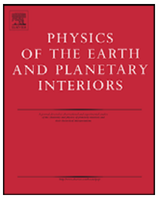
Physics of the Earth and Planetary Interiors 172 (2009) 257–267

Contents lists available at ScienceDirect

Physics of the Earth and Planetary Interiors

journal homepage: www.elsevier.com/locate/pepi



Density profiles of oceanic slabs and surrounding mantle: Integrated thermodynamic and thermal modeling, and implications for the fate of slabs at the 660 km discontinuity

Jibamitra Ganguly[a,*], Andrew M. Freed[b], Surendra K. Saxena[c]

[a] *Department of Geosciences, University of Arizona, Tucson, AZ, USA*
[b] *Department of Earth and Atmospheric Sciences, Purdue University, West Lafayette, IN 47906, USA*
[c] *CeSMEC, Center for the Study of Matter Under Extreme Condition, Florida International University, Miami, FL 33199, USA*

ARTICLE INFO

Article history:
Received 7 February 2008
Received in revised form
30 September 2008
Accepted 9 October 2008

Keywords:
Density
Transition zone
Mantle
Slab
Subduction

ABSTRACT

We have calculated the mineralogical properties of the Earth's mantle and the lithological units constituting the subducting oceanic slabs within a wide range of P–T conditions within the $CaO–FeO–MgO–Al_2O_3–SiO_2$ system, except for the basalt top-layer of a slab, for which the system is extended to include Na_2O. The mineralogical data are then converted, using the appropriate P–V–T relations, to bulk densities. The calculated adiabatic density vs. depth profile of the mantle between 200 and 725 km depths is in good agreement with geophysical and experimental data. The density data of the different compositional units are combined with calculated thermal structures for a variety of slab–mantle systems to construct equilibrium density profiles as a function of depth. The mean equilibrium densities of the slabs within the transition zone (400–660 km depth) are found to be ~0.04–0.05 g/cm^3 greater than those of the ambient mantle within the same depth interval. For the entire upper mantle, density differences between slabs and ambient mantle are slightly less, but the slabs still remain denser than the latter. At 670 km depth, slabs have lower density than the ambient lower mantle because of the commencement of perovskite forming reactions within the mantle, and displacements of these reaction boundaries to higher pressures within the slabs as a consequence of their negative P–T slopes. If perovskite forming reactions within slabs are hindered for kinetic sluggishness, then neutral buoyancy would be achieved when the slabs have penetrated ~100 km into the lower mantle. However, using the available data on the kinetics of spinel to perovskite plus periclase reaction, we conclude that the reaction would go to completion in a Peru-type young slab (41 Myr), and very likely also in a Tonga-type old slab (110 Myr), before these penetrated 100 km into the lower mantle. Thus, slabs should always remain negatively buoyant, and therefore continue to subduct through the lower mantle once it penetrates through the 660 km discontinuity. Despite a negative buoyancy force, a slab could deflect at the top of the lower mantle (660 km) because of factors resisting subduction, namely viscosity jump, low dip angle, slab roll back, and metastable persistence of olivine in cold slabs. If published scale model experiments represent realistic approximations of the factors affecting plate subduction, then according to our density data, any slab with a dip angle of $\leq$40–50° would bend at the 660 km discontinuity if there is a viscosity jump of at least by a factor of ~10 and trench migration. The basalt top-layer of a slab is denser than other slab components and the ambient mantle at all depths to 660 km, and therefore should continue to sink into the lower mantle, especially if a slab directly penetrates the 660 km barrier, instead of peeling off in the transition zone to form a "perched eclogite" or "piclogite" layer, as previously proposed. The harzburgite layer, which is sandwiched between denser basalt and lherzolite layers, faces greater resistance to subduction, especially in a young slab, and thus could significantly contribute to the deformation of a slab near the 660 km discontinuity.

1. Introduction

It is commonly accepted that the 660 (±10) km deep seismic discontinuity in the Earth's mantle, which defines the top of lower mantle, is primarily due to the ringwoodite (Rng) to Mg-perovskite (MgPv) plus magnesiowüstite (Mg–Wu) transformation.

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[*] Corresponding author. Tel.: +1 520 621 6006; fax: +1 520 621 2672.
E-mail address: ganguly@email.zrizona.edu (J. Ganguly).

0031-9201/$ – see front matter © 2008 Elsevier B.V. All rights reserved.
doi:10.1016/j.pepi.2008.10.005

Experimental data between 1000 and 2000 °C show the P–T slope of this transition to be negative, with estimates varying between −30 bar/K (Ito and Takahashi, 1989) and −13(±3) bar/K (Katsura et al., 2003; Fei et al., 2004). Since the interior of a subducting oceanic slab is colder than the ambient mantle at the same depth, the negative P–T slope of the transition implies Rng to MgPv plus Mg–Wu transformation at a greater depth within a slab than in the ambient mantle. This has led to the idea that the density of the slab at ~660 km depth is lower than that of the ambient mantle, thereby opposing or even preventing the slab's descent into the lower mantle (e.g. Ringwood, 1994; Davies, 1999). Seismic tomographic images, however, show both deflection of slabs at the 660 km depth and penetration to greater depths (e.g. van der Hilst et al., 1991; Grand et al., 1997). Understanding of balance of forces that control the fate of subducting slabs at the top of the lower mantle requires a comprehensive analysis of the density structure taking into account the effects of all important phase transformations in the slab–mantle system, rather than only the effect of Rng to MgPv + Mg–Wu transformation. Davies (1999) called attention to the possibility that the effect of the Rng to MgPv + Mg–Wu transformation on the buoyancy force of a slab may be significantly counterbalanced by other phase transformations with positive P–T slopes, and thus emphasized the need for calculations in the multicomponent and multiphase system. Also, inasmuch as the primary driving force for plate subduction comes from the negative buoyancy force on slabs (e.g. Forsyth and Uyeda, 1975; Hager, 1984), a comprehensive analysis of the buoyancy force taking into account the effects of thermal structure and mineralogical reactions in systems, which closely approximate the natural systems, constitutes an important step in our understanding of the dynamics of plate motions.

Ringwood (1982) suggested that an oceanic slab of ~80 km thickness is lithologically stratified with an upper ~6 km of basalt, followed successively downward by ~24 km layer of residual harzburgite, ~10 km layer of residual lherzolite and ~40 km layer of slightly depleted pyrolite. The last three components represent the residues left after different degrees of partial melting of pyrolitic mantle rock and extraction of basalt, with harzburgite being the most refractory component. Hoffmann and White (1982) proposed that the basaltic crust of a slab has a greater density than the mantle throughout its entire depth, and consequently it should separate from the associated refractory lithosphere and sink, perhaps all the way to the core-mantle boundary, thereby forming a layer of "degenerate crust". They envisioned this buried crustal layer to be the source of mantle plumes. Anderson (1989), on the other hand, suggested that the basaltic crust becomes lighter than the pyrolitic mantle at the 660 km discontinuity, and is thus trapped during subduction to form a "perched eclogite layer". This viewpoint was later supported by Irifune and Ringwood (1993) on the basis of their experimental data and calculation of the zero pressure density (1 bar, 298 K) of the assemblages observed in their experimental products at high P–T condition. However, using results from *in situ* determination of mineralogical transformations in basalt, Litasov et al. (2005) concluded that there is no density cross-over between the basaltic component of a cold slab and surrounding mantle at the 660 km depth, and consequently, the basaltic component should penetrate into the lower mantle.

In this work, we have taken a computational thermodynamic approach to the problem of density structure of the slab–mantle system within and around the transition zone, and present the details of density variation within the different lithological components as a function of temperature at pressures corresponding to 200–800 km depth, rather than zero-pressure densities as in the most earlier studies. We also calculate the thermal structures of selected slabs, and integrate the data with the results of the thermodynamic calculations to develop a comprehensive understanding of density variations within and around slabs of widely differing ages, vertical descent rates and subduction angles.

2. Thermodynamic calculations

2.1. Data and methodology

We take the compositions of the following rock types to represent the major element bulk chemistries of the different lithological units of an 80 km thick slab (Fig. 1): Grt-lherzolite from Kilborne Hole (KH), New Mexico (Takahashi, 1986), for the slightly depleted pyrolite (lowest 40 km) and residual lherzolite (10 km) layers, ophiolitic harzburgite (Brown and Mussett, 1981) for the residual harzburgite component, and the average between oceanic-basalt (Brown and Mussett, 1981) and mid-ocean ridge basalt (MORB; Ringwood and Irifune, 1988) for the top basalt layer of a slab. The major element compositional distinction between residual lherzolite and slightly depleted pyrolite is minor in terms of its effect on the density structures of these units. The bulk composition of the ambient mantle is assumed to be given by the average of three compositions, namely (a) the pyrolite model composition of Ringwood (1982), (b) mantle composition based on the assumption of average solar system abundance ratios for the whole Earth (Palme and O'Neill, 2003) and (c) composition of KLB-1 peridotite (Takahashi, 1986). The system CaO–FeO–MgO–Al_2O_3–SiO_2 (CFMAS) constitutes ~99 wt% of KH-lherzolite and ophiolitic harzburgite, ~98 wt% of assumed ambient mantle composition and ~95 wt% of the average of oceanic basalt and MORB. Thus, for the purpose of phase equilibrium calculations, the system CFMAS is considered to be an adequate representation of the major element compositional space of the different lithologic units, except for basalt. For the latter, we include Na_2O as an additional component. The Na_2O-CFMAS or NCFMAS constitutes ~97 wt% of the composition of the adopted basalt composition. All compositions are normalized within the respective subsystems, and summarized in Table 1. We mention at the outset that compositional variations within the proposed limits of the different lithologic units do not have any significant effect on the density structures of the slab–mantle system that are deduced below.

All thermodynamic data used in the calculations carried out in this study have been uploaded in a web page that can be accessed using the link http://cesmec.fiu.edu/data/thermodata01.txt. Except for Ca-perovskite (CaPv), the end-member thermodynamic properties are taken from the updated and internally consistent data base

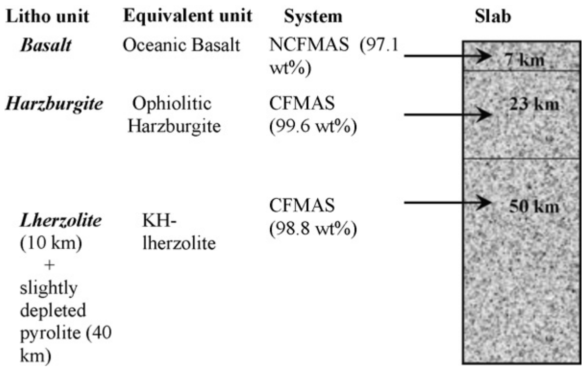

Fig. 1. Schematic representation of the different lithologic units that are assumed to constitute an approximately 80 km thick oceanic slab. The acronyms connected to the units indicate chemical systems that are used to represent the bulk compositions. The numbers beside the acronyms indicate the wt% that these systems constitute of the total chemical systems in the different units. N: Na_2O; C: CaO, F: FeO, M: MgO, A: Al_2O_3, S: SiO_2.