

# The cordierite- to spinel-cataclasite transition: Structure of the lunar crust

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**Abstract**—The moonwide magmatic processes which were responsible for the formation of the lunar crust and mantle cannot be understood without stratigraphic constraints. Some meteorite-excavated lunar highlands rocks contain this information in their mineral chemistry and stability relations, as do nodules in some basaltic rocks and kimberlites on the earth. Therefore, from an understanding of the physical-chemical conditions which are required to form certain nonmare mineral assemblages and their mineral chemistry, inferences can be drawn about their preexcavation stratigraphic location.

The cordierite- and spinel-cataclasites (olivine + plagioclase + high alumina orthopyroxene  $\pm$  spinel  $\pm$  cordierite) exhibit phase changes and considerable variations in their mineral chemistry at depths appropriate to the lunar crust. These have been calibrated by experiment in the system MgO-Al<sub>2</sub>O<sub>3</sub>-SiO<sub>2</sub> and extended to the natural lunar chemical system by solution thermodynamics. The results show that spinel cataclasites from Apollo 15 and 17 were located at depths greater than or equal to about 12 to 32 kilometers prior to excavation. An ultramafic spinel cataclasite clast in 15445,177 may have been located in the uppermost mantle of the moon.

From geochemical relationships between spinel cataclasites and the other members of the Mg-rich and anorthositic series given in plots of Mg/(Mg+Fe) in mafic minerals vs. An content of plagioclase, and from relative abundances of these rocks, a two layer crust of the moon can be modelled. In this model the uppermost stratigraphic unit consists of members of the anorthositic series, and is probably about 12 to 20 kilometers thick. The lower Mg-rich unit may constitute the greatest volume of the crust, and consist of rocks with roughly cotectic mineralogical proportions (i.e., norites, troctolites, spinel troctolites and their buffered low-K Fra Mauro and KREEP compositions).

## INTRODUCTION

Lunar highlands plutonic and volcanic rocks contain information about the moonwide process of crust and mantle formation. Yet after nearly 10 years of research, no durable and comprehensive model of this process exists. In part this is because there are few observational constraints which can be used to test models of lunar crust formation, the vertical and lateral petrologic structure of the crust being largely unknown.

Perhaps the many problems encountered by such models can best be appreciated by considering the progress made in understanding the formation of layered intrusions of the Earth such as Skaergaard and Stillwater. The papers contained in the American Journal of Science Jackson Memorial Volume (1980) report some of the geochemical, physical, and thermal processes which no single unifying

model has yet succeeded in integrating. That is, no model is currently able to predict the stratigraphy of rocks in layered intrusions. The problem then becomes compounded when similar attempts are extended to the lunar crust because of the absence of stratigraphic constraints which are routinely considered by geologists working on layered intrusions. The rocks which composed the lunar crust column are now meteorite-excavated surficial debris which may or may not have had cogenetic igneous histories.

Most models of lunar crust and mantle formation have been articulations of the magma ocean paradigm in which the entire thickness of a plagioclase-rich crust was a floated crystallization product (e.g., Wood *et al.*, 1970; Smith *et al.*, 1970; Wood, 1975; Drake, 1976; Solomon and Longhi, 1977; Longhi and Boudreau, 1979; Hubbard and Minear, 1975; Minear and Fletcher, 1978). The textures, mineralogy, and chemistry of this primordial crust were in most cases extensively modified by meteorite impact, thus obscuring the earliest record of the moon's history. However, some primordial material is believed to have survived to varying degrees the chemical and physical effects of meteorite bombardment. These are called pristine lunar highlands rocks (Warren and Wasson, 1977; Warner and Bickel, 1978). Such rocks, therefore, are likely to yield the most information about the nature and formation of the crust and mantle.

Pristine rocks define two distinct geochemical trends (Roedder and Weiblen, 1974; Warner *et al.*, 1976; Warren and Wasson, 1977; see Fig. 2): an anorthositic trend of varying  $Mg/(Mg+Fe)$  in the mafic minerals at essentially constant An content of plagioclase, and an Mg-rich trend (i.e., spinel-cataclasites, troctolites, dunite, norites) of complementary variations in these two parameters. The formation of these two suites of highland rocks can be modelled from the crystallization of a global magma ocean; the most detailed model has the ferroan anorthosites stratigraphically above the Mg-rich rocks (Longhi and Boudreau, 1979).

Some recent alternative models have tended to explain both pristine rock suites by formation from two separated magmas (Warner *et al.*, 1976; Warren, 1979; James, 1979; Binder, 1979); the ferroan anorthosites were formed as floated crystallization products of a lunar magma ocean whereas the Mg-rich rocks formed later as intruding plutons from below. Although a possible weakness in the "two magmas" hypothesis has been illustrated by analogy with similar mineralogical trends in the Stillwater Complex which crystallized from one magma (Raedeke and McCallum, 1979; DePaolo and Wasserburg, 1979), the model may account for a layered lunar crust similar in character to that proposed by Longhi and Boudreau (1979).

On the basis of melt-rock samples and their clasts from the Apollo 15 and 17 landing sites, Ryder and Wood (1977) constructed a layered model of the crust with features similar to those above. The uppermost stratigraphic horizons were inferred to consist of anorthosite and noritic anorthosite which grade downwards to more ferromagnesian-rich rocks with higher  $Mg/(Mg+Fe)$ . This model is consistent with the observation that the ferroan anorthosites are preponderant among pristine rocks from the Apollo 16 landing site which is the furthest removed from

large impact basins (Warren and Wasson, 1977). Conversely, the Mg-rich rocks are preponderant among pristine rocks from the Apollo 15 and 17 landing sites which, being near the rims of the Imbrium and Serenitatis impact basins, are the most likely to contain excavated material from the lower crust. A consensus, therefore, appears to be developing in which the anorthositic and Mg-rich suites of highland rocks are excavated samples of a layered crust.

Any layering of the lunar crust can have far-reaching consequences on the models themselves. For example, in the absence of any accurate estimate of the bulk composition of the crust, the compositional layering may range from dominantly anorthosite with minor variations in the plagioclase/mafic mineral ratio and  $Mg/(Mg+Fe)$ , to layering in which anorthosite is volumetrically overwhelmed by norite-troctolite, or low-K Fra Mauro material (Ryder and Wood, 1977; Charette *et al.*, 1977). In the latter case, "the incentive for calling on a lunar magma ocean disappears as no anorthositic layer of global proportions need be accounted for" (Walker, 1979). Indeed, the latter crustal profile could be modelled with or without a magma ocean. The interaction of convecting magma of a global ocean with the underside of anorthositic rockbergs, entrapment of mafic crystals in the advancing plagioclase flotation crust, lack of efficient gravitational separation caused by convection are all possibilities in the magma ocean concept (Minear and Fletcher, 1978; Longhi and Boudreau, 1979; Herzberg, 1978a). Alternatively, an Mg-rich layered crust with a profile similar to sections of the Stillwater Complex (Raedeke and McCallum, 1979) may also be the product of a lunar interior which underwent extensive partial melting. Indeed, the distinction between the two models may be somewhat misleading, the magma ocean requiring something to be partially or totally melted. If the products of a magma ocean differ from those of a partially melted lunar interior, these may be manifest in the kind of crust each may produce. The stratigraphy of each may differ and, accordingly, become identified as constraints on the competing hypotheses.

Assuming that the products of the two models will become understood, it is clear that the formation of the lunar crust will remain in a shroud of mystery unless its petrological and geochemical structure becomes known. We have directed our efforts at supplying such stratigraphic constraints by developing thermodynamic criteria which permit depth determinations to be made on various mineral assemblages. Of the highlands lithologies which, given proper calibration of the relevant equilibria, exhibit a range of mineral compositions and phase changes appropriate to the depths of the lunar crust, the spinel cataclasites are the best candidates (Herzberg, 1978a, 1979). Olivine-bearing norites, and olivine-two pyroxene anorthosites (e.g., 62275,4, Prinz *et al.*, 1973; Herzberg, 1979) are also subject to depth estimates upon calibration of the relevant equilibrium discussed in Herzberg (1979).

Depth estimates for spinel cataclasite clasts (olivine + plagioclase + high alumina orthopyroxene + spinel  $\pm$  cordierite) in 72435, 73263, and 15445 were previously made (Herzberg, 1978a) based on published mineral analyses and the best available thermochemical data at that time. It was noted, however, that such

estimates were subject to a possible error of about  $\pm 30$  kilometers, based on uncertainties in the thermochemical data. Accordingly, we have experimentally calibrated the equilibrium which bounds the T-P stability fields of spinel cataclasites and their low-pressure cordierite breakdown products. The results of these experiments are reported here and their application to the depth of origin of lunar spinel cataclasites discussed. Details of the experimental methods and thermodynamic analysis of the results are rather lengthy, and will be reported elsewhere (Herzberg, in preparation). Finally, we suggest some possible relationships between spinel cataclasites and the other members of the anorthositic and Mg-rich suites, and offer a stratigraphic model of the lunar crust.

### THE CORDIERITE- TO SPINEL-CATACLASITE BOUNDARY IN THE SYSTEM $\text{MgO-Al}_2\text{O}_3\text{-SiO}_2$

Ideally, the best way to experimentally establish the depth of origin of natural spinel cataclasite assemblages would be to “cook” the assemblage at the temperatures at which they equilibrated and the appropriate range of pressures (i.e., 1 to 3000 bars), and observe any breakdown products or changes in the mineral chemistry. However, spinel cataclasites being plutonic rocks which presumably cooled from magmatic temperatures, there are good reasons to expect temperatures of equilibration in the range of  $600^\circ\text{--}1000^\circ\text{C}$ . These temperatures, however, render direct experimental determination an impossible task due to the exceptionally low reaction rates in anhydrous silicate systems. An alternative method is to perform the experiments at high temperatures (e.g.,  $\geq 1200^\circ\text{C}$ ) and extrapolate. For natural chemical systems this approach is inadequate because the many chemical variables and equilibria (e.g., 7–8 important oxide components) preclude a thermodynamic basis for any extrapolation at this time. Linear T-P extrapolations of a limited set of high temperature data are invalid because changing configurational entropy terms due to order-disorder and solid solution effects (O’Hara *et al.*, 1971; Herzberg, 1978b) invariably result in pyroxene-bearing facies boundaries which are curves in T-P space.

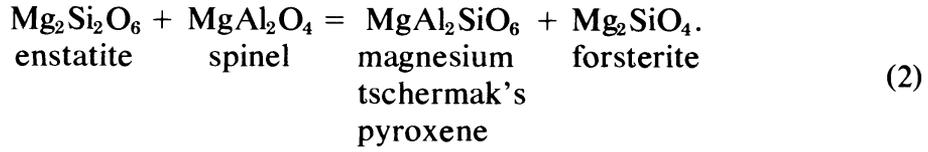
The method chosen here involves experimental determination of the cordierite-to spinel-cataclasite boundary in the simple system  $\text{MgO-Al}_2\text{O}_3\text{-SiO}_2$  which represents about 80–90 wt % of the relevant components in natural complex spinel cataclasites. The other important components are FeO,  $\text{Cr}_2\text{O}_3$ ,  $\text{TiO}_2$  and CaO. With certain testable assumptions, the effects of these extra components on the temperature and depth estimates are considered through solution thermodynamics.

In the system  $\text{MgO-Al}_2\text{O}_3\text{-SiO}_2$  the cordierite- to spinel-cataclasite boundary is represented by the univariant equilibrium (Herzberg, 1978a):

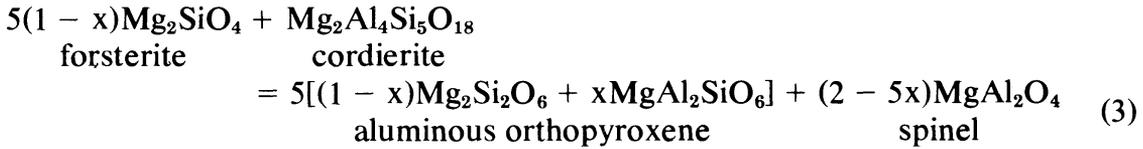


This is a simple end-member reaction which does not obviously show the aluminous nature of orthopyroxene. Orthopyroxene compositions buffered by for-

sterite and spinel vary with temperature and pressure according to the reaction (Dankwerth and Newton, 1978; Herzberg, 1978a):



A mass balance representation of the reaction forsterite + cordierite to aluminous orthopyroxene + spinel is a combination of reactions (1) and (2), and takes the form:



where  $x$  is the mole fraction of  $\text{MgAl}_2\text{SiO}_6$  in aluminous orthopyroxene [equivalent to  $(X_{\text{Al}}^{\text{Ml}})_{\text{opx}}$ , that is the mole fraction of Al in the Ml site of orthopyroxene].

The T-P locations of reaction (1) and the alumina content of orthopyroxene can be quantified in part by consideration of the free energy relation:

$$\begin{aligned} (\Delta G^\circ)_{\text{T,P,1,2}} &= (\Delta H^\circ - T\Delta S^\circ + P\Delta V^\circ)_{\text{T,P,1,2}} \\ &= -RT \ln K_{1,2} \end{aligned} \quad (4)$$

where the equilibrium constants:

$$K_1 = \frac{(a_{\text{Mg}_2\text{Si}_2\text{O}_6}^{\text{opx}})^5 (a_{\text{MgAl}_2\text{O}_4}^{\text{sp}})^2}{(a_{\text{Mg}_2\text{SiO}_4}^{\text{ol}})^5 (a_{\text{Mg}_2\text{Al}_4\text{Si}_5\text{O}_{18}}^{\text{cord}})} \quad (5)$$

and

$$K_2 = \frac{(a_{\text{MgAl}_2\text{SiO}_6}^{\text{opx}}) (a_{\text{Mg}_2\text{SiO}_4}^{\text{ol}})}{(a_{\text{Mg}_2\text{Si}_2\text{O}_6}^{\text{opx}}) (a_{\text{MgAl}_2\text{O}_4}^{\text{sp}})} \quad (6)$$

become for multisite mixing of cations:

$$K_1 = \frac{(X_{\text{Mg}}^{\text{M2}})_{\text{opx}}^5 (X_{\text{Mg}}^{\text{M1}})_{\text{opx}}^5 (\gamma_{\text{Mg}_2\text{Si}_2\text{O}_6})_{\text{opx}}^5 (X_{\text{Mg}}^{\text{TET}})_{\text{sp}}^2 (X_{\text{Al}}^{\text{OCT}})_{\text{sp}}^4 (\gamma_{\text{MgAl}_2\text{O}_4})_{\text{sp}}^2}{(X_{\text{Mg}}^{\text{OCT}})_{\text{ol}}^{10} (\gamma_{\text{Mg}_2\text{SiO}_4})_{\text{ol}}^5 (X_{\text{Mg}}^{\text{OCT}})_{\text{cord}}^2 (\gamma_{\text{Mg}_2\text{Al}_4\text{Si}_5\text{O}_{18}})_{\text{cord}}} \quad (7)$$

and

$$K_2 = \frac{(X_{\text{Al}}^{\text{M1}})_{\text{opx}} (\gamma_{\text{MgAl}_2\text{SiO}_6})_{\text{opx}} (X_{\text{Mg}}^{\text{OCT}})_{\text{ol}}^2 (\gamma_{\text{Mg}_2\text{SiO}_4})_{\text{ol}}}{(X_{\text{Mg}}^{\text{M1}})_{\text{opx}} (\gamma_{\text{Mg}_2\text{Si}_2\text{O}_6})_{\text{opx}} (X_{\text{Mg}}^{\text{TET}})_{\text{sp}} (X_{\text{Al}}^{\text{OCT}})_{\text{sp}}^2 (\gamma_{\text{MgAl}_2\text{O}_4})_{\text{sp}}} \quad (8)$$

In Eqs. (7) and (8),  $(X_a^b)_c^d$  refers to the mole fraction of cation  $a$  in site  $b$  of phase  $c$  to the power  $d$ , and  $(\gamma_e)_f^g$  refers to the activity coefficient of component  $e$  in phase  $f$  to the power  $g$ . The activity coefficients expressed in this way combine interactions of cations within and between energetically distinct sites (cf., Sack, 1979; Wood and Nicholls, 1978). It is a macroscopic formulation which does not allow the inter- and intrasite cationic interactions to be considered separately.

With the standard state taken to be the pure phases at the temperature and pressure of interest, all activity coefficients in the simple system are unity except those for the two pyroxene components. For the latter, evidence favouring a close approach to ideality (i.e.,  $\gamma \rightarrow 1$ ) has been suggested by Dankwerth and Newton (1978) on the basis of experimental and calorimetric data.

In this simple system, Eqs. (4) and (8) can be rearranged and quantified:

$$T = \frac{-P[0.18(X_{\text{Al}}^{\text{M1}})_{\text{opx}} - 0.005] - 4522}{\ln K_2 - 1.811}, \quad (9)$$

where

$$K_2 = \frac{(X_{\text{Al}}^{\text{M1}})_{\text{opx}}}{(1 - X_{\text{Al}}^{\text{M1}})_{\text{opx}}}. \quad (10)$$

In Eq. (9)  $P$  is in bars and  $T$  is in  $^{\circ}\text{K}$ . This equation was derived from Dankwerth and Newton (1978), and includes the entropy, enthalpy, and volume changes of reaction (2) and the effects of thermal expansion of the phases. Thus, for any value of  $(X_{\text{Al}}^{\text{M1}})_{\text{opx}}$  in this simple system at a designated pressure, the temperature can be calculated. Figure 1 shows the  $T$ - $P$  distribution of these isopleths of alumina from Eqs. (9) and (10).

Figure 1 shows also the experimentally determined temperatures and pressures at which reaction (1) occurs [or more descriptively reaction (3); Herzberg, in preparation]. These are the conditions at which the cordierite cataclasite assemblage (i.e., forsterite + cordierite + orthopyroxene  $\pm$  anorthite) reacts to the spinel cataclasite assemblage (i.e., forsterite + spinel + orthopyroxene  $\pm$  anorthite) in this simple system (note that anorthite is an extra phase which does not participate in any of the reactions, except in contributing small amounts of Ca to orthopyroxene). Combining Eqs. (4) and (8), the experimental data can be fitted to the equation:

$$P = \frac{RT}{-1.43} \left[ \frac{401.6}{T} - 1.839 - \ln K_1 \right] \quad (11)$$

where

$$K_1 = [1 - (X_{\text{Al}}^{\text{M1}})_{\text{opx}}]^5. \quad (12)$$

$R$  is the gas constant, and the value  $-1.43$  is the volume change of equilibrium (1) in  $\text{cal bar}^{-1}$ . At any temperature along the equilibrium boundary,  $K_1$  can be evaluated from the alumina isopleths of Fig. 1 or Eqs. (9) and (10).

Equation (11) permits an extrapolation of the cordierite- to spinel-cataclasite boundary to all temperatures as shown in Fig. 1. The accuracy of this extrapolation is tested by comparison with the experimental results of Seifert (1974) and Fawcett and Yoder (1966) on the same reaction in the water-saturated system by means of the hydration model of Newton and Wood (1979). Using this model, which is based on  $\text{H}_2\text{O}$  solubility data in Mg-cordierite (Mirwald and Schreyer, 1977), the location of the boundary in both the wet and dry systems satisfy the experimental data to within  $\pm 0.20$  kbar over the temperature range  $780^{\circ}$ – $1300^{\circ}\text{C}$ .

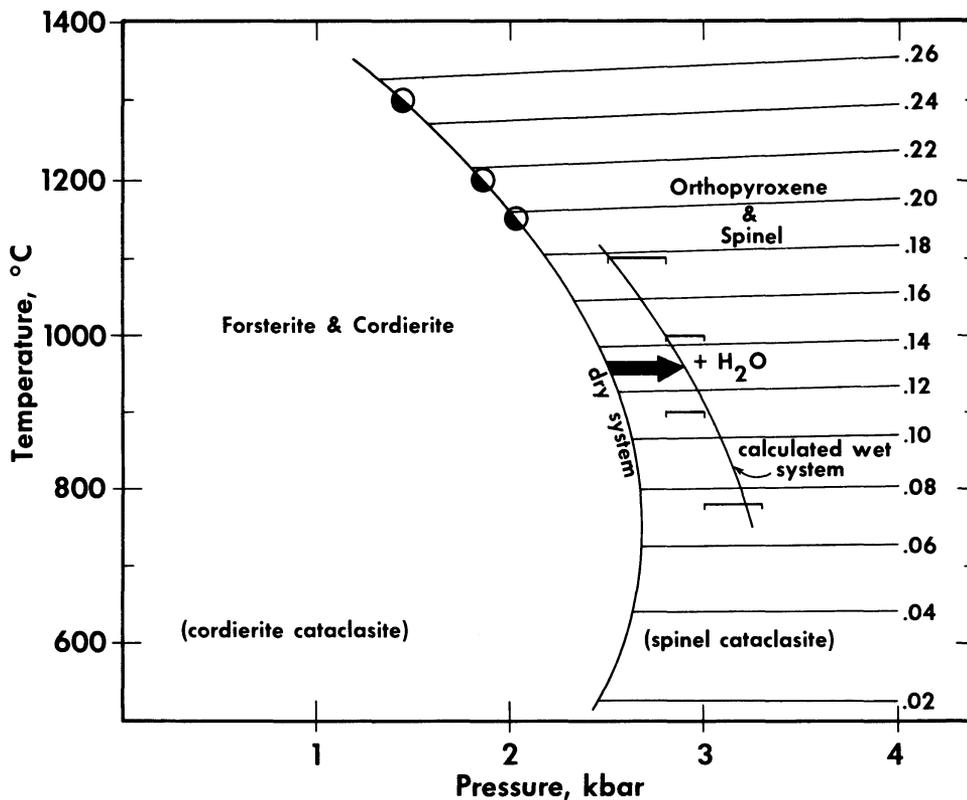


Fig. 1. The cordierite- to spinel-cataclasite boundary in the system  $\text{MgO-Al}_2\text{O}_3\text{-SiO}_2$  ( $\pm\text{H}_2\text{O}$ ). Half-filled circles are experimental results from Herzberg (in preparation), and extrapolated curve is from equation (11) in the text. Brackets are experimental data from Seifert (1974) and reversal data from Fawcett and Yoder (1966) in the wet system. Curve passing through these brackets is calculated from the dry reference state using the cordierite hydration model of Newton and Wood (1979). Numbered lines are isopleths of alumina in orthopyroxene  $[(X_{\text{Al}}^{\text{M1}})_{\text{OPX}}]$  from Dankwerth and Newton (1978) using Eq. (9) in the text.

### APPLICATION OF THE CORDIERITE- TO SPINEL-CATACLASITE BOUNDARY IN THE SYSTEM $\text{MgO-Al}_2\text{O}_3\text{-SiO}_2$ TO LUNAR SPINEL CATACLASITES

If a lunar spinel cataclasite assemblage was confined to the system  $\text{MgO-Al}_2\text{O}_3\text{-SiO}_2$  (+ CaO to form anorthite), it would be stable at some pressure greater than or equal to those shown in Fig. 1 (i.e., dry system) at any specified temperature. However, as reviewed by Ryder and Bower (1977), small but significant amounts of  $\text{Cr}_2\text{O}_3$  and FeO are present in the mafic phases. Because these two components partition preferentially into spinel, they can significantly expand the spinel stability field at the expense of the cordierite field (Herzberg, 1978a). Therefore, depth estimates must consider these extra components. To do so requires that the equilibrium constant  $K_1$  be calculated from the mineral analyses of the spinel cataclasite assemblage, and the pressure solved from Eq. (11). In this natural system,  $K_1$  must be evaluated from Eq. (7). However, because the activity coef-

ficients in Eq. (7) are not known, a first-order approximation of ideality can be made and tested (i.e.,  $\gamma$ 's remain equal to unity). Thus:

$$K_1 = \frac{(X_{Mg}^{M2})_{opx}^5 (X_{Mg}^{M1})_{opx}^5 (X_{Mg}^{TET})_{sp}^2 (X_{Al}^{OCT})_{sp}^4}{(X_{Mg}^{OCT})_{ol}^{10} (X_{Mg}^{OCT})_{cord}^2} \quad (13)$$

It is clear from inspection of Eq. (13) that the site occupancies for each mineral from their chemical analysis must be estimated. For orthopyroxene this involves construction of a structural formula as outlined previously (Herzberg, 1978a). It is known, however, that for orthopyroxenes in the system  $Mg_2Si_2O_6$ - $Fe_2Si_2O_6$ ,  $Fe^{2+}$  partitions preferentially into the larger M2 site (Virgo and Hafner, 1969; Saxena and Ghose, 1971). Not only is it safe to assume that this partitioning will be maintained for natural aluminous orthopyroxenes, but because of the lattice contraction effect of Al in the M1 site, it is possible that it will be enhanced somewhat relative to the simple binary system. To simplify the calculations, however, values for the site occupancies can be made by the approximation that  $Mg/(Mg+Fe)$  in M1 and M2 are equal to the whole pyroxene value. Although this approximation will result in slight underestimations in the calculated pressure (e.g., for 15445,177; 1.42 kbar instead of 1.43 kbar using the partitioning data in Virgo and Hafner (1969) and Saxena and Ghose (1971)), these are not significant.

Spinel site occupancies are simply calculated from the cation proportions:

$$(X_{Mg}^{TET})_{sp} = Mg/(Mg + Fe^{2+} + Mn); \quad (14)$$

$$(X_{Al}^{OCT})_{sp} = Al/(Al + Cr + Si + Ti). \quad (15)$$

These occupancies are those for a fully ordered normal spinel. Although some disordering is likely to occur and be similar for  $MgAl_2O_4$ - $FeAl_2O_4$  spinel solutions (Navrotsky and Kleppa, 1967) the boundary shown in Fig. 1 and evaluated from equation (11) considers this (Herzberg, in preparation).

Mg site occupancy for olivine consisting of two energetically similar sites is simply:

$$(X_{Mg}^{OCT})_{ol} = Mg/(Mg + Fe + Ca + Mn). \quad (16)$$

For spinel cataclasites which contain no cordierite, the Mg octahedral site occupancy [i.e.,  $Mg/(Mg + Fe + Mn)$ ] must be calculated for a hypothetical cordierite that would be in equilibrium with the other mafic phases at the temperatures and pressures of reaction (1). This was discussed previously (Herzberg, 1978a), and can be calculated from the relation:

$$\frac{(X_{Mg}^{OCT})_{cord}}{(1 - X_{Mg}^{OCT})_{cord}} = \frac{(X_{Mg}^{TET})_{sp}}{0.15(X_{Fe}^{TET})_{sp}} \quad (17)$$

where the value 0.15 is the Fe-Mg partition coefficient of the cordierite-spinel exchange reaction. From terrestrial occurrences and the lunar cordierite-spinel pair (Dymek *et al.*, 1976; Baker and Herzberg, in preparation) this value varies little for  $Mg/(Mg+Fe)$  of spinels greater than about 0.20. That is, this partition is affected little by changes in temperature and pressure. Large errors in this

value of 0.15 result in insignificant errors in the Mg site occupancy of cordierite. For example, using the average spinel analysis for 15445,177 in Table 1A,  $(X_{\text{Mg}}^{\text{OCT}})_{\text{cord}} = 0.96, 0.97,$  and  $0.98$  for the partition coefficients  $0.20, 0.15$  and  $0.10$  respectively.

Inspection of Eq. (11) shows also that the temperature of formation of spinel cataclasites must be known in order to solve for the pressure. The available selenothermometers which permit such temperatures to be estimated are the  $\text{Fe}^{2+}$ -Mg exchange between olivine and spinel [Evans and Frost (1975) minus about  $50^\circ\text{C}$  to correct for  $\text{Fe}^{3+} = 0$  from Medaris (1975)], olivine and orthopyroxene (Sack, 1980), and reaction (2) based on the alumina content of orthopyroxene buffered by olivine and spinel. The latter can be calculated from Eq. (9) by assuming ideal solid solutions and:

$$K_2 = \frac{(X_{\text{Al}}^{\text{M1}})_{\text{opx}} (X_{\text{Mg}}^{\text{OCT}})_{\text{ol}}^2}{(X_{\text{Mg}}^{\text{M1}})_{\text{opx}} (X_{\text{Mg}}^{\text{TET}})_{\text{sp}} (X_{\text{Al}}^{\text{OCT}})_{\text{sp}}^2} \quad (18)$$

It has been suggested that the experimental data of Fujii (1976), which results in a free energy equation similar to Eq. (9), can result in temperature underestimations of as much as  $390^\circ\text{C}$  (Grove *et al.*, 1979) based on experimental results on complex systems similar to spinel cataclasites. The experimental results of Mori (1977) also indicate that temperature underestimates will be made from experimental data in the system  $\text{MgO-Al}_2\text{O}_3\text{-SiO}_2$ . It was shown, however, that experiments on natural complex systems yielded results similar to those in the system  $\text{CaO-MgO-Al}_2\text{O}_3\text{-SiO}_2$  (Mori, 1977). Comparing the data of Mori (1977) to Eq. (9) indicates that such underestimates may be closer to about  $180^\circ\text{C}$  rather than  $390^\circ\text{C}$ . Regardless of the error, it appears that small amounts of Ca in orthopyroxene can substantially reduce the amount of alumina dissolved in orthopyroxene. This will require that  $(\gamma_{\text{MgAl}_2\text{SiO}_6})_{\text{opx}}$  of Eq. (8) be greater than unity. Little systematic data is available to quantify these activity coefficients as functions of composition, temperature, and pressure. Consequently, the temperature underestimates generated from Eq. (9) should be born in mind. This, however, does not create serious difficulties for estimating the depths of spinel cataclasites because the cordierite- to spinel-cataclasite boundary is largely a function of pressure (see Table 1).

Temperature and pressure estimates have been made from mineral chemistry data on 73263,1,11 (Bence *et al.*, 1974) and our data on 15445,177, 72435,30, and the cordierite-bearing spinel cataclasite in 72435,8. The latter has been briefly described by Dymek *et al.* (1976). Detailed petrographic work and extensive microprobe surveys for all three will be reported in Lunar and Planetary Science XI (Baker and Herzberg, in prep.). An example calculation is given in the Appendix for our representative data on 15445,177 listed in Table 1A. It is emphasized that the representative analyses must be truly representative of a statistically significant number of mineral chemistry data points. Only by such exhaustive surveys (e.g., 44 olivine, 24 pyroxene, and 58 spinel analyses of 15445,177) can foreign polymict material be detected (i.e., by Fe-Mg partition criteria; Baker and Herzberg, in prep.), inhomogeneities due to endogenic process (e.g., retro-

grade metamorphic effects) be assessed, and the original precataclasis mineral assemblage be reconstructed. These temperature and pressure estimates are given in Table 1.

Although in most cases the composition of olivine is highly uniform and homogeneous (e.g.,  $\text{Fo}_{92\pm 1}$  from 44 analyses), there is generally a small but significant range in composition of the orthopyroxenes and spinels. These are generally variations in  $\text{Al}/(\text{Al} + \text{Cr})$  in spinel and  $(\text{X}_{\text{Al}}^{\text{M1}})_{\text{opx}}$  which could be due to a number of endogenic processes. These include crystal-liquid interactions of the igneous precursors, kinetic effects of cooling from magmatic temperatures to those of the subsolidus, and interaction of the clasts with impact melt. Accordingly, there are a range of possible temperatures and pressures at which these rocks could have reached partial or complete equilibrium prior to excavation by meteorite impact. However, part of the T-P variation is due to insufficient information on the activity-composition relations of orthopyroxene and spinel. The latter, however, can be assessed on a very crude quantitative level as discussed below; that is,

**Table 1.** Temperature and pressure estimates of various spinel cataclasites

15445,177				
Selenothermometer	T(°C)	Depth (km)		
		I	II	III
1	700–800	≥26	≥26	≥12
2	700	≥26	≥26	≥12
3	810–1020	≥28	≥28	≥12
4	1230–1310	≥34	≥34	≥14

I: from lowest  $\text{Al}_2\text{O}_3$  in opx and highest  $\text{Mg}/(\text{Mg} + \text{Fe})$  and  $\text{Al}/(\text{Al} + \text{Cr})$  in spinel  
 II: from average  $\text{Al}_2\text{O}_3$  in opx and average spinel composition  
 III: from highest  $\text{Al}_2\text{O}_3$  in opx and lowest  $\text{Mg}/(\text{Mg} + \text{Fe})$  and  $\text{Al}/(\text{Al} + \text{Cr})$  in spinel  
 Depth is calculated from the lowest T if a range is indicated.

73263,1,11 (Bence <i>et al.</i> , 1974)				
Selenothermometer	T(°C)	Depth (km)		
		I	II	
1	750–1000	≥28	≥20	
2	750– 900	≥28	≥20	
3	860– 950	≥30	≥22	
4	1250–1290	≥36	≥24	

I: from opx 2 and spinel 11 (Table 5 of Bence *et al.*, 1974)

II: from opx 1 and spinel 10 (Table 5 of Bence *et al.*, 1974)

Depth is calculated from the lowest T where a range is indicated.

Table 1. (Continued)

72435,30			
Olivine (Fo <sub>73</sub> ), Orthopyroxene [Mg/(Mg + Fe) = 0.78, Al <sub>2</sub> O <sub>3</sub> = 3.83 wt %, Ti/Al = .14], Spinel [Mg/(Mg + Fe) = 0.57 - 0.64, Al/(Al + Cr) = .87 - .94], Plagioclase (An <sub>96</sub> )			
Selenothermometer	T(°C)	Depth (km)	
		I	II
1	1000-1200	≥32	≥12
2	800-1200	≥28	≥12
3	680- 810	≥26	≥12
4	1170-1230	≥32	≥12

I: from highest Al/(Al + Cr) and Mg/(Mg + Fe) in spinel  
 II: from lowest Al/(Al + Cr) and Mg/(Mg + Fe) in spinel

72435,8			
Olivine (Fo <sub>73</sub> ), Orthopyroxene [Mg/(Mg + Fe) = .75, Al <sub>2</sub> O <sub>3</sub> = 4 wt %, Ti/Al = 0.11], Cordierite [Mg/(Mg + Fe) = .84], Spinel [Mg/(Mg + Fe) = .45, Al/(Al + Cr) = 0.81], Plagioclase (An <sub>97</sub> ).			
Selenothermometer	T(°C)	Depth (km)*	
		I	II
1	700	-6	-12
2	no solution		?
3	950-1020	-10	-18
4	1290-1310	-16	-24

\* 72435,8 is a univariant mineral assemblage. In principle, a specific T and P can be determined.

I: from lowest Al<sub>2</sub>O<sub>3</sub> in opx

II: from highest Al<sub>2</sub>O<sub>3</sub> in opx

inferences can be made on geological grounds as to whether neglect of the activity coefficients in the calculations [i.e., Eqs. (7) and (8)] will result in pressure under- or overestimates.

Inspection of the results on 15445,177, 73263,1,11 and 72435,30 in Table 1 shows that all the clasts resided at depth in the lunar crust prior to excavation. Generally, a  $\pm 100^\circ\text{C}$  uncertainty in the temperature of equilibration generates an uncertainty of about  $\pm 2$  kilometers in the minimum depth of burial. In most cases, temperatures estimated from the Fe-Mg exchange equilibria are lower than those calculated from the aluminous orthopyroxene-olivine-spinel buffer reaction. This may be due in part to uncertainties in the calibration of these equilibria and the relevant activity coefficients. However, they may also represent real differences in the closure temperatures due to kinetic differences associated with reaction, diffusion, and nucleation. On equilibrium cooling from magmatic temperatures to say  $700^\circ\text{C}$ , in order for the aluminous orthopyroxene to continuously lower its alumina content (see Fig. 1) it must first be in contact with an olivine grain, diffusion must operate, and finally spinel (the recipient of alumina from orthopyroxene) must nucleate and grow. This is a complex process compared to

simple Fe-Mg diffusion between olivine/orthopyroxene and olivine/spinel co-existing pairs.

It should also be noted that the lowest pressure estimates result from use of the most iron- and chrome-rich spinels of any one clast. For the cordierite-bearing sample in 72435,8, which departs furthest from the MgO-Al<sub>2</sub>O<sub>3</sub>-SiO<sub>2</sub> system [i.e., lowest Mg/(Mg+Fe) in mafic phases] *negative* pressures are calculated. On geological grounds this is an absurd solution, which indicates that important activity coefficient terms arise for mineral compositions which depart to such an extent from the system MgO-Al<sub>2</sub>O<sub>3</sub>-SiO<sub>2</sub>. From Eq. (7) it can be shown that:

$$\frac{(\gamma_{\text{Mg}_2\text{Si}_2\text{O}_6})_{\text{OPX}}^5 (\gamma_{\text{MgAl}_2\text{O}_4})_{\text{SP}}^2}{(\gamma_{\text{Mg}_2\text{SiO}_4})_{\text{OL}}^5 (\gamma_{\text{Mg}_2\text{Al}_4\text{Si}_5\text{O}_{18}})_{\text{CORD}}} > 1 \quad (19)$$

and thus  $(\gamma_{\text{Mg}_2\text{Si}_2\text{O}_6})_{\text{OPX}}^5 (\gamma_{\text{MgAl}_2\text{O}_4})_{\text{SP}}^2 > (\gamma_{\text{Mg}_2\text{SiO}_4})_{\text{OL}}^5 (\gamma_{\text{Mg}_2\text{Al}_4\text{Si}_5\text{O}_{18}})_{\text{CORD}}$ . Indeed, a ratio of about 1.4 to 1.8 of Eq. (19) would result in lunar surficial pressures. If 72435,8 was in fact a deep-seated rock, the above activity coefficient ratio would have to be higher still. This result is consistent with expectations drawn from activity coefficients arising from reciprocal lattice effects in solid solutions containing two or more energetically nonequivalent sites (i.e., orthopyroxenes and spinels; Wood and Nicholls, 1978; Sack, 1980). That is, activity coefficients for orthopyroxene and spinel are likely to be significantly different from those for olivine and cordierite in which the cations Fe-Mg mix in two sites which are energetically similar.

The results for 72435,8 thus provide important information for assessing the error in depth estimates for the other spinel cataclasites. Because the mineral compositions in 15445,177, 73263,1,11 and 72435,30 approximate more closely the simple system MgO-Al<sub>2</sub>O<sub>3</sub>-SiO<sub>2</sub> than 72435,8, their ratios of activity coefficients in equation (19) should be less than 1.4 to 1.8, but remain greater than unity. This indicates that the minimum pressure estimates given in Table 1 are also underestimated somewhat, perhaps by 0 to 10 kilometers. We suggest, therefore, that the depths of  $\geq 12$  kilometers for 15445,177 determined from the most iron- and chrome-rich spinels are much too low, and the values  $\geq 26$  kilometers from the most aluminous spinels are more realistic. This is consistent with the most aluminous spinels being volumetrically the most abundant in 15445,177 (Baker and Herzberg, in prep.). Unfortunately, an error propagation analysis is precluded because sufficient information is not available on the activity-composition relations. However, with a suggested error of  $\pm 0.2$  kbar in the T-P location of reaction (1), the minimum pressure estimates in Table 1 may be accurate to about  $\pm 4$  km + (0 to 10 km).

### THE BEARING OF SPINEL CATACLASITES ON THE STRUCTURE OF THE LUNAR CRUST AND MANTLE

With the possible exception of the cordierite-bearing clast in 72435,8, the remainder of the samples considered appear to have had a deep-seated origin in the crust and/or mantle. Because only minimum pressure estimates can be made, it

is not possible to be more specific about their preexcavation stratigraphic location. That is, they may have resided in the middle or the deepest parts of the crustal column.

Although an uppermost mantle origin for some cannot be established from these equilibria alone, it also cannot be ruled out. Clasts in 15445 consist of olivines with forsterite contents averaging 92 (this work and review in Ryder and Bower, 1977) and modal proportions ranging from 40 to 61%. Section 15445,177 consists of the mode: 15% plagioclase, 44% olivine, 6–7% spinel, 2–3% orthopyroxene and 32% matrix material. Judging from the heterogeneities observed in other clasts from 15445 (Ryder and Bower, 1977), this mode is probably not representative of the entire clast. Regardless of this variation, these are some of the most ultramafic highland rocks yet reported.

Apollo 17 spinel cataclasites are generally more plagioclase-rich and lower in  $Mg/(Mg+Fe)$ . For 72435,8,30, the mineralogical proportions are: 83–89% plagioclase, 2–7% olivine ( $Fe_{73-75}$ ), 2–11% spinel, and 2–6% orthopyroxene. Bence and McGee (1976) have reported a mode of 57–67% plagioclase, 25–27% olivine ( $Fe_{90}$ ), 2–8% orthopyroxene, and 5–7% spinel with trace amounts of ilmenite and metal for two sections of 73263.

On the plot of  $Mg/(Mg+Fe)$  in mafic minerals and anorthite content of plagioclase shown in Fig. 2, the spinel- and cordierite-cataclasites which have been reported are located at the Mg-rich extremities of the Mg-rich and anorthositic suites. It is possible that their restricted distributions are a simple artifact of the limited number of samples that have been reported; future work on other clasts may show a broad distribution of spinel cataclasites in each suite. It should also be noted that the gap between the two trends is maintained.

Placing the spinel cataclasites in a stratigraphic context can be fraught with difficulties because it is not known how representative they are at depth in the crust. Indeed, they may be interleaved with other highland rocks of greater abundance, including pyroxene-free spinel troctolites. Their lateral distribution is also unknown. They may be of global extent, or they may be confined to the crustal columns below the Imbrium and Serenitatis Basins. It is also not yet possible to establish if there is any relationship between depth estimates for spinel cataclasites and the other members of the Mg-rich and anorthositic suites of Fig. 2. However, with all these uncertainties in mind, we propose the stratigraphic model shown in Fig. 3.

The fundamental assumption in the construction of this model is that the members of the two trends of Fig. 2 represent samples from two distinct lithologic units (e.g., Warren and Wasson, 1977), and their depths within each are reflected in their values of  $Mg/(Mg+Fe)$ . Figure 3 is similar in overall features to models which have been proposed previously (Ryder and Wood, 1977; Warren and Wasson, 1977) in that an anorthositic layer overlies a layer consisting of members of the Mg-rich trend. We would suggest that the anorthositic layer is between 12 and 20 kilometers thick, based on the depth values given in Table 1. That is, if spinel cataclasite in 72435,30 is a member of the anorthositic suite as Fig. 2 suggests, it must have been located at a depth greater than or equal to about 12 kilometers. However, spinel cataclasite in 73263,1,11, being a member of the

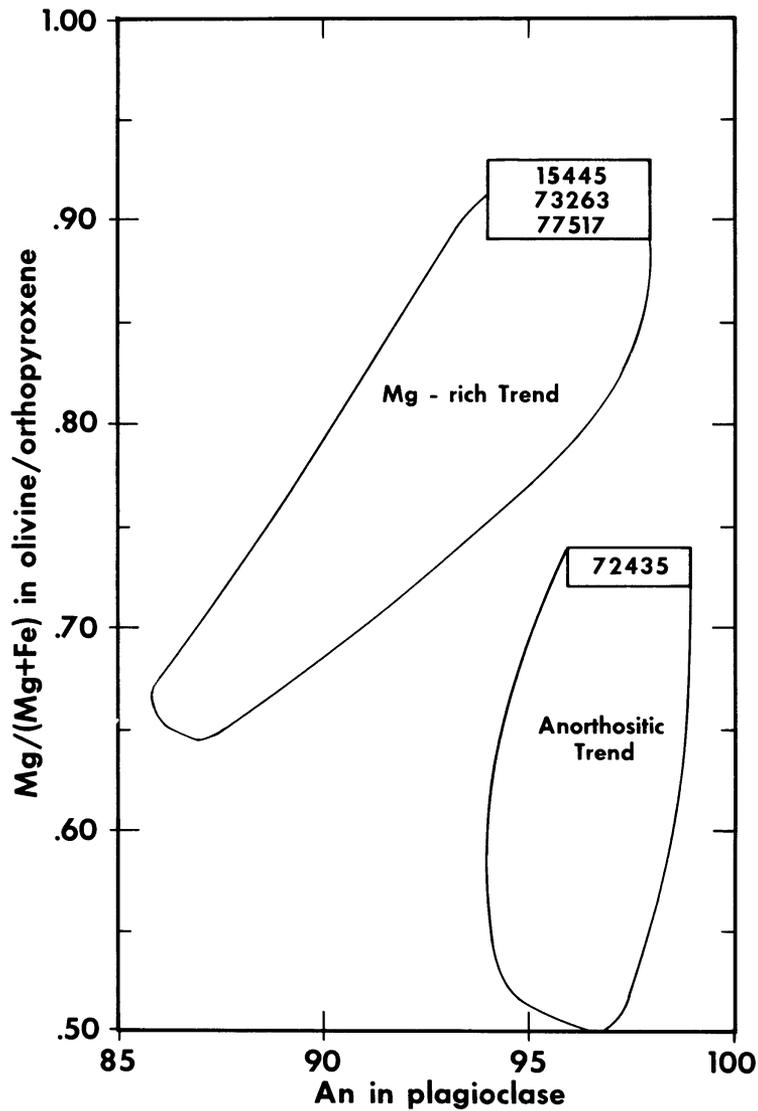


Fig. 2. Mg/(Mg+Fe) in mafic minerals vs. An content of plagioclase for pristine nonmare lunar rocks, after Warren (1979). Analyses for spinel cataclasites in 15445 and 72435 are from Baker and Herzberg (in preparation). Data for 73263 and 77517 are from Bence *et al.* (1974) and Warner *et al.* (1978).

lower Mg-rich series, was located at a depth greater than or equal to about 20 kilometers. Since the equilibria from which these estimates are made provide only minimum possible depths in the crust, the 12 to 20 kilometer range must also be considered a *minimum* thickness range for the anorthositic layer. Indeed, the seismic discontinuities at about 20 kilometers depth modelled by Goins *et al.* (1979) could reflect the interface of this two layer crust. That this interface is sharply defined rather than gradual is implied in Fig. 2. Accordingly, the boundary between these two layers given in Fig. 3 shows abrupt changes in the mineral chemistries and mineralogical proportions.

If the anorthositic layer exhibits stratigraphic variations in its mineralogical

abundances, we would suggest that 72435,30, being the most primitive member of the anorthositic series (Fig. 2), and having about 83–89% plagioclase, may constrain these variations to those shown in Fig. 3. That is, there may be a strong increase in the plagioclase content upward from the base of the anorthositic layer. This would be accompanied by the iron enrichment trend as given in Fig. 2. An iron-rich spinel horizon may or may not be present at the base of this layer, depending on how representative 72435 is. As with the Middle Banded Zone of the Stillwater Complex (Raedeke and McCallum, 1979), the anorthositic layer may consist of a number of lithologies, including norites and troctolites. Thus the actual volume of anorthosite *sensu stricto* may be less than that of the anorthositic layer itself. Heterogeneities and exceptions to the simplicity of Fig. 3 must be expected. Indeed, it is possible that the dunite 72415 was contained in the anorthositic layer, based on other equilibria discussed elsewhere (Herzberg, 1979).

The Mg-rich layer may constitute the largest volume fraction of the lunar crust, and consist of many lithologies whose minerals may be roughly in cotectic proportions. Plagioclase may occupy about 40–60% of this layer, with its volume fraction decreasing (perhaps highly irregularly) toward the crust-mantle boundary. Troctolites and norites, with perhaps their major element buffered low-K Fra Mauro and KREEP compositions (Walker *et al.*, 1979), may be the dominant lithologies of this lower layer, and hence the lunar crust as a whole. If the depth of excavation is reflected inversely in the abundance of a lithology and plagioclase content (a crude criterion because of polymict processes), the pleonaste spinel rocks, being relatively rare and having some of the lowest plagioclase contents, would be located at the deepest stratigraphic levels (e.g., 15445,177). Values for

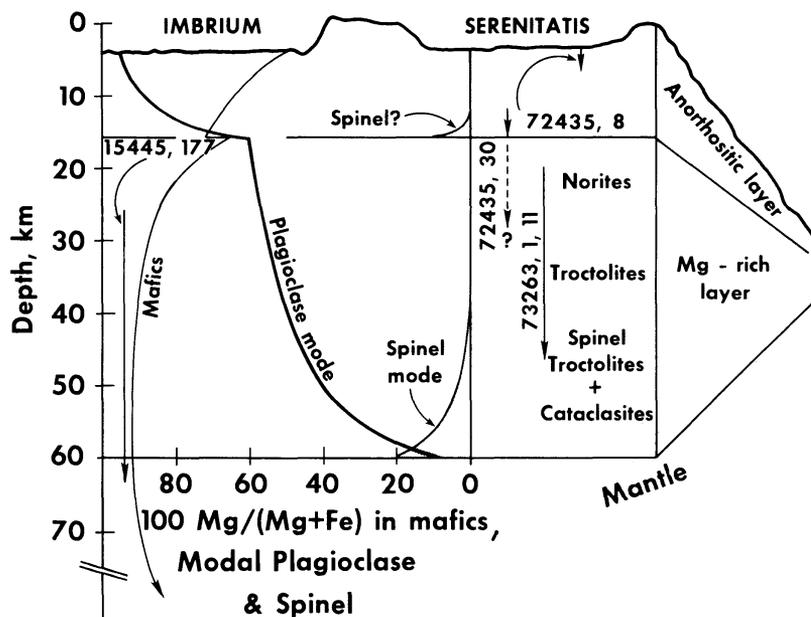


Fig. 3. Some possible stratigraphic features of the lunar crust and mantle. Vertical bars with spinel cataclasite clast numbers indicate their possible stratigraphic locations prior to excavation by meteorite impact.

Mg/(Mg+Fe) could range from 0.90–0.92 over a considerable thickness of the lower crust. As pointed out elsewhere (Walker *et al.*, 1975; Herzberg, 1978a), Mg/(Mg+Fe) would then have to decrease from the crust-mantle boundary to the depths of the mantle in order for there to have been a suitable source region for mare basalts.

## CONCLUSIONS

Nodules in some basaltic rocks and kimberlites represent a stratigraphy of the Earth's crust and upper mantle. Similarly, pristine highlands clasts of a plutonic nature represent stratigraphic sections of the lunar crust, and possibly the upper mantle. Obviously, the mode of transport of these deep-seated materials to the surfaces of each planet were different, being volcanic in the case of the Earth, and excavation by meteorite impact for the Moon.

Of the plutonic pristine highlands rocks, which are commonly the norites, troctolites ( $\pm$ spinel), and anorthosites, only the spinel- and cordierite-cataclasites (olivine + plagioclase + high alumina orthopyroxene  $\pm$  spinel  $\pm$  cordierite) exhibit phase changes and considerable variations in their mineral chemistry at depths appropriate to the lunar stratigraphic column. These have been calibrated by experiment in the system MgO-Al<sub>2</sub>O<sub>3</sub>-SiO<sub>2</sub> and applied to the natural lunar spinel- and cordierite-cataclasites by solution thermodynamics.

The results show that the cordierite + spinel cataclasite in 72435,8 was probably of near-surface origin, whereas spinel cataclasites in 72435,30, 73263,1,11 and 15445,177 were located at depths greater than or equal to about 12 to 32 kilometers. For the cordierite-free types, only *minimum* depths of burial prior to excavation by meteorite impact can be determined. This precludes the possibility of unequivocally identifying lunar upper mantle rocks from the equilibria of concern here; however, the ultramafic spinel cataclasite clast in 15445,177, which resided at depths greater than or equal to about 26 kilometers, is a logical candidate.

The highly magnesian spinel cataclasites in 15445 and 73263 are associated with troctolites and norites of the Mg-rich series of rocks in a plot of Mg/(Mg+Fe) in mafic minerals vs. An content of plagioclase. However, the Fe-rich spinel- and cordierite-cataclasites in 72435 are associated with the anorthositic members of this plot. All plot at the most Mg-rich extremities of both series. If it is assumed that the members of the two series represent samples from two distinct lithologic units of the lunar crust, and their depths within each are reflected in their values of Mg/(Mg+Fe), a two layer model of the crust can be modelled. The uppermost stratigraphic unit consists of the members of the anorthositic series, and is probably about 12 to 20 kilometers thick. The lower Mg-rich unit may constitute the greatest volume of the crust, and consist of lithologies with roughly cotectic mineralogical proportions (i.e., norites, troctolites, spinel troctolites, and their buffered low-K Fra Mauro and KREEP compositions). The boundary separating the two stratigraphic units may be sharp, rather than gradational.

**Acknowledgments**—Colin M. Graham at Edinburgh University and the Natural Environment Research Council of Britain are thanked for generously providing the experimental facilities. Part of this research was financed by NASA grant NGL 09-015-150 to J. A. Wood, and the remainder by the Lunar and Planetary Institute which is operated by the Universities Space Research Association under Contract No. NSR 09-051-001 with the National Aeronautics and Space Administration. Thanks go to Graham Ryder, Lewis Ashwal, Charles Bickel, and Jeffrey Warner for their reviews of the manuscript. Bob Newton from the University of Chicago is also thanked for his critical review of another manuscript (Herzberg, in preparation) within which the experimental data and thermodynamic analysis are discussed in detail. All support is gratefully acknowledged. This paper is Lunar and Planetary Institute Contribution No. 402.

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### Appendix

Representative mineral analyses in spinel cataclasite 15445,177 are given in Table 1A. The average orthopyroxene has the following structural formula:

M2	M1	TET
0.899 Mg	0.828 Mg	1.883 Si
0.077 Fe <sup>2+</sup>	0.071 Fe <sup>2+</sup>	0.115 Al
0.020 Ca	0.080 Al	
0.004 Mn	0.012 Ti	
	0.012 Cr	

Therefore,  $(X_{Mg}^{M2})_{opx} = 0.899$ ,  $(X_{Mg}^{M1})_{opx} = 0.828$ , and  $(X_{Al}^{M1})_{opx} = 0.080$ .

Determining the temperature at which this orthopyroxene was in equilibrium with olivine and the average spinel requires calculating  $K_2$  from Eq. (18). Thus

$$\begin{aligned}
 K_2 &= \frac{(X_{Al}^{M1})_{opx}(X_{Mg}^{OCT})_{ol}^2}{(X_{Mg}^{M1})_{opx}(X_{Mg}^{TET})_{sp}(X_{Al}^{OCT})_{sp}^2} \\
 &= \frac{(0.080)(0.924)^2}{(0.828)(0.833)(0.920)^2} \\
 &= 0.117.
 \end{aligned}$$

Substituting now into Eq. (9), T becomes 1145°K (872°C) at 1000 bars, and 1148°K (875°C) at 2000 bars. Using this temperature estimate, a pressure can now be determined by calculating first  $K_1$  of Eq. (13), and substituting into Eq. (11).

In order to calculate  $K_1$ , the hypothetical cordierite composition must be known. From Eq. (17) and the average spinel analysis:

$$(X_{Mg}^{OCT})_{cord} = 0.97$$

and thus

$$\begin{aligned}
 K_1 &= \frac{(X_{\text{Mg}}^{\text{M2}})_{\text{opx}}^5 (X_{\text{Mg}}^{\text{M1}})_{\text{opx}}^5 (X_{\text{Mg}}^{\text{TET}})_{\text{sp}}^2 (X_{\text{Al}}^{\text{OCT}})_{\text{sp}}^4}{(X_{\text{Mg}}^{\text{OCT}})_{\text{ol}}^{10} (X_{\text{Mg}}^{\text{OCT}})_{\text{cord}}^2} \\
 &= \frac{(0.899)^5 (0.828)^5 (0.833)^2 (0.920)^4}{(0.924)^{10} (0.97)^2} \\
 &= 0.27 \\
 \ln K_1 &= -1.3.
 \end{aligned}$$

Equation (11) now appears numerically:

$$\begin{aligned}
 P &\geq \frac{(1.987)(1148)}{-1.43} (-0.350 - 1.839 + 1.3) \\
 &\geq 1400 \text{ bars} \\
 &\geq 28 \text{ kilometers (i.e., where 1000 bars = 20 kilometers).}
 \end{aligned}$$

**Table 1A.** 15445,177: Representative mineral analyses<sup>1</sup>

			Orthopyroxenes			Spinel		
	Plagioclase	Olivine	Lowest-Al	Average-Al	Highest-Al	Lowest Cr/Cr+Al	Average Cr/Cr+Al	Highest Cr/Cr+Al
SiO <sub>2</sub>	44.22	41.00	55.38	54.54	54.30	0.07	0.09	0.25
Al <sub>2</sub> O <sub>3</sub>	35.95	0.02	4.09	4.80	6.03	62.90	61.86	57.91
TiO <sub>2</sub>	0.09	0.07	0.45	0.51	0.46	0.05	0.05	0.01
FeO	0.13	7.40	5.11	5.13	5.05	7.58	7.74	8.70
MnO	0.00	0.10	0.07	0.14	0.08	0.13	0.11	0.20
MgO	0.05	51.42	33.68	33.59	33.14	22.05	22.34	21.20
CaO	19.15	0.02	0.54	0.52	0.57	0.04	0.04	0.09
K <sub>2</sub> O	0.03	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.
Na <sub>2</sub> O	0.38	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.
Cr <sub>2</sub> O <sub>3</sub>	n.a.	0.00	0.37	0.40	0.52	7.32	7.89	11.97
Total	100.00	100.03	99.68	99.63	100.15	100.14	100.10	100.33
Si	2.043	.992	1.906	1.883	1.863	.004	.004	.008
Al	1.957	.000	.168	.195	.242	1.859	1.836	1.746
Ti	.004	.000	.012	.012	.012	.000	.000	.000
Fe	.004	.148	.148	.148	.145	.160	.164	.188
Mn	.000	.004	.004	.004	.004	.004	.004	.004
Mg	.004	1.855	1.730	1.727	1.695	.824	.840	.809
Ca	.949	.000	.020	.020	.020	.000	.000	.004
K	.000	—	—	—	—	—	—	—
Na	.035	—	—	—	—	—	—	—
Cr	—	.000	.012	.012	.016	.145	.156	.242
Total	4.996	3.000	4.000	4.000	3.996	2.996	3.004	3.000

<sup>1</sup> Oxides in wt. %; formula units based on eight oxygens for plagioclase, six for pyroxene, and four for olivine and spinel.

n.a. = not analyzed.