Supplementary material of 2007GL029515

1. Melting models

The dependence of degree of partial melting \( F \) on depth beneath a MOR is of primary importance for estimating the residual composition of an oceanic plate. The main parameter controlling the depth distribution of \( F \) is the so-called melt productivity function \( \gamma = \langle d F / d P \rangle _{S} \) [e.g. Asimow et al., 1999]. Many authors have estimated \( \gamma \) resulting from adiabatic decompression of an ascending mantle volume [see Langmuir et al., 1992; Asimow et al., 1999 for a review]. Usually a constant value is chosen, ranging between 10 - 20 % per GPa of pressure release. Several lines of evidence indicate that a more representative value for normal MORs would lie closer to the lower bound. Indeed, Klein and Langmuir [1987] and Langmuir et al. [1992], for example, have shown that 12 % GPa\(^{-1}\) seems to be a proper mean value for \( \gamma \). However, more recent thermodynamic simulations have shown that the productivity is far from constant along the melting regime, but instead it has at least two or three well defined paths with nearly constant \( \gamma \) [Asimow et al., 1999; 2001]. Asimow et al. [1999] highlight that the productivity for reversible adiabatic depressurization melting is not constant, but rather increases as melting proceeds. At pressures \( \gtrsim 1.0 - 1.5 \) GPa, \( F \) is unlikely to exceed 5 % in both batch and fractional melting regimes. Once the pressure decreases below this value, the productivity increases drastically until the clinopyroxene is totally consumed.

Fig. 1 shows a summary of the results presented by Asimow et al. [1999] and Asimow et al. [2001] for both batch and fractional melting models. The widely used quasi-linear model (i.e. with constant \( \gamma \), hereafter QL) is also included in this figure [Langmuir et al., 1992; Turcotte and Phipps Morgan, 1992]. There are two important differences between the non-linear (hereafter NL) and QL models. First, at comparable \( P_o \) (pressure at which melting begins), the QL gives significantly higher mean fractions of melting \( \bar{F} \) than the NL case, where \( \bar{F} \) is

\[
\bar{F} = \frac{\int_{P_o}^{P_f} F(P) \, dP}{P_o - P_f} \quad (1)
\]

and \( P_f \) the pressure at which melting stops. Since the crustal thickness is proportional to the product of \( \bar{F} \) and \( (P_o - P_f) \), the QL case predicts much thicker crusts. Second, because of the low productivity in the early stages of melting, as given by the NL model, there is a “tail” of low \( F \) that tend to decrease \( \bar{F} \) even more. These observations show that any given crustal thickness is generated by a melting path with much lower \( P_o \) according to the QL model than according to the NL model. For example, a \( \sim 6 - 7 \) km thick crust is obtained by the QL model with a \( P_o \sim 2.0 \) GPa, but requires a \( P_o \sim 2.75 \) GPa in the NL case. The preferred curve for the NL model used in this paper is shown in Fig. 1. Numerically, it is represented by two linear functions \( F_1 \) and \( F_2 \) in Fig. 1) intercepting at 1.25 GPa. This curve has \( P_f = 0.2 \) GPa and \( P_o = 2.75 \) GPa (i.e. \( \sim 90 \) km depth), produces an \( \bar{F} \) of around 7.2, and a crustal thickness of \( \sim 6.5 \) km. It is important to note here that, although the form of \( \gamma \) might vary from one work to another, the product of \( (P_o - P_f) \) and \( \bar{F} \) should be comparable in all cases.

2. Melt depletion/residual modes models

Niu [1997] presented the first quantitative melting model applied to abyssal peridotites. This model predicts that: (a) the relation between residual mineral modes and the extent of melting follows a quasi-linear trend, and (b) a simple CIPW norm scheme can be used to transform model residue compositions into normative mineral modes (see Figs. 5 and 12 in Niu, 1997). Results from this model differ from those obtained in isobaric melting experiments in that the former predicts that orthopyroxene contributes more than clinopyroxene to the melt during decompression melting, in contrast to what is observed in isobaric experiments and some model melting reactions [e.g. Walter 1999]. Niu [1997] ascribed these discrepancies to the complex polybaric nature of decompression melting beneath a MOR, which cannot be reproduced experimentally. Niu [1999] further showed that his model is consistent with modal observations in abyssal peridotites as well as with decompression melting models. We adopt the linear relationships presented by Niu [1997] as our preferred model to calculate residual modes as a function of extent of melting. However, we also use an experimental isobaric model [Baker and Stopler, 1994] to test the effects that different melting models have on density estimations (see Fig. 2 in the paper). It should be noted that this approach is only valid for a spinel-bearing peridotite, and it needs to be modified when applied to a peridotite within the garnet stability field. When this is the case, we use the experimental results of Lesher and Baker [1997] and Walter [1998], corrected for the slightly different parent source adopted in this work. The final melt depletion/residual modes relations used in this paper for both spinel and garnet peridotites are shown in Fig. 2.

3. General methodology, thermodynamic parameters, and sources of error

The general methodology used in LitMod [Afonso, 2006] to calculate the thermal structure, surface heat flow (SHF), and gravity/geoid anomalies follows that presented in Zeyen and Fernández [1994]. The methods to estimate the elevation, seismic velocities, density structure, total strength, and temperature-pressure-composition-dependent parameters such as the coefficient of thermal expansion, \( \alpha \), and thermal conductivity, \( k \), are all new or significantly modified. All relevant information can be found in Afonso [2006, http://wijia.ija.csic.es/gt/jd/jafonso/jafonso.htm]. A detailed up-to-date assessment of this methodology will be presented elsewhere (Afonso et al., manuscript in

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preparation). The code solves simultaneously for the heat transfer, thermodynamical, geopotential, isostasy, and rheological equations for a particular lithospheric structure with any given composition, and yields density, thermal and seismological structure, SHF, rheological strength, gravity anomalies, elevation, and geoid height. Predictions of geophysical observables are used to control the reliability of the proposed model.

For estimating the density of the lithospheric and sublithospheric mantle we first obtain the residual modes and chemical compositions of each constituent phase, as explained above. The density of each phase as a function of chemical composition, pressure, and temperature is calculated with an third-order Birch-Murnaghan EoS, corrected for thermal expansivity (hereafter TBMEOS).

The TBMEOS equation is written as

\[
P = \frac{3}{2} K_0(T_0) \left[ \left( \frac{P(T, P)}{P(T_0)} \right)^{7/3} - \left( \frac{P(T, P)}{P(T_0)} \right)^{5/3} \right]
+ \left( 1 + \frac{3}{4} K'_0(T_0, 0) - 4 \left[ \left( \frac{P(T, P)}{P(T_0)} \right)^{2/3} - 1 \right] \right)
\]

(2)

where \( K_0(T_0), \ K'_0(T_0, 0), \rho(T, 0) \) and \( \rho(T, P) \) are the temperature-dependent bulk modulus at \( P=0 \), its pressure derivative, the temperature-dependent density at \( P=0 \), and the temperature-pressure-dependent density, respectively. Following Angel [2000] and Pavese [2002], we assume \( K'_0(T, 0) \) independent of pressure and temperature, and

\[
K_0(T, 0) = K_0(T_0, 0) + \left( \frac{\partial K_0(T, 0)}{\partial T} \right) T
\]

(3a)

\[
\rho(T, 0) = \rho(T_0, 0) \exp \left( - \int_{T_0}^{T} \alpha(T) \, dT \right)
\]

(3b)

where \( K_0(T, 0), \rho(T, 0), \) and \( \alpha(T) \) are the isothermal bulk modulus at \( T = T_0 \) (usually 300 K) and ambient pressure, the density at \( T = T_0 \), and the coefficient of thermal expansion as a function of \( T \), respectively.

Within the uncertainties of current experimental techniques, HP-HT compressibility of most of the six upper mantle relevant phases (i.e. olivine, orthopyroxene, clinopyroxene, garnet, spinel, and plagioclase) have been successfully fitted with TBMEOS, assuming a linear variation of the bulk modulus with temperature [e.g. Zhao et al., 1995; Wang et al., 1998; Morishima et al., 1999; Angel, 2000; Pavese et al., 2001; Pavese, 2002; Nishihara et al., 2003; 2005; Kung et al., 2005, among others]. Orthoenstatite represents a potential exception [Flesch et al., 1998; Kung et al., 2004]. This mineral exhibits pressure effects that apparently can only be fitted with a fourth-order BMEOS. However, as shown in Fig. 3, the difference between the fourth- and third-order BMEOS is immaterial on mantle density determinations at pressures < 10 GPa given uncertainties in mineral phases and individual mineral moduli. Significant discrepancies are also found in the literature for the coefficient of thermal expansion (CTE) of orthoenstatite [e.g. Yang and Ghose, 1994], although recent experiments corroborate a temperature dependence of the form \( \alpha(T) = a + bT \), at least up to temperatures of about 1500 K [e.g. Jackson et al., 2003]. More detailed discussions on uncertainties in thermodynamic parameters can be found in Afonso et al. [2005] and Schutt and Lesher [2006]. In general, uncertainties in thermodynamic parameters translates into maximum density uncertainties \( \leq 10 \) kg m\(^{-3}\) in our models, and therefore do not affect the main conclusions of the paper. Thermodynamic parameters used in this study are listed in Fig. 4 (Table 1).

A comparison of our methodology with that of Hacker et al. [2003] was made using their spreadsheet [Hacker and Abers, 2004] with our calculated compositions at the same \( T-P \) conditions than in our model. It was found that the discrepancies between these two methodologies are generally within \( \pm 5 \) kg m\(^{-3}\) for the density of the aggregate (i.e. rock). This should not be surprising since both methodologies rely on solid mineral physics concepts and use similar (identical for some minerals) parameters for the present phases. The only significant deviation between the two approaches is the parameterization for the CTE, which is the main responsible for the observed discrepancies.

References


Figure 1. Degree of partial melting with depth for different melting models. The two dotted and two dash-dotted curves are isentropic melting models for fractional melting and batch melting at two different $P_o$, respectively [from Asimow et al., 1999; 2001]. The dashed line is the QL model of Turcotte and Phipps Morgan [1992]. The solid line is our preferred NL model. $F_1$ and $F_2$ are two linear functions used in the numerical implementation of the average NL model. Pressure in these functions is in GPa. Other relevant parameters are listed in the right upper corner.
Figure 2. Residual modes of spinel and garnet peridotites as a function of melt extraction. Red lines are the preferred trends adopted in this work. Ol = olivine, Opx = orthopyroxene, Cpx = clinopyroxene, Sp = spinel, Gt = garnet. A) Residual modes for spinel peridotite in vol %. Blue lines and experimental data at 1 GPa from Baker and Stopler [1994]; dashed lines enclose abyssal peridotite data from Niu et al. [1997]; black circles = Ol, black squares = Opx, black triangles = Cpx, black diamonds = Sp. B) Residual modes for garnet peridotite in vol %. Experimental data and associated uncertainty at 3.6 (black) and 3.0 (blue) GPa from Lesher and Baker [1997] and Walter [1998], respectively, corrected for the slightly different primitive composition adopted in this work. Unfilled diamonds represent fitting data obtained from the this correction but not observed in the experiments. Symbols as in (A) except for diamonds that now represent garnet.
Figure 3. Difference (in kg m$^{-3}$) in density predictions of orthoenstatite using a third-order Birch-Murnaghan EoS [Angel and Jackson, 2002] and a fourth-order Birch-Murnaghan EoS [Flesch et al., 1998]. For depths $\lesssim 300$ km ($\sim 10$ GPa) the density difference predicted by these two EoS is insignificant.

Flesch et al. (1998): $K_a=104$ GPa

$K'_a=10.9$

$K''_a=-1.6$ GPa$^2$

Angel & Jackson (2002): $K_a=105.8$ Gpa

$K'_a=8.5$
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<th>ρ</th>
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α(T) = a + b*T + c*T² + d*T⁴

Most values from compilations in Lee [2008], Matenlogne et al. [2005], and Schütz and Leeder [2006], except:

Fo, Fa: dK/dT from Stipp et al. [1998]; dK/dT and dG/dT of Fe (from Epp et al. [2004]); dK/dP of K (average from works in Table 4.1 of Poirier, 2006).

OEn: Adiabatic parameters from Kung et al. [2004]; isothermal values from Angel and Jackson [2002].

Di: dK/dT, dK/dP, dG/dT of K from Isaac et al. [2006];

Jd: dK/dT from Zhao et al. [1998];

Py: dK/dT and dG/dT from Gowanwalls [2005];

Gr: dK/dT from Pao et al. [2001].

HP-CEn: K and X extrapolated from Kung et al. [2004, 2005]

Plg: parameters for EoS from Angel [2004], assuming a composition Ab₅₃An₄₇; other parameters assumed.

* From Isaac et al. [2006]; † same as in Di; ‡ From Shinmei et al.[1999]; † same as in HP-CEn; * from Fei [1995] for Plg Ab₅₃An₄₇.

Linear combination for a composition Ab₅₃An₄₇.