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Contact metamorphism on 4 Vesta and the Petersburg polymict eucrite

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Abstract: Using equations for thermal diffusion, we model the temperature profiles in metamorphic aureoles surrounding 1 km and 3 km diameter spherical intrusions containing eucrite and diogenite magmas in the crust of 4 Vesta. The aureoles created by thermal diffusion experience temperatures significantly elevated above the regional thermal gradient for 10^3 to 10^4 years, which is comparable to the period of time required for the intrusions to crystallize. Temperatures of wall rock adjacent to a magma body intruded at mid-crustal levels can approach those in the lower part of the crust. In contrast, it has previously been suggested that thermal processes and, by implication, regional metamorphism could have affected 4 Vesta for a period of 10^6–10^8 years. Heating of underlying material by a lava flow or layer of impact melt of moderate thickness would have lasted for a period of 1–10 years. In an effort to obtain approximate cooling rates, we numerically fit theoretical diffusion profiles to the observed compositional profile of a 30-μm-wide Fe-rich alteration rim at the edge of a fragment of diogenitic orthopyroxene in the Petersburg polymict eucrite, which was affected by late-stage metamorphism. Using two sets of diffusion coefficients that are similar at 800°–1000℃, but very different at lower temperatures (*e.g.*, 500℃), the calculated diffusion profiles that most closely fit the compositional profile suggest cooling rates ranging from 0.12℃/yr (a period of ~4000 years) to 0.25℃/yr (~2000 years). These cooling rates are very similar to those expected for contact metamorphism, and, hence, the Petersburg meteorite may represent a polymict eucrite breccia derived from near an intrusive magma body. However, this interpretation is not unique as this alteration profile might also have been formed by regional metamorphism at low temperatures.

key words: contact metamorphism, polymict eucrite, Fe-Mg interdiffusion, orthopyroxene, 4 Vesta

1. Introduction

Volcanic and igneous rocks of the HED (howardite, eucrite, and diogenite) suite of meteorites display evidence of varying degrees of metamorphism, which previously has been attributed to either long-term regional metamorphism caused by an asteroidal thermal gradient (Yamaguchi *et al.*, 1996, 1997) or short-term metamorphism caused by burial of surface materials beneath a lava flow or a layer of hot impact ejecta (see Labotka and Papike, 1980; Fuhrman and Papike, 1981). In contrast, metamorphic processes on Earth also include contact metamorphism, heating of surrounding rock by intrusive magma bodies. Considering that heating of the asteroid 4 Vesta, the probable HED parent body (*e.g.*, McCord *et al.*, 1970; Gaffey *et al.*, 1989; Binzel and Xu, 1993), might have lasted as long as 100×10^6years (see Ghosh and McSween, 1998), similar processes possibly affected some HED materials. Characterizing the possible effects of contact metamorphism is particularly relevant, because the cumulate eucrites (*e.g.*, Moore County; Drake and Hostetler, 1978) display evidence of being the product of fractional crystallization of eucritic magma in small intrusive bodies (see Wilson and Keil, 1996). Time scale of contact metamorphism associated with such bodies would most certainly be intermediate between those of regional metamorphism and metamorphism associated with a surface lava flow or layer of impact melt. The goal of the present study is to mathematically describe the thermal effects of such contact metamorphism.

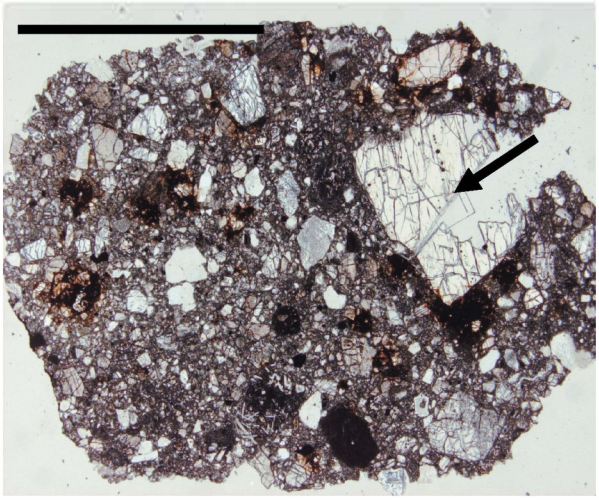
We also consider whether any previously described HED meteorites display mineralogical features consistent with this type of metamorphism. Several features suggest that the polymict eucrite Petersburg experienced intermediate degrees of post-brecciation thermal metamorphism that caused partial equilibration. Podosek (1972) and Podosek and Huneke (1973) determined that ^{39}Ar release patterns for Petersburg are similar to those expected for diffusive loss, possibly caused by metamorphism. Hewins (1979) suggested that compositions and textures of metal in the matrix of the meteorite were the result of metamorphism. Buchanan and Reid (1996) noted that the edge of a fragment of magnesian diogenitic orthopyroxene, which was 1.0 mm in diameter, was altered to more Fe-rich compositions to a distance of ~30 μm (Fig. 1). This alteration was the result of Mg-Fe interdiffusion between the brecciated matrix of the meteorite and the orthopyroxene fragment. Pyroxene fragments that are smaller in diameter than 50–60 μm in the brecciated matrix of the meteorite also have compositions that suggest equilibration with the bulk composition of the meteorite (Buchanan and Reid, 1996). For a more detailed petrographic and petrologic description of Petersburg, the reader is referred to Buchanan and Reid (1996). Similar metamorphic effects have been documented in the HED breccias Yurtuk and Frankfort (Labotka and Papike, 1980).

Hewins (1979) suggested that the metamorphism experienced by Petersburg resulted from deep burial beneath a lava flow or a hot layer of impact ejecta. Although this interpretation certainly is possible, Yamaguchi *et al.* (1996) noted that centers of 5 m thick lava flows cool to ambient temperatures in ~1 year. Hence, contact metamorphism is possibly a better candidate for this partial equilibration, because it is doubtful that such extensive metamorphism could have occurred in such a short period of time. However, it's also possible that long periods of shallow regional metamorphism at low temperatures might cause similar effects. Another goal of this study is to consider whether one of these mechanisms or both could have caused the partial equilibration of Petersburg.

2. Thermal diffusion calculations

To accomplish these goals, we first calculate, based on equations from Lovering

(a)



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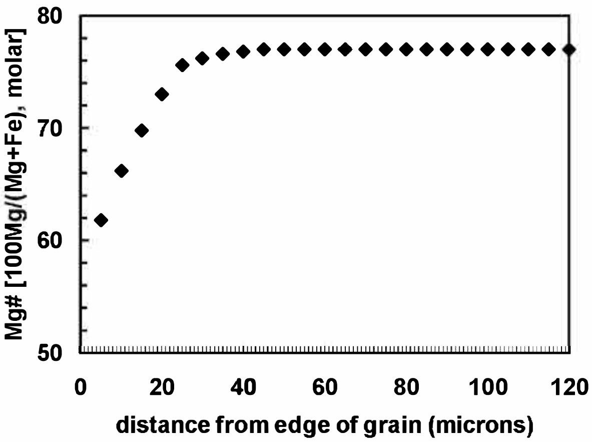
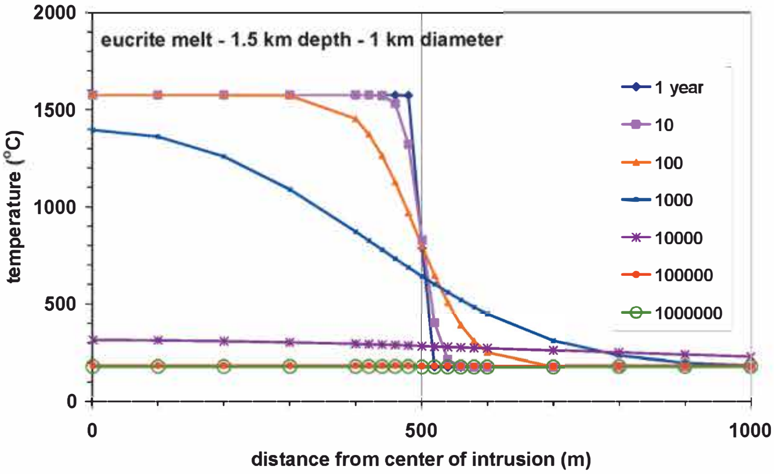


Fig. 1. (a) Photomicrograph taken with transmitted light of a thin section of the breccia of Petersburg with the large fragment of diogenitic orthopyroxene (arrow). Scale bar is 1.0 mm. (b) Traverse indicating compositional profile (Mg#) from edge to interior of grain in (a).

(1935) and Jaeger (1964), the thermal effects associated with the intrusion of medium-sized, late-stage magma bodies into a eucritic host that is probably similar to the crust of 4 Vesta (Fig. 2). In this study, we do not attempt to model the thermal effects associated with a large magma ocean with thickness comprising a majority of the crust (*e.g.*, 5 km of a 10 km thick crust). We instead limit our calculations to smaller 1 to 3 km diameter bodies, which are the most plausible for late-stage magmatism. We also do not consider the metamorphic effects associated with intrusion of thin dikes into the crust of 4 Vesta, because that subject was previously discussed by Wilson and Keil (1996).

Obviously, very little information is available about the shape of any late-stage



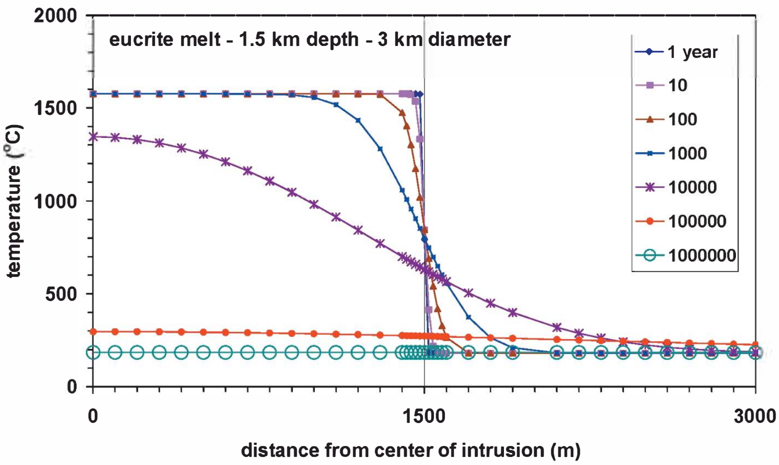
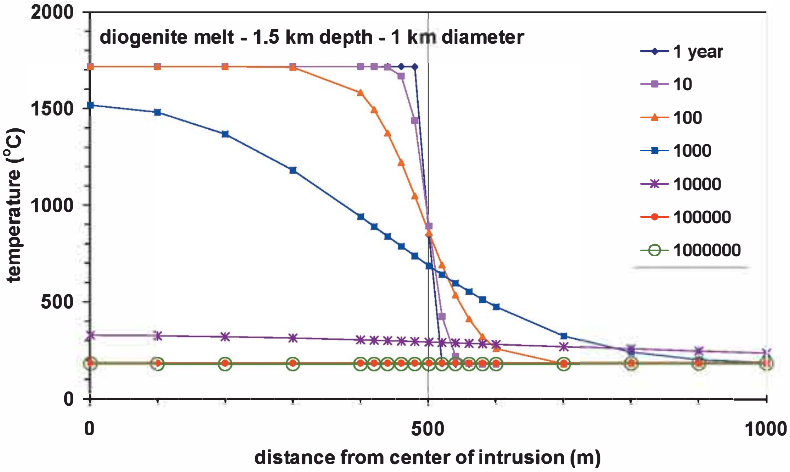


Fig. 2. Calculated models of the temperature effects in the metamorphic aureoles surrounding magma chambers that are modeled in this manuscript. Details of calculations are included in the text. Note: spherical magma chambers with 1 and 3 km diameters have radii of 0.5 and 1.5 km, respectively.

magma bodies on 4 Vesta. However, we assume spherical shape, because these calculations provide symmetrical temperature profiles and spherical bodies cool more slowly than other shapes. Length of time required for an intrusion to cool is dependent, among other factors, on the surface area/volume ratio, which is lowest for a sphere. Hence, the calculations are made for a spherical magma body contained in a uniform, infinite space with a constant initial temperature. We consider two compositional types of magma: eucrite (peritectic pyroxene-plagioclase basalt) and diogenite (orthopyroxene-normative). Initial temperature of the eucrite magma is assumed to be 1200°C, whereas the diogenite magma is assumed to have a higher initial temperature of 1300°C



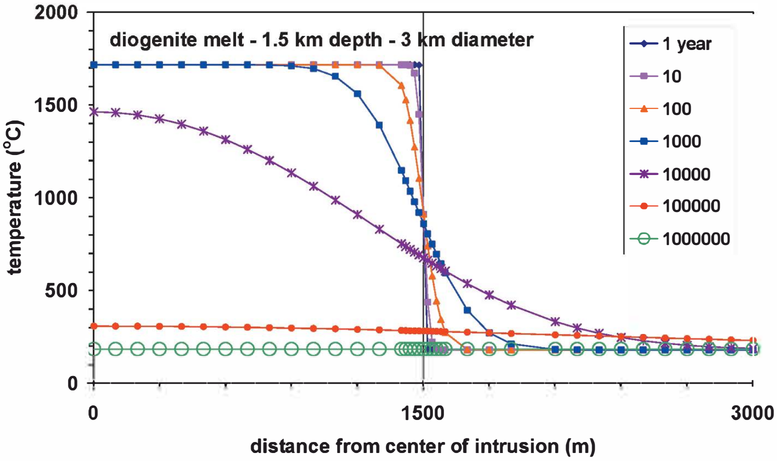


Fig. 2 (continued).

(Jurewicz *et al.*, 1993).

We assume for the crust of 4 Vesta a thickness of 10 km and a thermal gradient that is linear with depth from 0°C at the surface to 1200°C at the base of the crust (see also Yamaguchi *et al.*, 1996, 1997). One set of calculations is made for relatively shallow crustal levels of 1.5 km depth, which is the maximum depth possible for a spherical intrusion of 3 km diameter. In this case, a thin layer is assumed to insulate the magma chamber from the surface of the asteroid and the temperature of the wall rock is assumed to be 180°C. A second set of calculations is made for magma chambers located at an intermediate crustal depth, 5 km in a 10 km thick crust, and the temperature of the wall rock is assumed to be 600°C. We do not attempt to model a chamber at the bottom of the crust, because the assumed initial temperatures of the magmas (1200°C and 1300°C) are not very different from the temperature assumed at the base of

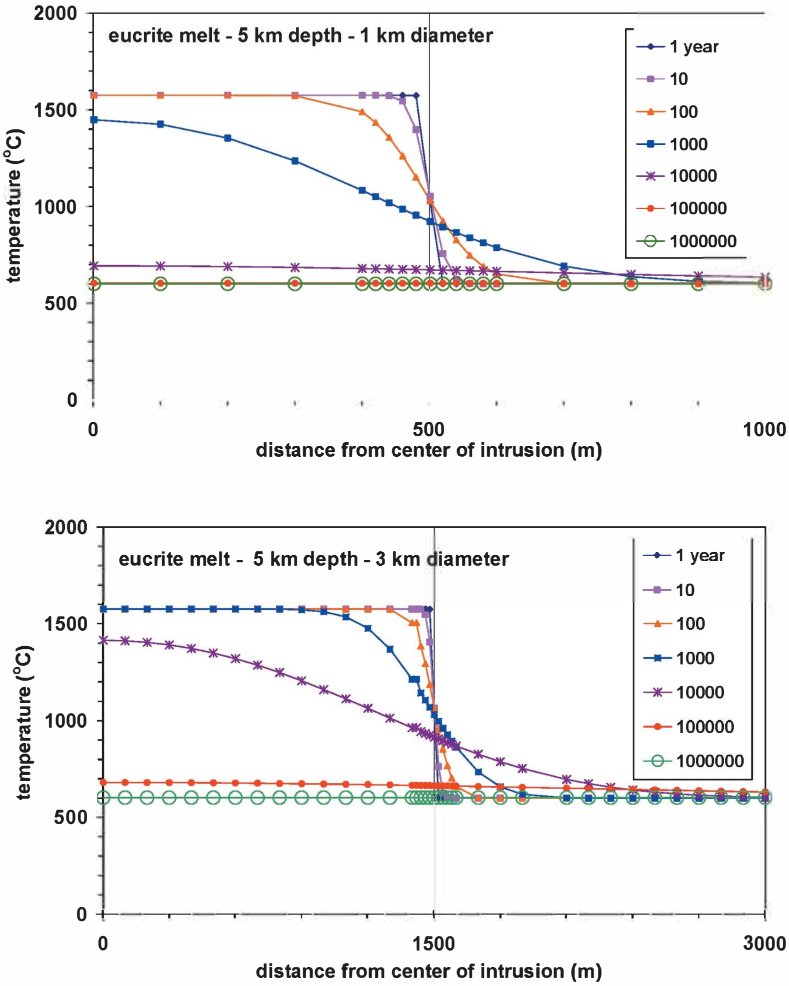
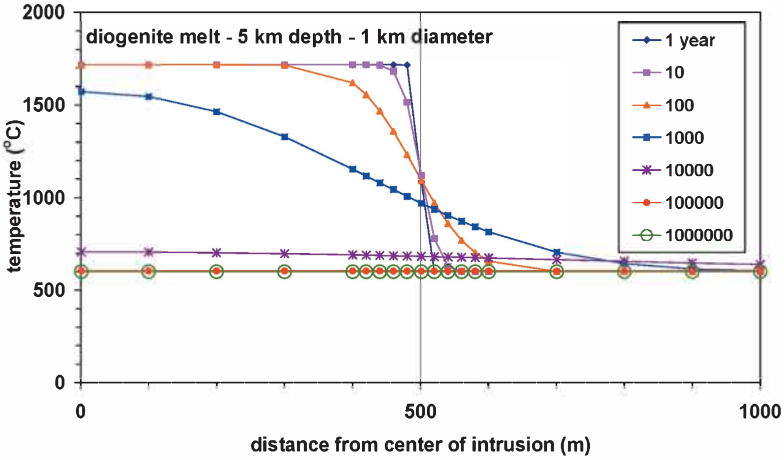


Fig. 2 (continued).

the crust (1200°C) (Yamaguchi *et al*., 1996, 1997). Temperature of the crust over the vertical height of each magma chamber was considered to be constant.

Calculations are based on the equation for thermal diffusion from a spherical body of radius r with an initial temperature θ_0 and an initial wall rock temperature of θ_w. Variations of this equation are found in Lovering (1935) and Jaeger (1964). Thermal diffusivity of the wall rock, h^2, is assumed to be 0.007 cm^2/s (Yamaguchi *et al*., 1997), which is appropriate for eucritic crystalline rock. Horai and Winkler (1979, 1980) determined that thermal diffusivity of brecciated versus unbrecciated lunar rock is affected by the abundance of microcracks. Hence, this value is probably also appropriate for lithified eucrite and howardite breccias. Nevertheless, we varied the thermal diffusivity from 0.01 to 0.001 cm^2/s for selected calculations and found that, because the square root of the variable is used, differences over this range are a maximum of an order



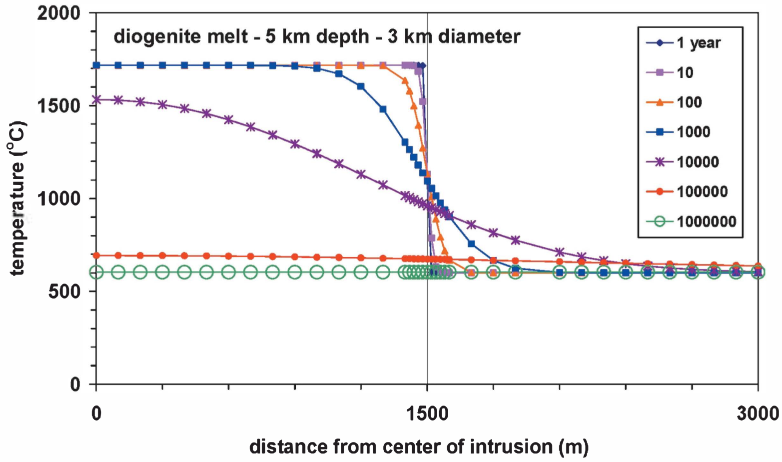


Fig. 2 (continued).

of magnitude and are small compared to differences between contact metamorphism and the other types of metamorphism (see also Yamaguchi *et al*., 1996, 1997).

For the temperature θ_{xt} at distance x from the center of the intrusion and at time t

$$\theta_{xt} = \theta_{\mathrm{w}} + \frac{\theta_0}{2}\left[I(a) + I(b) - \frac{2h\sqrt{t}}{x\sqrt{\pi}}\left(e^{-a^2} - e^{-b^2}\right)\right],$$

where $a = \dfrac{r-x}{2h\sqrt{t}}$, $b = \dfrac{r+x}{2h\sqrt{t}}$, and the function $I = \dfrac{2}{\sqrt{\pi}}\displaystyle\int_0^y e^{-y^2} dy$.

A variety of time periods subsequent to intrusion are considered, ranging from 1 year to 1 million years. Latent heat of crystallization is considered by approximating the equivalent initial temperature (T_0^*) as

$$T_0^* = T_0 + \frac{L}{c},$$

where T_0 is the initial temperature of the magma, L is latent heat of crystallization, and c is specific heat (Jaeger, 1964). Based on the discussion in Jaeger (1964), L for the eucrite melt is assumed to be 90 cal/g. For the diogenite (orthopyroxene-normative) melt, L is assumed to be 100 cal/g. The specific heat of both magmas is assumed to be 0.24 cal/g·°C (see, for example, Goranson, 1942, for a discussion).

Assimilation of wall rock adjacent to magma chambers is not considered. For a eucritic magma, calculations suggest that, despite consideration of latent heat of crystallization, temperatures of wall rock immediately adjacent (within 1 mm) to a 3 km diameter magma chamber at intermediate crustal levels will reach a maximum of 1087 °C. For a diogenitic melt in the same magma chamber, the maximum temperature in the adjacent wall rock will only reach ~1156°C. As the liquidus of a basaltic rock is close to 1200°C (Thy *et al*., 1999), only limited melting will occur. It must also be noted that assimilation of eucritic wall rock into a magma chamber containing eucritic or diogenitic magma will be difficult or impossible to recognize, because compositions of wall rock and magma are similar.

It must also be noted, however, that this method has limitations for calculations of temperatures close to the contact between the magma chamber and the wall rock; limitations for similar temperature calculations for sills and dikes are discussed in Jaeger (1964). The above equation has the characteristic of over-estimating the temperature at the contact between the magma chamber and the wall rock (Jaeger, 1964) and over-estimating the temperature within the magma chamber. In an actual magma chamber, the latent heat of crystallization is released over an extended period of time. Rather than a higher initial temperature, which has been modeled in these calculations, the latent heat of crystallization will continuously be released during the complete crystallization interval.

Temperatures at the contact between a 1 km diameter eucrite intrusion at a depth of 1.5 km and the surrounding wall rock are estimated to have reached a maximum of ~800°C. In comparison, the magma chamber represented by the Bushveld Complex of South Africa contained basalt to basaltic andesite liquids, compositionally similar to the eucrites (see discussion and references in Eales, 2002). Strata underlying this magma chamber include sedimentary rocks and volcanic rocks, which range from basalts to dacites (Buchanan *et al*., 2002). Engelbrecht (1990) and Johnson *et al.* (2003) estimate that the country rocks below the Bushveld Complex experienced temperatures of 750°–760°C. Hence, our calculations provide estimates similar to those suggested for the metamorphic aureoles surrounding terrestrial intrusions of similar compositions and emplacement environment. For the 1 km diameter magma chamber modeled by our calculations, wall rock experiences these maximum temperatures for hundreds of years. A larger volume of wall rock to a distance of 100 m from this magma chamber experiences temperatures of 400°–600°C for thousands of years, the approximate length of time required for complete crystallization of a magma chamber of this size. In contrast, maximum temperatures closer to 900°C are experienced for hundreds of years by wall rock immediately adjacent to a 1 km diameter orthopyroxene-normative magma body intruded at a depth of 1.5 km. Wall rock experiences temperatures between 450°C and 650°C as far as 100 m from this magma chamber for thousands of years.

In contrast, the maximum temperatures for wall rock in contact with a eucritic magma body of a similar diameter at intermediate crustal depths are as high as 1050°C for a period of hundreds of years. Wall rock to a distance of 100 m from this magma chamber experiences temperatures of 800°–900°C for a period of thousands of years. Maximum temperatures adjacent to an orthopyroxene-normative magma body of the same diameter at the same depth are closer to 1100°C for hundreds of years, whereas temperatures of 800°–950°C are experienced by wall rock as far as 100 m from the magma chamber for thousands of years. Hence, ironically, the effects associated with possible contact metamorphism on 4 Vesta are probably most significant at these intermediate depths. A combination of heating due to the regional thermal gradient and thermal effects associated with an intrusive magma body might raise the temperature of the wall rock from ~600°C to 800°–950°C. However, these increased temperatures only occur for periods of 10^3 to 10^4years, rather than the much longer periods of time associated with regional metamorphism (see Ghosh and McSween, 1998). Calculations were also made for magma chambers of 3 km diameter; these calculations also suggest time periods of 10^3 to 10^4 years (see Fig. 2).

3. Fe-Mg interdiffusion calculations for orthopyroxene

In an effort to determine whether the alteration zone at the edge of the fragment of orthopyroxene in Petersburg could have been caused by contact metamorphism, we calculated the cooling rate and time period that might generate a diffusion profile similar to the compositional zoning. Critical to this evaluation is whether this time period is similar to that suggested by the above calculations for contact metamorphism adjacent to a late-stage intrusion. Orthopyroxene is a common mineral both in terrestrial rocks and in meteorites, but there are fewer studies about Fe diffusion properties of orthopyroxene than those of olivine (*e.g.*, Miyamoto *et al.*, 2002). Miyamoto and Takeda (1994) and Ganguly and Tazzoli (1994) provided relevant Fe diffusion coefficients for orthopyroxene, which vary with Fe abundance, oxygen fugacity, and temperature. These diffusion coefficient data are graphically illustrated on the Arrhenius plot in Fig. 3.

Ganguly and Tazzoli (1994) evaluated the average Fe-Mg interdiffusion coefficient along the *c* and *b* axes of orthopyroxene as a function of temperature and composition as:

$$D_{Fe} = \exp(-12.8 + 0.060 C_{Fe}) \exp\left(-\frac{57.306}{RT}\right),$$

where D_{Fe}, C_{Fe}, R, and T are diffusion coefficient in cm^2/s, Fe composition in mol%, gas constant in $kcal \cdot mol^{-1} \cdot K^{-1}$, and temperature in K, respectively. The diffusion coefficients in Fe-bearing minerals such as olivine (Buening and Buseck, 1973), orthopyroxene (Ganguly and Tazzoli, 1994) and garnet (Chakraborty and Ganguly, 1991) have been reported to depend on oxygen fugacity, fO_2, and to vary approximately as $(fO_2)^{1/6}$. Hence, we modified this equation, which is valid at fO_2 ranging from the

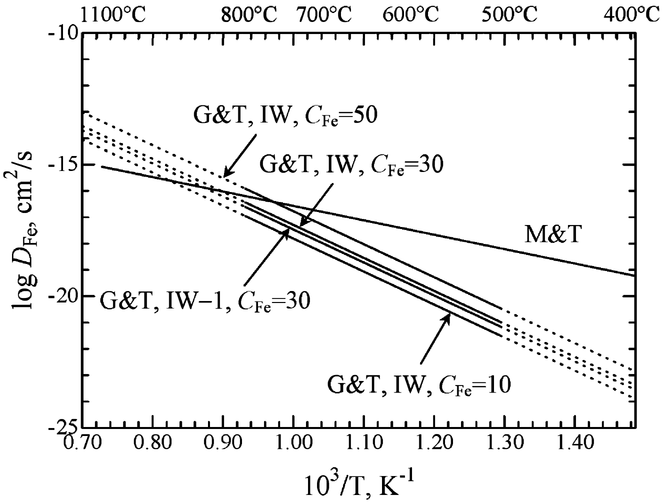


Fig. 3. An Arrhenius plot of the Fe-Mg diffusion coefficients in orthopyroxene by Ganguly and Tazzoli (G&T) (1994) and Miyamoto and Takeda (M&T) (1994). For G&T, the oxygen fugacity fO_2=IW and IW−1, and the abundance of Fe C_{Fe}=10, 30, and 50 mol% are calculated for comparison. The diffusion coefficient by G&T is valid between 500°C and 800°C (Miyamoto and Takeda, 1994).

iron-wüstite (IW) buffer to 0.8 log units above the IW buffer, by incorporating the fO_2 dependence as suggested by Miyamoto *et al.* (1986).

$$D_{Fe} = (fO_2)^{1/6}\exp(-15.3 + 0.060C_{Fe})\exp\left(-\frac{36.561}{RT}\right).$$

Miyamoto and Takeda (1994) estimated the diffusion coefficient in orthopyroxene to be 1×10^{-14} cm^2/s at 1200°C and the activation energy for Fe diffusion in pyroxene to be 100 kcal/mol (>1100°C) and 25 kcal/mol (<1100°C). Using their estimates, the Fe-Mg interdiffusion coefficient in orthopyroxene is expressed as:

$$D_{Fe} = 7.9\times10^{-12}\exp\left(-\frac{25}{RT}\right).$$

Although the diffusion coefficient by Miyamoto and Takeda (1994) is not reported to depend on C_{Fe} and fO_2, the values are very close to those of Ganguly and Tazzoli (1994) for $fO_2 \approx$ IW and $C_{Fe} \approx$ 10–50 mol% at relatively high temperatures of ~800°C.

Jurewicz *et al.* (1993) experimentally determined the redox conditions under which HED meteorites probably formed. For our calculations, we assumed similar redox conditions (*i.e.*, fO_2 of 1 log unit below the IW buffer) for metamorphism. The fO_2 versus temperature relationships reported by Eugster and Wones (1962) for the IW buffer are

$$\log(fO_{2(IW)}) = -\frac{27215}{T} + 6.57.$$

We evaluated the C_{Fe} and fO_2 dependence of these diffusion coefficients. The effect of fO_2 is very small, whereas the effect of composition is greater (Fig. 3). We also compared the two sets of diffusion coefficients by Ganguly and Tazzoli (1994) and Miyamoto and Takeda (1994). In the temperature range ~1000° to ~800°C, values for both sets of diffusion coefficients are similar. However, at 500°C, the value suggested by Miyamoto and Takeda (1994) is nearly three orders of magnitude higher than that of Ganguly and Tazzoli (1994) (Fig. 3). We calculated the compositional profile resulting from Fe-Mg interdiffusion during cooling over the temperature range generally indicated by the thermal diffusion calculations above (850° to 400°C; we chose 400°C as the lower limit because diffusion is so slow compared to 850°C that it becomes negligible over periods of 10^3 to 10^4 years), using the two sets of diffusion coefficients. The diffusion coefficients of Ganguly and Tazzoli (1994) give the best fit profile at a cooling rate of 0.12°C/yr (a period of ~4000 years), and the diffusion coefficients of Miyamoto and Takeda (1994) give the best fit profile at a cooling rate of 0.25°C/yr (~2000 years). Results for the calculations using the diffusion coefficients of Ganguly and Tazzoli (1994) are compared in Fig. 4 with EPMA analyses acquired at the edge of the diogenitic fragment in Petersburg (Fig. 1a).

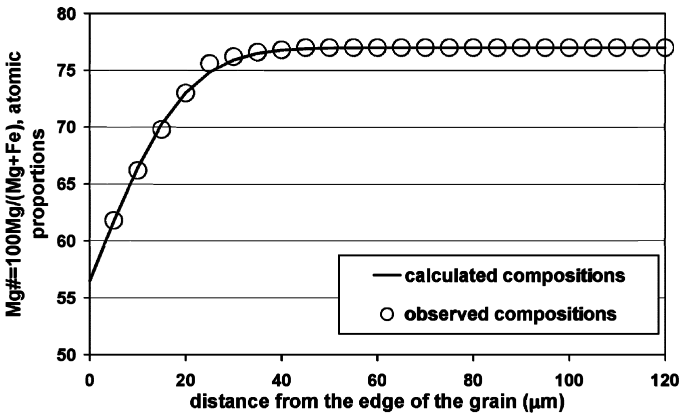


Fig. 4. Calculated (solid curve) and observed (open circles) zoning profiles for Mg# at the edge of a fragment of orthopyroxene in the Petersburg breccia. Calculated profile is obtained with the diffusion coefficient of Ganguly and Tazzoli (1994) for fO_2=IW−1 and cooling from 850°C to 400°C.

4. Discussion and conclusions

Contact metamorphism on 4 Vesta, as modeled in this study, could not alone have caused the high degrees of metamorphism of the equilibrated eucrites (*e.g.*, Juvinas). However, for shallow magma chambers, temperatures in an aureole surrounding the intrusion could be >200°C higher than the regional thermal gradient. When combined with regional metamorphism at intermediate levels in the crust, contact metamorphism could have generated maximum temperatures that are similar to those in the lower parts of the crust.

These metamorphic aureoles may seem like relatively minor volumes of eucritic crustal material. However, the volume of an aureole that extends 100 m from a magma chamber that is 1 km in diameter is ~3.812×10^8 m^3 compared to a volume of ~5.236×10^8 m^3 for the magma chamber. An aureole of 20 m width around the same magma chamber has a volume of 6.538×10^7 m^3 and is more than 10% of the volume of the magma chamber. Volumes of rock affected by the contact metamorphism modeled in this study are comparable to, but somewhat greater than, the volumes of associated intrusive magmatic rocks. Ratio of intrusion volume to volume of country rock that experienced significant metamorphism (defined here as temperatures of >400°C for periods of thousands of years) for a 1 km diameter magma chamber filled with eucritic magma at a depth of 1.5 km is 0.9996. The ratio for the same magma chamber filled with diogenitic magma is 0.7112. For a 3 km diameter magma chamber filled with eucrite magma at 1.5 km depth, the ratio is 0.9687 and for the same magma chamber filled with diogenitic magma is 0.7297.

Time scale is the most significant difference between contact metamorphism, regional metamorphism, and metamorphism associated with a hot layer of lava or impact melt of moderate thickness. Contact metamorphism for the intrusive magma bodies modeled in this study would have occurred for 10^3 to 10^4 years compared to 10^6 to 10^8 years for regional metamorphism (see Ghosh and McSween, 1998). A lava flow of moderate thickness extruded onto the surface of 4 Vesta will cool in 1–10 years, which Yamaguchi *et al.* (1996, 1997) suggests is inadequate to cause significant metamorphism.

Modeling of Fe-Mg interdiffusion for compositional zoning at the edge of the orthopyroxene fragment in Petersburg suggests this breccia may have experienced metamorphism comparable to that expected in a contact metamorphic aureole. Despite significant differences in the suggested diffusion coefficients at low temperatures (~500°C), the cooling rates calculated with the results of Ganguly and Tazzoli (1994) and Miyamoto and Takeda (1994) for the temperature range 850°–400°C are similar. These cooling rates are within the range expected for wall rock of a magma body intruded into the crust of 4 Vesta.

Could this compositional zoning have been caused by regional metamorphism at low temperatures? To evaluate this suggestion, we calculated the time period required to generate this profile with temperatures decreasing from 200° to 100°C and from 400° to 100°C, using the above algorithms. These temperature ranges are appropriate for relatively shallow regional metamorphism. For the temperature decrease from 200° to 100°C, the required time period is 10^{10} years. However, for the temperature decrease from 400° to 100°C, the required time period is 6×10^6 years, which is within the estimated range of Ghosh and McSween (1998) for thermal processes on 4 Vesta. Hence, this compositional profile could possibly have formed by slow cooling associated with regional metamorphism under the right conditions. In our model, a maximum temperature of 400°C represents a depth of 3.3 km within the crust of the asteroid.

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