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NATURE AND EVOLUTION OF THE METEORITE PARENT BODIES: EVIDENCE FROM PETROLOGY AND METALLURGY

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Petrologic and metallurgical properties of the meteorites
that specify or limit their depth of equilibration in the
parent bodies are noted. Origin of the structure of palla-
sites is discussed in detail. The pallasitic structure
could have formed stably at the core/mantle interfaces of
internally melted small planets, where the weight of sunken
olivine cumulate layers submerged the lowermost olivine
crystals in underlying molten metal. However, the weight
of the cumulate layer would also deform the olivine crystals
so extensively as to destroy the pallasitic structure, ex-
cept in the smallest parent bodies (< ~10 km radius). It
appears that melting and differentiation (to produce palla-
sites, irons, achondrites) occurred in an early generation
of small planetesimals, but final cooling of the meteoritic
material occurred in larger bodies.

All the evidence from petrological and metallographic studies of meteorites indicates
that they evolved and to some degree equilibrated at relatively shallow depths in planets.
Most of the evidence from these disciplines does not testify to the total size of the
planets, though there is one item of evidence that I believe does constrain the dimension
of certain of the parent planets to be very small; most of the present paper will be de-
voted to a discussion of this point.

So far as evidence constraining depths of origin (not planetary sizes) is concerned,
at the simplest level one can cite the ophitic textures of eucritic achondrites, which are
reproduced in terrestrial circumstances only by volcanic rocks that crystallized in surface
flows or at very shallow depths in feeder conduits. Evidence from textures, microcraters
(Brownlee and Rajan, 1973), and fossil tracks (Wilkening, Lal, and Reid, 1971) in other
types of Ca-rich achondrite point to their evolution in regoliths, at the surface of one or
more bodies.

The meteorites contain no minerals inconsistent with equilibration at very low pressures
apart from the occurrence of diamond in the Cañon Diablo iron and in the ureilites (a type
of achondrite), the formation of which is clearly attributable to shock pressures upon im-
pact rather than sustained pressure at depth in a planet (Lipschutz and Anders, 1961;
Lipschutz, 1964). On the other hand, minerals that *would* be produced by high pressures are
not observed. Figure 1 shows a phase diagram relevant to chondrites and mesosiderites.
The mineral assemblages of both meteorite classes fall in the left (low-pressure) field;
i.e., plagioclase is stable rather than spinel or garnet. The diagram cannot be used to
make a quantitative estimate of pressures for these meteorites because it does not include
the effect of Na and Fe^{2+}, which are important components of the meteoritic systems. The
abundance of aluminum in the M1 sites of orthopyroxene in ordinary chondrites is vanish-
ingly small (0.1 ± 0.1%), which at the apparent temperature of equilibration (~850°C)
indicates very low pressures. Interestingly, the Al_{M1} content of orthopyroxene in a
mesosiderite (Patwar) is somewhat higher (0.4%; Weigand, 1975). Because of the approxi-
mations mentioned, however, these values cannot be interpreted quantitatively.

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Fig. 1. Stable mineral assemblages in the system $CaO-MgO-Al_2O_3-SiO_2$, as a function of pressure and temperature (Obata, 1976). Fo = forsterite, Opx = orthopyroxene, Cpx = clinopyroxene, An = anorthite, Sp = spinel, Ga = garnet. Numbers on dashed lines represent the atomic percent. Al in six-fold coordinated sites in ortho-pyroxene.
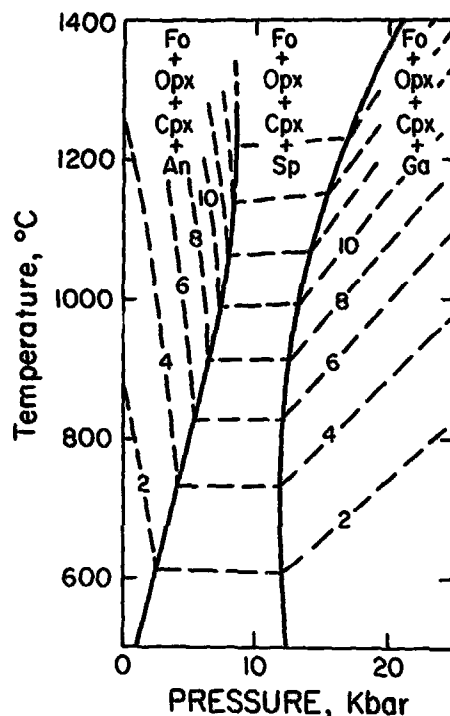
Perhaps the most valuable evidence for depth of equilibration of meteorites is pre-served in the metallic minerals they contain. The rates at which metal-bearing meteorites cooled through the temperature interval $600°-400°C$ in their parent planets can be estimated from the nature of Ni diffusion gradients preserved in their metal alloys (Wood, 1964; Goldstein and Ogilvie, 1965). The slower the cooling rate, the lower the temperature (and hence the higher the Ni content of the γ alloy) when diffusion was immobilized. Character-istic cooling rates for a number of meteorite classes are summarized in Figure 2 (Wood, 1967; Goldstein, 1969; Powell, 1969; Buseck and Goldstein, 1969). This figure also sug-gests possible cooling sites for the meteorite types, by displaying cooling rates as a function of depth in planets of asteroidal dimension. The cooling rate calculations assume a uniform initial temperature of $1000°C$ and chondritic long-lived radioactivity, but do not take account of redistribution of heat sources by igneous activity (Wood, 1967).

Many writers have concluded that the meteorites came from differentiated, concentric-ally layered parent bodies, structura,ly analogous to Earth. Iron meteorites would repre-sent the cores of these bodies. Others have advocated parent planets with "raisin bread" structure, meaning that relatively small zones of metallic Ni,Fe were dispersed at all depths in them. Urey (e.g., 1963) cites curious reentrant cavities on the surface of the Goose Lake iron meteorite (Henderson and Perry, 1958) and the relatively large abundance of pallasites among meteorites as evidence that the surface/volume ratio of iron masses in the parent planets was high, therefore the iron masses occurred in "raisins." (Pallasites are stony-iron meteorites that consist of roughly equal amounts of coarse olivine ($\sim$1.0 cm) and metallic Ni,Fe. The olivine crystals are in close-packed array, with metal filling the spaces between them. Clearly the solid olivine crystals accumulated stably in this configu-ration while the metal was molten (i.e., in the temperature range $1600°-1400°C$). Properties of pallasites are reviewed by Mason (1963) and Buseck (1977)). Wasson (1972) notes that the

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Meteorite classes

Depths and pressures in
small planets

R =
500 km   R =
200 km   R =
100 km   R =
50 km

Cooling rate at 500°C, °C/10⁶ yr

1000

IIc
irons

IVa3
irons

100

IVa2
irons

IIIa2
irons

IVb
irons

IVa1
irons

Ordinary
chondrites

10

IIIa1
irons

I
irons

IIIb
irons

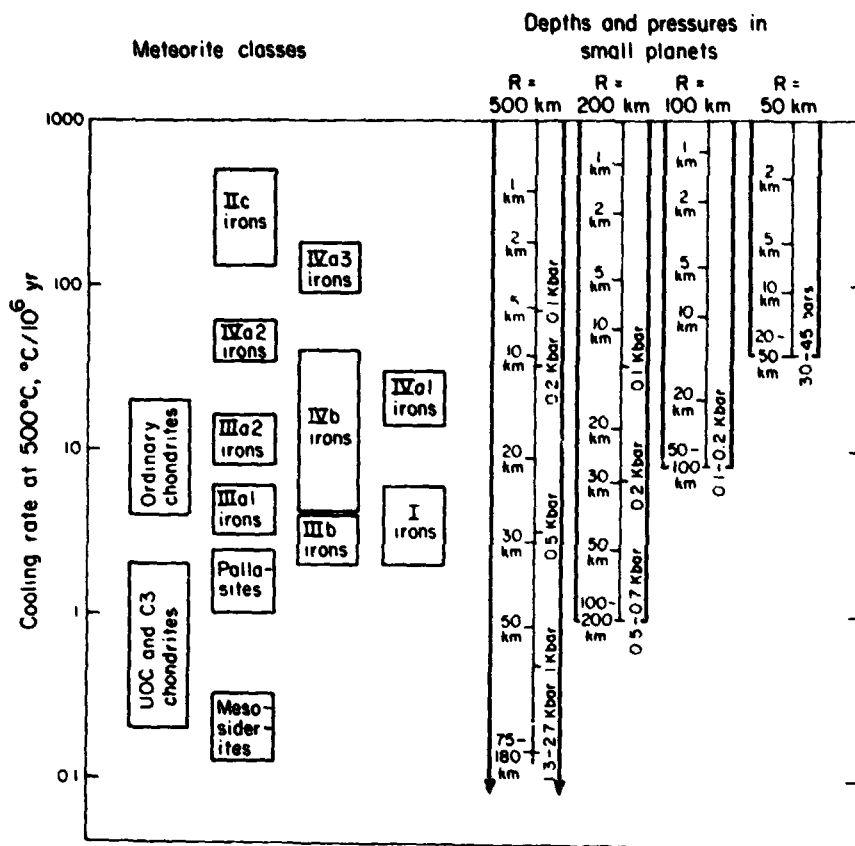UOC and C3
chondrites

1

Palla-
sites

Meso-
sider-
ites

0.1

Fig. 2. Left, cooling rates of metal-bearing meteorite classes. Right,
depths at which cooling would have occurred at these rates, in four
hypothetical planets of asteroidal dimension. Depths >180 km in the
500 km body would not cool to 500°C in 4.6×10^9 yr.

wide range of cooling rates of some geochemically coherent classes of iron meteorit s,
taken at face value, indicate that the latter evolved in the same parent planet but in dis-
crete bodies at widely varying depths, *i.e.*, in "raisins." (Cooling rates of 33 Group IIIa
irons range from 1.5° to 10°/10⁶ yr; 23 IVa irons, 7°-80°/10⁶ yr; Goldstein (1969).)

However, a detailed consideration of the chemistry and cooling rates of Group IIIa
irons (Figure 3) makes it appear likely that meteorites from two sources, each with cooling
rates uniform to within the uncertainty of the method, are lumped in this group. Group IVa
can be similarly decomposed into two or three uniformly-cooling components. Figure 2 re-
flects the subdivision of these two groups. Further, the concept of "raisins" does not
really make the Goose Lake cavities any easier to understand, nor does the model when exam-
ined in detail help to account for the apparently unstable mixture of high- and low-density
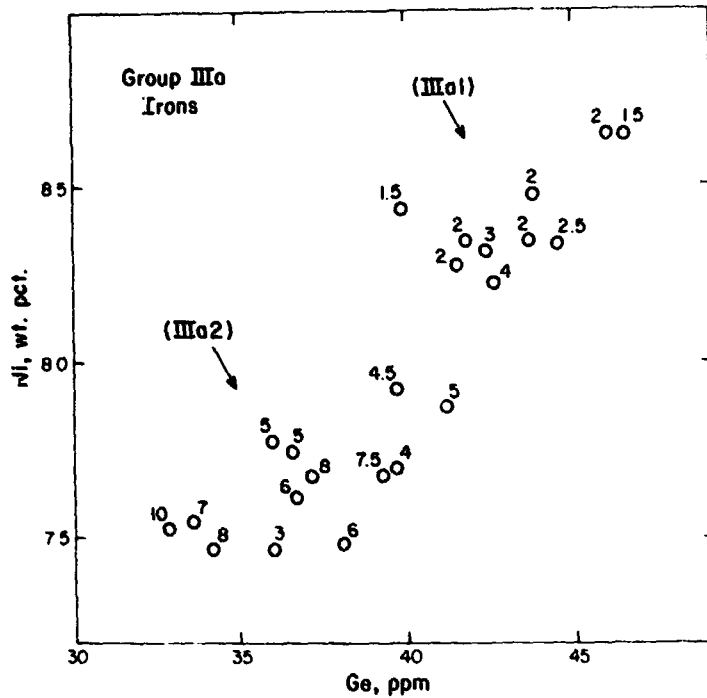components that constitutes pallasites. The "raisin bread" model and others involving

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Fig. 3. Ni and Ge contents and metallographic cooling rates of
Group IIIa iron meteorites. The number beside each data point
is the cooling rate, in degrees $C/10^6$ yr. All meteorites common
to the studies of Wasson and Kimberlin (1967) and Goldstein (1969)
are plotted. Two subgroups, probably representing discrete
sources, are indicated.

dynamic processes (e.g., Scott's (1977) concept of intrusion of molten metal into olivine
cumulates) do not square well with the orderly, close-packed structure of most pallasites.

There is an aspect of melting and differentiation of planetary interiors that would
have produced pallasitic material as a gravitationally stable layer at the core/mantle
interface. An olivine cumulate layer immersed in mafic silicate magma would press down on
the interface between magma and differentiated molten metal/sulfide with a weight propor-
tional to (1) the thickness of the cumulate layer, (2) the difference in density between
olivine and magma, and (3) the local value of g. This would submerge the lowermost olivine
crystals a certain distance into the molten metal/sulfide (Figure 4a). The depth of sub-
mergence would be greater if magma were free to erupt to the surface of the hypothetical
planet, meaning that (1) above would embrace all the unmelted substance of the planet whose
density was greater than that of the magma (Figure 4b). The amount of pallasitic material
that can be formed by the mechanism of Figure 4b can be calculated by assessing the downward
weight of solid silicates between R_2 and R_3 and requiring the upward buoyant forces of solid
olivine crystals between R_1 and R_2 to equal this value:

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$$\int_{R_1}^{R_2} \Delta\rho \text{metal}, g(r) dr = \int_{R_2}^{R_3} \Delta\rho \text{olivine}, g(r) dr \qquad (1)$$
$$\quad\;\text{olivine} \qquad\qquad\qquad \text{magma}$$

Results are shown in Figure 5, for a range of possible core sizes. The relative volume of pallasitic material produced is independent of the absolute size of the planet. In principle, the volume of pallasitic material that would form in an interrally melted planet (relative to pure Ni,Fe metal) is substantial, larger than the ratio of pallasites to irons in museum collections.

Absolute values of the weight exerted downward by cumulate olivine at R_2 can be estimated, by assessing one side of Equation (1) as a function of R ($= R_3$). These are 4, 16, 63, and 390 bars, in planets of total radius 50, 100, 200, and 500 km respectively, for reasonable compositions and internal configurations. These are small stresses, but olivine is extremely weak at the high temperatures of molten iron, and even small directed stresses cause it to yield by the mechanism of power-law creep (Ashby and Verrall, 1977).
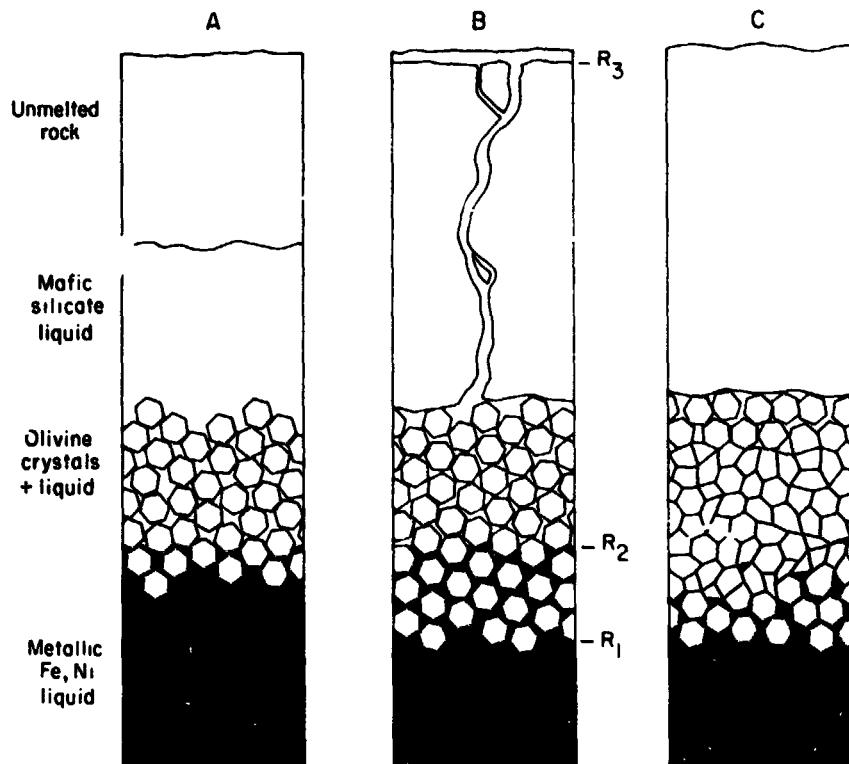

Fig. 4. Radial columns in an internally melted small planet. A: the simple case in which the melted zone is enclosed by an integral shell of unmelted rock. B: the more realistic case where the unmelted shell fractures and founders, resting on cumulate olivine in the melted zone; silicate melt is erupted to the planetary surface. C: the effect of deformation of olivine in the pallasitic layer is to squeeze inter-cumulus liquids out of it, closing the olivine cumulate layer into pure dunite.
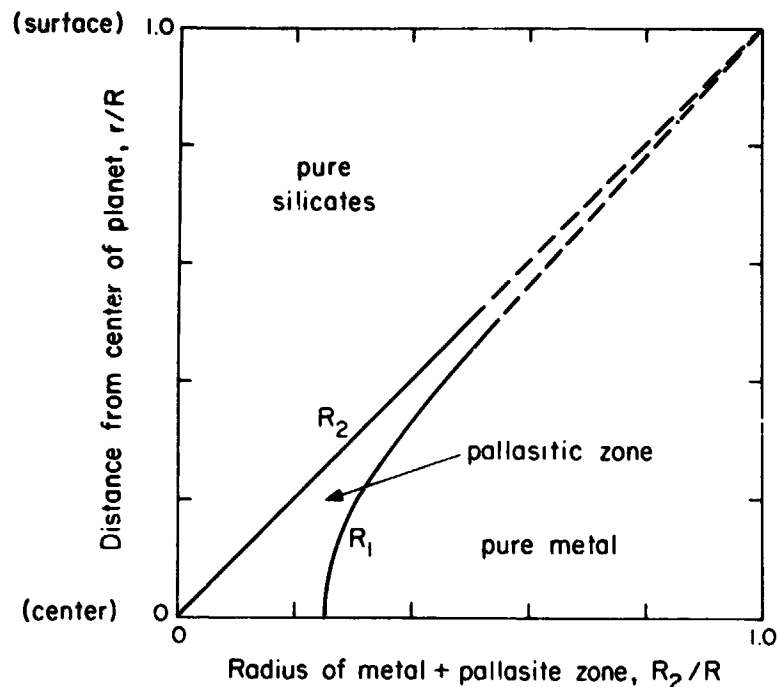
Fig. 5. Positions of the R_2 and R_1 leve'? (which define the thickness of the pallasite zone), as a function of the tutal size of the metal + pallasite zone (the core). All values are relative to the overall radius of the planet (R); the relationships are independent of absolute size. Dashed portions of curves correspond to unrealistically large cores, larger than would be produced by total melting and differentiation of ordinary chondrites.

Total deformation (γ, where $\gamma = 1$ corresponds to the shear strain needed to deform a right angle to a 45° angle) equals the strain rate ($\dot{\gamma}$) times the time (ΔT) needed to cool to solidification. Taking

$$\sigma \simeq (5 \times 10^{-6})R^2$$

(σ = directed stress in Kbar, R = planet radius in km; from the above calculations),

$$\dot{\gamma} = (4.6 \times 10^4)\sigma^{2 \cdot 91}$$

for olivine in the temperature and shear stress regime of interest (Ashby and Verrall, 1977), and

$$\Delta T \simeq (2 \times 10^3)R^2$$

from cooling rate calculations made by the author (ΔT in years; assuming a site at depth = $0.5R$, and cooling through the temperature range 1600°-1400°C), the relationship

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between planetary dimension and total deformatioı experienced by pallasitic olivine is found to be

$$\gamma = (3.4 \times 10^{-8})R^{7.82} \tag{2}$$

This relationship is plotted in Figure 6 (uppermost curve).

Deformation great enough to obliterate the characteristic pallasite geometry would be experienced by olivine crystals at the core-mantle interfaces of planets larger than ~ 10 km radius. Olivine deformation in these circumstances would have the effect of squeezing molten metal out the bottom of the pallasitic layer and molteɴ mafic silicate out the top of the overlying layer (Figure 4c), resulting in a stable layer of virtually pure dunite.

Since Equation (2) applies only to level R_2, and the directed stresseₛ and γ taper to zero at R_1, it might appear that only the upper portion of the pallasitic layer was in danger of obliteration in large planets. However, the dashed curves of Figure 6 make it clear that no significant portion of the pallasitic layer in a planet much larger than 10 km radius would survive destruction.
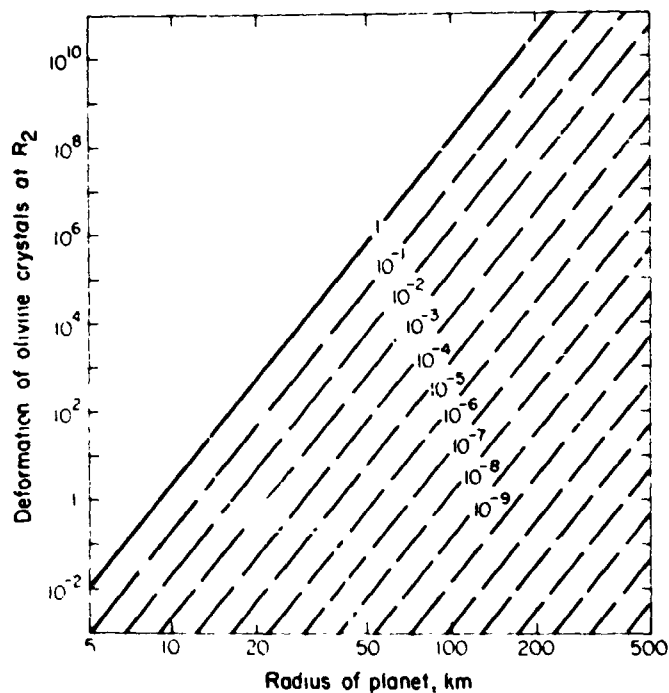

Fig. 6. Total deformation experienced by pallasitic olivine during cooling of a small planet, as a function of the planet's dimension. The curve labeled 1 represents olivine ıt the R_2 interface, where stresses are greatest; the 10^{-1} curve appliₑₛ at a level 10^{-1} of the distance between R_1 and R_2, and so forth.

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Deformation experienced in >100 km planets (such as Figure 2 appears to require as a cooling site for pallasites) is excessive by such a large factor ($>10^8$) that even generous allowance for the uncertainties attached to the estimate does not make it possible to reconcile the circumstances of formation with the circumstances of cooling of pallasites. *It appears inescapable that the pallasites solidified in small ($\sim$'J km) bodies, but these subsequently must have joined larger (>100 km) bodies before il cooling through 500°C occurred.* The s ze and energy requirements and the timing of the first generation of bodies are consistent with the Goldreich-Ward mechanism of planetesimal formation by gravitational instability of a dust disk within the primordial nebula (^{19} and the ^{26}Al content of early solar system material (Lee *et al.*, 1977). Presumably the second generation was accumulated by planetesimal encounters over a longer time period, after tne dissipation of the nebula.

Thus it appears likely that the (second generation) parent meteorite planets did in fact have "raisin bread" structures, as a consequence of their assembly from smaller differentiated bodies; but there is no reason to expect that the "raisins" in any particular second-generation body were geochemically related. It is quite possible that most or all iron meteorites went through a similar two-stage history, being melted and differentiated in a small body then cooling as a "raisin" in a larger planet, but there is no obvious way to test this. The iron meteorites may have formed as cores interior to pallasitic layers, or their first generation bodies may have been too large to permit the survival of significant amounts of pallasitic material.

ACKNOWLEDGMENT

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DISCUSSION

GROSSMAN: I would like to know if the boundaries in Figure 1 change depending on the proportions of CaO, MgO, etc.? I am wondering whether the particular section you are showing is sensitive to these proportions.

WOOD: If by proportions you mean changing the modal forsterite to anorthite ratio, no, the boundaries do not change. Al' the isopleths of Al in the M1 site of orthopyroxene and the boundaries shown in Figure 1 are valid for the mineral assemblages listed, irrespective of the mineral proportions.

GROSSMAN: I wonder if you ought to stretch the point to include up to 12 Kbars on the spinel minerals?

WOOD: It is important to first specify what sort of spinels you are referring to. Mg and Al-rich spinels in equilibrium with olivine and two pyroxenes would indeed reflect very high pressures. In this case plagioclase would no longer be stable with olivine. However, chromite-type spinels can be in equilibrium with olivine, plagioclase, and two pyroxenes at very low pressures (*i.e.*, 0-1 Kbar at about 850°C). The Mg-rich spinels of chondritic Ca, Al-rich inclusions coexist with minerals other than those given in Figure 1; the figure is not applicable to spinel + melilite + fassaite, etc., assemblages.

ZELLNER: Apparently you must have at least one parent body of 500 km diameter.

WOOD: That is what is implied in Figure 2, where the diameter jumps from 200 to 500 km. But I am not sure you couldn't achieve the lowest cooling rates in somewhat smaller bodies.

ANDERS: Isn't it still true that the very lowest cooling rates, below a degree per million years, are a paradox, being found in some of the unequilibrated ordinary chondrites that are volatile-rich and which seem to have had a fairly gentle temperature history? To get these low cooling rates, one has to go fairly deep inside a large body. The second paradoxical category is mesosiderites, where I think you pointed out that they would not cool to a reasonable temperature in the entire age of the solar system.

WOOD: Yes, that is right, if you make the assumption that the mesosiderites have resided at the same position in their parent bodies since the time they were formed. If you are willing to explore more exotic possibilities of relocation of volumes of meteoritic material, and to go through several generations of parent bodies, then these things can be worked out.

ANDERS: Isn't it possible that the method works reliably for irons but occasionally malfunctions for mixtures of silicate and iron? I think your cooling rates for ordinary chondrites certainly are very plausible. But those for the other types of meteorites such as unequilibrated chordrites, mesosiderites and pallasites (again mixtures of stone and iron), all come out suspiciously lower. I wonder if there isn't something in the method that goes sour.

WOOD: I personally don't believe so. The fact is, the first data that indicated a very low cooling rate for a mesosiderite was not gotten by me or Powell on the basis of isolated metal grains. It was gotten by Short on the basis of a nodule in a mesosiderite that displayed integral Widmanstätten structure, and on the basis of the old classical cooling rate method he came up with that slow cooling rate. The work was never published, although it appeared in an abstract for a Meteoritical Society meeting. Short had a small piece of an iron meteorite there; the cooling rate he derived from it should be as reliable as the cooling rates gotten for octohedrites.

ANDERS: I think there is one very important clue that you omitted which can lead to a totally different conclusion. Namely, quite a number of pallasites have olivines at the top and at the bottom and a clear channel of metal in the middle. This coniiguration is not in hydrostatic equilibrium. In my 1964 review I pointed out one way to get this is to have an olivine zone overlying a core and then mix these two by shock. Obviously such configurations could not have persisted if the materials had remained molten for a long time. It is only the momentarily liquid state after shock, followed by instantaneous freezing, that allows this to survive.

WOOD: I didn't have time to address all aspects of pallasites. There are some properties of pallasites and some types of pallasites that have clearly been affected by dynamic situations, where there was an injection or intrusion of molten metal. Indeed, the fact that some pallasites consist of olivine fragments rather than nicely rounded crystals bespeaks some sort of violent event. The fact is, however, a great many, perhaps most of the pallasites, do consist of rounded olivines in nicely ordered, close-packed arrays, as was shown by no less than Lord Rayleigh. To that you would have to add a caveat to the effect that some of these parent bodies suffered shocks or stresses or techtonic events during the crystallization of their cores that allowed still molten metal to intrude regions that had begun to solidify. The regions of olivine crystals at the two ends of the particular specimen that you speak of are in close-packed array. The violent event that injected metal between them could not really have been the same event that filled in the spaces between these nicely ordered arrays of olivine crystals. The crystals just wouldn't stay together; they would scatter out.

VEVERKA: Two points. First, a crucial question is what is the effective g? We can reduce g by considering rotation. In fact, if the original bodies had very rapid rotation periods (~2 hrs), then g would have been close to zero, or at least very small throughout the body. Thus for rapidly rotating bodies deformation may have been small even at great depth. As long as g wasn't exactly zero, gravitational settling could still be invoked as Goles, Fish and Anders showed in 1960. Second, a discussion of the uncertainties in the exponent of σ, which is a component of Equation (2), is needed. Two aspects need to be considered. First, in the actual laboratory measurements is the exponent determined to ±0.1, ±0.5, or what? Second, how well can one extrapolate the exponent to the present case? The argument about the need for two generations of "parent" bodies hinges in part on the high exponent of σ. Could the exponent be reduced to 5? Would that change the conclusions? The argument that two generations of parent bodies are needed to account for the pallasites hinges on Equations (1) and (2). In Equation (1) it is essential to make sure that the correct g is being used. In Equation (2) it is essential that the correct exponent is being used.

WOOD: Let me address the second point first. The relationship between stress (σ) and the strain rate ($\dot{\gamma}$) of olivine is based on experimental data by Durham and Goetze (Plastic flow of orientated single crystals of olivine. *J. Geophys. Res.* 82, 5737-5753, 1977), at 1600°C in the stress range 0.2-0.5 Kbar. This corresponds to conditions at the core/mantle interface of planets of approximate total radius 200-300 km. Outside this range, the stress-strain relationship depends upon formulae based on the theoretical analysis of creep mechanisms by Ashby and Verall (1977), which is fitted to the data of Durham and Goetze and also to the data of other authors at lower temperatures and

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higher stresses. The exponent of σ found by Durham and Goetze is actually higher than the value used by me (3.6 ± 0.3, versus 2.91). This is because my formula is based on a curve I fitted to Ashby and Verall's extrapolation (in an attempt to embrace the stress range 0.01-1 Kbar), rather than to Durham and Goetze's data. I am not really qualified to critically evaluate Ashby and Verall's modeling, but note that if I went with the experimental data instead, the exponent in Equation (2) would be even higher than it is, and the point this paper attempts to make would be even more compelling. The experimental data only goes down to a planetary radius of $\sim$200 km, but even a modest extrapolation of the experimental stress-strain relationship to smaller dimensions suffices to exclude the possibility that the pallasites crystallized in planets as large as their cooling rates indicate. The effects of rotation are interesting to consider. Centrifugal acceleration would tend to offset gravity most beneath a planet's equator, and there the pallasitic structure might be preserved in arbitrarily large planets if they were spinning fast enough ($i.e.$, at just less than the stability limit). The centrifugal acceleration would go to zero at the poles of the planet, though, and there the pallasitic structure would be vulnerable. However, if Equation (2) is valid, the deformative stresses that would act in a nonrotating planet are excessive by such a large factor ($>10^8$, as noted toward the end of the article) that it seems unlikely pallasites could be saved from collapse by rotation. First, if the centrifugal acceleration were less than gravitational acceleration by even as much as 10^{-8}, even pallasites beneath the equator would be doomed. This leaves an extremely narrow window of rotation velocities that would do the job, since the planet would fly apart if the centrifugal acceleration exceeded the gravitational acceleration by very much. Second, even if rotation exactly offset gravity at the equator, one would only have to go to latitudes $>\sim10^{-8}$ radians above and below the equator to find the component of centrifugal acceleration no longer adequate to preserve the pallasites.