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Cr spinel and chromite as petrogenetic indicators in ordinary chondrites: Equilibration temperatures of petrologic types 3.7 to 6

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Abstract–Many equilibrated ordinary chondrites contain (besides chromites of constant composition) Cr spinel with a large spread in Cr/(Cr + Al) ratios. They occur mainly as large grains in chondrules rich in mesostasis, preventing complete equilibration in Cr/Al but not in Fe/Mg. This partially equilibrated Cr spinel turned out to be particularly useful for the selection of an appropriate olivine/spinel thermometer and for the determination of equilibration temperatures. The main results are:

1) The H3.7 to 3.8 and the L3.7 to 3.8 chondrites analyzed show temperatures of 625 to 680 °C;

2) Equilibrated chondrites show a range of olivine/Cr-spinel temperatures between 700 and 820 °C, and the same average temperatures for type 4 to 6 (number of analyzed meteorites in brackets): H4 (9) 766 °C, H5 (7) 774 °C, H6 (3) 775 °C, L4 (5) 752 °C, L5 (4) 754 °C, L6 (1) 754 °C. These temperatures are interpreted as equilibration temperatures. One indication is that the measured isotherms are straight lines down to low Cr/(Cr + Al) ratios, which have a higher Fe/Mg interdiffusion coefficient than grains with high ratios. And there is no correlation of measured temperature with grain size of Cr spinel.

3) Chromites sensu stricto show temperatures about 50 to 100 °C lower than Cr spinel, and a correlation with grain size. This is a closure temperature established during cooling and in situ crystallization.

These results can best be interpreted by a "rubble pile" model of parent body evolution. This model cannot explain, however, the absence of type 4 to 6 chondrites with temperatures as low as for type 3.7 to 3.8.

INTRODUCTION

Ordinary chondrites (OC) are primitive agglomerates. Their main constituents (chondrules, matrix, and metal grains) were never molten together. Nevertheless, they experienced thermal alterations, which led to recrystallization, chemical equilibration and textural integration (e.g., Dodd 1969). Binns (1967) distinguished primitive, intermediate, and crystalline chondrite types by their texture. Van Schmus and Wood (1967) devised a more elaborate system: unequilibrated or type 3 chondrites with variable Fe/Mg ratios in the mafic silicates, and equilibrated ones (type 4, 5 and 6) with constant silicate compositions and increasing textural integration. The latter system has been universally adopted and is now used for the classification of OCs. The thermal conditions under which this metamorphism took place can be studied from the chemical exchange equilibria between minerals. This was attempted mainly by the determination of the Ca, Fe, and Mg distribution between ortho- and clinopyroxene for which geothermometers exist. The first such attempt by Bunch and Olsen (1974) using a thermometer by Kretz (1963) was unsuccessful, they could only give an estimated range of 750 to 950 °C for type 6 chondrites. Dodd (1981) used this estimate and a temperature of 400 °C for type 3 chondrites derived from Ni profiles in taenite (Wood 1967) to propose a temperature range for type 4 (600–700) and type 5 (700–750) by interpolation between these two corner values. But he states: "the temperatures experienced by type 6 chondrites are poorly known and those of types 4 and 5 are virtually unknown" (Dodd 1981). The two pyroxene thermometer was later refined (Lindsley 1983) and applied by Olsen and Bunch (1984) to LL6 chondrites. The result of 830 to 860 °C lies within the range of their earlier estimate. The two pyroxene thermometer could be applied only to type 6 chondrites, however, because the pyroxenes are not yet completely equilibrated in type 4 and 5

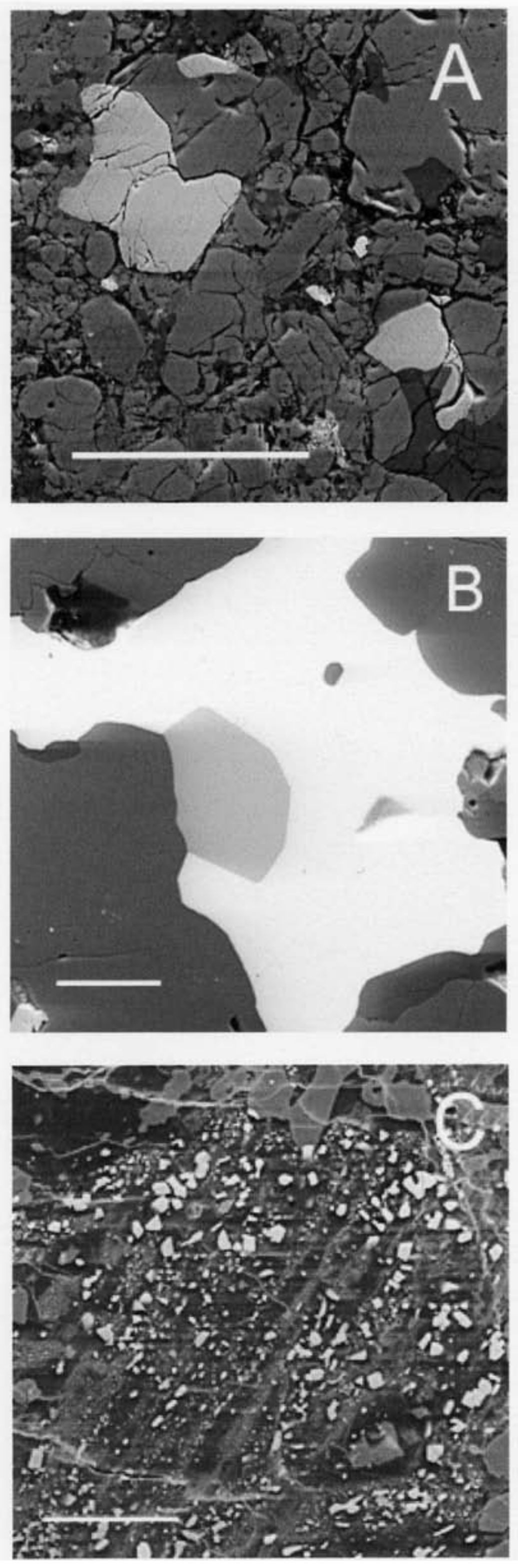

Fig.1. Chromite in ordinary chondrites (SE pictures, scale bars 50 mm). a) H5 Holbrook; irregular grains (medium grey) in silicates (dark grey). b) L5 Ehole; chromite grain with ideomorphic faces against metal (white). c) H6 Daraj 016; small grains (white) dispersed in the mesostasis of a chondrule.

(McSween and Patchen 1989). They and Reid and Dussaud (2000) tried to determine the equilibration temperature of type 4 and 5 from orthopyroxene equilibria alone, which react faster than clinopyroxene (Jones 1998). Unfortunately, these opx temperatures lie generally about 100 to 150 degrees lower than the cpx temperatures derived for the same meteorite (McSween and Patchen 1989). Thus, the equilibration temperature of type 4 and 5 are still poorly constrained.

The thermal conditions experienced by the ordinary chondrites are important clues for the size and structure of the chondrite parent body. Most parent body models need the abundance and equilibration temperature of the different petrologic types for their calculations, and most of them use the estimated temperature ranges by Dodd (1969, 1981) mentioned above (Miyamoto et al. 1981, Göpel et al. 1994, Trieloff et al. 2003), which are also quoted in many review articles (e.g., McSween et al. 1988; Keil 2000). Because of the problems encountered with the pyroxene thermometer, another exchange reaction has been used more recently, the Fe/Mg equilibrium between olivine and chromite (Wlotzka 1985; Johnson and Prinz 1991; Sack and Ghiorso 1991b; Kessel et al. 2002; Calafiore 2002). The advantage of this reaction is that it is faster and can be applied to type 4 to 6 chondrites, as well as to partially equilibrated type 3. The purpose of this paper is to investigate the possibilities of this exchange reaction in detail, to compare different olivine/spinel thermometers, and to discuss the results with respect to the thermal history of ordinary chondrites.

Chromite is a minor but regular constituent of ordinary chondrites (Tschermak 1885). Besides the main components FeO and Cr_2O_3, it contains minor amounts of MgO and Al_2O_3. Fe and Mg can be exchanged with the Fe, Mg silicate minerals and in equilibrated chondrites the Fe/(Fe + Mg) ratio in chromite correlates with the same ratio in olivine, that is, it increases from the H to the L and LL groups of chondrites (Bunch et al. 1967)

Besides homogeneous chromites, some equilibrated chondrites also contain Cr spinel with a large spread in Cr/(Cr + Al) ratios. Paradoxically, these unequilibrated, or better, only partially equilibrated Cr-spinel grains turned out to be particularly useful for the selection of an appropriate thermometer and for the determination of equilibration temperatures.

METHODS

Polished thin sections of chondrites were studied with an optical microscope in reflected and transmitted light. Special attention was paid to Cr-spinel grains, which differed from normal chromites by their lower reflectivity and/or their red-brown transparency. These grains, normal chromites, and Fe, Mg silicates were analyzed by electron microprobe by standard procedures. Chemically analyzed natural chromite, spinel, and olivine (Jarosewich et al. 1980) were used as standards.

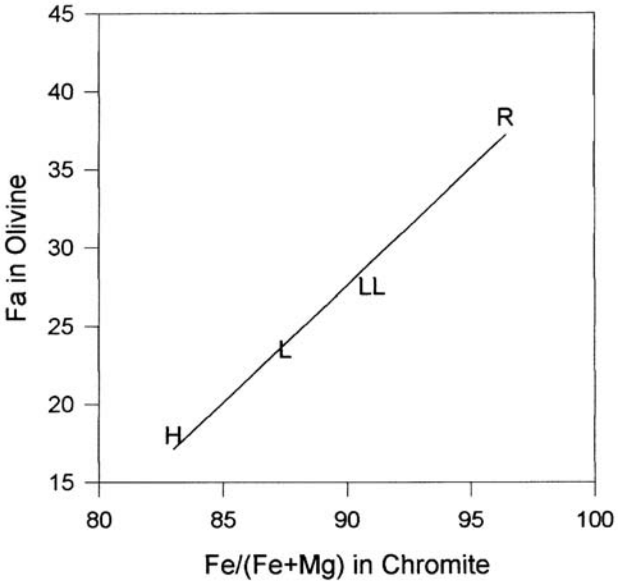

Fig. 2. Average Fa in olivine versus average mole% Fe/(Fe + Mg) in chromite for the H, L, LL, and R chondrite groups (data from this work).

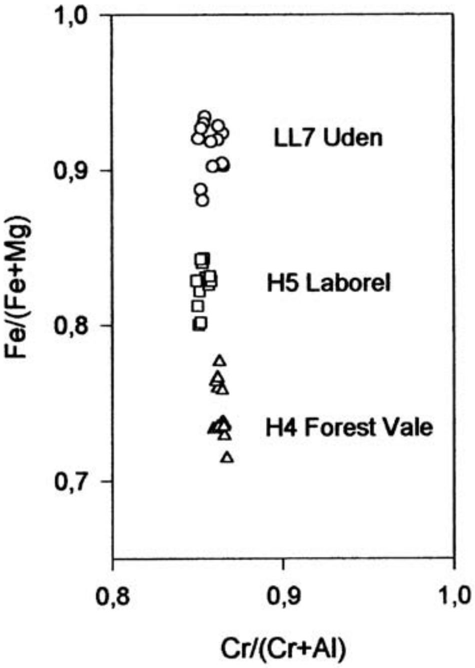

Fig. 3. Atomic ratio Fe/(Fe + Mg) versus Cr/(Cr + Al) in chromite grains of two H and one LL chondrite.

Occurrence and Composition

Chromite

Ordinary chondrites contain 0.3% Cr (Palme et al. 1981). This would yield about 0.8 weight% chromite, whereas only 0.27% chromite were found by point counting in ordinary chondrites (Keil 1962). The rest of Cr is contained as minor element in silicates, mainly in clinopyroxene. It contains up to 0.5% Cr_2O_3 (Jones 1998). A survey of polished thin sections shows that chromite grains occur in two main forms. The first are interstitial grains in the matrix. They are xenomorphic against silicates (Fig. 1a) and show ideomorphic shapes only sometimes when in contact with troilite or metal (Fig. 1b). Another minor part of the chromite is found as small grains disseminated in the mesostasis of chondrules (Fig. 1c). An extensive description of chromite types in chondrites was given by Ramdohr (1967).

In contrast to type 4 to 6, type 3 chondrites contain only little chromite. The same observation was made by Dodd (1969). Here the element Cr is found in the mesostasis of chondrules, e.g., 0.4% Cr_2O_3 in H3 Sharps chondrules (Wlotzka 1988), and in silicate minerals (about 0.4% and 0.9% Cr_2O_3 in olivine and low-Ca pyroxene, respectively, in LL3.0 Semarkona [Jones and Scott 1989]), and in Cr spinel (see below).

A survey of the chemical composition of chondritic chromites was given by Bunch et al. (1967). They noticed a relation of the Fe, Mg content of chromites to that of the olivines in the H, L, and LL groups of equilibrated chondrites. Figure 2 shows such a diagram confirming their analyses. Table 1 lists chromite analyses for individual equilibrated chondrites and the average composition for the H, L, and LL groups. The increase in Fe/(Fe + Mg) from H to LL chondrites is evident. In Table 1 also chromites of deviating compositions (mainly in the Fe/(Fe + Mg) ratio) are included, e.g., from H4 Forest Vale or from shock-reheated L6 St. Catarina and Rio Limay. These were especially sought for, so that their number in Table 1 is not representative for their abundance. A more detailed analysis of chromite grains in a given chondrite shows a narrow range in Cr/(Cr + Al) values, and a larger range in Fe/(Fe + Mg), see Fig. 3.

Figure 4a shows the same diagram as Fig. 2, but for individual chondrites, showing the main group clusters, but many outliers. It is noteworthy, that the intergroup relations shown in Fig. 2 are not found within the H and L groups, i.e., no correlation between Fa in olivine and the Fe/(Fe + Mg) ratio in the chromite. This stands in contrast to the good correlation of Fe/(Fe + Mg) between olivine and pyroxene within the H and L group, respectively (Fig. 4b). It will be shown below that the lack of this correlation between olivine and chromite within a chemical group does not imply disequilibrium between them in an individual chondrite. It rather reflects different equilibration temperatures. The overlapping Fe/(Fe + Mg) ratios of chromites between the H and L groups makes it impossible to assign chondrites to one of these groups from the chromite composition alone, as this was attempted, e.g., for fossil chondrites (Bridges et al. 2004).

The main minor element in chromite besides Al and Mg is titanium. Its value increases from the H over the L to the LL group (Table 1). Ti is also found in ilmenite in chondrites, and its content in the chromite depends probably on the amount of ilmenite present. Many Cr-spinel grains show also lamellar

Table 1. Average composition of chromites from H, L, and LL chondrites.

Name	Type	MgO	Al_2O_3	TiO_2	V_2O_3	Cr_2O_3	MnO	FeO	ZnO	Fe#	Cr#
Elsinora	H4	3.45	6.65	2.11	0.62	55.6	0.85	29.9	0.29	83.0 ± 1.2	84.9 ± 0.5
Foster	H4	3.06	5.96	1.46	0.69	59.5	0.83	29.1	0.39	84.2 ± 2.3	87.0 ± 1.4
Kesen	H4	3.33	6.48	1.66	0.63	57.8	1.03	28.2	n.d.	82.6 ± 0.8	85.7 ± 0.5
Beddgelert	H4	3.27	6.66	1.93	0.60	57.7	0.78	28.1	n.d.	82.8 ± 1.0	85.3 ± 1.0
Richardton	H5	3.36	5.86	2.00	0.72	57.8	0.93	28.3	0.56	82.5 ± 1.4	86.9 ± 1.0
Tiffa 005	H5	3.74	6.89	1.83	0.67	57.1	0.93	28.4	0.25	81.0 ± 0.7	84.8 ± 0.3
Ehole	H5	3.47	6.69	2.23	0.64	57.9	0.88	28.9	0.33	82.4 ± 1.3	85.3 ± 0.4
Guarena	H5	3.44	6.86	2.19	0.65	57.0	0.81	28.7	0.30	82.4 ± 1.0	84.8 ± 0.4
DaG 739	H5	3.70	6.83	2.05	0.65	54.9	0.92	29.1	0.28	81.5 ± 1.6	84.4 ± 0.3
Laborel	H5	3.44	6.58	2.14	0.63	57.1	0.90	29.1	0.33	82.6 ± 1.3	85.3 ± 0.4
Morro do Rocio	H5	3.19	7.60	1.36	0.69	56.1	0.93	28.5	0.37	83.4 ± 1.2	83.2 ± 0.6
Daraj 117	H6	3.34	6.68	2.21	0.63	56.8	0.78	29.8	0.28	83.4 ± 1.1	85.1 ± 0.7
DaG 747	H6	3.36	6.57	2.31	0.65	56.7	0.84	29.4	0.30	83.1 ± 1.0	85.3 ± 0.2
Mean H chondrites		3.40	6.64	1.96	0.65	57.1	0.88	28.9	0.28	82.7	85.2
Std. deviation		0.18	0.41	0.29	0.03	1.1	0.07	0.6	0.14	1	1
Not included in the mean											
Beaver Creek	H4	5.54	6.99	2.30	0.58	57.7	0.75	26.2	n.d.	72.7 ± 2.5	84.7 ± 0.4
Forest Vale	H4	5.28	6.15	2.18	0.60	58.5	0.77	26.7	0.11	74.0 ± 1.7	86.5 ± 0.2
DaG 310	H5/6	4.48	6.76	2.32	0.65	57.8	0.81	27.6	0.40	77.6 ± 1.8	85.2 ± 0.3
Bjurböle	L4	2.18	5.72	2.14	0.67	56.9	0.65	30.7	0.33	88.8 ± 0.6	87.0 ± 0.4
DaG 323	L4	2.35	6.16	2.04	0.65	57.3	0.74	30.7	0.38	88.0 ± 1.0	86.2 ± 0.5
Holbrook	L5	2.61	5.94	3.28	0.62	55.3	0.65	31.3	0.24	87.1 ± 0.7	86.2 ± 0.4
Baszkówka	L5	2.76	5.99	2.82	0.74	56.2	0.65	30.4	0.32	86.1 ± 0.6	86.3 ± 0.3
DaG 952	L6	2.77	6.00	2.98	0.79	55.7	0.56	30.3	0.44	86.0 ± 2.2	86.2 ± 0.4
NWA 848	L6	2.46	5.58	2.76	0.74	54.9	0.83	32.1	0.32	88.0 ± n.d.	86.9 ± n.d.
Mean L chondrites		2.52	5.90	2.67	0.70	56.1	0.63	30.9	0.34	87.3	86.5
Std. deviation		0.21	0.19	0.44	0.06	0.8	0.08	0.60	0.06	1	0.3
Not included in the mean											
Haxtun	L4	3.49	4.86	1.69	0.63	59.3	0.69	28.4	0.23	82.0 ± 1.8	89.1 ± 3.5
HaH 219	L4	5.08	4.45	1.76	0.64	60.5	0.66	26.7	0.15	74.7 ± 1.0	90.1 ± 0.7
NWA 767	L4	1.55	5.62	2.62	0.83	56.8	0.74	31.7	0.38	92.0 ± 0.6	87.2 ± 0.1
Etter	L6	3.80	5.29	3.05	0.54	58.7	0.66	26.9	n.d.	79.9 ± 1.3	88.2 ± 0.3
Rio Limay	L6	5.53	6.32	2.58	0.64	57.1	0.58	27.1	0.11	73.3 ± 2.0	85.9 ± 0.8
Bhola	LL3–6	1.83	5.55	3.43	0.68	55.9	0.50	31.5	n.d.	90.6 ± 0.6	87.1 ± 0.5
Krähenberg	LL5	2.09	5.77	3.12	0.80	56.6	0.46	30.6	n.d.	89.1 ± 1.5	86.8 ± 1.3
Vishnupur	LL6	1.77	5.31	4.29	0.60	55.0	0.51	31.9	n.d.	91.0 ± 0.8	87.4 ± 0.7
Uden	LL7	1.72	5.44	2.75	0.60	55.7	0.57	32.3	0.24	91.4 ± 1.6	87.3 ± 0.5
Mean LL chondrites		1.85	5.52	3.40	0.67	55.8	0.51	31.6	–	90.5	87.2
Std. deviation		0.14	0.17	0.57	0.1	0.56	0.04	0.62		0.9	0.2

n.d.: not determined; Fe#: mole% Fe/(Fe + Mg); Cr#: mole% Cr/(Cr + Al).

exsolutions of ilmenite, e.g., in ALHA77262, as shown in Fig. 6. Vanadium and zinc are always present in more or less constant amounts.

Cr Spinel

As shown above, chromites in equilibrated chondrites have constant compositions with a narrow range of Fe/(Fe + Mg) and Cr/(Cr + Al) ratios. The chromites deviating from the average compositions listed in Table 1 have also constant compositions. But some chondrites in addition contain grains with variable Cr/(Cr + Al) and Fe/(Fe + Mg) ratios. All such grains with variable and usually lower than normal Cr/(Cr + Al) ratios are called here Cr spinel. They occur mainly in type 3 chondrites, but also in several type 4 and 5, and occasionally also in type 6 chondrites. They are found more frequently in H than in L chondrites, and rarely in LL chondrites.

In type 3 chondrites, Cr-spinel grains are mostly small and occur almost exclusively inside chondrules. Fig. 5 shows all Cr-spinel grains found in one thin section of L3.7 Julesburg, showing the varieties of shapes. Most of them fit into the texture of the silicates, indicating that they formed in situ. Others look corroded and could be relict grains formed elsewhere. The grains found in type 4 are larger (Fig. 6), and in type 5 and 6 only large grains occur in large chondrules rich

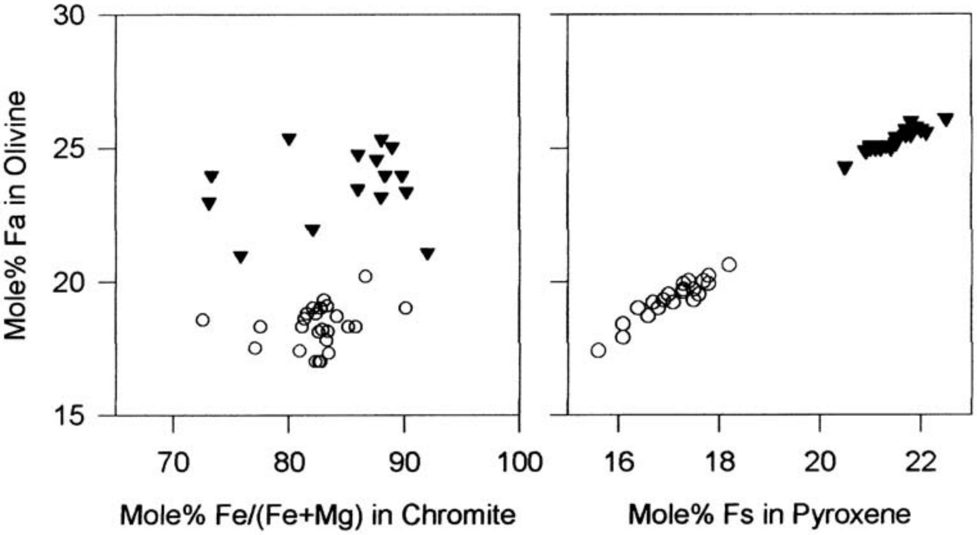

Fig. 4. Left: mole% Fe/(Fe + Mg) in coexisting chromite and olivine for individual H (open circles) and L (triangles) chondrites (data from this work). Right: olivine Fa versus pyroxene Fs for a suite of H and L chondrites (data from analyses of the Acfer chondrites by T. Geiger and A. Bischoff, in Wlotzka 1994).

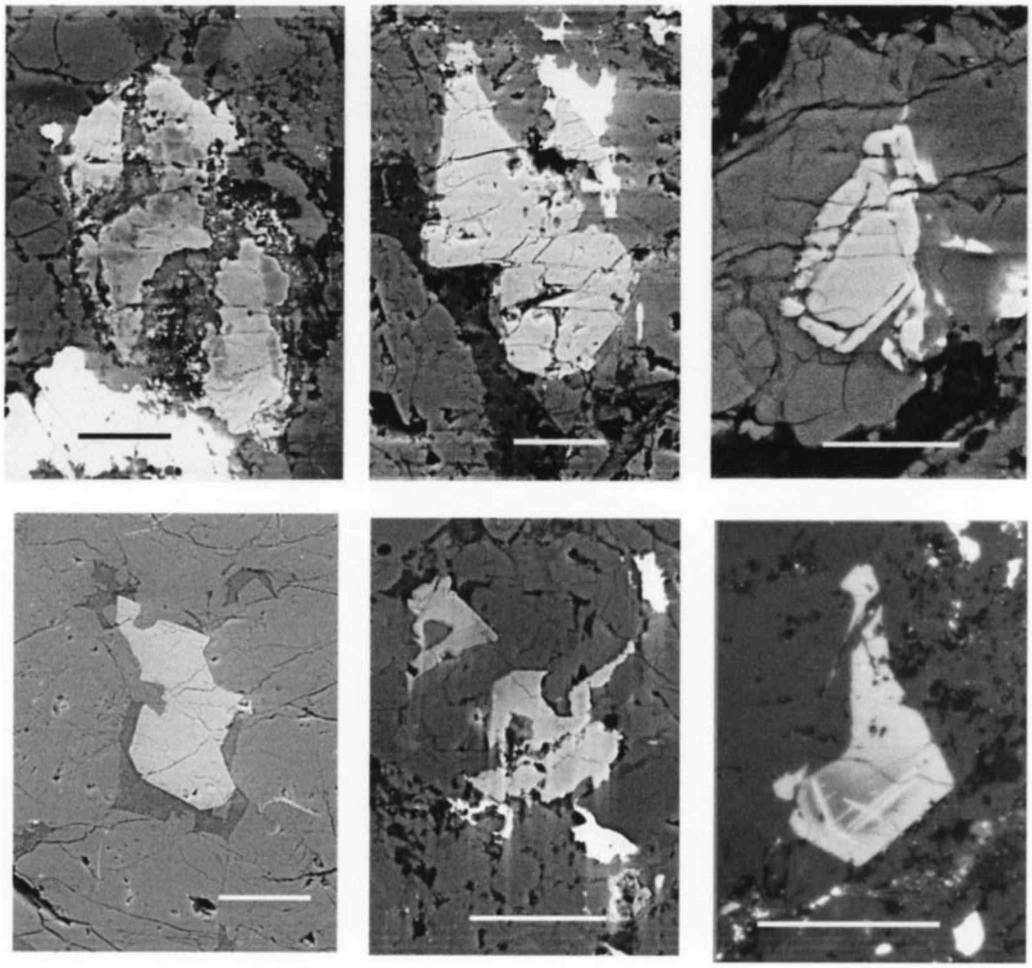

Fig. 5. Cr-spinel grains found in one thin section of L3.7 Julesburg. SE pictures, scale bar 20 µm.

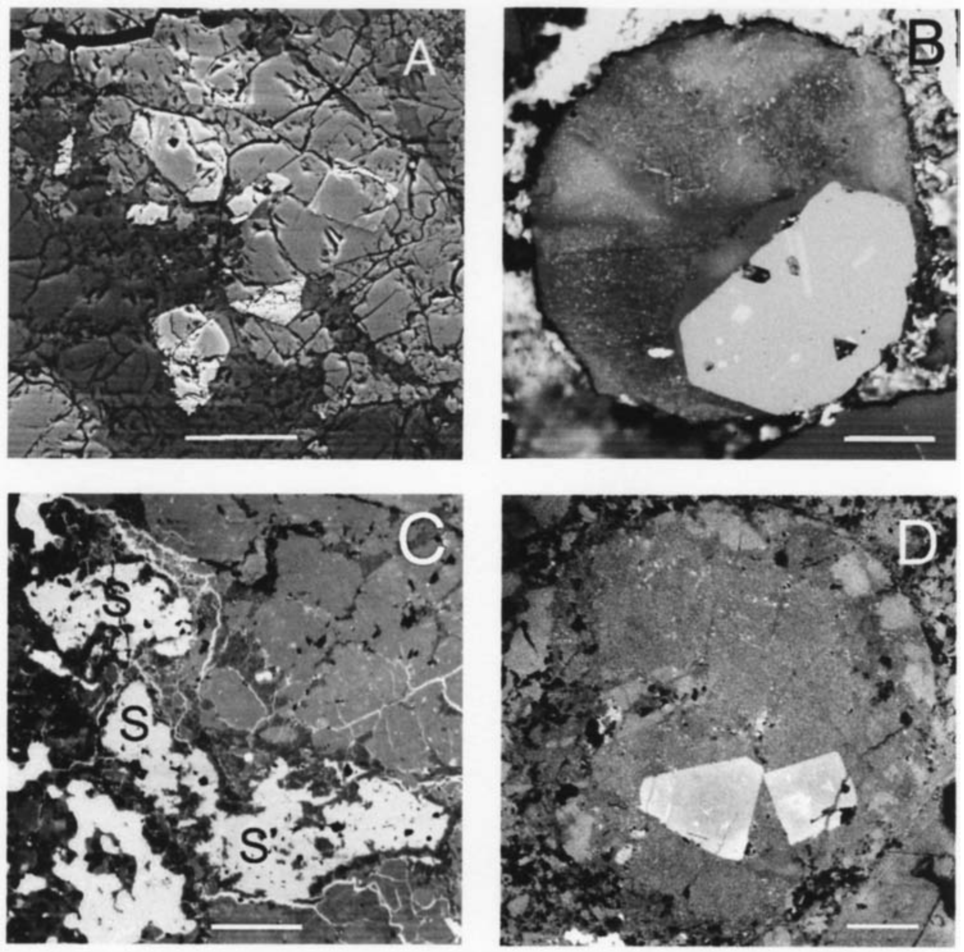

Fig. 6. Cr-spinel grains in type 4 chondrites. Reflected light, scale bar 20 µm. a) L4 MAC 87306: small euhedral grains in olivine and mesostasis of a chondrule. b) H4 Forest Vale: large crystal in a cryptocrystalline chondrule. c) Irregular grain in H4 ALHA77262, marked S. D: L4 Bjurböle: euhedral grains in a chondrule.

in mesostasis hindering Cr and Al diffusion (Fig. 7). Cr-spinel grains are usually zoned in Cr/(Cr + Al) and Fe/(Fe + Mg) and these ratios increase towards the exterior of the grains. In special cases, where spinel grains are close to a chondrule margin, the Cr/(Cr + Al) and Fe/(Fe + Mg) ratios are the highest in the crystal face adjacent to this margins, lower in the parts facing the chondrule interior. Figure 8 shows an example from H5 Flandreau: a large spinel grain occurs 50 microns away from the rim of a 0.7 mm chondrule. It shows an Fe/(Fe + Mg) and Cr/(Cr + Al) gradient with enrichments towards the chondrule surface. Also, the large Cr-spinel grain in a chondrule in H5 Richardton (see Fig. 7a) shows an increase in Cr/(Cr + Al) towards the chondrule interface with the chondrite matrix (Fig. 9). These enrichments suggest that Fe and Cr were introduced from the chondrite matrix by diffusion on the parent body, which leads ultimately to equilibrated chromite compositions. More rare are large, more or less rounded Cr-spinel grains in the matrix (Fig. 10). Characteristically, these grains are usually intergrown with and surrounded by a feldspatic rim, and thus isolated from the chondrite matrix, similar to the grains inside chondrules. They form a transitional group to the Al, Cr-rich objects described below.

Table 2 lists examples of Cr-spinel analyses, in order to compare them with the equilibrated chromites. Fe/(Fe + Mg) and Cr/(Cr + Al) vary in these Cr spinels, but in equilibrated chondrites they are always correlated (Fig. 11). Noteworthy is that Ti is lower than in the chromites, and that it correlates with the Cr and Fe contents. Zn is usually higher than in the chromites. Heterogeneous Cr-spinel grains in equilibrated chondrites were reported before, e.g., in H4 Ybbsitz (Brandstätter et al. 1985), in H5 Morro do Rocio (Fredriksson and Wlotzka 1985), in the L4 RC071 and the L5 RC072 by McCoy et al. (1991), or in L6 Los Martinez (Brearley et al. 1991).

Al, Cr-Rich Objects

Chromite and Cr spinel occur also in special Al- and Cr-rich inclusions (Bischoff and Keil 1984). These consist mainly of Cr-rich spinel and plagioclase. The shape varies

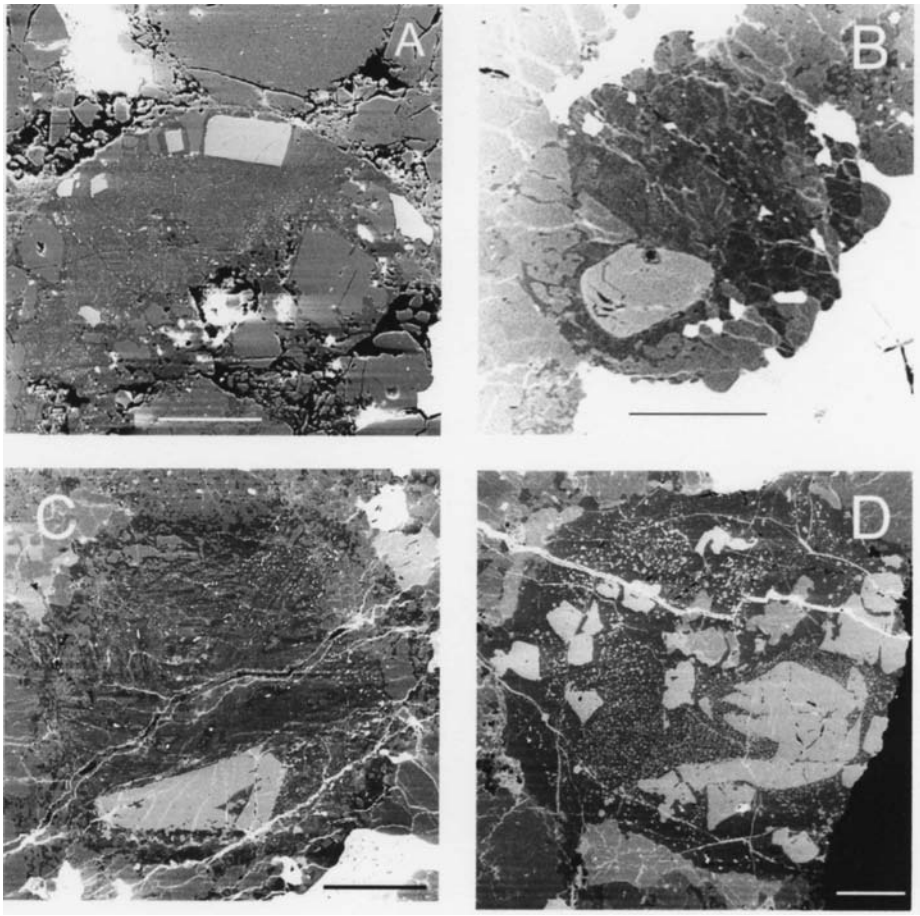

Fig. 7. Cr-spinel grains in chondrules of type 5 and 6 chondrites. SE pictures, scale bar 100 µm. a) H5 Richardton. b) H5 Acfer 206. c) H6 Daraj 016. d) L6 Dar al Gani 952.

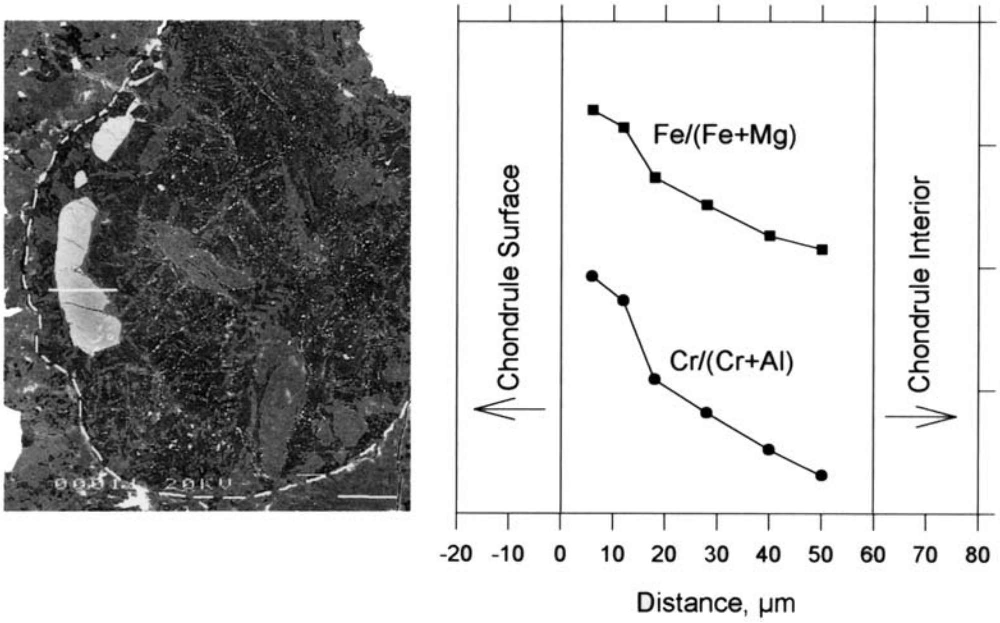

Fig. 8. Left: Cr-spinel grain (light grey) in a large, mesostasis-rich chondrule in H5 Flandreau. SE picture, scale bar 50 µm. Right: Cr/(Cr + Al) and Fe/(Fe + Mg) profiles along the white line through the Cr-spinel grain.

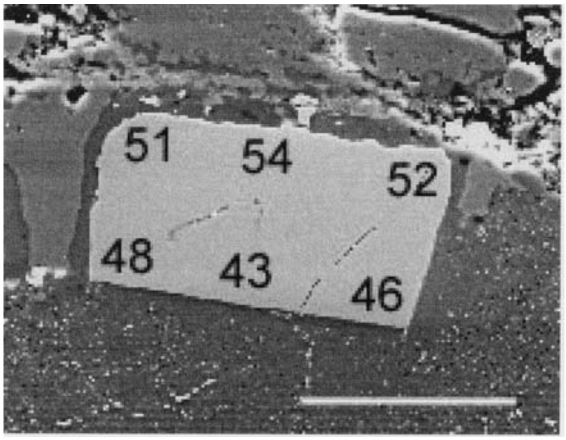

Fig. 9. Cr-spinel grain from a chondrule in H5 Richardton (shown in Fig. 7a). The chondrule boundary is just above the grain. SE picture, scale bar 50 µm. The numbers are mole% Cr/(Cr + Al), showing the enrichment towards the chondrule/matrix boundary.

from massive, sometimes chondrule-like to fine intergrowths of spinel and feldspatic mesostasis (Fig. 12). Christophe Michel-Lévy et al. (1995) described nuggets enriched in refractory metals from them. A special form are the "chromite-plagioclase assemblages" described as shock indicators by Rubin (2003). It is not the purpose of this paper to discuss the origin of the different forms of these objects. They were encountered in some of the meteorites studied (Forest Vale, Beddgelert) and were usually found to be equilibrated in Fe, Mg with olivine of the host meteorite, so that they could be used for olivine/Cr-spinel thermometry.

Olivine

The Fe, Mg exchange partner of chromite and Cr spinel are olivine and pyroxenes. The thermometers developed for this exchange are mainly based on olivine because its structure is simpler than that of the pyroxenes. Olivine is more than 100 times more abundant than chromite and Cr spinel in chondrites, so that its composition does not change during an equilibration with the latter minerals. In equilibrated chondrites, Cr-spinel grains are mainly found inside large chondrules where they are surrounded by feldspatic mesostasis. Usually olivine grains are also present inside these chondrules. In nearly all cases these olivines and olivine grains elsewhere in the meteorite had the same constant composition, as expected for an equilibrated chondrite. An exception is H5 Morro do Rocio, where some grains of different composition were found (Fredriksson and Wlotzka, 1985). Here the composition of the main constant olivines were taken for the temperature calculation. In type 3.7 to 3.8 chondrites olivine has a variable composition. As an example the olivine Fa contents in Cr spinel-bearing chondrules of L7 Julesburg are shown together with the isotherm diagrams below (Fig. 17). In the type 3 chondrites, the average composition of the olivines coexisting with Cr spinel in a chondrule were taken for temperature calculations.

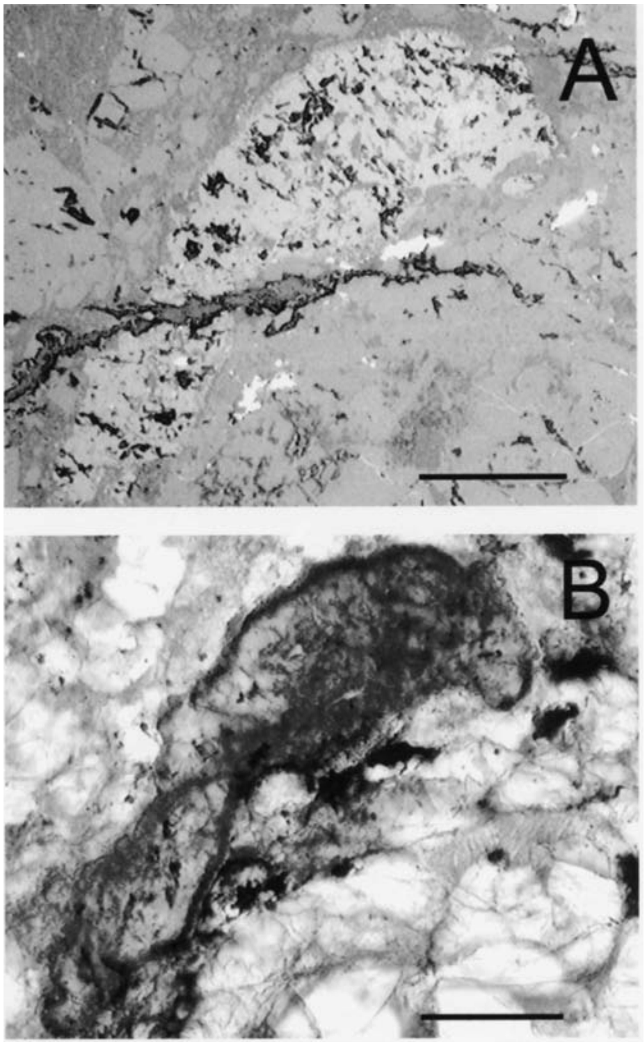

Fig. 10. Elongated, curved Cr-spinel grain in the matrix of H4 Forest Vale; scale bar 50 µm. a) Reflected light. b) Transmitted light. The dark rim is due to enrichment in Cr and Fe.

Usually the spread in Fa values encountered was small (see Fig. 17) and a choice between rim or central values not critical, as a difference in 1 mole% Fa will change the calculated temperature, e.g., in the case of Julesburg, by only 16 °C.

THERMOMETRY

The Fe, Mg exchange between Cr spinel and olivine was studied by several authors. It was found that the distribution coefficient $K_D = (\mathrm{Mg/Fe})_{\mathrm{olivine}}/(\mathrm{Mg/Fe})_{\mathrm{spinel}}$ at a given temperature depends on Cr/(Cr + Al) of the coexisting Cr spinel. Ln K_D shows a linear correlation with Cr/(Cr + Al). The position and slope of these lines depends on the equilibration temperature. Experiments around 1200 °C and natural mineral assemblages in metamorphic rocks confirmed the fact that under isothermal conditions straight correlation lines between ln K_D and the Cr/(Cr + Al) ratio exist.

Table 2. Composition of Cr spinel from H and L chondrites. Analyses with the minimum and the maximum Cr# (which show also the minimum and maximum Fe#) are given.

Name	Type	MgO	Al_2O_3	TiO_2	V_2O_3	Cr_2O_3	MnO	FeO	ZnO	Fe#	Cr#
Tieschitz	H3.6	21.6	67.1	0.46	0.12	0.18	0.06	9.16	0.36	19.3	0.2
		16.8	62.6	0.21	0.19	2.51	0.11	15.4	0.58	34.5	2.6
Dhajala	H3.8	17.5	65.7	0.35	–	1.98	–	13.7	n.d.	30.5	2.0
		2.63	9.02	2.23	0.75	54.3	1.37	29.2	n.d.	86.2	80.2
Clovis No. 1	H3.9	19.1	67.0	0.25	0.20	0.19	2	13.1	0.50	27.8	0.2
		4.65	17.1	0.88	1.00	48.9	0.88	25.9	0.40	75.8	65.7
ALHA77262	H4	15.7	52.2	0.19	0.16	15.01	0.27	16.4	0.62	36.9	16.2
		4.79	8.51	1.84	0.55	56.5	0.75	26.4	0.45	75.5	81.7
Forest Vale	H4	16.5	59.2	0.04	0.14	6.09	0.01	14.4	0.22	32.9	6.5
		6.40	14.2	1.21	0.60	52.2	0.70	25.2	0.16	68.9	71.2
Daraj 020	H5	12.3	34.1	0.34	0.32	30.6	0.54	20.6	0.50	48.6	37.6
		4.93	9.48	1.40	0.65	54.8	0.98	27.1	0.25	75.5	79.5
Simmern	H5	11.7	36.7	0.27	0.29	30.6	0.84	19.3	0.33	48.2	35.8
		6.01	11.08	3.36	0.49	50.0	1.05	27.3	0.66	71.8	75.2
Richardton	H5	10.8	33.0	0.47	0.21	34.4	0.49	20.5	0.51	51.2	41.2
		8.64	26.0	0.56	0.25	40.5	0.59	23.0	0.48	59.9	51.1
Daraj 016	H6	13.6	44.8	0.18	0.11	22.1	0.49	19.0	0.54	43.9	24.8
		5.35	11.5	1.96	0.67	51.1	0.84	27.7	0.26	74.4	74.9
Suwahib (Buwah)	L3.7	16.6	57.8	0.10	0.13	8.14	0.06	13.6	1.10	31.5	8.6
		4.38	9.37	0.16	0.39	56.5	0.90	26.2	0.29	77.0	80.2
Julesburg	L3.7	12.5	50.9	0.36	0.1	15.1	0.33	19.4	2.45	46.5	16.6
		2.88	12.4	1.76	0.78	50.7	0.86	29.7	0.76	85.2	73.7
Acfer 047	L4	11.2	41.0	0.23	0.17	24.9	0.51	22.6	0.47	53.1	28.9
		4.8	12.5	1.13	0.34	51.8	1.21	27.9	0.33	76.5	73.6
Bjurböle	L4	13.3	59.2	0.19	–	8.23	0.25	17.7	1.15	42.9	8.5
		5.16	18.0	1.29	–	44.9	0.76	29.0	0.93	75.9	62.6
Acfer 307	L5	11.8	47.7	0.17	0.14	17.3	0.38	22.0	1.44	51.1	19.6
		3.51	8.71	0.70	0.76	56.0	1.06	29.7	0.56	82.6	81.2
Dar al Gani 952	L6	9.37	31.8	0.95	0.12	32.7	0.26	24.3	0.28	59.3	40.8
		4.56	12.12	2.44	0.49	50.9	0.49	28.8	0.07	78.0	73.8

Fe#: mole% Fe/(Fe+Mg). Cr#: mole% Cr/(Cr + Al).

Conversely, a straight correlation line in such a diagram shows that the coexisting minerals in this rock equilibrated under isothermal conditions.

Several geothermometers have been developed for this reaction. The high temperature end was approached by equilibration experiments at 1200 °C, whereas the lower temperature end at 700 °C was constructed from natural assemblages of metamorphic rocks (Evans and Frost 1975). The application of such thermometers to chondritic chromites or Cr spinels is straightforward, because they are rather pure solid solutions of the end members and contain only negligible amounts of trivalent Fe (due to the presence of Fe also in metallic form) and only low contents of Ti and other interfering elements (see Table 2). On the other hand, these thermometers were up to now only seldom used for chondritic meteorites, because the different calibrations given in the literature differ strongly, and the right choice is difficult.

Figure 13 gives 700 °C olivine/spinel isotherms for a number of different calibrations, they differ mainly in the slope of the correlation between Cr/(Cr + Al) and ln K_D. The projection point for OC chromites in this diagram lies above some and below others of these lines. The corresponding temperature varies between 600 and 750 °C and it is difficult to decide which calibration may be best suited for these rocks. In this situation, the Cr-spinel grains with a larger range of Cr/(Cr + Al) ratios can be used to select the right calibration.

Evaluation of Different Olivine/Cr-Spinel Geothermometers

Figure 13 shows also Cr-spinel data from the H5 chondrite Flandreau. It is evident that the Cr-spinel data show a linear correlation in this diagram, but define a steeper slope than most proposed isotherms. They fall close to those of

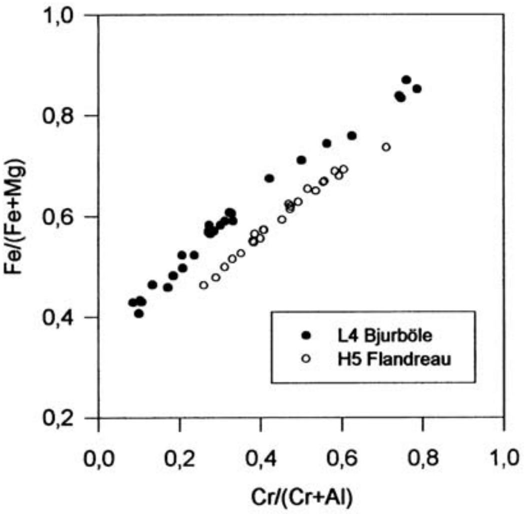

Fig. 11. Correlation of Fe/(Fe + Mg) with Cr/(Cr + Al) in Cr-spinel grains of L4 Bjurböle and H5 Flandreau.

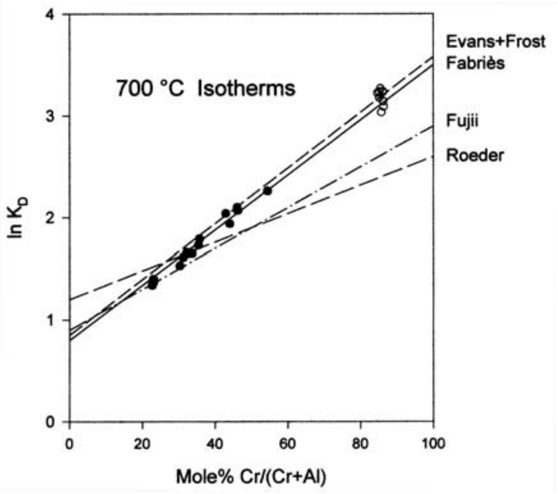

Fig. 13. Olivine/spinel 700 °C isotherms as calibrated by Evans and Frost (1975), Fabriès (1979), Fujii (1977) and Roeder et al. (1979), together with Cr spinel (closed circles) and chromite (open circles) data from H5 Flandreau.

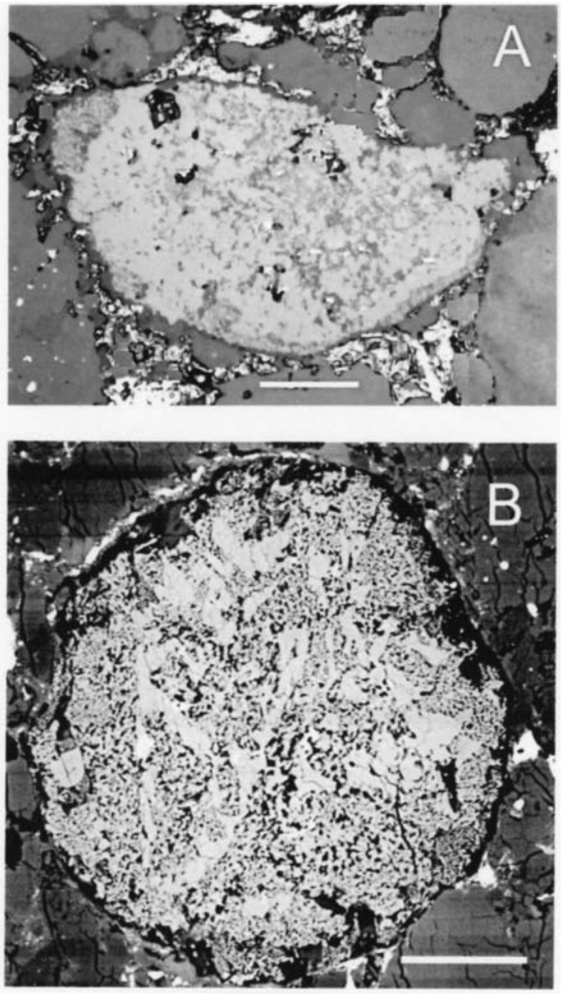

Fig. 12. Cr-rich objects, scale bar 50 µm. a) H4 Forest Vale, chondrule fragment? Reflected light. b) Cr spinel + plagioclase chondrule in H4 Beddgelert, SE picture.

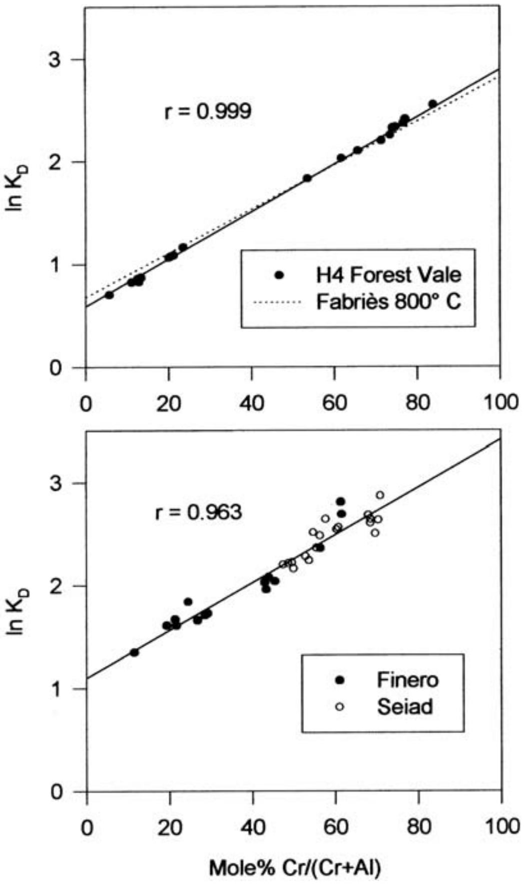

Fig. 14. Top: olivine/spinel 800 °C isotherm (dotted line) by Fabriès (1979) and data for Cr-spinel grains from H4 Forest Vale and correlation line (solid line). Bottom: isotherm diagram for Cr-spinel grains from granulitic metaperidotites of Finero and Seiad (Medaris 1975), which were used by Fabriès for calibration of the 700 °C isotherm.

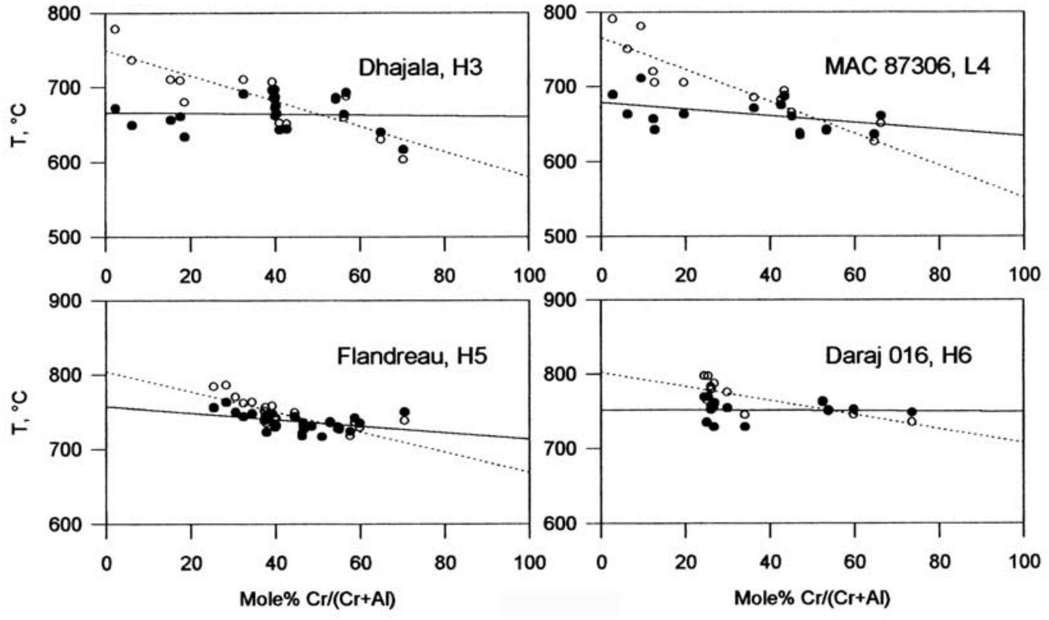

Fig. 15. Cr/(Cr + Al) ratio in Cr spinel versus temperature *T* calculated after the original Fabriès thermometer (open symbols, dotted line), and the modified one (closed symbols, solid line).

Evans and Frost (1975) and of Fabriès (1979). This correspondence indicates Fe/Mg equilibrium between Cr spinel and olivine, and that these isotherms can be successfully applied here. The advantage of the Cr-spinel data with variable Cr/(Cr + Al) is, that the Fe/Mg equilibrium can be seen over a wide stretch of the isotherms, and thus a specific thermometer can be selected with more confidence than for a single point at constant Cr/(Cr + Al) on one of these isotherms.

Fabriès (1979) used empirical data for the construction of his thermometer. For the 700 °C isotherm he used analyses of Cr spinel from metaperidotites by Medaris (1975), the 700 °C being inferred from the mineral equilibria of the granulitic facies. A 1200 °C isotherm was constructed from several natural and experimental data, i.e., Kilauea Iki pumices (Evans and Wright 1972), spinel lherzolite equilibrated at 1200 °C and 16 kb (Mori 1977), lunar basalt 14310 equilibrated at 1230 °C and 5 kb (Green et al. 1972). He gives a thermometer equation and a modification of it this was used in the following temperature determinations.

Figure 14a compares the 800 °C isotherm of Fabriès with data for Cr spinel from the H4 chondrite Forest Vale and shows that these data define a somewhat steeper line. Such a deviation was noticed also for Cr spinel from most other chondrites measured. The 800 °C isotherm was therefore recalculated according to the Forest Vale correlation line, and the other isotherms between 700 and 1200 °C according to an $1/T$ relation. This adjustment seems justified, because the Forest Vale correlation is much better (r = 0.999) than that for the 700 °C isotherm by Medaris (r = 0.963), which Fabriès used for his calibration (Fig. 14b). The reason is that in the

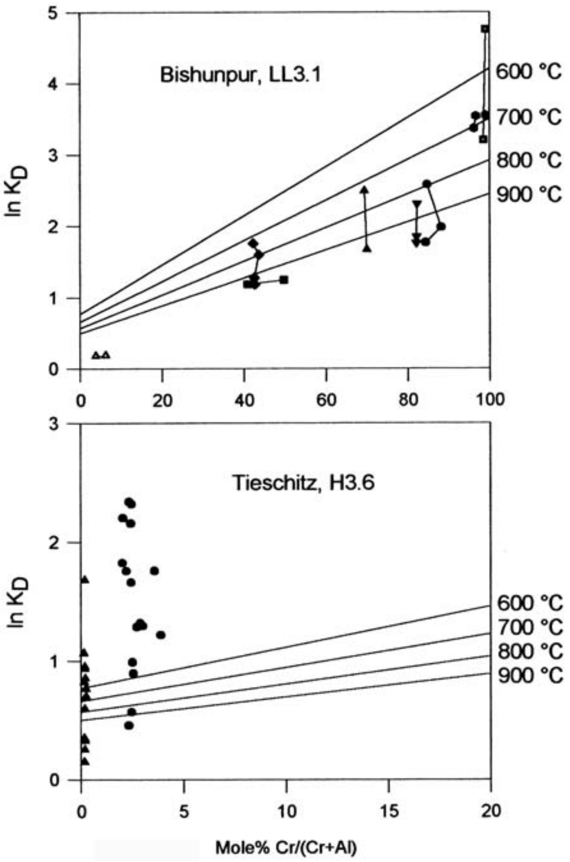

Fig. 16. Isotherm diagram mole% Cr/(Cr + Al) versus ln K_D for spinel grains in LL3.1 Bishunpur and H3.6 Tieschitz. Analyses from one chondrule are marked by the same symbol, and for Bishunpur connected by lines.

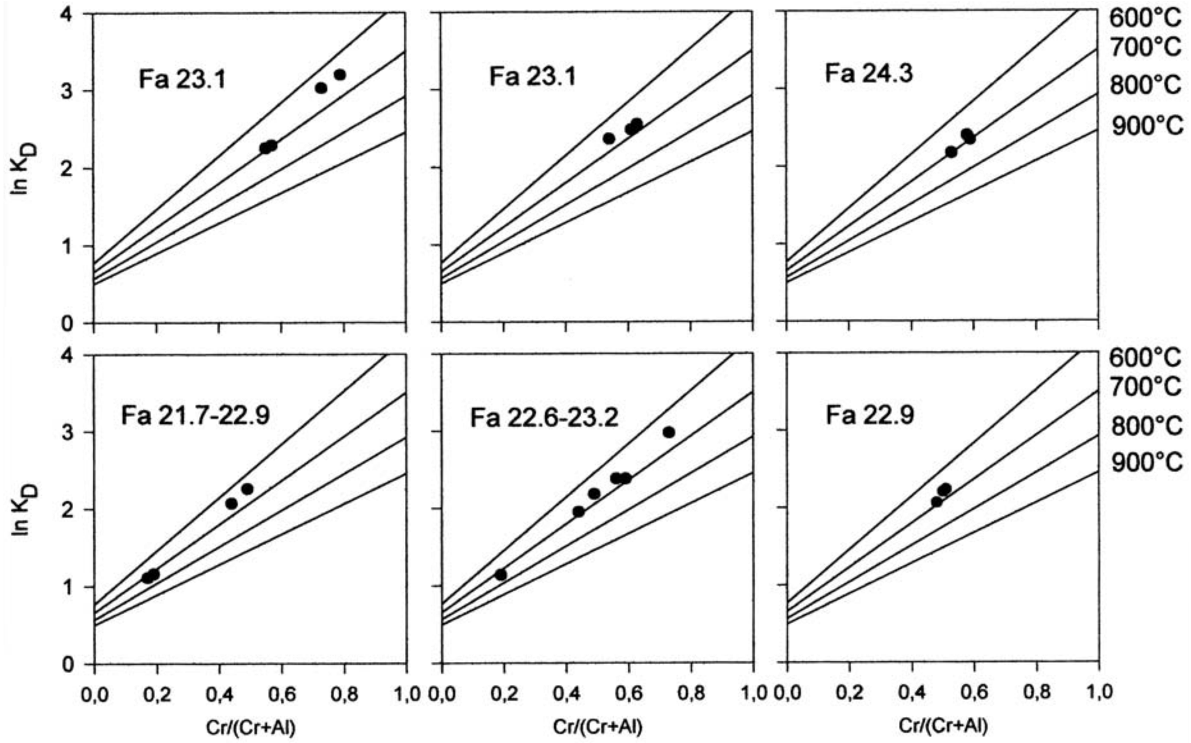

Fig. 17. Isotherm diagram for Cr-spinel grains in L3.7 Julesburg. Each diagram represents spinel grains from one chondrule and gives the Fa value for this chondrule.

Forest Vale chondrite spinel grains with different Cr/(Cr + Al) ratios occur in the mm-range within the same thin section, and were thus exposed to identical thermal conditions. In contrast, the data by Medaris (1975) were taken from rock samples collected in different regions of a large metamorphic unit, where temperature gradients can occur.

The new thermometer equation is the following:

$$T(°C) = \frac{5099 \times Y_{Cr} + 917}{\ln K_D + 0.282 + 2.402 \times Y_{Cr}} - 273 \qquad (1)$$

with Y_{Cr} = mole ratio Cr/(Cr + Al) in spinel; distribution coefficient $K_D = (Mg/Fe)_{olivine}/(Mg/Fe)_{spinel}$.

The temperatures derived from this equation differ from those derived from Fabriès original one mainly at low Cr/(Cr + Al) ratios. Figure 15 shows that these temperature are more consistent also for other chondrites at other temperatures than Forest Vale: the Fabriès temperatures increase with decreasing Cr/(Cr + Al) ratio, whereas the modified version yields constant temperatures throughout. The average temperatures from both equations differ only slightly, but the new equation yields lower standard deviations.

The advantage of this graphic thermometer solution over purely numerical calculations which were developed more recently (Sack and Ghiorso 1991a) is that the results can be vividly seen over the whole range of Cr/(Cr + Al) ratios. It is thus possible to evaluate the degree of equilibration at once, especially since the Fe-Mg interdiffusion coefficient in spinel varies with this ratio (see below).

It is not possible, however, to define the absolute temperature with this thermometer alone. For the metamorphic temperatures found in chondrites, it depends critically on the calibration for the low temperature region, which is not accessible by experiments. It will be discussed later, if a comparison with other, independent calibrations can solve this problem. The temperature results presented below should therefore be mainly regarded as relative ones.

RESULTS

Cr-Spinel Temperature

The modified thermometer equation was applied to Cr-spinel grains and olivines from chondrites of different petrologic types. The unequilibrated chondrites LL3.1 Bishunpur and H3.6 Tieschitz give no isotherm correlations (Fig. 16), only a row of decreasing ln K_D values at about constant Cr/(Cr + Al). These values are inherited from the cooling of molten chondrules and no equilibration took place on the parent body. In the partially equilibrated type 3.7 to 3.8 chondrites we see a different picture. Figure 17 shows as an example isotherms for a number of Cr-spinel grains in individual chondrules of L3.7 Julesburg (see Fig. 5), together with the Fa values found in these chondrules and used for calculation of the K_D values. The Cr-spinel grains are either in direct contact with olivine or are included in a feldspatic mesostasis containing also olivine. The olivine composition

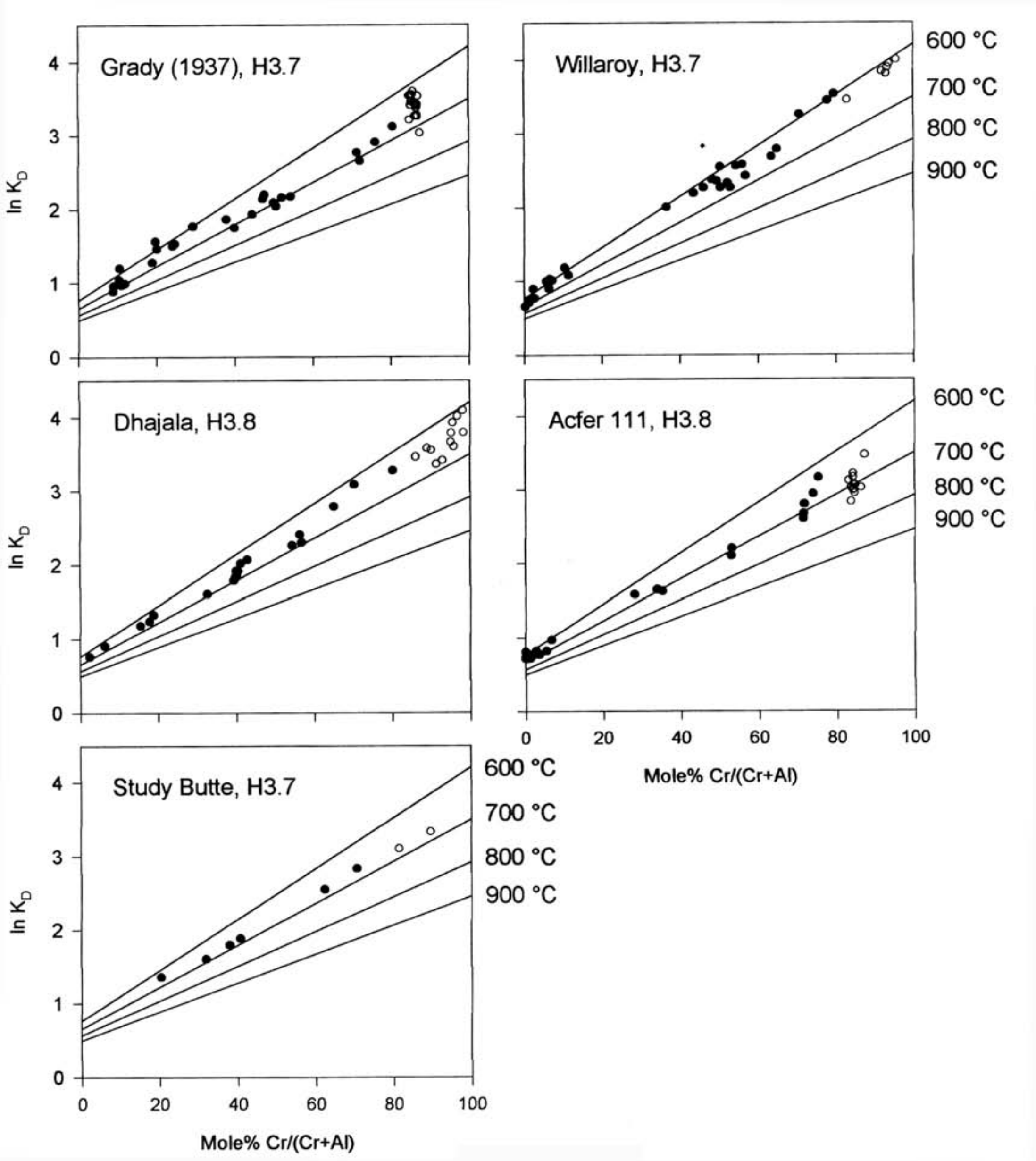

Fig. 18. Isotherm diagram for Cr spinel (solid symbols) and chromite (open symbols) for H3.7 to 3.8 chondrites.

may vary from grain to grain, but in Julesburg all Fa values measured are close to the average of 22.6 found by Graham (1993). Variations in a given chondrule are small and the mean Fa value in the chondrule where the Cr spinel occurred was taken for the temperature calculation. It is evident that these grains define isotherms and all grains in Julesburg were equilibrated at the same temperatures. Similar isotherm correlations were found for a number of other H and L 3.7 and 3.8 chondrites (Figs. 18 and 19).

The temperatures calculated for type 3.7 to 3.8 range from 625 to 683 °C, with an average of 660° ± 20 for H and 653° ± 21 for L chondrites. The two H3.9 Bremervörde and Clovis No. 1 have a significantly higher temperature, 712 and 745 °C, respectively, which already overlap with the type H4 range (Fig. 20). ALH 85045, classified as L3.8 according to its TL sensitivity, is an outlier; its temperature is in the L4 to 6 range. Its limited range in olivine composition at Fa 22 to 25 (Grossman 1994) puts it close to the L4 group. No trace of an igneous temperature can be found in these chondrules. One would expect to see differing temperatures for individual chondrules and also higher temperatures than 650 °C.

All H chondrites of type 4 to 6 studied show good isotherm correlations (Figs. 21–23). It is evident that a major temperature difference exists between petrologic type H3.7 to 3.8 and the equilibrated types H4, 5 and 6: the latter show an about 100 °C higher temperature: 763, 774, and 775 °C,

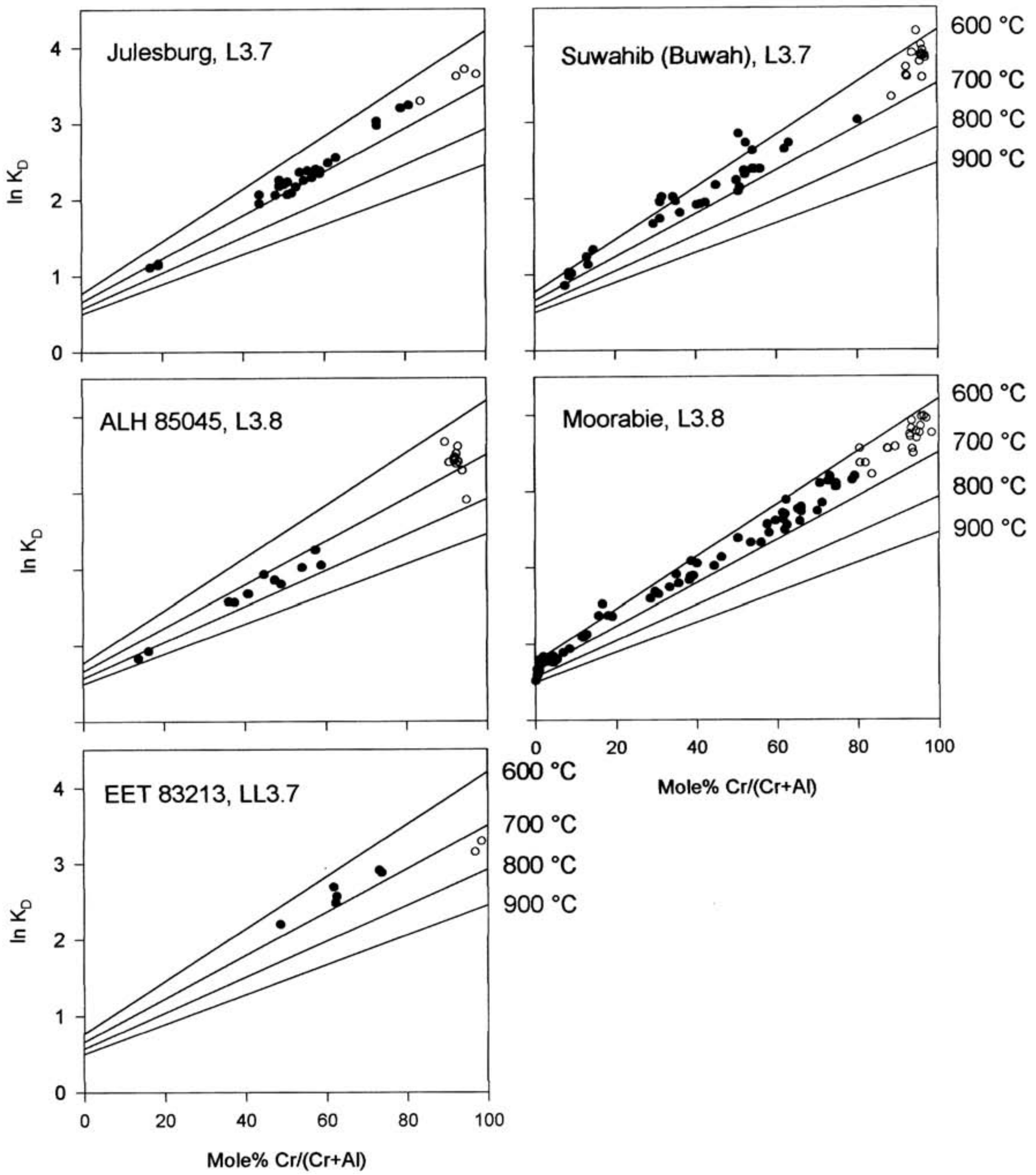

Fig. 19. Isotherm diagram for Cr spinel (solid symbols) and chromite (open symbols) for L3.7–3.8 and one LL 3.7 chondrite.

respectively (Table 3). A range of 70 to 100 °C exists in a given group, and the ranges for the equilibrated types (including type H3.9) overlap. Within type H5 and H6 two groups seem to form with about 750 and 800 °C, respectively (see Fig. 26 below). More data are needed, however, to verify such a grouping. The main result is that the petrologic types H4, 5, and 6 experienced essentially the same temperatures.

The temperature range displayed within the petrologic groups is a real effect and not caused by uncertainties in the measurements, so that equilibration temperatures can be distinctly different for individual chondrites of the same type. This can be seen in Table 3 and by comparing the different isotherms, e.g., for H5 Flandreau (737 °C) and H5 Dar al Gani 739 (815 °C), see Fig. 22.

In the L group we see the same behavior. Type L3.7–3.8 (Fig. 19) lie here 100 °C lower than type L4, 5, and 6 (Figs. 24 and 25), which have identical average equilibration temperatures: 752, 754, and 754 °C, respectively (Table 4). The range within the L4 group is about 80 °C, if we ignore the low value of 664 °C for MAC 87306. The latter temperature is in the type 3 range, but this L4 meteorite has a range in pyroxene composition of Fs 8 to 21 (Grossman 1994) and is

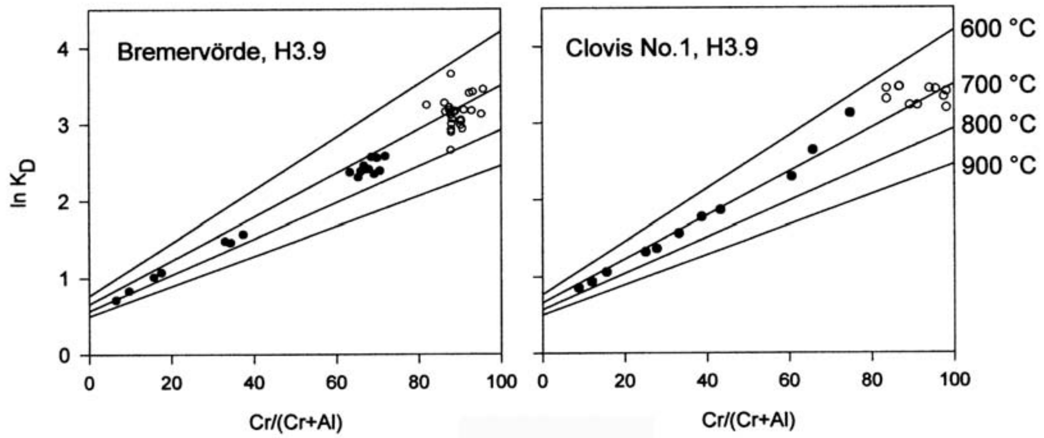

Fig. 20. Isotherm diagrams for Cr spinel (solid symbols) and chromite (open symbols) for H3.9 chondrites.

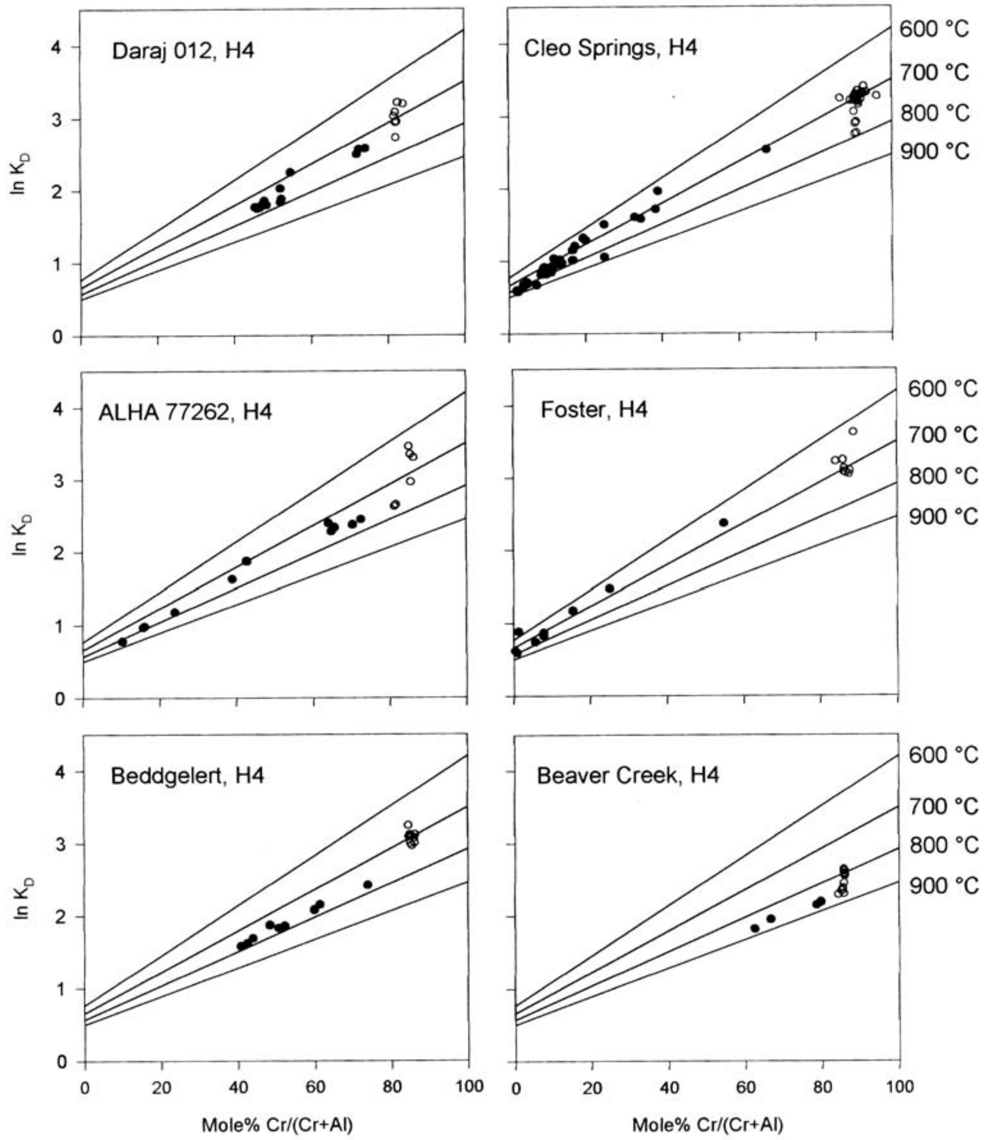

Fig. 21. Isotherm diagrams for Cr spinel (solid symbols) and chromite (open symbols) for H4 chondrites.

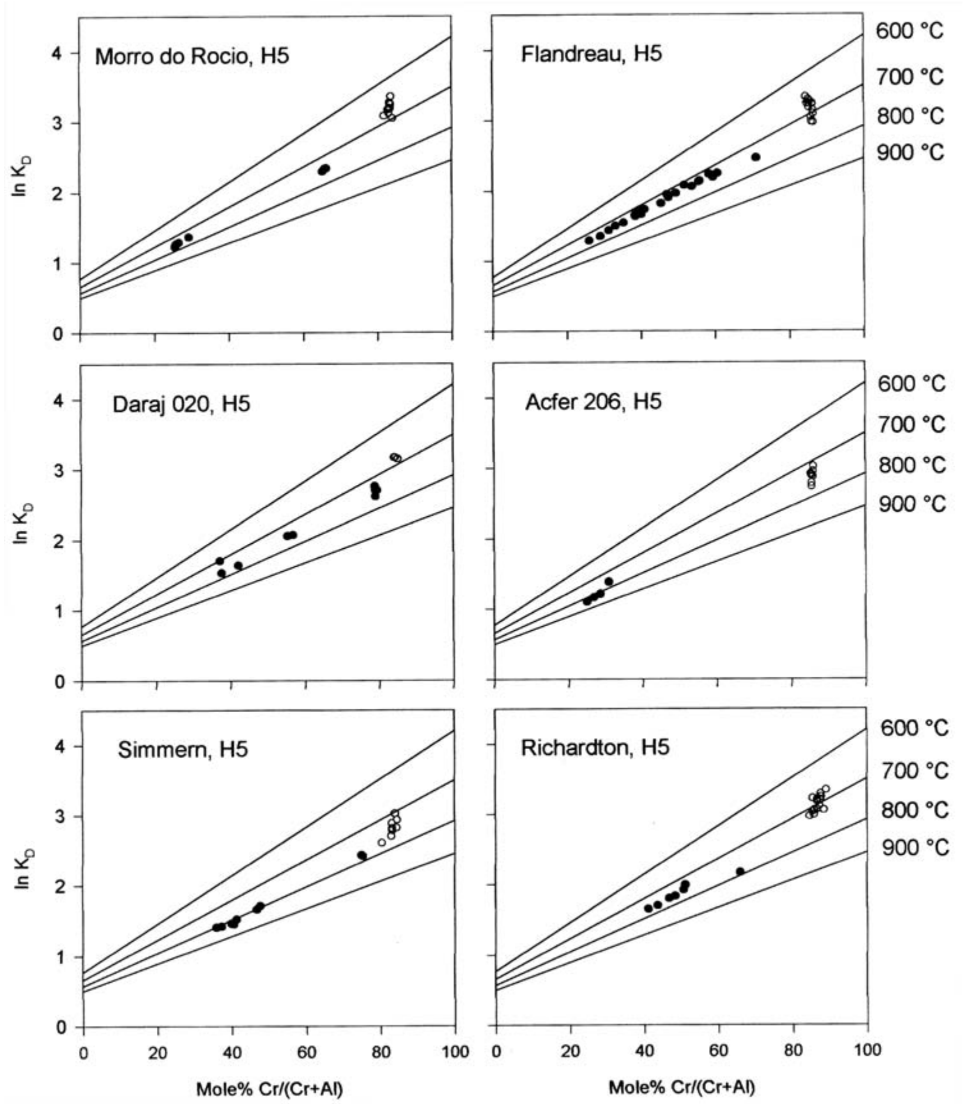

Fig. 22. Isotherm diagrams for Cr spinel (solid symbols) and chromite (open symbols) for H5 chondrites.

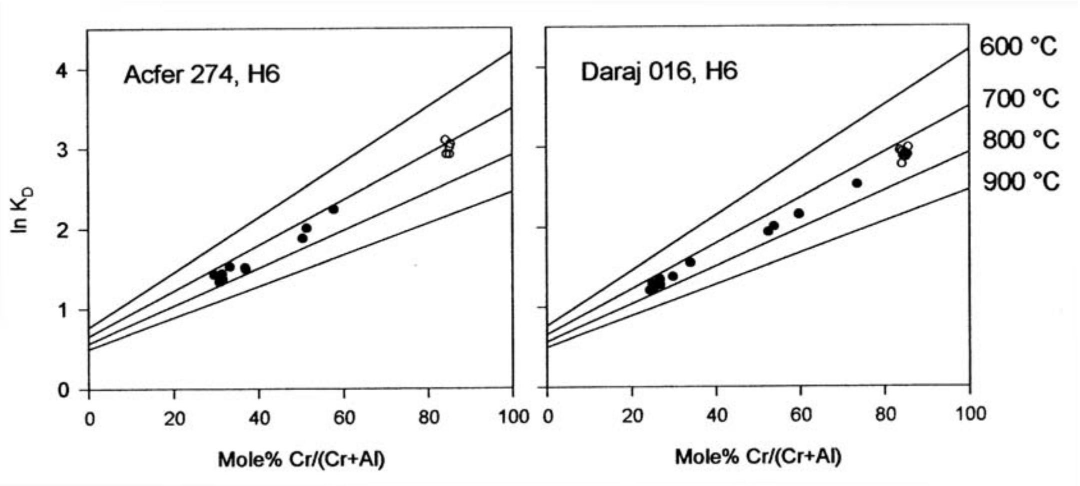

Fig. 23. Isotherm diagrams for Cr spinel (solid symbols) and chromite (open symbols) for H6 chondrites.

Table 3. Average Cr-spinel and chromite temperatures *T* (°C) for H chondrites.

Name	Type	Cr spinel *T* (°C)	Std. dev.	Chromite *T* (°C)	Std. dev.
Willaroy	H3.7	625	30	619	–
Grady (1937)	H3.7	670	30	653	–
Dhajala	H3.8	664	22	644	23
Acfer 111	H3.8	670	36	699	24
Study Butte	H3.7	673	13	n.d.	
Mean (5)	H3.7–3.8	660	20	654	29
Clovis No.1	H3.9	712	31	689	34
Bremervörde	H3.9	745	24	711	36
Mean (2)	H3.9	728	16	700	11
Foster	H4	722	52	690	28
Gobabeb	H4	726	42	n.d.	
Daraj 012	H4	750	25	697	27
Cleo Springs	H4	760	63	710	25
ALHA77262	H4	762	30	670	32
ALH 88010	H4/5	768	16	700	16
Beddgelert	H4	773	11	703	15
Forest Vale	H4	804	12	815	15
FRO 8417	H4	825	20	n.d.	
Beaver Creek[a]	H4	863[a]	5	818[a]	28
Mean (9)	H4	766	31	707	46
Flandreau	H5	737	12	687	15
Daraj 020	H5	746	20	682	3
Morro do Rocio	H5	754	9	668	18
Richardton	H5	754	19	695	15
Simmern	H5	803	16	740	20
Acfer 206	H5	808	28	736	12
Dar al Gani 739	H5	815	5	713	20
Mean (7)	H5	774	31	703	27
Acfer 274	H6	750	22	713	14
Daraj 016	H6	752	13	733	10
Andura	H6	824	8	n.d.	
Mean (3)	H6	775	34	723	10

[a]Reheated, not included in the mean. See text.

thus less equilibrated than the other L4 chondrites measured here. The range within type L5 is even larger, about 130 degrees (Fig. 26). The average temperatures of type L4 to 6 are 20 degrees lower than those of H4 to 6, but in view of the range observed within each type, this difference is not significant. Table 5 and Fig. 26 give a summary of the temperatures obtained for H and L chondrites.

Only few results exist for LL chondrites. One LL3.7 is included in Table 4, its temperature of 676 °C is the same as found for H and L 3.7 and 3.8 chondrites. I did not find any Cr-spinel occurrences in equilibrated LL chondrites. Krot et al. (1992) described Cr spinel in a chondrule from LL6 Melnikovo. They give only average analyses for chromite and the Cr spinel, from these the temperatures in Table 4 were calculated. The Cr-spinel temperature of 777 °C is close to the average of H6 and L6 chondrites (Table 5), and the chromite temperature of 684 °C the same as found for a number of LL chondrites (Table 6).

Shocked and Reheated Chondrites

Reheating by shock or other means can be seen in the olivine/Cr-spinel temperature. Examples are shown in Table 6: the L6 chondrites Etter, Rio Limay, and Santa Catarina were shocked to stage S6 and their chromites register temperatures of 841, 888, and 894 °C, respectively, about 100 to 200 degrees higher than other L chondrites. Thus, their Fe/(Fe + Mg) ratio becomes as low as in average H chondrites (Table 1). Another example is the H4 chondrite Beaver Creek. Hutchison et al. (1981) concluded from its metal texture that it was reheated to above 800 °C. Taylor et al. (1987) found an unusually high metallographic cooling rate of 5000°/Ma, two orders of magnitude higher than for other H4 chondrites. Its

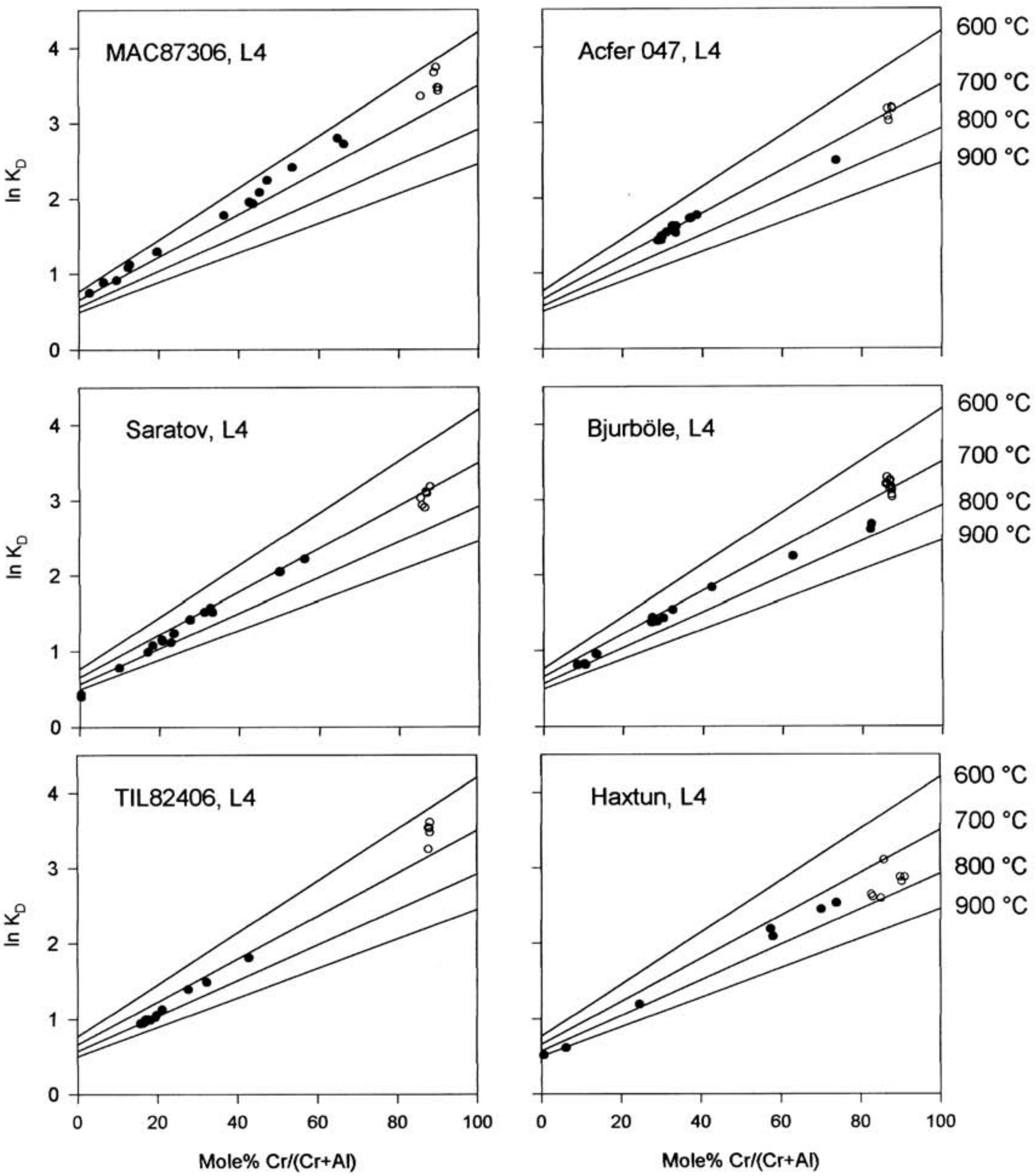

Fig. 24. Isotherm diagrams for Cr spinel (solid symbols) and chromite (open symbols) for L4 chondrites.

Cr-spinel temperature of 863 °C lies 100 degrees above the average for H4 chondrites and outside their range (Table 3). It was therefore not included in the mean value for H4 chondrites.

Chondrites which were reheated to form shock melts show also high temperatures (Table 6). The H5 chondrite Ramsdorf was reheated so that metal and sulfide melted (Begemann and Wlotzka 1969). Small euhedral chromite grains are found in a feldspatic matrix replacing the normal chondritic matrix (Fig. 27a). Their composition differs markedly from that of average H chondrite chromites and register a temperature of 967 °C. Another shock melted L chondrite, Muckera 007, has small newly formed chromites, which seem to have formed by sintering of many tiny grains (Fig. 27b). They yield an olivine/spinel temperature of 882 °C. The isotherm diagrams for Ramsdorf and Muckera 007 (Fig. 28) show a larger scatter of points and look clearly different from the normal chondritic isotherms. Their high temperature equilibrium was preserved by rapid cooling. Begemann and Wlotzka (1969) calculated from the Ni-enriched rims in taenite crystals formed from metal-sulfide melts, that proto-Ramsdorf cooled down from 1200 to about 900 °C in less than 29 days.

The chromite-plagioclase assemblages described as

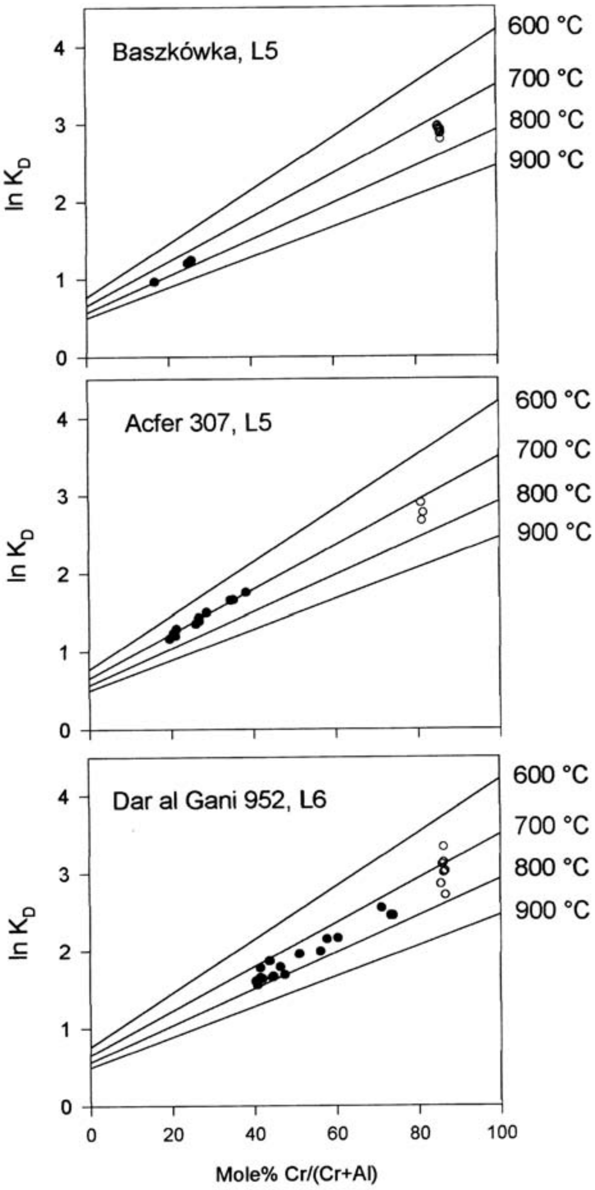

Fig. 25. Isotherm diagrams for Cr spinel (solid symbols) and chromite (open symbols) for L5 and L6 chondrites.

shock indicators by Rubin (2003) were also found in this work, and often their chromites are enriched in Al and Mg compared to normal chromites. These Cr spinels can also be used for thermometry, and usually the temperature derived from them is the same as from other Cr-spinel grains in the same meteorite. An example is the H4 chondrite Beddgelert, which contains Cr-spinel/feldspar chondrules (Fig. 12). The temperature derived in this case is within the normal H4 range (Table 3). If these Cr-spinel chondrules formed by local shock melting as inferred by Rubin (2003), it must have taken place early in the history of the parent body so that Beddgelert shared the normal thermal history of other H4 chondrites.

Chromite Temperature

The position of the chromites on the isotherm diagrams (Figs. 18–25) is different for unequilibrated and equilibrated chondrites, respectively. In most type 3 chondrites, they continue the spinel trend, and the temperatures calculated from them is the same as derived from Cr spinel (Tables 3 and 4). In some, the chromites show the type 4–6 trend discussed below (Grady [1937], Acfer 111, ALH 85045). Noteworthy is also the fact that Cr/(Cr + Al) ratios in chromite in type 3 chondrites can reach high values of up to 98 mole% (see Dhajala and Moorabie). In the case of type 4, 5, and 6 chondrites, the chromites do not follow an isotherm, but are arranged at a constant Cr/(Cr + Al) ratio over a temperature range, and the average is usually 20 to 70 °C lower than the corresponding Cr-spinel temperature (Table 3 and 4). Exceptions occur, e.g., for H4 Forest Vale, where the chromite temperature is the same as the Cr-spinel temperature, indicating faster than usual cooling. Thus, the chromite temperature is a closure temperature and does not give the real equilibration temperature of a chondrite. An indication for this is the correlation of the measured temperature with the grain size of the chromite (Wlotzka 1987), as discussed below.

The olivine/chromite temperature was also measured in a number of equilibrated chondrites, in which no Cr spinel was found (Table 6). These temperatures are in the same range as for the Cr-spinel-bearing chondrites, and the H-group average of 720 ± 19 °C is within the error limits the same as for these. The L group average is slightly lower. This result shows that the Cr-spinel-bearing chondrites do not differ in their thermal history from those without Cr spinel. Table 6 gives also some olivine/chromite temperatures for LL chondrites. Their average is the same as that for L chondrites. Of special interest is that the value for LL7 Uden is also close to the group average of 688 °C, because McSween and Patchen (1989) found a significantly higher cpx temperature (1150 °C) for Uden than for a number of LL6 chondrites (about 900–1000 °C). If Uden experienced such a high equilibration temperature, its later cooling history was apparently similar to other LL chondrites.

DISCUSSION

Mg, Al spinel formed either during early condensation in the solar nebula or crystallized from chondrule melts. Both may have played a role as formation processes. The ideomorphic spinel grains in the chondrule mesostasis formed most probably from a melt. Cr and Fe entered the spinel later by diffusion from silicates and chondrite matrix. Krot et al. (1993) have discussed the possibility to form Cr-rich objects by condensation under special conditions already in the nebula. The Cr and Fe enrichment in the spinel grains towards the chondrule surface (Figs. 8 and 9) argues for a parent body

Table 4. Average Cr-spinel and chromite temperatures T for L chondrites and two LL chondrites.

Name	Type	Cr spinel T (°C)	Std. dev.	Chromite T (°C)	Std. dev.
Suwahib (Buwah)	L3.7	637	44	634	23
Moorabie	L3.8	640	68	637	18
Julesburg	L3.7	683	25	655	9
ALH 85045	L3.8[a]	759[a]	39	682[a]	37
Mean (3)	L3.7–3.8	653	21	642	9
EET 83213	LL3.7	676	46	n.d.	
MAC 87306[a]	L4[a]	664[a]	21	646[a]	19
Acfer 047	L4	711	18	702	11
Saratov	L4	725	39	705	25
Bjurböle	L4	751	42	690	12
TIL 82406	L4	780	32	648	19
Haxtun	L4	793	51	777	10
Mean (5)	L4	752	31	704	42
Reliegos	L5	700[b]	–	n.d.	
Acfer 307	L5	711	19	714	16
Baszkówka	L5	778[c]	14	734	9
RC 072	L5	827[d]		n.d.	
Mean (4)	L5	754	52	724	10
Dar al Gani 952	L6	754	32	727	34
Melnikovo	LL6	777[e]	–	684[4]	–

[a] Outlier, not included in the mean, see text. n.d.: not determined.
[b] Using spinel analyses from Lozano and Martin-Crespo (2004).
[c] Using spinel analyses from Maruyama (2001) and Borucki and Stępniewski (2001).
[d] Using spinel analyses from McCoy et al. (1991).
[e] Using spinel and chromite analyses by Krot et al. (1992).

Table 5. Average and range of olivine/Cr-spinel and olivine/chromite temperatures for H and L group chondrites (Data from Tables 3 and 4).

Type	Number	Cr spinel T (°C)	Range	Chromite T (°C)	Range
H3.7–3.8	5	660 ± 20	625–673	654 ± 29	619–699
H3.9	2	728 ± 16	712–745	700 ± 11	689–711
H4	9	766 ± 31	722–825	707 ± 46	670–815
H5	7	774 ± 31	737–815	703 ± 27	668–740
H6	3	775 ± 34	750–824	723 ± 10	713–733
Mean[a] H4–6	19	770 ± 32	722–825	707 ± 36	668–815
L 3.7–3.8	3	653 ± 21	637–683	642 ± 9	634–655
L4	5	752 ± 31	711–793	704 ± 42	648–777
L5	4	754 ± 52	700–827	724 ± 10	714–734
L6	1	754	–	727	–
Mean[a] L4–6	10	753 ± 39	700–827	712 ± 35	646–727

[a] The means are not the average of the 3 petrologic types, but of all individual values.

process. But whatever the initial formation process, the Cr/(Cr + Al) to Fe/(Fe + Mg) correlation was completed on the parent body, as shown by the isotherm correlations of the distribution coefficient K_D (the distribution of Fe and Mg between olivine and spinel) with the Cr/(Cr + Al) ratio of spinel. This process starts in the partly equilibrated type 3.7 to 3.8 chondrites (Figs. 18 and 19) and continues in type 4 to 6. These "equilibrated" chondrites show constant Fe/(Fe + Mg) ratios in olivine, also inside the large mesostasis-rich chondrules which contain Cr-spinel grains. The Fe-Mg interdiffusion coefficients in Cr spinel are equal to (for Cr-rich spinel) or ten times higher (for Al-rich ones) than that of

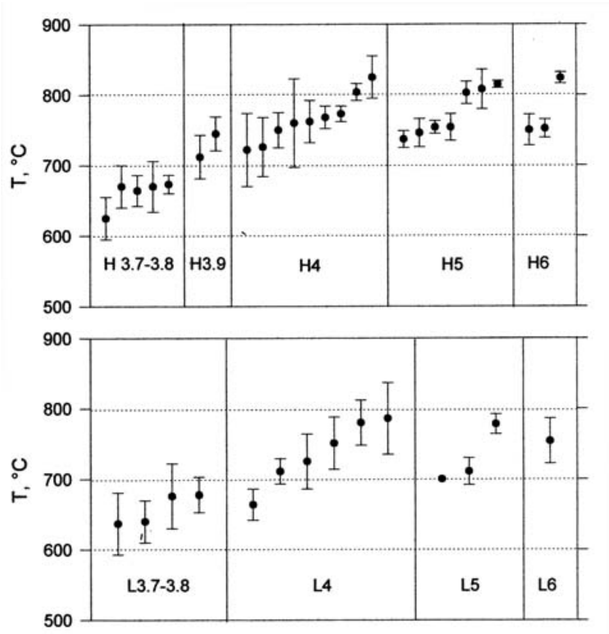

Fig. 26. Olivine/Cr-spinel temperatures for H and L chondrites of different petrologic types. Data from Tables 3 and 4 (including LL3.7 EET 83213 in the L3s).

olivine (Ozawa 1984). It follows that Fe/Mg in Cr spinel should also be in equilibrium with olivine. This is confirmed by the straight isotherms shown by all equilibrated chondrites (Figs. 21 to 25) and the constant temperatures found for different spinel occurrences in a given chondrite. The temperature range is that of metamorphism, no trace of inherited igneous temperatures can be found, which would also cause a larger scatter of the data. Later reheating by shock events can lead to the formation of new high-temperature Fe/Mg equilibria. These can be preserved, if the cooling was fast enough, as shown for some examples above.

The equilibration of Cr/Al in the spinel proceeds more slowly because of the lower diffusion coefficients of the trivalent ions (Ozawa 1983). It leads ultimately to constant chromite compositions, only large spinels which are isolated by Al-rich mesostasis from the chondrite matrix survive as heterogeneous grains. Anhedral, interstitial chromites formed also by solid state crystallization in the matrix of the equilibrated types 4 to 6, they are absent in type 3.6 or lower. This is also shown by the increase in grain size of chromite grains from type 4 to type 6 (Bridges et al. 2004).

The first main result of this work is that, for the first time, a temperature which can be interpreted as an equilibration temperature could be determined for the "unequilibrated" type 3.7 and 3.8 chondrites. Before, only estimates for this temperature could be given (e.g., Miyamoto et al. 1981). This temperature of 650 °C is at the same time an upper limit for the lower type UOCs.

The second result are the 100-degree temperature difference between type 3.6 to 3.8 and the equilibrated types 4 to 6 (Table 5 and Fig. 26). This difference is well outside the range of standard deviations in a given group. The type H3.9 are intermediate between H3.7 to 3.8 and type H4, although their temperature ranges overlap (Table 3).

Table 6. Olivine/chromite temperatures of chondrites in which no Cr spinel was found, and of shock heated chondrites.

Name	Type	*T* (°C)
Elsinora	H4	694 ± 27
Tiffa 005	H5	711 ± 9
Guarena	H5	711 ± 13
Laborel	H5	703 ± 16
Holbrook	H5	724 ± 6
Dar al Gani 325	H5	722 ± 9
Dar al Gani 746	H5	717 ± 19
Dar al Gani 310	H5/6	766 ± 10
Dar al Gani 747	H6	710 ± 14
Daraj 117	H6	704 ± n.d.
Mean (10)	H group	716 ± 19
Dar al Gani 323	L4	685 ± 18
NWA 767	L4	646 ± 13
Ehole	L5	712 ± 11
NWA 848	L6	710 ± n.d.
HaH 219	L4	864[a]
Mean (4)	L group	688 ± 27
Bhola	LL3–6	692 ± 17
Krähenberg	LL5	697 ± 37
Vishnupur	LL6	686 ± 21
Uden	LL7	676 ± 34
Mean (4)	LL group	688 ± 8
Shock-heated chondrites		
Etter, S6	L6	841 ± 19
Rio Limay, S6	L6	888 ± 25
Sta. Catarina, S6	L6	894 ± 31
Ramsdorf, S6	L5	967 ± 42
Muckera 007	L shock melt	882 ± 29

[a] Outlier, not included in the mean.

The third result is that type 4, 5, and 6 chondrites show a range of equilibration temperatures, which is the same for all three types. Other recent temperature determinations confirm the fact that types 4 to 6 equilibrated at about the same temperature. Reid and Dussaud (2000) used the opx/cpx thermometer and found a range of temperatures from 800 to 1000 °C, which is the same for type LL4, 5, and 6. Kessel et al. (2002) applied the olivine/Cr-spinel thermometer of Sack and Ghiorso (1991a) to chromites from ordinary chondrites and found also similar temperatures (650–770 °C) for types H4 to H6.

Closure versus Equilibration Temperature

The question is whether the measured element distributions were frozen in at the closure temperatures for

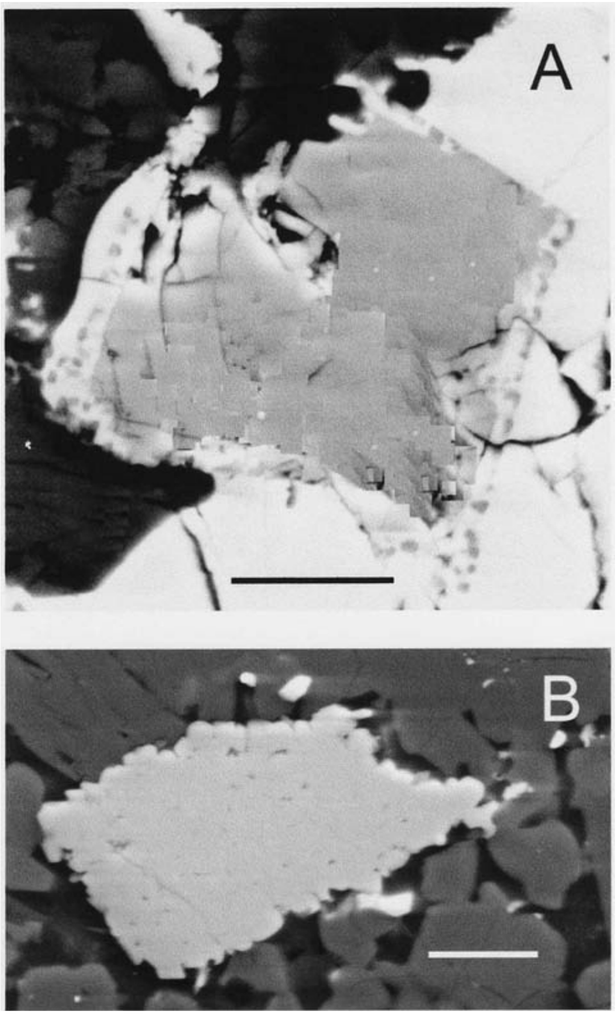

Fig. 27. Chromite in shock-melted chondrites. SE pictures, scale bar 10 µm. a) L5 chondrite Ramsdorf. b) L chondrite Muckera 007.

diffusion during continuous cooling or represent equilibration temperatures at an isothermal stage, at which the chondrites stayed for a longer time. Two lines of evidence argue for a real equilibration stage:

1. The straight isotherms over a wide range of Cr/(Cr + Al) ratios, because the interdiffusion coefficient for Fe and Mg in spinel varies with this ratio. It is about 10 times higher at low Cr/(Cr + Al) ratios than at high ones (Ozawa 1984), so that these spinel grains should close at lower temperatures than grains with a higher ratio. The isotherm plots and Fig. 15 show that there is no significant variation of temperature over a large range of Cr/(Cr + Al) values. This effect turns out to be an unique advantage of the olivine/Cr-spinel thermometry with grains of variable Cr/(Cr + Al), because it allows to distinguish between an isothermal stage and a closure temperature. And the graphical presentation makes the difference visible at a glance. (Note that the deviation of Cr-spinel data from Forest Vale from the Fabriès isotherms shown in Fig. 14 goes into the opposite direction and cannot be the effect of different closure temperatures).

2. The measured equilibration temperature is independent of the grain size of Cr spinel, an example is shown in Fig. 29. During cooling small Cr-spinel grains or grain rims will register lower temperatures than large grains and grain centres. Ozawa (1984) used such zoning patterns for the calculation of cooling rates of peridotites and he found that isothermal stages produce “plateau temperatures” independent of spinel grain size. To be sure, Cr/(Cr + Al) and Fe/(Fe + Mg) ratios show strong zoning in the chondritic Cr-spinel grains, as seen in all the isotherm plots, but the temperature calculated from these ratios are constant in a given grain and a given chondrite. In contrast to this, chromite grains (Fig. 29) do show such a correlation of grain size and temperature (Wlotzka 1987). Such a correlation was also observed in chromite grains of H chondrites by Kessel et al. (2002).

The lower temperature registered for type 3 chondrites fits also into this picture. These chondrites are only partly equilibrated with respect to the Fe, Mg distribution between the Fe, Mg silicates. Thus, they either did not reach as high a temperature as the equilibrated types 4 to 6, or they cooled faster than these from similar temperatures. In the latter case they should show a higher closure temperature than type 4 to 6, which is not the case. The lower equilibration temperature agrees with their lower degree of equilibration.

There is no contradiction between the fact that these type 3 chondrites are “unequilibrated” and the formulation of an “equilibration” temperature. The subtypes 3.7 to 3.9 are already partially equilibrated by an Fe, Mg exchange between the Fe, Mg silicates, and this applies also to the Cr spinel. They are equilibrated in Fe/(Fe + Mg), even if Cr/(Cr + Al) is still variable. The equilibration is not complete, which is also demonstrated by the larger scatter in the ln K_D versus Cr/(Cr + Al) diagrams (Figs. 18 and 19). Nevertheless, the temperature range at which this partial equilibration took place can be derived from the isotherm diagrams.

The same argument applies also to the apparent contradiction between heterogeneous Cr spinel and equilibration of Fe, Mg in these spinels. Ozawa (1983) observed the same phenomenon in peridotites from the Miyamori ultramafic complex and explained it by the fact, that the Cr-Al interdiffusion in spinel is considerably slower than Fe-Mg interdiffusion. In addition, the Cr-spinel grains inside chondrules are surrounded by an Al-rich feldspatic mesostasis and thus isolated from the matrix. This is particularly evident in the type 5 and 6 chondrites, where they occur inside large chondrules with a long diffusion path for Cr^{3+} (see Fig. 7).

Comparison with Other Olivine/Cr-Spinel Thermometers

It was shown above that a modified version of the Fabriès thermometer gives the best fit to the Cr-spinel data of different chondrites and this thermometer was therefore preferred over other graphic solutions for the calculation of

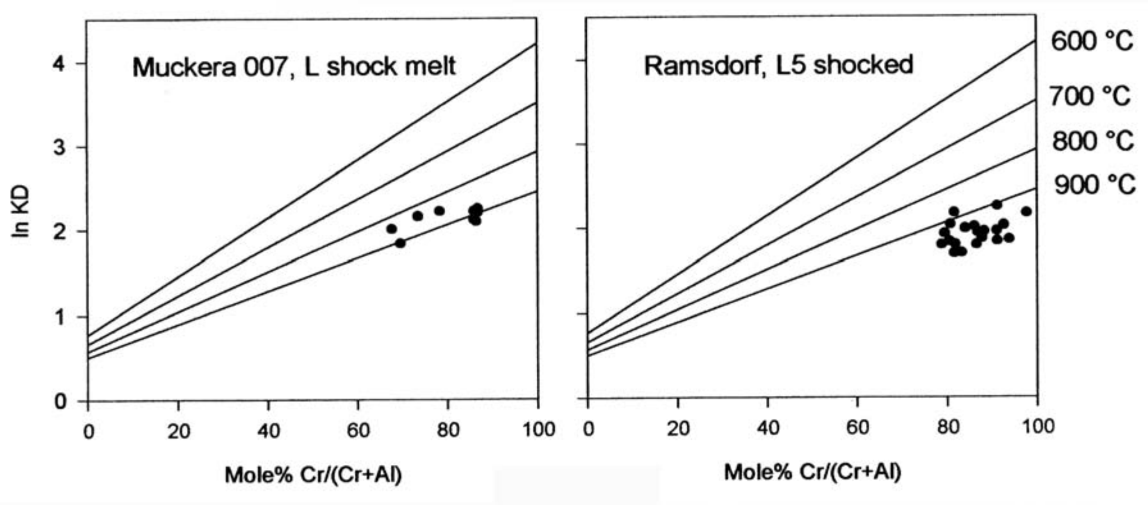

Fig. 28. Isotherm diagrams for chromites in shock-melted L chondrites Ramsdorf and Muckera 007.

equilibration temperatures. Ozawa (1983) applied several olivine/Cr-spinel geothermometers to analytical data from two peridotite complexes. He came also to the conclusion that Fabriès calibration gave the most consistent results. On the other hand, Johnson and Prinz (1991) selected arbitrarily the calibration by Engi (1983) for the evaluation of equilibration temperatures of type II chondrules in carbonaceous and ordinary chondrites, and got rather low temperatures of about 600 °C from chromites of L4 and L5 chondrites.

More recently, other olivine/Cr-spinel thermometers have been published. The one by Sack and Ghiorso (1991a) uses mainly new thermodynamic data. Table 7 shows a comparison between temperatures calculated using their thermometer and the modified Fabriès thermometer. It can be seen that there is good agreement between the two. Sack and Ghiorso (1991b) have also applied their thermometer to chromite analyses of chondrites taken from the literature (without differentiating the petrologic types). They get an average of 720 ± 40 °C for Type 4 to 6 chondrites (Table 8), in good agreement with the chromite temperature of 714 and 718 °C for H and L chondrites, respectively, from this work (Table 5). As shown above, however, the use of chromites gives lower values than those derived from Cr spinel, and not the real equilibration temperature. The Sack and Ghiorso (1991a) thermometer was also applied by Kessel et al (2002, 2004) to chromite analyses of H and L chondrites. They got also similar equilibration temperatures for type H4, 5 and 6, 680 to 770 °C, average 724 °C, in agreement with the results of this work. For L chondrites they found a lower maximum temperature for type 4, 664–672 °C, than for type 5 and 6, 767–773 °C, but they measured only one chondrite of each type. Calafiore (2002) applied also several olivine/chromite thermometers to LL chondrites and found averages of 680, 650, and 705 °C for LL4, LL5, and LL6 chondrites, respectively (Table 8). These temperatures are the same as found here for chromites of LL chondrites (Table 6).

Table 7. Comparison of olivine/Cr-spinel temperatures (°C) calculated with different thermometers.

Meteorite	Type	Modified Fabriès	Sack and Ghiorso	O’Neill and Wall
Flandreau	H5	737°	765°	619°
Acfer 274	H6	750°	748°	659°
Acfer 047	L4	711°	735°	626°
Acfer 307	L5	711°	732°	631°

O’Neill and Wall (1987) published another olivine/spinel thermometer using only thermodynamic data and no empirical calibration, but little experience exists with its application. Temperatures calculated by their method lie about 100 °C lower than the temperatures of this work and of Sack and Ghiorso (Table 7).

Comparison with Other Geothermometers and the Absolute Temperature

Several authors have used the distribution of Ca and Mg between ortho- and clinopyroxene (the opx/cpx thermometer) for the determination of chondrite equilibration temperatures. Table 8 gives an overview. This thermometer yields temperature from both the Ca content in opx and in cpx. Olsen and Bunch (1984) applied the thermometers by Lindsley (1983) and by Kretz (1982). They found cpx temperatures of 830 ± 20, 870 ± 35, and 860 ± 30 °C for H6, L6, and LL6 chondrites, respectively, from the Lindsley thermometer, whereas Kretz’ calibration yielded unsatisfactory high temperatures. McSween and Patchen (1989) using also the Lindsley thermometer, determined somewhat higher temperatures, 900 to 960 °C for LL6 chondrites. They found that cpx in type 4 and 5 is not fully equilibrated, so that a cpx temperature cannot be given for them. Reid and Dussaud (2000) attempted to determine two-pyroxene temperatures for type 6 as well as for type 4 and 5 chondrites. They found

Table 8. Temperatures in °C measured with different geothermometers for equilibrated chondrites of different petrologic types. Figures in parentheses give the number of meteorites analyzed.

Thermometer	H4	H5	H6	L4	L5	L6	LL4	LL5	LL6	Reference
Cpx			(5) 830 ± 20			(6) 870 ± 35			(3) 860 ± 30	Olsen and Bunch (1984)
Cpx Opx								Cluster at (4) 675–750	(3) 900–960 (3) 800–900	McSween and Patchen (1989)
Opx							(5) 830–980	(7) 800–900	(14) 850–1000	Reid and Dussaud (2000)
Cpx Opx			(2) 970–990 (2) 880–910			(1) 1000 (1) 915			(1) 1000 (1) 920	Slater and McSween (2002)
Plagioclase			(3) 725–740			(2) 810–820			(1) 800	Nakamuta and Motumura (1999)
Olivine/ chromite	(11) 660–800			(6) 660–750			(4) 650–730			Sack and Ghiorso (1991b)
	average 720[a] ± 40									
Olivine/ chromite	(4) 650–770 average 724			(1) 630–670	(1) 640–770	(1) 650–730	(1) 650–670	(2) 630–720	(1) 670–770	Kessel et al. (2002)
Olivine/ chromite							(2) 630–740 avg. 680	(4) 570–695 avg. 650	(6) 640–800 avg. 705	Calafiore (2002)
Olivine/ chromite	(25) 670–815 average 710 ± 35			(12) 650–730 average 710 ± 35			(4) 680–700 average 690 ± 10			This work
Olivine/ Cr-spinel	(19) 720–825 average 770 ± 30			(10) 700–830 average 750 ± 40					(1) 780	This work

[a] The values from Sack and Ghiorso were calculated from analyses taken from the literature, without differentiating between petrologic types.

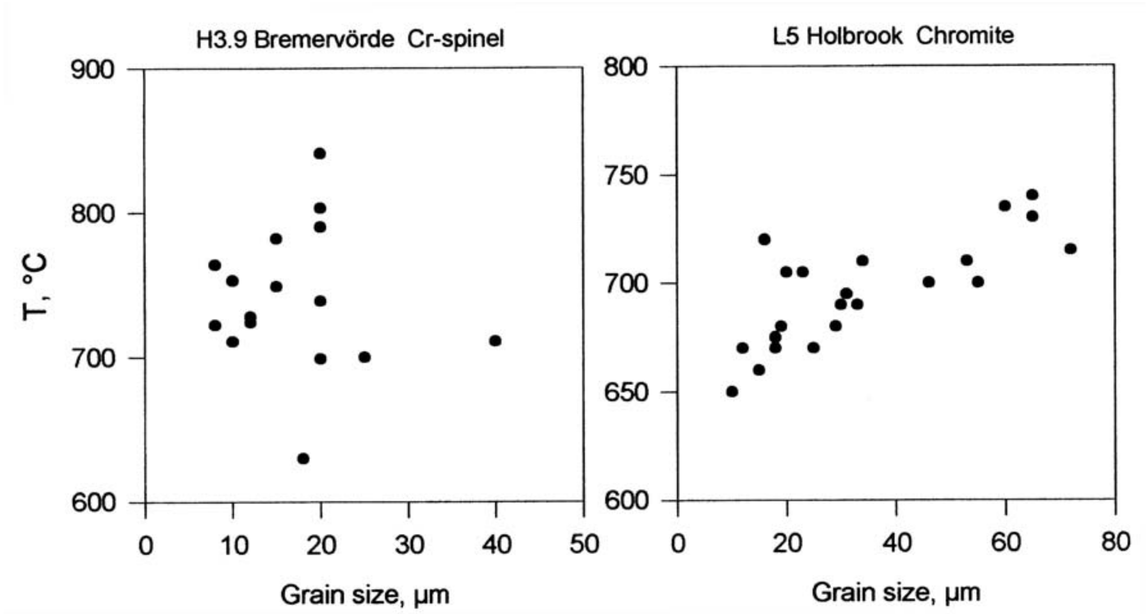

Fig. 29. Olivine/spinel temperature versus grain size for Cr spinel from H3.9 Bremervörde and chromite from L5 Holbrook.

variable Ca contents in both opx and cpx even in type 6 chondrites, and tried to use the most frequent Ca values (the "plateau value") in a given meteorite for thermometry. They derived opx temperature ranges of 830 to 980 °C for LL4, 800 to 900 °C for LL5, and 850 to 1000 °C for LL6 chondrites, where the range given is not the uncertainty (it is ±50 degrees), but the difference in temperature found for different meteorites. These temperatures as well as the other two-pyroxene thermometer results (Olsen and Bunch 1984, McSween and Patchen 1989) are 100 to 200° higher than the olivine/Cr-spinel temperatures found in this work, although both temperature ranges overlap. One possible explanation is the difference in closure temperature, pyroxenes are known to close at higher ones than olivine/spinel (Sack and Ghiorso 1991b). It is also possible that the difference between the two temperature values are, at least partly, due to differences in the absolute temperature calibration. Both thermometers have probably an uncertainty of about 50 °C. It is also possible that the pyroxene temperatures are too high, because cpx did not reach equilibrium even in type 6 chondrites. One indication is the variable Ca content mentioned above (Reid and Dussaud 2000). Another aspect are the about 100° lower temperatures found by McSween and Patchen (1989) from opx alone. McSween and Patchen (1989) favour the higher values of the cpx thermometer, because of its higher precision and the suspicion, that opx may have inherited Ca from earlier protoenstatite precursors. However, Jones (1998) showed that equilibration of minor elements in opx of LL chondrites is faster than in cpx, so that opx is homogenous in type 4 to 6, whereas cpx is not completely equilibrated even in type 6. This means that the lower temperatures recorded by opx may be closer to the equilibration temperature. McSween and Patchen (1989) found that "a pronounced cluster of type 5 chondrite orthopyroxenes at 675–750 °C probably represents an equilibration point". This value is close to the 750 to 770 °C found in this work as average values for H4–H6 and L4–L6 chondrites.

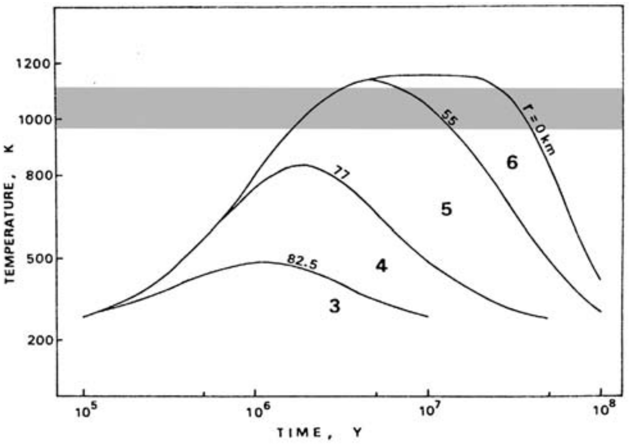

Fig. 30. Temperature distribution at different depths (noted as radius on the temperature curves) in the H chondrite parent body model (r = 85 km) by Miyamoto et al. (1981). Numbers 3 to 6 between the curves refer to petrologic types. The shaded area gives the olivine/Cr-spinel temperature range for H4, 5, and 6 chondrites from this work.

Nakamuta and Motomura (1999) developed another approach for type 6 chondrites. It uses the Al, Si ordering in feldspar for the determination of its crystallization temperature from chondrule glass. Their maximum temperature of 734 ± 9 °C for three H6 chondrites is similar to the equilibration temperatures of this work, the maximum of 814 ± 6 °C for two L6 chondrites is somewhat higher (Table 8). Chamot and McSween (2000) and Slater and McSween (2002) have questioned the validity of the plagioclase

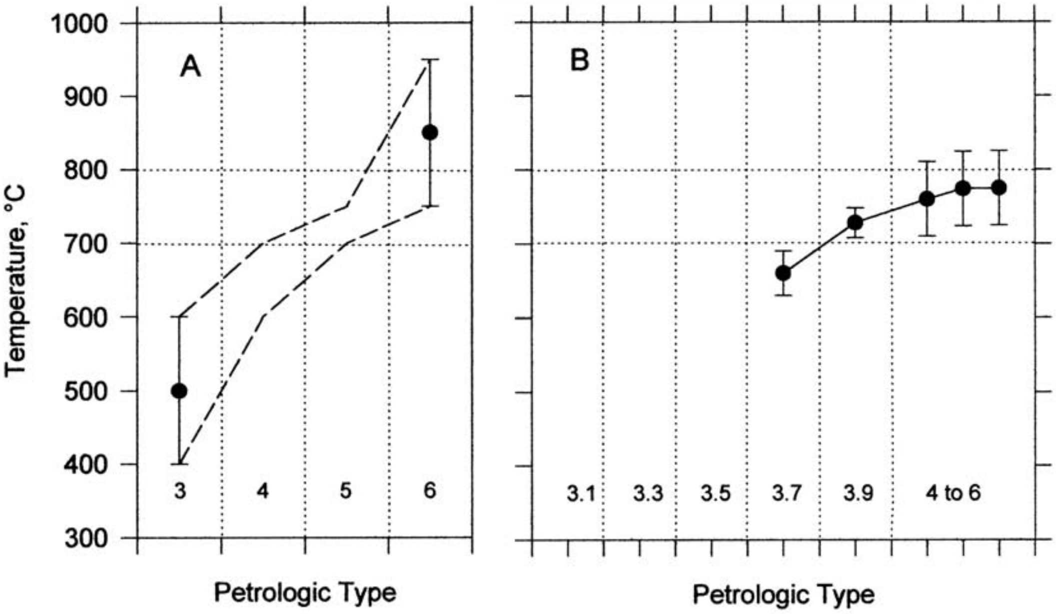

Fig. 31. a) Estimates of equilibration temperatures for type 6 and type 3 chondrites by Dodd (1981) and interpolation proposed (not measured) for intermediate types. b) Olivine/Cr-spinel temperatures from this work, on a scale more appropriate to the changes observed. The error bars give the observed range for each type.

thermometer. Slater and McSween (2002) got about 100 degrees higher opx/cpx temperatures than Nakamuta and Motomura (1999) found by the plagioclase thermometer for the same meteorites.

The question whether the Cr-spinel/olivine temperatures calculated with the modified Fabriès thermometer are close to the absolute temperature was already mentioned above. Unfortunately, the calibration at these low, metamorphic temperatures is not accessible by equilibration experiments. A possible way is the comparison with the results of other thermometers. The Sack-Ghiorso calibration is completely independent from the empirical calibration of Fabriès. The agreement in the temperatures obtained by the two (Table 7) suggests that these temperatures may be close to the absolute one. It is also noteworthy that three independent reactions yield the same equilibration temperature of about 750 °C: the Ca content of opx (McSween and Patchen 1989), the Si-Al ordering in plagioclase (Nakamuta and Motomura 1999), and the Fe-Mg distribution between olivine and Cr spinel (this work). This suggests that this temperature is close to the real equilibration temperature of ordinary chondrites and close to the absolute temperature. Still, most of the conclusions reached below refer only to relative temperature differences.

The equilibration temperature is not necessarily the maximum temperature seen by these chondrites. But for the unequilibrated chondrites of type 3.7 to 3.8 it cannot have been much higher for a longer time, because at about 100° more they would have been equilibrated to type 4. In the case of type 4 to type 6 chondrites, it was argued above that the olivine/Cr-spinel temperatures represent an isothermal stage. This occurred probably close to the maximum temperature. Parent body models for chondrites show a broad maximum in the temperature curves, that is, the chondrites had an isothermal stage around this maximum temperature. As an example the H chondrite parent body model by Miyamoto et al. (1981) is shown in Fig. 30. They used an upper limit of 880 °C for type 6, and 230 °C as well as an abundance of 8.5% for type 3 chondrites, internal heating by ^{26}Al, and thermal diffusivity as measured for non-weathered chondrites. The broad maximum in the central part of the body is evident. Other features of this model are not compatible with the olivine/Cr-spinel temperature distribution. This "onion shell" parent body model has shells of decreasing temperature corresponding to the petrologic types 6 to 3. Figure 30 shows the temperature curves from their model compared to the range of olivine/Cr-spinel temperatures found here. If we interpret these temperatures as equilibration temperatures, then they do not fit into this model. Also a combination of different maximum temperatures and cooling rates could not produce equal closing temperatures for all petrologic types (or shells in this model), as each shell has a different cooling rate in this temperature region.

It was pointed out by Rubin (2003, 2004) that practically all type 5 and 6 chondrites were shocked and later annealed, even if their final visible shock stage may be as low as S1. In this way, the temperature range observed here for all petrologic types could be the result of random shock heating events. Of special interest are therefore the results for H4 Forest Vale and H5 Richardton (Table 3). These meteorites show undisturbed Pb/Pb (Göpel at al. 1994), Ar/Ar (Trieloff et al. 2003), and Cr/Mn systems (Polnau and Lugmair 2001) and correspondingly high absolute ages of more than 4550 Ma. We can therefore assume, that their olivine/Cr-spinel temperatures were also not disturbed by later heating events and take these as "pristine" equilibration temperatures. The temperature of 804 °C for Forest Vale is the second highest measured in the H4 group (Table 3). Moreover, Forest Vale shows an atypical behavior, namely similar chromite (815 °C) and Cr-spinel temperatures (804 °C). This chromite temperature lies about 100 °C higher than the average H4 chromite temperature and indicates faster than normal cooling, so that its temperature history differs from that of typical H4 chondrites. In view of this result, it seems unfortunate that Trieloff et al. (2003), who inferred also a high cooling rate for Forest Vale, took this as representative for the H4 group. H5 Richardson, on the other hand, shows with 750 °C a temperature close to the average of the H5 group. We can conclude that the high temperature shown by Forest Vale is not due to late shock heating events and probably also not the range of 100 degrees observed in the H4 group.

CONCLUSION

I have shown above that type 4 to 6 chondrites have a range of olivine/Cr-spinel temperatures, and, contrary to expectation, the same average temperatures (Table 5). If we take these temperatures as equilibration temperatures, these features are not compatible with a conventional "onion shell" parent body (Fig. 30). They can best be understood by a two-stage model of asteroid evolution, which was developed because no relation between petrologic type and metallographic cooling rates of chondrites could be found (Scott and Rajan 1981; Grimm 1985; Taylor et al. 1987): the chondrites equilibrated at different discrete temperatures in small planetesimals or in one larger parent body where a temperature gradient produced increasing degrees of equilibration and textural integration (the type 4 to 6 chondrites); a later breakup and new accumulation, while still hot, produced a "rubble pile" structure where rocks of different types were mixed and stored together in a new temperature regime. Here they developed the olivine/Cr-spinel equilibrium according to the temperature at the new position in the rubble pile, similar for originally different petrologic types and different for similar types at different new positions. The result will be a range in temperatures for all petrologic types and, due to the random distribution, the same average for all types. The maximum temperature reached in the new rubble pile body will be about 825 °C, the maximum temperature registered for type 4, 5, and 6 (see Table 5). The textural integration reached by type 4, 5, and 6 will not be reversed at lower temperatures, but type 4 and 5 could continue to recrystallize into higher types, if they came into higher temperature regions than before. This was apparently not the case, as we find type 4 chondrites up to 825 °C. This would imply that type 5 and 6 formed originally at higher temperatures than this. On the other hand, the 650 to 660 °C found for type 3.7 to 3.8 must be their original equilibration temperature, as we do not find higher temperatures for these.

There remains, however, one problem with this model: the absence of low equilibration temperatures, as low as for the type 3.6 to 3.8, for type 4 to 6 chondrites. In the rubble pile model, all chondrite types were randomly distributed in their new parent body after they had equilibrated to different degrees and acquired their characteristic textures in a different body. Unequilibrated type 3 chondrites, which got inside the body with higher temperatures, may have equilibrated to higher types and only survived as such in the outer, cooler parts. This way, we would get the temperature distribution that we find now. But why do we not have type 4 to 6 chondrites with the same low temperature of 650 °C as the type 3.7 to 3.8, at which an equilibration of olivine and Cr spinel could obviously still take place? They should have got into the outer, cooler parts of the rubble pile as easily as the type 3s. This remains an open question and makes the rubble pile model problematic. We observe here some kind of decoupling of the thermal history of type 3 and the type 4 to 6 chondrites: whereas features of type 4 to 6 can best be understood by the rubble pile model (a 100° temperature range and similar average equilibration temperatures), type 3.7 to 3.8 need an undisturbed location in a surface shell of lower temperature where no types 4 to 6 were present (no low-temperature type 4 to 6).

We see that the conventional onion shell parent body as well as the rubble pile model run into problems explaining all features of the olivine/Cr-spinel temperature distributions. Another possible scenario is offered by the model of Yomogida and Matsui (1984), which takes into account the strongly differing thermal diffusivities of initially loose and later sintered matter. This results in a nearly isothermal interior of bodies with a 20–50 km radius, the final temperature depending mainly on the size of the body, with a thin outer layer of lower temperature. They conclude that the different petrologic types must have developed in separate bodies of different size. However, their result would fit to the equal temperatures for type 4 to 6 found in this work, and the lower temperature of type 3.7 to 3.8. The increasing textural integration seen in type 4 to 6 may then be due to small temperature differences (about 30°, within the average standard deviation of the measured temperatures), or to increasing times of recrystallization (the innermost regions becoming hot earlier than the outer). The range of temperatures observed for each petrologic type may be caused by other variable parameters as primary texture and grain size. Different contents of volatiles may also have played a role (Kessel et al. 2002), higher contents facilitating diffusion and recrystallization.

If we follow this model, we can conclude from the similar temperatures measured for type 4, 5 and 6 chondrites that the main step in chondrite metamorphic evolution occurs within type 3, namely the Fe-Mg equilibration between the mafic silicates. This fact has already been realized by Sears et al. (1980) from the differences in induced thermoluminescence within the type 3s, which lead them to introduce the subtypes 3.0 to 3.9 with increasing degree of equilibration. Grossman and Brearley (2005) have shown that type 3.0 and 3.1 can be further subdivided in 3.00 through 3.15. The equilibrated types 4 to 6 differ only in progressive textural integration, which apparently takes place in the same temperature range, a longer time may lead to higher petrologic types. It seems unfortunate that the van Schmus-Wood petrologic classification (van Schmus and Wood 1967) uses the integers 4, 5, and 6 for these different textures, thus overemphasizing their significance. A more appropriate classification scheme seems to be that of Binns (1967), which distinguishes only three types: primitive, intermediate, and crystalline, which corresponds to type 3, 4, and 5 + 6, respectively, in the van Schmus-Wood system.

In Fig. 31, the temperature scheme envisioned by Dodd (1981) for the different petrologic types is shown. Dodd (1981) started with estimates of 750 to 900 °C for type 6 and 400 to 600 °C for type 3 and proposed intermediate temperatures evenly distributed between these corner values for type 4 and 5 (Fig. 31a). No measurements for type 4 and 5 were available at that time. This scheme is still cited in many review articles, e.g., McSween et al. (1988), Keil (2000), and the assumed correlation of temperature with petrologic type widely accepted as real (e.g., Trieloff et al. 2003). From the results of this work, and following the work of Sears et al. (1980) and Grossman and Brearley (2005), a different picture must be drawn (Fig. 31b). It uses an extended scale for type 3.0 to 3.9 and a reduced scale for type 4 to 6 (which would better be called 4.1 to 4.3). In this figure, only the relative temperature relations between types 3, 4, 5, and 6 are significant; the absolute values may be uncertain.

The merits of the van Schmus-Wood scheme for the petrologic characterization of ordinary chondrites are indisputable. But its limitations shown here should be kept in mind and taken into account in future models of chondrite evolution. The olivine/Cr-spinel thermometry is a useful tool for deciphering the thermal history of individual chondrites. It should be applied especially to "unequilibrated" chondrites, because they usually contain Cr-spinel grains of variable composition and no other method is available for measuring their equilibration temperature. It may help to clarify the relation of type 3 to type 4, 5, and 6 and possible transitions between type 3.8 and 4. It can also be used to select undisturbed, "normal" meteorites for the determination of radiogenic ages and cooling rates, as shown above for the negative examples Beaver Creek and Forest Vale.

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