Time-dependent interaction between subduction dynamics and phase transition kinetics

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Accepted 2009 March 14. Received 2009 March 14; in original form 2008 March 21

SUMMARY

Numerical subduction models are presented that account for the kinetics of the olivine–spinel transformation and the equilibrium transformation of spinel into perovskite and magnesiowustite. Temperature-, depth- and stress-dependent rheology is used. Buoyancy due to temperature and the olivine, spinel, and magnesiowustite/perovskite phases leads to free subduction into the transition zone, and retarded penetration into the lower mantle. A dynamic feedback mechanism is found between the different buoyancy forces, the kinetics of the phase transition and the release of latent heat. This mechanism modulates the ‘parachute’-effect of the metastable olivine (MO) in a subducting slab and leads to characteristic temporal variations of the depth of MO and, with a time delay, variations of the subduction velocity. Time periods are of the order of 3–4 Myr. This mechanism allows to explain variations of the depths of the deepest earthquakes in otherwise similar subduction zones. Further conclusions are as follows. The maximum depth of MO increases roughly linearly with lithospheric thickness, and the maximum depth may reach as deep as 720–750 km. Different subduction zones may represent different stages of ‘velocity–depth phase loops’ of MO.

Key words: Numerical solutions; Phase transitions; Dynamics of lithosphere and mantle.

1 INTRODUCTION

Tomographic images indicate that some slabs penetrate straight into the lower mantle whereas others flatten and stagnate within the transition zone (van der Hilst et al. 1991). One example is the Northern Tonga subduction zone where the slab deflects in the transition zone (between about 400–700 km depth) before continuing into the lower mantle (van der Hilst 1995). Another example of slab flattening is the Japan subduction zone, in which the slab extends horizontally for about 900 km, thereby, deflecting the 660 km—discontinuity downward to as much as 700 km (Li et al. 2000). A systematic investigation of some physical causes, such as Clapeyron slope, viscosity contrast, or dip angle influencing slab stagnation in the transition zone has been carried out by Torii & Yoshioka (2007). In that study, a kinetically delayed olivine–spinel transformation has not been considered, which is the issue of this study.

The transition of the abundant mineral (Mg, Fe)$_2$SiO$_4$ in the upper mantle from the $\alpha$-phase (olivine) to the $\beta$-phase (wadsleyite) and to the $\gamma$-phase (ringwoodite) may be kinetically inhibited because of the low temperature conditions in the subducting lithosphere (Rubie & Ross 1994; Sung & Burns 1976). It has been proposed that this metastable olivine might be the cause for flattening and the delayed subduction of cold slabs into the lower mantle (Okal & Kirby 1998; Tetzlaff & Schmeling 2000). How conclusive is the observational evidence on the presence of metastable olivine at greater depth? Indications for metastable olivine within the transition zone immediately above the 660 discontinuity come from observations of seismic anisotropy (olivine is significantly more anisotropic compared to spinel). Examples of such observations include Northeast Asia (Liu et al. 2008) and the lower transition zone of the Tonga–Fiji subduction region (Chen & Brudzinski 2003). Within the same region low seismic velocities within the flat lying deep seismicity zone have been interpreted by the presence of metastable olivine or volatiles (Brudzinski & Chen 2003). Late arrivals of $P$-waves at northern Marianas indicate evidence for a 25 km wide metastable olivine wedge at 590 km depth (Kaneshima et al. 2007). Pankow et al. (2002) obtained better fit of their shear wave amplitude patterns if they assumed a metastable wedge within the Kurile subduction slab. The double seismic layer observed in the Tonga subduction zone may be due to the stress distribution associated with a metastable olivine wedge at 460–590 km (Guest et al. 2004). However, the kinetically delayed phase transformation of metastable olivine into spinel has also been questioned by Wiens (2001) as being an appropriate mechanism for deep earthquakes because of its narrow lateral extent.

Experimental data of phase kinetics of the olivine $\rightarrow$ wadsleyite $\rightarrow$ ringwoodite transition (Rubie & Ross 1994; Mosenfelder et al. 2001) have been used in kinematic subduction models to determine the depth extent of the metastable olivine wedge as a function of thermal parameter (Kirby et al. 1996; Devaux et al. 1997; Mosenfelder et al. 2001) or for particular subduction zones (Jing et al. 2002).
The thermal parameter $\phi$, which is given by the product of the vertical component of the subduction velocity and the age of the subducting plate at the trench, is a measure of the depth of penetration into the mantle of a specific isotherm. Assuming steady state, such kinematic subduction calculations show that above a threshold value of $\phi$ of about $5000 \pm 2000$ km the maximum depth of the metastable olivine increases roughly linearly with $\phi$. Assuming that deep earthquakes are related to stresses associated with the transformation of metastable olivine into spinel, the deepest earthquakes should correlate with the thermal parameter in a similar way as the kinetic models mentioned above predict. However, compilations of such earthquake data by Kirby et al. (1996) and Gorbatov and Kostoglodov (1997) show a significantly different behaviour: within a limited range above a threshold thermal parameter of about 3000 km the depths of deepest earthquakes cover a broad depth range, that is, some subduction zones with a certain $\phi$-value may exhibit maximum depths of 450 km, others with the same $\phi$-value occur at 700 km depth. All subduction zones with $\phi$-values larger than 6000 km exhibit very deep earthquakes. It has been noted by Devaux et al. (1997) and Mosenfelder et al. (2001) that this is difficult to reconcile with the predicted depth-dependence of metastable olivine. In this paper, we approach this problem by emphasizing the time-dependent behaviour of formation of metastable olivine.

Whereas subduction is usually regarded as a rather steady process, recent analyses of backarc basin formation, volcanic ages of subduction related volcanism and earthquake distributions suggest that the convergence rate at trenches may strongly vary with time on timescales of several tens of Myr down to at least 5 Myr (Sdrolias & Müller 2006; Spicak et al. 2007). The dynamics and temporal behaviour of cold subducting slabs are mainly controlled by negative thermal buoyancy forces and by buoyancy anomalies resulting from slab mineralogy because of the density contrasts of the different mineralogical phases. Under equilibrium conditions the phase boundaries of the exothermic transitions $\alpha \rightarrow \beta \rightarrow \gamma$ are elevated in a cold slab resulting in negative buoyancy forces. If $\alpha$-olivine persists as a metastable phase in the high pressure stability-fields, positive buoyancy forces will reduce or even overcome these negative buoyancy forces. Sufficient volumes of metastable olivine may be expected to inhibit further penetration into the lower mantle in addition to the positive buoyancy forces resulting from the endothermic transition of the $\gamma$-phase to perovskite and magnesiowustite. If, on the other hand, metastable olivine is transformed into spinel at great depth, higher amounts of latent heat are released. This excessive heat might influence the penetration behaviour of the slab into the lower mantle. Thus, the investigation of the influence of the metastable olivine (MO) and the effect of latent heat on the shapes and the dynamics of slabs subducting through the transition zone and eventually into the lower mantle is of particular interest.

Schmeling et al. (1999) considered the behaviour of a self-consistent subducting slab penetrating the upper mantle and observed an increasing influence of the MO on the dynamics with lithospheric age. They found that MO formed in slabs with ages greater than approximately 70 Myr and that a significant slowing down of subduction velocity occurs for ages greater than 100 Myr. The metastable wedge of olivine acted as a low-density ‘parachute’ (Kirby et al. 1996; Marton et al. 1999; Schmeling et al. 1999) resulting in a characteristic temporal behaviour with typical periods of 10 Myr. The maximum depth of metastable olivine did not exceed 550 km in their models, partly because no slab penetration beneath 660 km has been allowed for.

Tetzlaff & Schmeling (2000) included both phase boundaries—olivine (ol) $\rightarrow$ spinel (sp) (410 km) and sp $\rightarrow$ perovskite (pv) + magnesiowustite (mw) (660 km). They investigated the influence of the MO-wedge and the behaviour of the subducting oceanic lithosphere near the boundary between upper and lower mantle with particular emphasis on the slowing down of subduction and the question whether MO could retard or prevent slabs from penetrating into the lower mantle. Both, Schmeling et al. (1999) and Tetzlaff & Schmeling (2000), however, simplified the kinetic equation that could describe the disequilibrium transformation of MO into spinel. They obtained relatively large amounts of MO, and maximum depths up to 770 km (sometimes isolated MO fragments reached even 800 km). In contrast to these former models we will solve the kinetic equation explicitly and consider the effects of (depth-dependent) latent heat release. Particular emphasis will be put on the time-dependent coupling between slab dynamics and kinetics of MO and the interaction with the sp–pv transition.

### 1.1 Reaction kinetics

Our simplified phase diagram for the composition of ($\text{Mg}_{0.8},\text{Fe}_{0.2})_2\text{SiO}_4$ is based on thermodynamic data of Akao et al. (1989). We approximate both the exothermic transitions of olivine ($\alpha$) to modified spinel ($\beta$) to spinel ($\gamma$) by one single transition regime. The $\alpha \rightarrow \beta \rightarrow \gamma$ and the $\beta \rightarrow \gamma$ transition are represented by two sets of contour lines. We included the results of extrapolations of high pressure experiments (Rubie et al. 1990), which suggest that below a temperature $T_1$ the transition of ol $\rightarrow$ sp is kinetically delayed and will not take place on the subduction timescale, whereas between $T_1$ and $T_2$ the transformation takes place but is still kinetically delayed. For temperatures greater than $T_2$ the transformation takes place under equilibrium conditions. Based on the experiments by Rubie et al. (1990), Tetzlaff & Schmeling (2000) used $T_1$ and $T_2$ values of 600 and 650°C, respectively. Below $T_2$ metastable olivine transforms into spinel by a disequilibrium transformation, which could be described by a kinetic equation suggesting an exponential increase of the $\alpha$-fraction. The kinetic of this reaction can be described by the Avrami equation (Cahn 1956)

$$\xi = 1 - \exp(-kt^n),$$

where $\xi$ is the volume fraction transformed, $t$ is time, and $k$ and $n$ are assumed to be constants. See also Table 1 for the meanings and the values of all variables and parameters. Rubie & Ross (1994) suggest that transformation of olivine to spinel in relatively cold slabs is independent of a range of nucleation rate models and is controlled only by the growth rate of the high pressure phase. Under the assumption that all nucleation occurs instantaneously (nucleation rate $= 0$)

$$\xi = 1 - \exp[-2S \int_0^t \dot{\xi}(\Theta) d\Theta].$$

where $\dot{\xi}$ is the growth rate of the product phase at time $\Theta$, $t$ is the reaction time, and $S$ is the grain boundary area per unit volume. For equidimensional grains (tetrakaidecahedra), $S = 3.35/d_g$, where $d_g$ is the average grain diameter (Rubie et al. 1990). The growth rate of the product phase in an interface controlled polymorphic phase transformation is given by

$$\dot{\xi} = k_0 T \exp\left[-(\Delta H_a + PV_a)/RT\right] \left[1 - \exp(-\Delta G_i/RT)\right],$$

where $k_0$ is a constant, $T$ is absolute temperature, $\Delta H_a$ is activation enthalpy, $V_a$ is the activation volume, $P$ is the pressure, $\Delta G_i$ is the free energy change of reaction, and $R$ is the gas constant.
For equilibrium phase transition (i.e. at \( T \)) the influence of the dynamic pressure region) the convective derivative of where \( \Delta H_\alpha \) is the activation enthalpy \( 350 \, \text{kJ mol}^{-1} \) and \( \Delta G_r \) is the free energy change of reaction \( 1.3 \times 10^{-5} \, \text{m}^3 \, \text{mol}^{-1} \).

\[ \frac{d}{dt} = \frac{\partial \xi}{\partial t} \left( \frac{D T}{\partial z} + v_z \frac{\partial \xi}{\partial T} \right), \]

where \( \Delta H_{T,p} \) is the enthalpy change at \( T \) and a reference pressure \( P_0 \) (usually 1 bar), and \( \Delta V_{T,P} \) is the molar volume change at the \( P \) and \( T \) conditions, where the olivine spinel transition takes place. As \( \Delta V_{T,P} \) varies within the depth- and temperature-range of interest only by a few per cent, eq. (6) may be written to first order to give the depth-dependent latent heat \( L \)

\[ L = L_0 + L_z z \]

with \( L_0 \) and \( L_z \) are given in Table 1, \( z \) is the depth (positive downward).

The endothermic phase transition of \( \text{sp} \rightarrow \text{pv} + \text{mw} \) is approximated by a linear increase of the \( \text{pv} + \text{mw} \) fraction with depth. The kinetics of the endothermic phase transition of \( \text{sp} \rightarrow \text{pv} + \text{mw} \) are not known. Therefore, we assume an equilibrium transition, which may be approximated by a linear increase of the \( \text{pv} + \text{mw} \) fraction with depth within the depth interval. In case of metastable olivine entering the \( \text{pv} + \text{mw} \) field, it is assumed that olivine first transforms into spinel via the kinetically delayed transition as described above, but the resulting spinel immediately transforms into perovskite +
magnesiowustite assuming equilibrium transition. This kind of behavior has also been implicitly assumed in previous slab models with a metastable wedge (e.g. Okal & Kirby 1998).

2 MODEL

The system of equations for thermal convection—conservation of mass, momentum, and energy—is solved in two dimensions in the extended Boussinesq and infinite Prandtl number approximation.

The non-dimensional momentum equation in the stream function formulation is given in two dimensions

\[
\left( \frac{\partial^2}{\partial x^2} - \frac{\partial^2}{\partial z^2} \right) \psi + 4 \frac{\partial^2}{\partial x^2} \psi \frac{\partial^2}{\partial z^2} \psi + \frac{\partial}{\partial x} \left( \frac{\partial \psi}{\partial x} \right) \frac{\partial}{\partial z} \frac{\partial \psi}{\partial z} = Ra \frac{\partial^2}{\partial x^2} - Rcp \frac{\partial \xi_{sp}}{\partial x} - Rcp \frac{\partial \xi_{pp}}{\partial x},
\]

(8)

with the buoyancy terms on the right-hand side of the equation. \(x', z'\) are the coordinates, \(\psi'\) is the stream function (\(v'_x = \frac{\partial \psi}{\partial x}, v'_z = -\frac{\partial \psi}{\partial z}\)), \(v'_x, v'_z\) are the velocities in horizontal and vertical direction, \(Ra = \frac{\omega_{eq} \Delta h^3}{\kappa \eta}\) is the thermal Rayleigh number, \(Rc_{pp} = \frac{\Delta I_{eq} \Delta h^3}{\kappa \eta}\) and \(Rc_{sp} = \frac{\Delta I_{eq} \Delta h^3}{\kappa \eta}\) are the compositional Rayleigh numbers of the simplified ol→sp, respectively sp→pv + mw mantle transitions, \(\alpha\) is the thermal expansivity, \(\rho_0, \Delta \rho_{pp}, \) and \(\Delta \rho_{sp}\) are the reference density and the ol–sp and sp→pv + mw density contrasts, respectively, \(g\) is the gravity acceleration, \(\Delta T\) is the scaling temperature, \(h\) is the total layer thickness, \(k\) is the thermal diffusivity, \(\xi_{sp}\) and \(\xi_{pp}\) are the fractions of the sp and pv + mw phases, respectively (see also Table 1). The density contrasts have been chosen assuming that only the olivine component (60 per cent) of the mantle material undergoes the phase transformations and are in general agreement with recent data based on seismological observations (Shearer & Flanagan 1999). The dynamic viscosity \(\eta' = \frac{1}{2} \frac{1}{2} \exp \left( \frac{\xi_{+} + \xi_{-}}{n} \right) \sigma^{1/3} \) is assumed to be temperature, pressure, and stress dependent according to a power-law rheology (Chopra & Paterson 1984) \(\left(A = 10^{46} \text{ MPa}^{-n} \text{ s}^{-1}, E_a = 535 \text{ kJ mol}^{-1}, z\right)\) is the depth, \(V'_a = 1.3 \times 10^{-4} \text{ m} \text{ mol}^{-1}, T\) is the absolute temperature, \(\sigma\) is the second invariant of the stress tensor, \(n = 3.6\). The range of viscosity variation in our model is restricted to \(10^{21}\) to \(10^{27}\) Pa s. The rheology law we used is an over-simplification, particularly in the transition zone and in the lower mantle. The role of spinel and modulated spinel is very poorly constrained (Riedel & Karato 1997; Karato 1996). Subducted slabs are likely to have complicated rheological structures in the transition zone due to combined effects of crystal structure, grain size, and temperature on rheology (Karato 1996). Recent geodynamic observations suggest that the subducted slabs are rather weak and highly deformable. Karato (1996) showed viscosity profiles of subducting slabs with different temperature profiles and different subducting velocities in which the effect of grain size reduction due to the ol–sp phase transition is considered. In these models the viscosity inside the subducting lithosphere is found to vary between \(10^{21}\) and \(10^{23}\) Pa s. The rheology of the lower-mantle minerals is poorly known due to the lack of direct experimental studies under high pressures and temperatures. It has been argued that the lower-mantle rheology depends strongly on whether perovskite or magnesiowustite controls the rheology, because the rheology of the two materials is likely to be very different at the same homologous temperature (Karato 1989).

Because of this lack of knowledge of rheological properties of mantle minerals at greater depths we used a simplified rheological law, which is based on laboratory data for the upper mantle, but that is controlled by additional parameters (truncation viscosities and an activation volume) to match other geophysical observations. The formation of a highly viscous slab is allowed, too high viscosities are avoided by an upper truncation viscosity. The lower value of \(10^{21} \text{ Pa s}\) has been chosen to represent the average upper-mantle viscosity according to analyses of postglacial rebound data. The activation volume was adjusted to result in an appropriate viscosity increase in the lower mantle, as required from dynamic inversions of geoid data (e.g. Hager 1984, 1991) or from matching hotspot tracks by advecting plumes (Steinberger & O’Connell 1998).

Within the upper few tens of kilometres of the model, a depth-dependent plastic rheology is superimposed by a rheology based on frictional sliding, described by Byerlee’s law, to account for the weakening effect of a cold brittle upper lithosphere. The importance of this weakening zone for decoupling the slab from a free slip model surface has been discussed by Enn et al. (2005).

The non-dimensional heat equation is given by (non-dimensional quantities are primed)

\[
\frac{DT'}{Dt} + v'_i Dv'_i = \nabla \cdot (\kappa \nabla T') + \frac{Di}{Ra} \sigma_{ij}' \frac{\partial v'_i}{\partial x^j} + H'_L,
\]

(9)

where the non-dimensional latent heat release is determined by the non-dimensional version of eq. (4)

\[
H'_L = \frac{L_{ol}}{c_p \Delta T} \frac{Dk_{sp}}{Dt} + \frac{L_{sp}}{c_p \Delta T} \frac{Dk_{pp}}{Dt},
\]

(10)

with the latent heat term \(L_{ol} = L_0 + L_z\) of the ol→sp and \(L_{sp}\) of the sp→pv + mw transitions, respectively. \(\xi_{sp}\) and \(\xi_{pp}\) are the fractions of transformed spinel and perovskite, respectively. And \(t\) is the time, \(T\) is the absolute temperature, \(Di = \frac{gh}{c_p}\) is the dissipation number, \(c_p\) is the heat capacity, and \(\sigma_{ij}\) is the stress tensor. In the case of equilibrium thermodynamics (for \(T > T_2\)) the convective derivative terms can be rewritten as

\[
\frac{D\xi_{sp}}{Dt} = \left( \frac{\partial \xi_{sp}}{\partial T'} \right) \frac{DT'}{Dt} + \frac{\partial \xi_{sp}}{\partial z'} v'_z
\]

(11)

where the subscripts \(k\) stands for either sp or pv.

It is interesting to explore the physical meaning of the depth-dependence of the latent heat in a convecting system. This can be done by scaling the depth-dependent latent heat in eq. (10) by the equilibrium latent heat given by (e.g. Christensen 1998) \(L_{eq} = \gamma \Delta \rho T\rho^2\), where \(\gamma = D P/\Delta T\) is the Clapeyron slope, \(\Delta \rho\) the density contrast between the two phases, \(\rho\) the mean density of the two phases, and \(T\) the absolute temperature. Then the non-dimensional rate of gain or release of latent heat of the olivine–spinel transition (first term in eq. (10)) reads

\[
H '\bigg|_{\text{el-sp}} = \frac{Rc_{sp} D \gamma T' \left(1 + L'_s(z' - Z'_{eq}) \right)}{Ra} \frac{Dk_{sp}}{Dt},
\]

where \(L'_s = hl_{eq}/L_{eq}\) is of the order 1, \(z'_{eq}\) is the non-dimensional depth of equilibrium phase transition, \(Rc_{eq}\) is the phase change Rayleigh number (in which the density constraint \(\Delta \rho\) occurs), and \(\gamma'\) is the Clapeyron slope scaled by \(\rho \gamma /\Delta T\). This temperature dependent scaling is thermodynamically equivalent to extracting the temperature dependence from the enthalpy terms of eq. (6). Let us assume a steady state convection cell. At equilibrium phase transformation \((z' = z'_{eq})\), the gain and loss of latent heat in the down- and upwelling limbs is different because \(T'\) is different. Thus, the system experiences a net cooling effect, in addition and similar to the net cooling due to the difference in adiabatic heating and cooling in the two flow limbs. This net cooling effect scales with \(Di\) in the same way as adiabatic cooling does. This net cooling is balanced by dissipative heating associated with mechanical work, which is also scaled with \(Di\). In case of non-equilibrium phase changes with
metastable olivine ($z' > z'_{eq}$ in the downwelling, $z' = z'_{eq}$ in the upwelling), the phase transition is delayed in the cold downwelling, and takes place at greater ambient pressure. Thus, more energy is released due to the work done by compression at a higher pressure level. The temperature increase in the cold slab is stronger than in the equilibrium case. The net cooling of the convection cell is reduced compared to the equilibrium case, and as a consequence, dissipation of mechanical work compensating the net cooling, is also reduced. This reduction is accomplished by a slowing down of the flow due to the buoyancy forces of the metastable olivine. The models carried out here are far from steady state. As we are not interested in the differences between up- and downwinnings, the temperature dependence of latent heat as indicated in eq. (10), or eq. (12) is neglected.

The reaction kinetic eq. (2) is modelled in the Lagrangian reference frame, that is, following material particles along their flow path lines. This is done by defining a marker field ($900 \times 400$ markers) within the whole box, which is advected with the flow by using a Runge–Kutta scheme fourth order (see e.g. Schmeling et al. 2007). Each marker carries the field $\xi$, whose evolution is determined by integrating eq. (2) along the marker path. Additionally the total derivative of $\xi$ is used to obtain the rate of latent heat (eq. 4) for every marker. The $\xi$- and $H_z$-fields known at the marker positions, are then interpolated to the FD-grid by linear interpolation.

Using a finite difference code (see e.g. Schmeling & Marquet 1991) the equations are solved with a spatial resolution of 11 km for the stream function, viscosity, and phase fractions and with a four times higher resolution for the temperature.

Our model is 1320 km deep with an aspect ratio of 3. The boundary conditions are free slip at all boundaries, thermally insulated sides and a bottom heat flux of 20 mW m$^{-2}$. The initial temperature field of the model is divided into two parts (according to Christensen 1996): In the left part of the box ($0 \leq x \leq 2244$ km) we have superimposed a cold layer, representing the lithosphere with a conductive temperature gradient, with a typical mantle adiabat. In the right part of the box the adiabatic temperature profile is continued to the surface (cf. Schmeling et al. 1999). Accordingly the surface temperature is fixed at $0^\circ$ C for $0 < x < 2244$ km and 1500$^\circ$ C for $x > 2244$ km. We are interested on the evolution of the slab after the phase transition. Thus, we neglect the effects of an overriding plate, which we assume to be small in comparison to the slab pull forces. Furthermore backarc basins produce only very thin lithospheres. The thickness of the cold layer corresponds to the initial ages of 33, 49, 90, 111, 131, and 165 Myr applying the cooling half-space model.

3 RESULTS

3.1 Selected temperature–depth paths

To study the general kinetic behaviour of the olivine–spinel transition, we first solve the kinetic and grain growth eqs (2, 3) along selected P–T paths (i.e. depth–T paths). First, two simple depth–T paths are prescribed (cf. Rubie & Ross 1994), and second we choose the depth–T paths of three selected markers from one of our dynamic model.

For the two prescribed paths initial temperatures of 470$^\circ$ C (P-T-I) and 670$^\circ$ C (P-T-Th) at 350 km depth are chosen, respectively. A background temperature gradient of 0.7$^\circ$ C km$^{-1}$ is assumed, in agreement with thermal models (Helffrich et al. 1989), which suggest 0.7$^\circ$ C km$^{-1}$ for the coldest part of a subducting slab between 350 and 700 km depth. Each path is calculated with and without the effect of latent heat by modifying the temperature accordingly with the heat release. Thus, the thermal feedback between latent heat and kinetics is consistently accounted for.

For the colder path (P-T-I) (470$^\circ$ C at 350 km depth) (Fig. 1a) the growth rate is small and a significant delay of the phase transition is observed. Olivine is transformed into spinel at significantly greater depths than under equilibrium conditions. For the case with latent heat the phase transition takes place within only a few kilometres (470–517 km Fig. 1b), whereas without latent heat the transition covers a larger depth interval (470–580 km). The strong overexponential increase of volume fraction of spinel for the case with latent heat clearly demonstrates thermal feedback mechanism due to latent heat. For the warmer path (P-T-Th) the phase transition lies closer to equilibrium phase boundary (373–385 km) (Fig. 1b). Due to the smaller amount of the (depth-dependent) latent heat the phase induced temperature jump at the phase boundary is smaller compared to the cold P–T path (Fig. 1a).

As the previous depth-T paths are quite idealized and do not include the effect of heat conduction within the slab or slab dynamics, we now show depth-T paths of three markers of two fully dynamic subduction models. In these models the initial lithospheric thickness is 110 km and corresponds to a trench age of 131 Myr. In one model the latent heat is assumed constant, in the other model it is depth-dependent. A detailed description of these models follows in Section 3.2. The initial positions of the three markers denoted by M1, M7, and M10 are 1584, 1848, and 2112 km away from the left side of the box and at 66, 52.8, and 39.6 km depth. M1 represents a high temperature depth-T path. The phase transition takes place close to thermodynamic equilibrium at a depth range of 360–470 km. The release of latent heat warms up the marker gradually. The depth-dependence of the latent heat has almost no effect on the case of the phase transition. The temperatures are almost identical within the depth range of the phase transition of M1. However, the colder parts of the slab (M7, M10) show a stronger temperature increase of more than 100$^\circ$ C (Fig. 2a) in case of depth-dependent latent heat. The phase transition is kinetically delayed and takes place at greater depth (M7: 480–650 km, M10: 600–730 km). At the early stage of the phase transition ($\xi < 0.1$) the rate of release of latent heat is small and hardly influences the kinetics. At higher $\xi$ the feedback between kinetics and release of latent heat lead to a runaway process and the temperature steeply increases by more than 100$^\circ$ C. Due to this runaway process the phase transition is completed within a narrow depth range (Fig. 2b). Fig. 2(a) also shows the effect of conduction of (latent) heat perpendicular to the slab: The depth-dependent latent heat model produces a higher amount of heat at greater depth in the central part of the slab (solid curves of M7, M10). This heat adds up to the conductive heating of the cold central part of the slab, thereby, reducing the slab-perpendicular heat flux into the slab. This reduction is equivalent to an additional conductive heat flux due to enhanced latent heat out of the deep slab. Therefore, the M1-temperature curves with (solid) and without (dashed) depth-dependent latent heat are different at greater depth even though the corresponding $\xi$-curves are similar and they have experienced the identical latent heating.

3.2 Subduction with and without depth-dependent latent heat

Fig. 3 shows the typical behaviour of a subducting lithosphere with a kinetic olivine–spinel transition with constant latent heat. The
Time-dependent interaction

Figure 1. (a) Depth–T paths for a kinetic phase transition with and without latent heat (P-T-h and P-T-l) and two different initial temperatures. (b) Volume fraction of transformed spinel for the cases shown in (a).

The lithosphere has an initial thickness of 110 km, corresponding to a trench age of 131 Myr. This model will be referred to as M110kin. The ol–sp transition is shown as contour lines of constant spinel fraction originally at a depth range between 450 and 550 km. After about 10 Myr after onset of subduction the ol–sp boundary starts to rise and generates additional negative buoyancy accelerating the subduction rate. Together with the thermally induced slab pull forces these forces generate tensile stresses within the slab section between surface and the ol–sp boundary. These slab pull forces are counteracted by trench resistance and bending resistance force, thus the slab thins and is stretched. Part of the trench resistance stems from the fact that the subduction zone retreats and migrates into the (more realistic) $T = 0^\circ$ C surface boundary condition region. It should be noted that this slab stretching occurs although the slab is still very cold, stress concentrations within the cold slab core reduce the power-law viscosity to values as low as $1.6 \times 10^{21}$ Pa s during the stage of maximum stretching, whereas at later stages it remains still as low $3 \times 10^{23}$ Pa s. In the cold central part of the slab the phase transformation is kinetically delayed, and a MO-wedge forms ($\xi_{sp}$ contour lines are deflected downwards). This olivine wedge generates positive buoyancy forces, slowing down the subduction velocity of the slab. This retarding effect significantly increases as the slab encounters the endothermic sp − pv + mw phase boundary at 660 km depth. The MO-wedge thins by slab-perpendicular migration of the kinetic phase transformation zone and pinches off after about 14.5 Myr. As a result, a remaining isolated MO-blob remains near the 660 km boundary. Actually the formation of this blob is a consequence of the early stage of the slab model: during the earliest stage of subduction the free, cold edge experienced a short period of thickening. Therefore in the subsequent sections, we...
disregard this part of the MO-blob when determining the depth of the MO-wedge.

In the model M110kin the positive buoyancy forces due to both the less dense MO and the downward-deflected endothermic sp $-$ pv + mw boundary are not sufficient to prohibit sinking of the slab into the lower mantle. Conductive heating of the deep cold parts of the slab drive the ol$-$sp phase transition kinetics and reduce the amount of MO, in particular within the pinched region. Buoyancy of the residual amount of MO within the leading edge of the slab continues to contribute to the bending and flattening of the slab near 660 km depth. The maximum penetration depth of the MO of model M110kin reaches 730 km after about 14 Myr, and 690 km after a second phase of acceleration (after about 17 Myr), not accounting for the deep isolated blob of MO as mentioned above. As will be discussed, subsequently, these depths are maximum depths as M110kin does not account for depth-dependent latent heat.

In the previous model (M110kin) latent heat was assumed constant leading to a temperature jump of 48°C at the exothermic
ol–sp transition and to a negative temperature jump of $-49^\circ$C at the endothermic sp $- pv + mw$ transition. In the following model M110Lat the effect of depth-dependent latent heat according to eq. (10) will be considered. The evolution of a subducting slab with a kinetic phase transition with depth-dependent latent heat (model M110Lat) differs from model M110kin (constant latent heat) after the first occurrence of MO (Fig. 4). As increased latent heat accelerated phase transition kinetics, the size of the MO-wedge of M110Lat is significantly smaller compared to M110kin at the time it reaches the spinel–perovskite phase boundary. However, the arrival

Figure 3. Model M110kin. Evolution of subduction of a 131 Myr old 110 km thick lithosphere. The colour and a few black contour lines show the temperature field. The white lines (between 410 and 570 km depth in the undisturbed mantle) are lines of constant sp-fraction from 10 to 90 per cent, thus depict the ol–sp transition zone. The pink lines (between 620 and 660 km depth in the undisturbed mantle) are lines of constant pv + mw fraction, thus depict the sp $- pv + mw$ transition zone.
Figure 4. Same as Fig. 3 but for depth-dependent latent heat (M110Lat).
time of the MO-wedge of M110Lat is delayed (around 13.6 Myr) compared to model M110Kin (around 13 Myr). Obviously, the effect of slowing down the M110Lat-slab by thermal buoyancy associated with the higher amount of latent heat is stronger than slowing down the M110Kin-slab by a larger MO-wedge. After about 14 Myr the MO-wedge is strongly thinned within the transition zone region. While after 14 Myr M110Kin still contains large amounts of MO near the spinel–perovskite boundary, in M110Lat almost all MO is transformed to spinel at that depth, even within the leading edge of the slab. After 14.7 Myr the deep MO of M110Lat is completely transformed into spinel. Thus the maximum depth of the MO-wedge has risen to about 550 km. In comparison, at the same time M110Kin still has a MO-blob sitting at 740 km depth directly above the sp–pv boundary, and the tip of the pinched MO-wedge is at 590 km depth, that is, significantly deeper than for M110Lat. Upon further subduction the tip of the MO-wedge of M110Lat reverses and moves downwards again reaching a depth of 650 km after 16.9 Myr. After this stage the MO-wedge retreats upwards to 610 km depth after 17.1 Myr, followed by a phase of further diminishing. Altogether, the effect of depth-dependent latent heat is characterized by a significant reduction of MO, both in volume and depth extent. Nevertheless, occasionally maximum depths of MO-material below 650 km are still possible.

3.3 Temporal behaviour of subduction velocity and depth of metastable olivine wedges

The previous section demonstrated the complex behaviour of the MO due to the dynamics of a subducting slab, and the importance of latent heat. In this section, we show that this behaviour can be described by a dynamic feedback process leading to a cyclic behaviour, which can be described in a velocity–depth phase diagram.

The temporal behaviour of our models is controlled by thermal buoyancy, kinetics of the ol–sp phase transition, and the sp–pv phase transition in thermodynamic equilibrium. Taking the model of a subducting 131 Myr old lithosphere of 110 km thickness including kinetic phase transition and depth-dependent latent heat as a reference model (M110Lat), we investigate the temporal behaviour of the root velocity in 410 km depth (410-velocity) of the model and the maximum depth of the MO-wedge. The first phase of subduction of the cold lithosphere down the first contact with the ol–sp boundary is characterized by thermally induced negative buoyancy. As the cold slab reaches the transition zone, first the moderately cold frontal part of the slab starts to interact with the ol–sp phase boundary. The exothermic ol–sp phase transformation takes place in thermodynamic equilibrium, and the ol–sp boundary is lifted upward. The resulting negative buoyancy forces accelerate the slab velocity and the 410-velocity up to 4.8 cm yr\(^{-1}\) (Fig. 5b). During this phase subsequently colder parts of the slab enter the phase transition region, and the kinetically delayed phase transition deforms the upper boundary of the phase transition region downwards. Within the coldest part of the slab the width of the phase transition region is drastically reduced from about 410–550 km at thermodynamic equilibrium to about 10 km width after 11.3 Myr (Fig. 5a). Due to the kinetical delay of the phase transformation in the cold part of the slab a MO-wedge forms and penetrates deeply into the transition zone, reaching 700 km after 14 Myr. Due to the positive buoyancy of this MO and, somewhat later, the increasing interaction with the endothermic phase boundary sp – pv + mw at 660 km depth, the subduction velocity (here the 410-velocity) starts to drop from 4.8 to about 3 cm yr\(^{-1}\) at 13.6 Myr. After 13.6 Myr the 410-velocity remains approximately constant for a period of more than 1 Myr and the MO transforms progressively into spinel at great depth by kinetic phase transformation aided by latent heat. Consequently, the depth of the MO-wedge rises again (simultaneously the deep isolated MO-blob disintegrates). During this phase trench roll back with about 2.8 cm yr\(^{-1}\) is observed (Fig. 4, second diagram from top). As the amount of MO decreases during this phase, the retarding buoyancy forces decrease and the slab accelerates slightly again until 16 Myr. Consequently, the depth of the MO-wedge increases again.

It is interesting to compare this behaviour to slab dynamics without kinetically delayed ol–sp transformation (model ‘nokin,’ see also Tetzlaff & Schmeling, 2000, for such a comparison). In the nokin-model (Fig. 5b, dashed) the slab accelerates to significantly higher values (>5.5 cm yr\(^{-1}\)) during the first phase, which results in fast deployment of cold material near the 660-boundary. This cold material strongly inhibits further subduction due to the inhibiting effect of the exothermic phase boundary, resulting in a second phase with a much stronger slowing down of subduction, down to values of 0.5 cm yr\(^{-1}\). Comparing M110Kin (without depth-dependence of latent heat) with M110Lat (with depth-dependence of latent heat) shows that the principal temporal behaviour of the depth of MO and of the subduction velocity are similar (Figs 5 and 6). In M110Kin the MO-wedge subducts into greater depth (750 km) than in M110Lat. Whereas in M110Lat the subduction velocity reaches an almost constant value between 13.5 and 16 Myr, the larger MO-blob at the leading edge of the M110Kin-model continues to slow down the slab and no quasi-steady state is reached for this period.

4 DISCUSSION

4.1 Velocity–depth phase loops of metastable olivine wedges

To better illustrate the feed back between dynamics and kinetic phase transformation during subduction, the 410-velocity and the maximum depth of the MO have been plotted as phase loops for some further models, which have a 33, 90, and 131 Myr lithosphere prior to subduction (Figs 7, 8, and 9). The maximum depth of MO is given again as the maximum depth of 0.9 and 0.1 per cent spinel fractions. Both the models M90xxx and M110xxx (where xxx is 33, 90, and 131) show a profound reduction of MO, both in volume and depth extent. Nevertheless, occasionally maximum depths of MO-material below 650 km are still possible.
As a general observation we note that subsequent loops seem to be shifted towards slower velocities and shallower MO-depths. Obviously, progressive subduction is associated with increasing interaction with the high viscous lower mantle and the whole process of subduction is decelerated \cite{Ennsetal2005}. However, we note that the amplitude of the second loops of each model with loops does not decreases with respect to the previous loop. This observation is an indication of the inherent time dependence of subduction of old (90 Myr or older) lithosphere modulated by the varying depth of the MO-wedge. Comparing the loops of models with depth-dependent and not depth-dependent latent heat shows that the magnitude of the loops is not reduced by the depth-dependent latent heat, only the position of the loops is shifted towards slightly higher velocities, and the maximum depth of MO prior to each loop is somewhat smaller. The inherent time dependence seen in the models can be understood as follows. During the fast, but decelerating phase I cold MO is transported deeply into the transition zone and transforms at great depth. Cold, non-transformed material is effectively advected downward, the slab immediately around and below the deep MO is colder than it would be without nearby MO. As a result, also the sp–pv boundary is deflected downward, retarding the subduction even more. During phase II advection is retarded and the MO transforms into spinel and the slab and its surroundings get warmer compared to phase I. This effect is increased especially for the case of depth-dependent latent heat. The cooler parts from phase I continue to slowly sink and increase to interact with the sp–pv boundary. This is the phase where flattening occurs, and penetration into the lower mantle is hampered. Depending on the timescale of advection, the warmer parts of the slab from phase II approach the sp–pv boundary during phase III and influence the equilibrium phase transition sp–pv. Due to a reduction of the downward deflection, penetration into the lower mantle is accelerated, and the slab speeds up. At the same time, MO starts to enter the transition zone again, delaying the release of latent heat, advecting cold material into the transition zone, thereby, effectively cooling the slab. After the warmed material has passed the sp–pv boundary, the situation is similar to phase I (see also the sketch in Fig. 10). Thus, the interplay between the subduction through the transition zone, kinetic phase transformation, and temperature-dependent deflection of the sp–pv boundary contains three different timescales: heat diffusion near the sp–pv

\begin{figure}
\centering
\includegraphics[width=\textwidth]{figure5.png}
\caption{Top: Maximum depth of MO (dashed: 0.9, solid: 0.1) of model M110Lat as a function of time. Bottom: Velocity in 410 km depth of M110Lat as a function of time as compared to the velocity (dashed) obtained for a model without kinetically delayed ol–sp transformation (nokin-model).}
\end{figure}
boundary and laterally within the slab, advection due to sinking of the slab and delay of phase transformation. As a consequence, periodic behaviour of subduction and penetration into the lower mantle is observed. Longer runs (which require modified boundary conditions to ensure a continuously generated lithosphere) are needed to proof this hypothesis. Clearly the model with a thin lithosphere M55xxx (where xxx is either kin or Lat) do not exhibit the characteristic loop behaviour discussed above (Fig. 7). As the slab encounters the phase transition, subduction velocity increases, and the phase boundary is deflected downward, but the resulting MO-wedge is only small or not existent.

It is interesting to look at the differences between the time-dependence of the models with and without depth-dependence of latent heat. Although models M90kin and M90Lat are quite similar, inspection of Fig. 9 shows that the sense of rotation of the first loop is different for the M110Lat and the M110kin models. The clockwise rotation of the M110kin loop is due to continued slowing of the slab even after MO has risen to shallower depth. This decelerating

Figure 6. Top: Maximum depth of MO (dashed: 0.9, solid: 0.1) of model M110kin as a function of time. Bottom: Velocity in 410 km depth of M110kin as a function of time.
From the above discussion one may ask whether the difference between the cases with constant and with depth dependent latent heat is due to the depth dependence or due to the difference in absolute value of latent heat. Figs 8 and 9 show that the first part of the loop (phase III) are almost identical and the differences occur at the end of phase I (i.e. when deep portions of MO transform at great depth). We conjecture that if the latent heat would be higher but constant for all depth, also phase II would look differently and it is more difficult that large amounts of MO would reach great depth.
Thus, the effect of depth dependence is to ensure the transport of MO into greater depth with lower latent heat production at shallow depth, but maximizing the latent heat effect at greater depth as the MO interacts with the sp–pv interface.

4.2 Maximum depth of metastable olivine

The maximum depth of the MO obtained for each time-dependent run increases with the thickness d (or age) of the lithosphere (Fig. 11). The increase is linear for a depth (thickness) range 550–700 km (55–100 km), and flattens for greater thickness as the MO does not deeply penetrate into the lower mantle. The effect of depth-dependent latent heat is to reduce the maximum depth by about 50 km. The linear part of the obtained curves can be approximated by the equation $z_{MO_{max}} = c_1d + c_2$ with the constants $c_1 = 3.7$ or 3.3 and $c_2 = 340$ or 350 km for the cases with constant or depth-dependent latent heat, respectively.

It is interesting to note that the relationship between $z_{MO_{max}}$ and $d$ is approximately linear, that is, does not scale in the same way as the depth of a specific isotherm scales with $d$ in case of free
subduction without phase changes. The maximum depth of a certain temperature within a subducting slab scales with the product \( v_{\text{sub}} t \) (McKenzie 1969), the age \( t \) scales with \( d^2 \), and \( v \) scales with the Stokes velocity \( v_{\text{stokes}} \) of a free slab. This velocity scales with \( S \), a representative surface of the sinking slab (Guillou-Frottier et al. 1995). \( S \) scales either with \( d^2 \) or with \( d \cdot L \) where \( L \) is a length scale associated with the length of the slab or lateral spatial constraints (e.g. thickness of mantle layer). For simplicity we assume that \( v_{\text{stokes}} \) scales with \( d^n \), where \( n \) lies between 1 and 2, thus, \( v \) scales with \( t^2 \). This scaling with \( n \) close to 2 is seen, for example, in the dynamic models of free slabs without MO by Tetzlaff & Schmeling (2000) or Capitanio et al. (2007). This implies that the depth of a specific temperature isotherm in a free slab should scale with \( d^{n+2} \). In contrast, we do not see such a scaling for \( z_{\text{MOMax}} \) in the present models. We explain this behaviour (1) a slowing down of subduction by the ‘parachute effect’ (Kirby et al. 1996; Tetzlaff & Schmeling 2000), which effectively reduces the above mentioned \( n \)-value to values between 0 and 1, and (2) the inherent time-dependence due to the dynamic feedback between the buoyant MO, the subduction velocity and the interaction with the 660 km discontinuity: during

Figure 9. Top: Velocity in 410 km depth of M110kin as a function of maximum depth of MO (dashed: 0.9, solid: 0.1). Bottom: Velocity in 410 km depth of M110Lat as a function of maximum depth of MO (dashed: 0.9, solid: 0.1).
Figure 10. Sketch of the different phases (I–IV) of the phase loops shown in Figs 8 and 9 and described in Section 4.1. c: cold region, w: warm region, $F_{th}$: thermal buoyancy force, $F_{MO}$: buoyancy force of MO, $F_{660}$: buoyancy force at the 660 km discontinuity.

Figure 11. Age of lithosphere as a function of maximum depth of MO (dashdot: models using a simplified phase diagram to approximate the kinetic equation, solid : Mxxxkin, dashed: MxxxLat).
the decelerating phase I due to MO-buoyancy and interaction with the 660 discontinuity (see previous section) thermal diffusion accommodated by release of latent heat prevents deep penetration of MO and keeps $z_{\text{MOmax}}$ below a non-linear increase with $d$. Thus, kinematical models which disregard the interaction with the 660-discontinuity and the associated episodicity may overestimate the depth of MO.

The results can also be compared with the earlier results of Tetzlaff & Schmeling (2000) who used a simplified phase diagram to approximate the kinetics of the olivine–spinel transition. In those models $z_{\text{MOmax}}$ was somewhat overestimated for slabs of moderate age. However, as the interaction with the 660-discontinuity was accounted for the maximum depth of old slabs was only slightly overestimated, thus the conclusions in that paper are still valid. A comparison of our results with the models of Mosenfelder et al. (2001) shows that our $z_{\text{MOmax}}$ values are greater than theirs, and the formation of a MO-wedge occurs at smaller thermal parameters. The reason for this difference is a different choice of kinetic parameters (e.g. $k_0$ and $\Delta H_f$), associated with a 5 times faster growth rate of the spinel phase in their models compared to our models and the models of Rubie & Ross (1994).

4.3 Application to the Earth

Our models predict that the maximum depth of MO may exceed the equilibrium depth of the $\text{sp} - \text{pv} + \text{mw}$ boundary for lithospheric ages greater than 90 Myr (constant latent heat) or 100 Myr (depth-dependent latent heat). Maximum depths exceeding 750 or 720 km seem to be possible. Such depths are consistent with the depths of the deepest earthquakes observed at 700 km (Kirby et al. 1996). Therefore, we suggest that the transformation of MO bodies embedded in a $\text{pv} + \text{mw}$ lower mantle and the associated changes in the stress field might be the cause for these earthquakes. Our models predict a pronounced time dependence of the maximum depth of MO and the subduction velocity following characteristic loops as discussed in the previous sections. As mentioned there, this prediction is verified for the first few loops, but we anticipate a high potential of oscillatory time dependence also for later stages. If this is the case, then for a given lithospheric age, different maximum depths of MO are possible, depending on the actual position on the loop. Loops of old lithospheres cover, for example, a deeper depth range of 715–595 km (first loop of 165 Myr old lithosphere) than younger lithospheres, for example, 650–515 km (first loop of 90 Myr lithosphere). This is consistent with observed maximum depths of earthquakes for lithospheres with different age (Fig. 12). In this figure the depth of the deepest MO of models with different lithospheric ages (with and without depth-dependence of latent heat) is compared with the deepest earthquakes at different sections of various subduction zones (data taken from Gorbatov and Kostoglodov 1997). The vertical extent of the first loops (cf. Figs 8 and 9) are also indicated by the dashed vertical bars. Subsequent loops would extend these bars to even shallower depths.

First of all, the deepest earthquakes of the Mariana, Tonga, and Chile extend as deep as the maximum depth of the modelled MO, following roughly the predicted age dependence. The deepest earthquakes of Japan, Kuriles, and Kamchatka reach as deep as the shallow part of the first or later loops of the corresponding numerical models. Depth variations among different sections of the Mariana and the Chile subduction zones are roughly 1/2 of the depth extent of one MO-loop, whereas for Tonga all sections have earthquakes in agreement with the maximum depth of the corresponding loops. From this picture one may speculate that different subduction zones or even different sections within one subduction zone may represent different stages of the corresponding depth–velocity loops.

If this is the case, our models predicts Tonga and Chile (non-detached) to be in phase I, Kurile and Japan in phase III, and Marianas and Chile (detached) in phase II (although a combination of III and I is also possible). The Chile (non-detached, detached) region might be in the special situation of phase II, in which there exists an isolated deep pocket of seismically active MO, whereas at intermediate depth the MO has already been transformed into spinel (cf. Fig. 3, $\tau = 13.6$ Myr), resulting in a seismic gap within the tomographically visible slab within the upper part of the transition zone (Fukao et al. 2001).

It is interesting to compare our results with observed temporal variations of subduction rates. Typical temporal variations of convergence rates at various trenches have timescales of the order of several tens of Myr down to at least 5 Myr (Sdrolias & Müller 2006).

![Figure 12. Depth of the deepest earthquakes as a function of age at the trench Gorbatov 1997, data are shown for subductions zones with a thermal parameter greater than 3000 km only, dashed: depth of the deepest MO of Mxxxkin, solid: depth of the deepest MO of MxxxLat.](https://academic.oup.com/gji/article-abstract/178/2/826/625708)
Shorter timescales have not been resolved yet. For example, during the last 50 Myr the Tonga–Kermadec subduction zone appears to show a 20 Myr period on which short-term variations (one phase of speed up 27 Myr ago and one that started 5 Myr ago) are superimposed (Sdrolias & Müller 2006). According to the same authors the Mariana subduction rate seems to have slowed down in the last 5 Myr, and Chile subduction shows a deceleration during the last 15 Myr. The characteristic time variations predicted by our models are of the order of 3–4 Myr. In comparison, the characteristic times of fully dynamic subduction models without phase transitions (Enns et al. 2005) are of the order of 5 Myr to more than 20 Myr for slabs with various stiffnesses. The time dependence in those models is due to the dynamic interaction of the slab with the rheological interface at 660 km depth, whereas in our models this behaviour is additionally modulated by the kinetics of the phase change. Thus, we may interpret the observed long-term subduction rate variations being influenced by the rheological interaction between slab and the 660 discontinuity, whereas the phase kinetics might be invoked to explain temporal variations on shorter timescales. If this is true, the recent slowing down of the Mariana and Chile subduction zones is consistent with the decelerating phase II of a MO depth–velocity loop as discussed in the previous sections. However, the accelerating (‘phase II’) northern Tonga case associated with the deepest earthquakes (‘phase I’) does not fit into this simple picture. Probably the flattening of the dip angle associated with the 3-D slab geometry near the northern end of the Tonga trench modifies the depth–velocity relation.

The strong time dependence of subduction combined with the associated time-dependent release of high amounts of latent heat in the depth-dependent latent heat cases leads to an interesting prediction: Different parts of the same slab may have experienced different intensities of latent heat release, thus along-slab temperature variations of the order of 100 K might be expected for the deeper parts of the slab. Although other effects such as folding near the 660 km boundary and subsequent sinking of the folded slab into the lower mantle (Enns et al. 2005) might also lead to characteristic temperature variations, the above effect due to episodic release of different amounts of latent heat might still be visible by seismic tomography and might, at least partly, explain apparently disconnected or necking regions often seen within the uppermost part of the lower mantle (see e.g. the tomographic images of Fukao et al. (2001) for the Java, Mariana, Tonga, Japan subduction zones). It should, however, be noted that such along-slab temperature variations are a second-order effect compared to the temperature contrast associated with the slab itself. It should be noted that the present models have assumed dry conditions. From laboratory experiments it is known that the effect of water speeds up the phase kinetics proportional to the OH-content to the power of 3.2 (Hosoya et al. 2005). Those results indicate that a water content of more than 500–1000 weight ppm H2O could significantly rise the maximum depth of metastable olivine.

5 CONCLUSION

Dynamic subduction models have been presented that account for the kinetics of the olivine–spinel transformation and the equilibri um transformation of spinel into perovskite and magnesiowustite. A dynamic feedback mechanism is found between buoyancy due to metastable olivine, temperature and deflection of the spinel–perovskite/magnesiowustite phase boundary, kinetics of the phase transition and the release of latent heat. This mechanism modulates the ‘parachute’-effect of the metastable olivine in a subducting slab and leads to characteristic episodic subduction behaviour, namely temporal variations of the depth of metastable olivine and, with a time delay, variations of the subduction velocity. This mechanism allows to explain why in some subduction zones earthquakes occur as deep as 700 km, whereas in other sections of the same subduction zones, or in other subduction zones of similar age the deepest earthquakes are considerably shallower. Further conclusions are

(1) Subduction velocity variations due to this feed back mechanisms have time periods of the order of 3–4 Myr. Amplitudes are consistent with observations.

(2) The maximum depth of metastable olivine increases roughly linearly with lithospheric thickness (Smax = c1 d + c2 with the constants c1 = 3.7 or 3.3 and c2 = 340 or 350 km for the cases with constant or depth-dependent latent). The maximum depth may reach as deep as 720–750 km.

(3) The increase of maximum depth of metastable olivine with increasing lithosphere age is in good agreement with the depth of the deepest earthquakes.

(4) Different subduction zones may represent different stages of ‘velocity–depth phase loops’ of metastable olivine.

(5) Depth-dependent latent heat leads to an upward shift of the metastable phase transition by about 50 km.

(6) Depth-dependent latent heat release during episodic subduction may produce along-slab temperature variations of the order of 100 K (superimposed on the steady state temperature variations).

ACKNOWLEDGMENTS

We gratefully acknowledge the discussions with D. Rubie and J. Mosenfelder and a constructive review by J. van Hunen. We also acknowledge the support by the Deutsche Forschungsgesellschaft (DFG grant Schm872/6-2).

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