chromian diopside, and plagioclase are given in Table 4.

Analysis of 12 olivine grains in GRV 021663 shows that trace and minor elements are all higher in this meteorite than reported in other winonaites, acapulcoites, and lodranites (Fig. 4) (see table 1d of Floss 2000; table 2a of Floss et al. 2008). While the concentration of elements such as Na, P, Y, and Zr show significant variations in olivine of this meteorite (e.g., Na = 89.6–1787 ppm, P = 54.4–303.4 ppm, Y = 0.089–0.364 ppm, and Zr = 0.123–7.55 ppm), Sc, Ti, V, and Mn contents do not. As with other studies of winonaite olivine (e.g., Floss et al. 2008), REEs were found to occur in quantities below detection limits, and are not discussed further.

Figure 5a displays the ranges in REE concentration in orthopyroxene in GRV 021663 and other winonaites, acapulcoites, and lodranites (Floss 2000; Floss et al. 2008). The CI-normalized REE pattern for GRV 021663 orthopyroxene displays a heavy REE-enriched pattern with a negative Eu anomaly similar to that displayed by the winonaite, HaH 193, but richer in overall abundances than the other winonaites, acapulcoites, and lodranites (Floss 2000; Floss et al. 2008). Compared with other winonaites, acapulcoites, and lodranites, orthopyroxene in GRV 021663 has a slightly flatter pattern that is less depleted in total REE. A negative-Ce anomaly is also observed in GRV 021663 orthopyroxene. It is possible that the elevated REE contents relative to the other winonaites and the Ce anomaly reflect the weathered nature of this sample (cf. Floss and Crozaz 1991; Crozaz et al. 2003). The abundances of minor elements in GRV 021663 orthopyroxene are also higher than determined in other winonaites. The P and Zr content in orthopyroxene varies by factors of ~145 (range 41–5541 ppm) and ~229 (range 1.07–245.3 ppm), respectively. The highest P and Zr content was detected at the same point in one orthopyroxene grain. We cannot rule out the possibility that these high trace element values represent contamination, perhaps in the form of minute inclusions of phosphate or a Zr-rich phase in orthopyroxene, although we do not observe evidence for this in backscattered electron images.

In contrast, the abundances of trace and minor elements in chromian diopside from GRV 021663 fall within the range of data for the other winonaites presented in Floss et al. (2008). Data are slightly higher than reported for diopside from most of acapulcoites and lodranites (Floss 2000). REE patterns for chromian diopside in this meteorite display a bow-shaped profile with a strong, negative Eu anomaly; these nearly overlap with the data presented for winonaite NWA 1463 (Fig. 5b), which has been considered the most primitive winonaite examined to date (Benedix et al. 2003).

In contrast, GRV 021663 plagioclase displays a strongly light REE (LREE)-enriched pattern, with a strong positive Eu anomaly; the REEs heavier than Eu are below the detection limits of the ICP-MS method utilized (Fig. 5c). Compared with data on the other winonaites, the REE pattern for plagioclase in GRV 021663 is somewhat flatter (i.e., it displays less LREE-enrichment), falling to the higher end for the winnonaite group.

Table 4. ICP-MS determined minor and trace element concentrations (averages and ranges, in ppm except where noted) in olivine, orthopyroxene, clinopyroxene, and plagioclase for GRV 021663.

	Olivine (n = 12)	Orthopyroxene (n = 12)	Clinopyroxene (n = 12)	Plagioclase (n = 3)
Na	504 (89.6–1787)	751 (595–1187)	5594 (5246–5868)	6.3 (6.1–6.4) (%)
P	121 (54.4–303.4)	585 (41–5541)	98 (35–152)	77 (66–99)
Ca (%)	0.12 (0.07–0.20)	1.03 (0.96–1.06)	14.4 (12.7–15.2)	2.8 (2.7–2.8)
Sc	4.2 (3.7–4.9)	14 (12.8–15.5)	76 (70–82)	1.6 (1.3–1.8)
Ti	149 (85–206.8)	1422 (1369–1482)	4612 (4205–4815)	386 (354–419)
V	78 (75.2–82.3)	104 (101–109)	407 (368–424)	2.9 (1.7–5.6)
Mn	3871 (3729–4013)	4102 (3998–4301)	2682 (2592–2807)	85 (19–212)
Sr	0.15 (b.d.–0.39)	0.13 (b.d. –0.30)	5.5 (5.0–6.0)	105 (102–108)
Y	0.26 (0.098–0.36)	2.2 (2.03–2.39)	23 (21–24)	0.20 (b.d. –0.23)
Zr	1.9 (0.12–7.6)	22 (1.07–245.3)	36 (31–38)	1.32 (0.47–2.55)
Ba	0.21 (b.d.–0.62)	0.16 (b.d.–0.30)	0.20 (b.d.–0.86)	34 (32–35)
La	0.079 (b.d.–0.21)	0.062 (b.d.–0.097)	2.4 (2.2–2.6)	1.2 (1.19–1.22)
Ce	0.15 (b.d.–0.49)	0.095 (0.042–0.16)	10 (9.3–10.5)	1.8 (1.7–1.9)
Pr	0.026 (b.d.–0.058)	0.022 (b.d.–0.032)	1.7 (1.5–1.8)	0.12 (0.093–0.16)
Nd	0.20 (b.d.–0.27)	0.16 (b.d.–0.29)	9.5 (8.6–10.2)	0.23 (b.d.–0.23)
Sm	b.d.	0.11 (b.d.–0.13)	3.0 (2.7–3.3)	0.19 (b.d.–0.19)
Eu	b.d.	0.013 (b.d–0.013)	0.058 (0.043–0.075)	0.80 (0.77–0.82)
Gd	b.d.	0.16 (b.d.–0.23)	4.1 (3.6–4.6)	b.d.
Tb	b.d.	0.029 (0.019–0.035)	0.67 (0.59–0.74)	b.d.
Dy	0.062 (b.d.–0.089)	0.28 (0.22–0.33)	4.5 (4.0–4.9)	0.083 (b.d.–0.083)
Ho	0.013 (b.d.–0.014)	0.08 (0.072–0.099)	0.90 (0.83–0.97)	b.d.
Er	0.047 (b.d.–0.075)	0.28 (0.22–0.31)	2.4 (2.2–2.6)	b.d.
Tm	b.d.	0.047 (0.031–0.057)	0.31 (0.27–0.34)	b.d.
Yb	0.089 (b.d.–0.114)	0.35 (0.31–0.42)	1.9 (1.6–2.1)	b.d.
Lu	0.017 (b.d.–0.033)	0.059 (0.050–0.067)	0.26 (0.23–0.27)	b.d.

Note: b.d. = below detection limit.

DISCUSSION

Classification of GRV 021663

Kimura et al. (1992) presented several mineralogical and petrographic differences to distinguish between the winonaite and acapulcoite meteorite groups. For example, winonaites are characterized by olivine with lower MnO and FeO contents, lower Cr_2O_3 and Na_2O contents in chromian diopside, and higher anorthite content of plagioclase, etc. However, this subdivision was based on a very limited data set of just three winonaite (Yamato [Y]-74025, Y-75300, and Y-75305) and two acapulcoite (Y-74064 and Allan Hills [ALH] 77081) samples. With the discovery of many new winonaite examples, these criteria appear to be much less robust; for example, the winonaite, Tierra Blanca, contains olivine with MnO contents of up to 0.7 wt%, while the Cr_2O_3 and Na_2O content of chromian diopside exceeds that of its counterparts in the acapulcoites (Yugami et al. 1998). Moreover, some winonaites contain plagioclase that is less anorthitic than found in the acapulcoite group (Yugami et al. 1998; Hutchison 2004). Although the consensus is that winonaite meteorites are more reduced than acapulcoites, exceptions do exist; Zag(b) is a winonaite, with fayalite and ferrosilite contents of olivine and orthopyroxene as high as 19.4 and 25.7, respectively (Grossman 2000). Another contradictory example is the acapulcoite, ALHA81187/84190, whose olivine fayalite content is 4.2 (McCoy et al. 1996). Hence, the classification of winonaite and acapulcoite meteorites based solely on the composition of their silicate mineralogy is somewhat risky.

The approximate chondritic mineralogy and chemical composition, as well as textural evidence for extensive metamorphism, indicate that GRV 021663 is a primitive achondrite meteorite that could belong to either the winonaite or acapulcoite subgroups. As winonaites and acapulcoites have similar mineral compositional ranges and textural features, they are not easily distinguished. The following features, however, suggest that GRV 021663 is in fact a winonaite:

1. K-feldspar is present as an accessory mineral in GRV 021663. Acapulcoites do not typically contain K-feldspar (FRO 93001 is one exception; McCoy et al. 1996; Yugami et al. 1998; Patzer et al. 2004).
2. The composition of plagioclase is $Ab_{74.5}Or_4An_{21.5}$; this correlates well with other winonaites (Hutchison 2004).

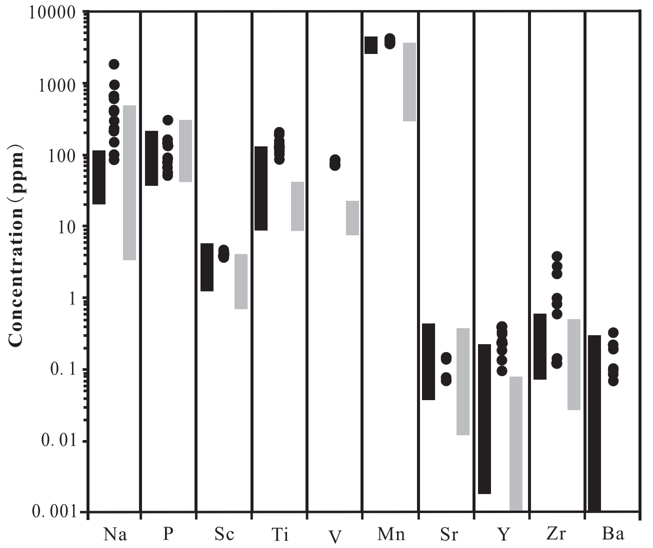

Fig. 4. Trace element concentrations in olivine of GRV 021663 (black circles), winonaites (gray bars), acapulcoites, and lodranites (black bars for the last two groups). Data are from Floss (2000) and Floss et al. (2008).

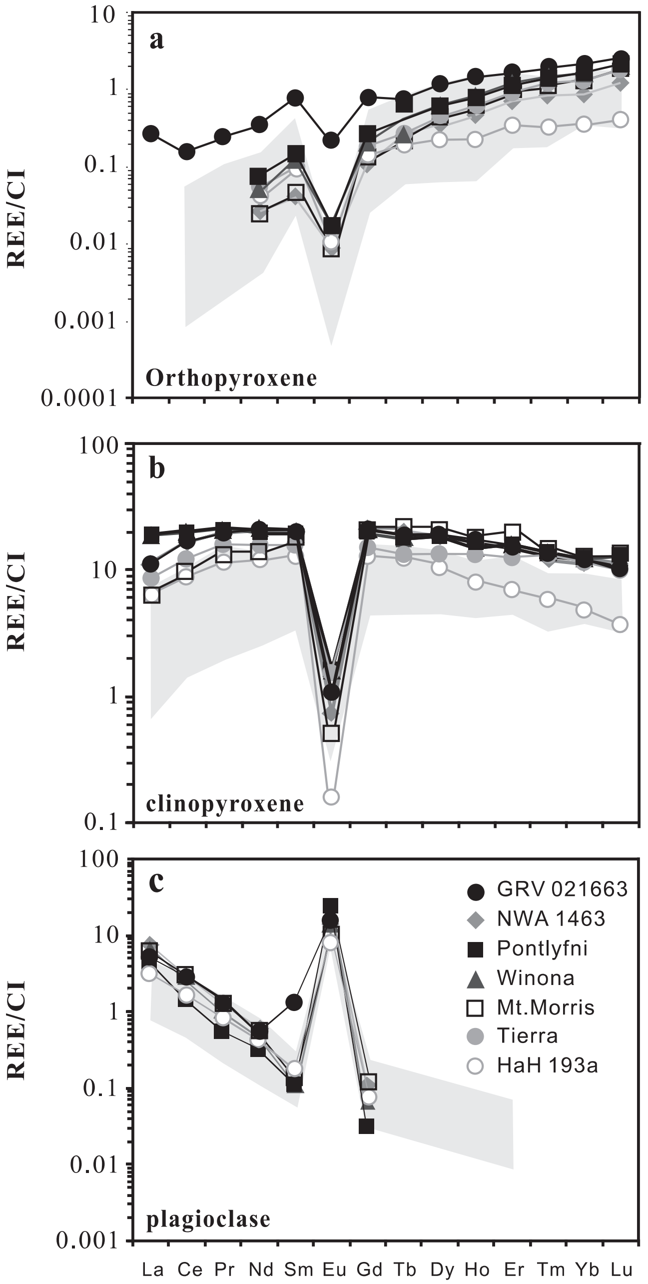

Fig. 5. Plot of average CI chondrite-normalized REE patterns for: (a) orthopyroxene, (b) clinopyroxene, and (c) plagioclase from GRV 021663 and other winonaite examples. The gray-shaded areas show the ranges for acapulcoites and lodranites. Data are from Floss (2000) and Floss et al. (2008).

3. Compared with the majority of acapulcoites, the FeO content of olivine (Fa = 5.4), pyroxene (Fs = 4.7), and diopside (Fs = 2.4) of GRV 021663 is low (see Fig. 6). Taenite is also absent in this meteorite, indicating that GRV 021663 is more reduced than the majority of the acapulcoites (McCoy et al. 1996; Yugami et al. 1998; Patzer et al. 2004). Furthermore, Cr-enrichment in troilite also indicates that GRV 021663 formed under more reducing conditions than the acapulcoites (Lorenz et al. 2003).
4. The P content of Fe-Ni metal in GRV 021663 is 0.76 wt%, while troilite contains 1.26 wt% Cr. This is comparable to data from the winonaite, Sahara 02029, in which Fe-Ni metal contains 0.6 wt% P and troilite contains 2.1 wt% Cr (Russell et al. 2004). In the metal of other winonaites and acapulcoites, the content of P always falls below the detection limit of EPMA, while metal in lodranite meteorites contains P in the range of 0.03–0.08% (see table 2e of Yugami et al. 1998).
5. The most important parameter for the classification of these meteorite groups, however, is whole rock, oxygen isotopic composition. Figure 3 highlights the range in oxygen isotopes for the acapulcoites, lodranites, winonaites, and silicates inclusions present in IAB and IIICD irons, respectively (Clayton and Mayeda 1996). The studied GRV 021663 sample has had three independent laser fluorination runs, such that the data are true replicates. The determined values of δ^{18}O are +7.57‰, +7.40‰, and +7.53‰, respectively, and differ from that of the acapulcoites (+2.54‰ ~ +5.06‰; Clayton and Mayeda 1996; Irving et al. 2007). Moreover, the winonaites (~+3.26‰ to +5.8‰; Mayeda and Clayton 1989; Clayton and Mayeda 1996) as a group are richer than the acapulcoites in terms of ^{18}O content. The most important parameter is Δ^{17}O. The Δ^{17}O range of

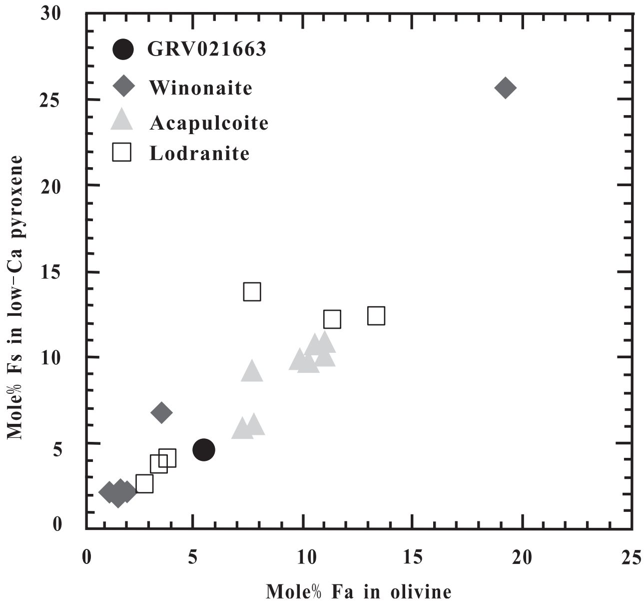

Fig. 6. Mean Fs in low-Ca pyroxene versus mean Fa in olivine for GRV 021663, winonaites, acapulcoites, and lodranites. Data are from Yanai and Kojima (1991), McCoy et al. (1996), Yugami et al. (1998), Yanai (2001), and Patzer et al. (2004).

acapulcoites and winonaites is −0.85‰ to ∼−1.5‰ and −0.35‰ to ∼ −0.78‰, respectively (Mayeda and Clayton 1980, 1989; Clayton and Mayeda 1996; Floss et al. 2003; Russell et al. 2003, 2004; Irving et al. 2007). The average value of Δ^{17}O of GRV 021663 is −0.385‰. This value is similar to that of the winonaites.

We also find that the values of δ^{18}O of GRV 021663 are very similar to the iron IIICD (thought to originate from the same parent body to the winonaites) MaltehÖhe (+7.21‰). Clearly, GRV 021663 lies within the range of winonaites, IAB and IIICD parent body and is clearly more ^{16}O-depleted than the acapulcoites. As such, we suggest that GRV 021663 should be reclassified as a winonaite.

Partial Melting of the Meteorite

The winonaites have experienced extensive thermal metamorphism, resulting in equigranular textures, abundant 120° triple junctions in most meteorites, and at least limited partial melting, as is evidenced from the presence of Fe-Ni metal and troilite veins (Benedix et al. 1998). Evidence in favor of silicate partial melting is, by contrast, limited. The Winona and Mt. Morris (Wisconsin) meteorites display coarse-grained domains that are olivine-rich and plagioclase- and pyroxene-poor, suggesting that these areas may represent residues after partial melting (Benedix et al. 1998). At the same time, there is additional lithological evidence from Pontlyfni in the form of localized areas enriched in plagioclase and clinopyroxene, which may have formed as a result of in situ crystallization of a basaltic partial melt (Benedix et al. 1997, 1998). It has also been suggested (Benedix et al. 1998) that the olivine-rich areas in Winona and Mt. Morris represent a dunitic, partial melt residuum from another region of the winonaite parent body, and that these were incorporated into the meteorite host-rocks by impact mixing. On the Earth, however, these textural relationships could also be explained by the entrainment of mantle xenoliths (the dunite domains) in a migrating basaltic melt. Pontlyfni, however, is more akin to type 6 chondrites, in that relict chondrules are present; this meteorite exhibits widely varying grain sizes, textures, and mineralogy. The estimated temperatures of equilibration for Pontlyfni of 975 °C, based on two-pyroxene geothermometry (Graham et al. 1977), are only high enough to form a Fe-Ni-S eutectic (Kullerud 1963; Kubaschewski 1982; Brandes and Brook 1998). Basaltic partial melting is thought to initiate at approximately 1150 °C in an ordinary chondritic system (Taylor et al. 1993; Jurewicz et al. 1995), implying that silicates in Pontlyfni may have never been involved in melting. Thermodynamic constraints, however, and evidence for highly variable temperatures resulting in Fe-Ni-S eutectic and/or basaltic partial melting on a very localized scale, are consistent with the idea that the winonaite-IAB iron parent body may have experienced collisional fragmentation and reassembly following peak metamorphic temperatures (Benedix et al. 2005). Floss et al. (2008) measured the trace element compositions of individual plagioclase, pyroxene, and olivine grains in six winonaites that are representative of all textures and mineralogies observed in this meteorite group. Unfortunately, while petrographic evidence in these samples support partial melting of metal and silicate materials, the trace element data did not show any clear evidence for or against the partial melting of silicates in these meteorites.

Grove Mountains 021663, as with other winonaites, has experienced extensive modification by heating (i.e., the presence of equigranular textures and abundant 120° triple junctions; Fig. 1c). This meteorite has also experienced Fe-Ni-S-P eutectic melting—a distinguishing feature of GRV 021663 is the extensive network of troilite and ‘‘rusty’’ veins penetrating the entire rock (Fig. 2a). The majority of metal grains, and their weathering products, are enclosed by either troilite or schreibersite (see Figs. 2b–d), textures suggestive of melting of metal, troilite, and schreibersite, and that a Fe-Ni-S-P eutectic melt was formed in this meteorite (Esbensen and Buchwald 1981; Rubin et al. 1999). Furthermore, the low Ni contents of kamacite may be evidence for the migration of a Fe-Ni-S eutectic melt because as metallic Fe-Ni is

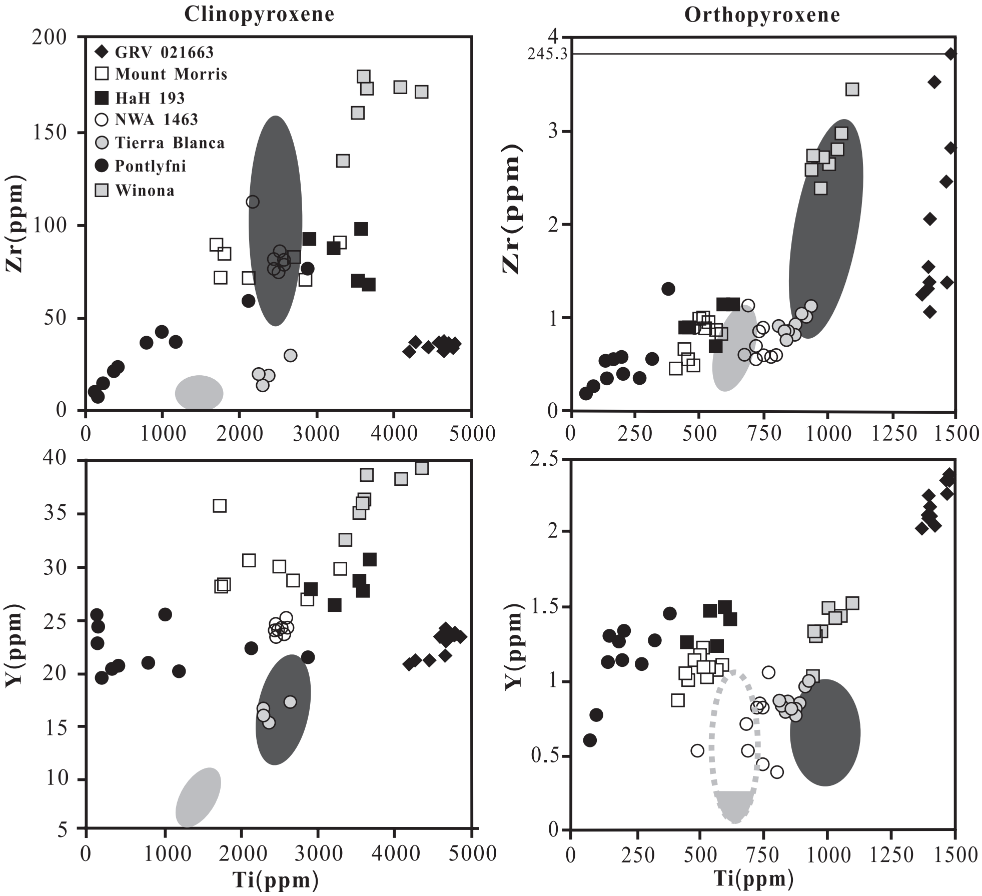

Fig. 7. Plot of concentrations of the incompatible elements, Zr and Y versus Ti in clinopyroxene and orthopyroxene from GRV 012663 and other winonaites. Black areas represent fields of acapulcoites, and gray areas represent fields for typical lodranites. The data for other winonaites, acapulcoites, and lodranites derive from Floss (2000) and Floss et al. (2008). Note that the Zr concentration in one analysis point for orthopyroxene in GRV 021663 is 245.3 ppm.

heated above the eutectic temperature of Fe-Ni-S, any metallic melt lost to the system would have been enriched in Ni and S (Kullerud 1963). As a result, a residuum would be enriched in Fe, and depleted in Ni and S.

While GRV 021663 is petrologically distinct from many of the winonaite meteorites, it is very similar to the olivine-rich lithologies found in Winona (Benedix et al. 1998; Floss et al. 2008). Moreover, these meteorites share similar trace element contents and behavior (see Figs. 7 and 8). The abundance of olivine (~56 vol%) and paucity of plagioclase (~1.4 vol%) perhaps indicate removal of a basaltic partial melt from this meteorite (McCoy 1994; Benedix et al. 1998). At the same time, GRV 021663 hosts single feldspar (plagioclase and K-feldspar) crystals that poikilitically enclose mafic silicate grains (see Figs. 1d and 1f), and that could be indicative of an interstitial basaltic partial melt. Local euhedral, mafic silicate inclusions within feldspar further suggest that these crystallized from a melt (Figs. 1d and 1f). Moreover, the presence of magmatic inclusions (such as K-rich feldspar, olivine, diopside, and plagioclase) in orthopyroxene (Fig. 1e) and K-feldspar along fractures between other mafic silicate minerals (Figs. 1c and 1d) are additional evidence in favor of partial melting and melt migration, at least on a local scale (Folco et al. 2006). REE patterns for plagioclase from GRV 021663 lie at the upper end of the rather uniform range exhibited for all other winonaite samples (Floss et al. 2008) and both the acapulcoites and lodranites (Floss 2000). REE concentrations for elements Gd-Lu are below detection limits in the analyzed plagioclase grains.

Do Trace Element Mineral Compositions Record Silicate Partial Melting?

Acapulcoites and lodranites constitute two related groups of primitive achondrites. The two groups have similar mineralogies, mineral compositions, thermal histories, cosmic ray exposure ages, and oxygen isotopic compositions, and these data suggest that acapulcoites and lodranites are rocks from a common parent body. Systematic differences in incompatible trace element

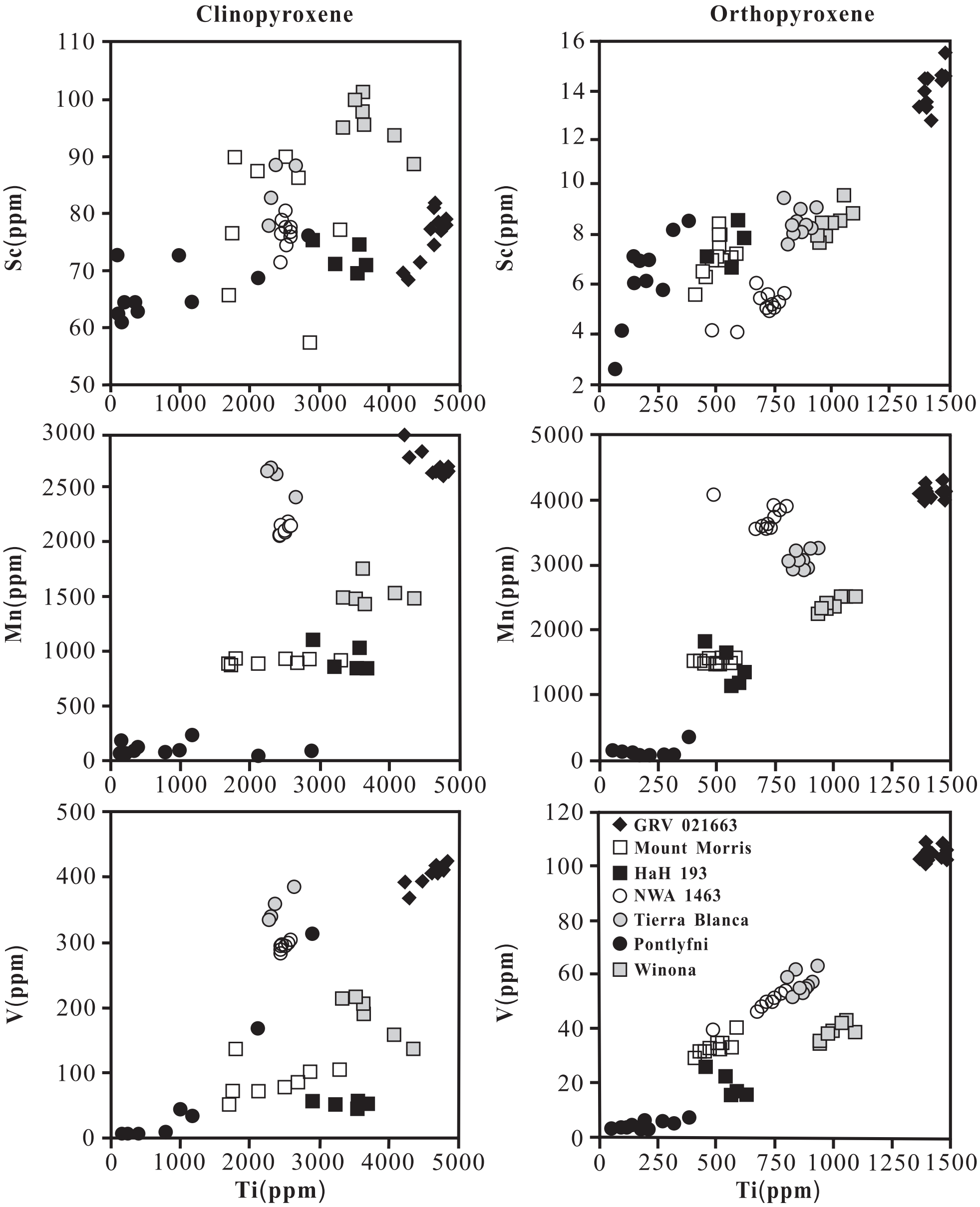

Fig. 8. Plot of the concentrations of compatible elements Sc, Mn, and V versus incompatible element, Ti in clinopyroxene and orthopyroxene from GRV 012663 and other winonaites. The data for other winonaites derive from Floss et al. (2008).

content in both ortho- and clinopyroxene, however, have been cited as evidence that the lodranites can be considered to represent residual material left over from the partial melting of acapulcoite-like precursors (e.g., Clayton and Mayeda 1996; McCoy et al. 1996, 1997b; Floss 2000; Rubin 2007 and references therein).

It has been suggested previously that partial melting may also play a role in generation of winonaite meteorites (Benedix et al. 1998, 2000; Floss et al. 2008). However, petrological and incompatible trace element data are somewhat conflicting, reflecting the complicated histories experienced by the winonaites on their parent body (brecciation, mixing of lithologies, thermal metamorphism). Petrographic evidence suggests that the relationship of GRV 021663 with other winonaites may be similar to that of the lodranites and acapulcoites. First of all, GRV 021663 is richer in olivine, but poorer in both metal and troilite, and displays a paucity of plagioclase, as compared with other winonaites; chromite and phosphate are also lacking in this sample. Secondly, it contains some textural evidence suggestive of the partial melting of silicates in its make-up; for example, single plagioclase crystals poikilitically enclosing mafic silicate grains. Thirdly, the composition of kamacite in

GRV 021663 is poorer in Ni and richer in P than for kamacite present in other winonaites. These characteristics are all found in the majority of lodranite meteorites and do not occur in any acapulcoite (Kimura et al. 1992; Nagahara 1992; McCoy et al. 1996, 1997a; Mittlefehldt et al. 1998; Patzer et al. 2004). This implies that GRV 021663 may have experienced a similar evolutionary process to that of the lodranites, such that this meteorite could represent a residue from the partial melting of a winonaite-like precursor or parent body. While incompatible trace element data for lodranite pyroxenes are shown to have retained a record of silicate mineral partial melting, experienced on their parent body (Floss 2000), Figs. 5a and 5b, which display the REE patterns of GRV 021663 pyroxenes, do not provide obvious evidence in favor of partial melting of silicates in this sample. Other incompatible trace elements in pyroxene in GRV 021663, such as Ti, Zr, and Y which can record evidence of depletion in a partial melting residuum (Mittlefehldt and Lindstrom 2000; Floss et al. 2008), provide little information for or against partial melting of silicates. Figure 7 is a plot of Zr versus Ti, and Y versus Ti in GRV 021663 pyroxenes and, for comparison, the other winonaites for which data are available. When compared with these other examples, GRV 021663 pyroxene has the highest Ti abundance, while the contents of Zr and Y are lower in clinopyroxene, but higher in orthopyroxene. These values are at the high end for the winonaite group and comparable to values for Winona (Fig. 7). While the studied trace element abundances in minerals present in GRV 021663 do not provide conclusive information concerning the partial melting processes in the formation of this meteorite, compositional data for Ti, Zr, and Y indicate that orthopyroxene could represent a component formed from silicate partial melts. An argument has been made that such trends in trace element data could be indicative of silicate partial melting from a common precursor, where Pontlyfni is a residue and Winona the incompatible-element-enriched partial melt (Floss et al. 2008). Perhaps both Winona and GRV 021663 represent melt material from an intermediate protolith, with the other winonaites representing mixtures of residue and melt. Floss et al. (2008) suggested that NWA 1463 could be a suitable candidate for such a protolith because it is the least metamorphosed of the winonaite meteorites.

Figure 8 is a plot of Sc, Mn, and V, elements that are compatible in pyroxene, versus Ti. With the exception of Sc, that is somewhat lower in GRV 021663 clinopyroxene compared with other winonaites, the concentrations of these elements in clinopyroxene and orthopyroxene of GRV 021663 are among the highest in this group. That is to say, Mn and V (and to a lesser extent Sc) display near identical incompatible trace element behavior to Ti. It is clear therefore that the compatible and incompatible trace elements have not retained any significant record of silicate mineral partial melting in GRV 021663. Alternatively, GRV 021663, as with the other winonaites, experienced significant impact brecciation and mixing of lithologies, followed by inhomogeneous thermal metamorphism on its parent body, that may have led to trace element re-equilibration between the different lithologies (cf. Floss et al. 2008).

If partial melting is important in the origin of GRV 021663 and the other winonaites on their parent body, a heat source is needed for this to have been effective. Several workers have suggested that impacts are the major geological process to have affected the parent asteroids of the primitive achondrite meteorites (e.g., Rubin 1995). Furthermore, several authors have argued that metamorphism and partial melting igneous processes are solely the result of collisional processes on an asteroid (e.g., Wasson and Kallemeyn 2002). However, the studies of Stöffler et al. (1991) and Keil et al. (1997) have shown that while impacts can produce localized melts and some whole-rock melts, impacts cannot have been the heat source for the generation of the primitive achondrite meteorites. In these studies, many lines of evidence point against impact as a mechanism, for example, that the bulk of melt generated during a collision is ejected from the impact crater. As impact melting is thought to be an unrealistic mechanism to explain the evolution of the winonaites group, an alternative and more plausible mechanism is likely related to heating processes internal to the parent asteroid, such as radioactive decay of short-lived isotopes, or perhaps electromagnetic induction related to the meteorites interaction with the Sun (Keil et al. 1997).

SUMMARY

Grove Mountains 021663 can be characterized as a member of the winonaite primitive achondrite meteorite group, based upon its mineralogical, textural, stable isotopic, and mineral trace element similarities to the other winonaites. The degree of weathering and shock effects for this meteorite are W3 and S2, respectively. At the same time, this meteorite displays some notable differences, which expand the compositional field of the winonaites: (1) a high abundance of olivine, but lower orthopyroxene content and a paucity of plagioclase, phosphate, taenite, and chromite are also absent; (2) most of the metal grains (or their weathering products) are enclosed by troilite and/or schreibersite; (3) single plagioclase crystals poikilitically enclose silicate grains, some of which are well crystallized; (4) the presence of magmatic inclusions (such as K-rich feldspar, olivine, diopside, and plagioclase) in orthopyroxene, and K-feldspar along fractures between other mafic silicate minerals; (5) kamacite has low Ni (3.01 wt%) and high P (0.76 wt% P) concentrations; (6) a mean olivine Fa (5.4 mole%) value that exceeds the mean orthopyroxene Fs (4.7 mole%) value; and lastly (7) the whole rock oxygen isotopic composition is very ^{16}O-depleted.

Grove Mountains 021663 has experienced thermal metamorphism and appears to have been heated to approximately 1100 °C, resulting in the formation of both a Fe-Ni-S-P eutectic and, potentially, basaltic partial melts. Melts have migrated from the source areas, leaving a residuum that tends to be depleted in minerals with a low melting temperature; such as, troilite, kamacite, and plagioclase. The trace element content of pyroxenes in this meteorite, however, does not record clear information concerning the partial melting of silicate minerals. Similar conclusions have been presented for other studied winonaite meteorites (e.g., Floss et al. 2008).

If partial melting played an important role in the origin of this meteorite and the other winonaites or their parent body it is likely related to heating processes internal to that parent, such as radioactive decay of short-lived isotopes, or perhaps electromagnetic induction related to the Sun. While melting is possible from large impacts, the effects have been shown to be ineffective in generating the heat necessary for the kinds of partial melting invoked for genesis of lodranites and acapulcoites (see Keil et al. 1997 and references therein for a detailed discussion). As such, impact melting is also thought an unrealistic mechanism in the evolution of the winonaite group.

The study of GRV 021663 therefore has provided and will continue to provide important and valuable information concerning both its origin and the subsequent evolution of the parent body to the winonaites.

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