ABSTRACT

A finite element method is applied to model the thermal structure of the subducted Pacific plate and overlying mantle wedge beneath the southern part of the Kamchatka peninsula. A numerical scheme solves a system of 2D Navier-Stokes equations and a 2D steady state heat transfer equation.

A model with isoviscous mantle exposed very low temperatures (~ 800ºC) in the mantle wedge, which cannot account for magma generation below the volcanic belt. Instead, a model with strong temperature-dependent viscosity shows a rise in the temperature in the wedge. At a temperature of more than 1300ºC beneath the active volcanic chain, melting of wedge peridotite becomes possible. Although the subducting slab below the Kamchatka peninsula is rather old (~ 70 Myr), some frictional heating (μ = 0.034) along the interface between the subducting oceanic slab and the overlying Kamchatka peninsula lithosphere would be enough to melt subducted sediments. Dehydration (> 5 wt% H2O release) occurs in the subducting
slab because of metamorphic changes. As a consequence, hydration of the mantle wedge peridotite might produce melt, which may rise to the base of the continental crust as diapir-like blobs.

Considering that melting processes in the subducting plate generate most of the volcanic material, we developed a dynamic model which simulates the migration of partially melted buoyant material in the form of blobs in the viscous mantle wedge flow. Blobs with a diameter of 0.4 - 10.0 km rise to the base of the continental lithosphere within 0.002 - 10 Myr depending on the blob diameter and surrounding viscosity.

The thermal structure obtained in the model with temperature dependent viscosity is used to estimate seismic P-wave velocity anomalies (referenced to PREM) associated with subduction beneath Kamchatka. A low velocity zone (~ - 7% velocity anomaly) is obtained beneath the volcanic belt and a high velocity anomaly (~ 4%) for the cold subducted lithosphere. These results agree with seismic tomography results from P-wave arrivals.

**Keywords:** Kamchatka subduction zone, thermal models, mantle wedge flow, blobs, tomographic imaging.
GENERAL ASPECTS OF SUBDUCTION ZONE DYNAMICS

Mantle plumes are frequently assumed to come from the transition zone or from the core-mantle boundary. There is another place inside the Earth from where these thermal or chemical instabilities might develop: the top layer of the subducting slabs where strong dehydration and melting is expected. This chapter aims to study the subduction magmatism through numerical modeling of the mantle wedge thermal structure and magma transport with a diapir model.


The object of the present study is to model one of the oldest and most interesting subduction zones; the Kamchatka subduction zone (KSZ), in which thermal structure is poorly known up to now. Kamchatka thermal structure is analyzed in models with constant viscosity (isoviscosity) and strong temperature-dependent viscosity of the asthenosphere.

A distinctive feature of the thermal regime associated with subduction zones is the inverted thermal gradient just above the slab in the mantle wedge where the most important and intensive chemical and thermal exchanges occur. In this region the material transferred from the oceanic slab to the mantle wedge enriches the source region of arc lavas with incompatible elements and volatiles (Plank and Langmuir, 1993; Stolper and Newman, 1994; Johnson and Plank, 1999; Eiler et al., 2000). An influx of volatiles from metamorphosed oceanic crust and sediments triggers partial melting of peridotite above the subducted slab (Tatsumi, 1986;

The latest insights revealed by new developed high-resolution dynamical models for subduction zones with 0.5 billion markers of Gerya and Yuen (2003) show that the hydration and related melting introduce a source of chemical buoyancy on the 1 - 10 km thick layer on top of the subducting slab. These plumes, with diameter up to 10 km, penetrate the hot mantle wedge with velocities of ~ 10 cm/yr.

Various mechanisms were proposed for rapid magma transport from the mantle wedge toward the surface: through channelized network (channels diameter of 1 - 100 m) (Spiegelman and Kelemen, 2003), through fractures and shear zones in the mantle (Shaw, 1980), through a fractal tree network (very thin channels with 1 - 4 mm diameter) (Hart, 1993) and the dike transport mechanism in the mantle (Rubin, 1993). The model of Shaw (1980) proposes fracture zones inside the mantle wedge. The presence of such fracture system is unlikely to exist in a very hot mantle wedge (> 1300°C). The mantle wedge peridotite behaves completely ductile for temperature above ~ 700°C. Spiegelman and Kelemen (2003) provide a model capable to explain the chemical and spatial variability of lava samples taken from mid ocean ridges. The geodynamics of subduction zones is quite different from the one of mid ocean ridges, basically the strong convection beneath the volcanic arcs represents one of the main differences. The applicability of this model with magma transport velocity of ~ 1.5 m/year through channels with diameters < 100 m proposed by Spiegelman and Kelemen (2003) needs to be further investigated in the presence of strong convecting systems as the mantle wedge beneath volcanic arcs. Although the other models (Hart, 1993; Rubin, 1993) predict fast magma transport toward the surface through very narrow channels,
they do not take into account the presence of strong mantle wedge convection which affect the continuation of these channels. For this reason, in the present study, for magma generation and propagation in the southern Kamchatka subduction zone, concept of positive buoyant diapirs (blobs) that migrate in the mantle wedge flow induced by the subducting slab is applied. We used for the modeling the numerical scheme of Manea et al. (2004-B).

Partial melting of fluid-saturated peridotite at low temperature on top of the subducted slab may generate H\textsubscript{2}O-rich partial melt. In order for melt to progress, ascending blobs have to be efficiently heated as they move in the mantle wedge (Gaetani and Grove, 2003). Conductive thermal evolution of these diapirs as they ascend toward the base of the continental crust is modeled.

Mantle wedge viscosity is an important factor that controls the migration of diapirs in the mantle wedge. Dissolved H\textsubscript{2}O and the partial melt might lower mantle viscosity below 10\textsuperscript{18} Pa s (Kelemen et al., 2003). Moreover, Hirth and Kohlstedt (2003) showed that experimentally measured viscosities for olivine at upper mantle pressures and temperatures are in the range 10\textsuperscript{17} - 10\textsuperscript{21} Pa s. Thermal modeling is applied to investigate the possibility of melting slab fluid-saturated sediments (wet solidus from Nichols et al., 1994) and mantle-wedge wet peridotite (wet solidus from: Mysen and Boettcher, 1975; Wyllie, 1979), and dehydration along the slab-wedge interface beneath Kamchatka.

The main constraint on mantle-wedge temperature distribution could be P-wave seismic tomography. Zhao et al. (1992, 1997) observed a low-velocity zone (-6% P-wave velocity anomaly) beneath the Tonga and NE Japan volcanic arcs, which might be related to upper-mantle melting below the base of the continental crust. Gorbatov et al. (1999) detected a low-velocity zone (-7% P-wave velocity perturbation) beneath the active volcanic chain in Kamchatka. Tomography images of Northeast Japan (Tamura et al., 2002; Zhao, 2001) reveal a limited low velocity region that is in contact directly with the slab surface (-2 - 3% velocity anomaly) at depths of 80 - 100 km.
Following Karato (1993), the temperature dependence of seismic wave velocities can be used to estimate velocity anomalies from thermal models. An agreement between the velocity perturbation beneath the volcanic arc (shape and magnitude) observed in the seismic tomography anomaly and the tomography anomaly estimation from thermal modeling might be an indication of a satisfactory estimate of the thermal regime in the mantle wedge.
The Kamchatka peninsula has one of the most active volcanic chains in the world and is a dynamic convergent margin where the Pacific plate (PAC) subducts beneath the North-American plate. The PAC subducts westward at a dip angle of ~55° from 50°N to 54°N, and at a rate of ~ 7.8 cm/year (Gorbatov et al., 1999). The age of the subducting plate along the Kamchatka trench varies between 70 and 100 Myr. The dip of the Wadati-Benioff zone varies along the KSZ from ~ 55° in the south to ~ 35° in the north. Heat flow data (Smirnov and Sugrobov, 1979; 1989-A,B) suggest that the thermal thickness of the subducted plate is up to 75% smaller in the north of the peninsula then in the south. Thus the effective age (thermally defined) is less there, because of the geological age (Renkin and Sclater, 1988). In order to avoid the effect of oceanic plate rejuvenation in the north of the KSZ, the present paper explores a 2D profile normal to the trench in the southern part of Kamchatka (Fig. 1), far from this important thermal anomaly. The seismicity and structure of the KZS was studied in detail by Gorbatov et al. (1997). The volcanic front, from 50°N to 54°N, is trench-parallel and corresponds to a depth of the subducted slab of about 90 to 140 km.

Southern Kamchatka is separated into an eastern active volcanic chain and the western inactive tectonically and volcanically Sredinny Range (Fedotov & Masurenkov, 1991). Numerous active and inactive volcanoes, which form the Eastern Volcanic Front (EFV), are situated above the subducting slab at depth where partial melting in the mantle wedge is induced by fluids from slab dehydration.

Observations of a slab-melt chemical signature are mainly restricted to the northern volcanoes as Sheveluch, which is the only active volcano with an adakite magma signature. The tectonic reconstruction of Kamchatka-Aleutian corner (Park et al., 2002) strongly suggests that the extinct subduction zone (just beyond the edge of the slab near the Aleutian junction) involved a shallow-dipping young slab. The melting of subducted slab and fluids from slab dehydration produced an
adakites type of arc volcanism. However, in southern Kamchatka no adakites were found, suggesting that in this region the slab is not melting. The lack of this style of volcanism would be a very important constraint on the subducting slab thermal structure. While melting of subducted basaltic crust is unlikely to occur beneath southern Kamchatka, there are evidences of sediment-melt chemical signature in Kamchatka, although this contribution is small (< 1%) compared with other volcanic arcs (Park et al., 2002).
MODELING PROCEDURE

A system of 2D Navier-Stokes equations and the 2D steady state heat transfer equation are solved for the south Kamchatka cross section (Fig. 1) using the numerical scheme proposed by Manea et al. (2004-B). The strong temperature-dependence of viscosity imposed in the present modeling, corresponding to diffusion creep of olivine, has the following form:

\[ \eta = \eta_0 \cdot e^{\frac{E_a}{R \cdot T_0}} \cdot \left[ \frac{T_0}{T} - 1 \right] \]  

(1)

where:
\( \eta \) - mantle wedge viscosity (Pa s),
\( \eta_0 \) - mantle wedge viscosity at the potential temperature \( T_0 \) (1.0\( \times \)10\(^{20}\) Pa s),
\( T_0 \) - mantle wedge potential temperature (1,450°C),
\( E_a \) - activation energy for olivine (300 kJ/mol) (Karato and Wu, 1993),
\( R \) - universal gas constant (8.31451 J/mol \(^{\circ}\)K),
\( T \) - temperature (°C).

A finite-element grid extends from 25 km seaward of the trench up to 375 km landward of it, and consists of 12,000 triangular elements with higher resolution in the tip of the wedge (Fig. 2). A benchmark with various grid resolution of the present numerical scheme (Manea et al., 2004-B) confirms that a numerical error of less than 5°C is introduced in the thermal models when 12,000 triangles are used. The lower edge of the grid follows the shape of the subducting plate upper surface at a constant distance of 100 km (Fig. 3). The model consists of five thermo-stratigraphic units as follows: upper continental crust, lower continental crust, oceanic lithosphere, oceanic sediments, and mantle wedge.
A summary of the thermal parameters used is presented in Table 1 (compilation from: Peacock and Wang, 1999; Smith et al., 1979; Vacquier et al., 1967). The continental crust in Kamchatka is divided into two layers: the upper crust (0 - 15 km depth) and lower crust (15 - 35 km depth). These depths are consistent with values inferred from 1D tomographic inversion by Gorbatov et al. (2000). A recent paper of Levin et al. (2002) shows a Moho depth range of 30 - 40 km across the entire Kamchatka peninsula. In this chapter we used a constant Moho depth of 35 km. The shape and dip of the subducting plate beneath the volcanic arc are constrained by earthquake hypocenter distribution. A 1.5 km-thick sediment layer is included in the model (Dickinson, 1978; Selivestrov, 1983).

The upper and lower boundaries are maintained at constant temperatures of 0ºC at surface and of 1,450ºC in the asthenosphere, respectively (Fig. 3). The left, landward vertical boundary condition is defined by an 18.5ºC/km thermal gradient for the continental crust. Below the 35 km depth, the left boundary condition is represented by a low thermal gradient of 5.5ºC/km down to the depth of 180 km. Beneath 180 km depth no horizontal conductive heat flow is specified. Underneath the Moho (35 km), for the left boundary, corresponding to the mantle wedge, zero traction is assumed. At the intersection between the subducted slab and the left boundary, the velocity of the subducting slab is assumed.

The right, seaward boundary condition is a one-dimensional geotherm calculated for the oceanic plate by allowing a half-space to cool from zero age to the oceanic plate age at the trench. This geotherm is obtained using a time-dependent sedimentation history (Wang and Davis, 1992) and assuming a constant porosity-depth profile of the sediment column with a uniform sediment thickness of 1.5 km at the trench.

Since the Stokes equations are applied only in the mantle wedge, the region between Moho and the slab surface is entirely involved in the flow induced by the subducting slab. This is consistent with the conclusion of Levin et al. (2002), suggesting that the mantle lithosphere beneath Kamchatka is actively deforming.
In terms of displacement, the velocity of the oceanic plate is taken with respect to the continental plate. Thus the convergence rate of 7.8 cm/year between the PAC and North American plates is used for the KSZ (Renkin and Sclater, 1988). Although we assume the motion of the subducting slab as parallel to the dip, the slab can have an additional downward directed velocity component due to slab bending or slab rollback. On the other hand, slab can also have upward velocity component in case of swallowing of the subduction angle with time due to the slab unbending. In our models we are dealing with a steady state model and therefore, do not consider additional velocity component due to the slab bending/unbending. The PAC age at the trench is 72 Myr according the estimation of Gorbatov and Kostoglodov (1997).

We consider two different models: the first has constant viscosity (isoviscosity) in the mantle wedge and the second one has strong temperature-dependent viscosity (diffusion creep of olivine). In the second model, the system of equations becomes strongly nonlinear. To deal with this difficulty Picard iterations are applied. In order to achieve a convergent solution a cut-off viscosity of $10^{24}$ Pa s for temperature less than 1,000 ºC is used.

A long-term continuous sliding between the subducting and continental plates along the thrust fault should produce frictional heating. We introduced into the models a small degree of frictional heating using Byerlee’s friction law (Byerlee, 1978). Frictional heating cease at a depth of 35 km, where the oceanic plate and the mantle wedge came into first contact (Fig. 3). We impose this depth limitation of frictional heating because the contact between Moho (35 km) and the slab surface represents a maximum extent where interplate earthquakes might occur. The tip of the mantle wedge is subject of mantle serpentinization which has a completely ductile behavior and therefore decoupling the subducting and overriding plates. The location of serpentinized mantle wedge tip is critical because it controls the down-dip extension of the interplate earthquakes (Manea et al., 2004-A).

Volumetric shear heating is calculated as follows:
\[ Q_{sh} = \frac{\tau \cdot v}{w}, \]

where:

- \( Q_{sh} \) - volumetric shear heating (mW/m\(^3\)),
- \( \tau \) - shear stress:
  \[ \tau = 0.85 \cdot \sigma_n \cdot (1 - \lambda) \quad \text{for} \quad \sigma_n \cdot (1 - \lambda) \leq 200 \text{MPa} \]
  \[ \tau = 50 + 0.6 \cdot \sigma_n \cdot (1 - \lambda) \quad \text{for} \quad \sigma_n \cdot (1 - \lambda) > 200 \text{MPa} \]
- \( \sigma_n \) - lithostatic pressure (MPa),
- \( \lambda \) - the pore pressure ratio, (PPR - the ratio between the hydrostatic and lithostatic pressures. \( \lambda \leq 1 \). The maximum value, \( \lambda = 1 \), means no frictional heating),
- \( v \) - convergence velocity (7.8 cm/year),
- \( w \) - the thickness of a thin element layer (500 m) along the plate-interface,

where frictional heating is formulated as body-heat source.

Based on the velocity field obtained in the case of temperature dependent viscosity, a dynamic model for the blob tracers is applied. A blob moves under the action of drag, mass, and buoyancy forces in the mantle wedge stationary velocity field generated in the previous model. The description of the modeling approach is given in (Manea et al., 2004-B). The trajectories of positively buoyant blobs (\( \Delta \rho = 200 \text{ kg/m}^3 \)) with diameters between 0.4 and 10 km are calculated for different values of wrapping viscosity, \( \eta_w \) (\( 10^{14} - 2.10^{17} \text{ Pa s} \)). Very low wrapping viscosity around the blobs might be a consequence of viscous heating between the surrounding mantle and the blob (Gerya and Yuen, 2003). The total rise time that the blobs require to reach the base of the continental crust is also estimated.

During migration through the mantle wedge the blobs are heated first because of the inverted thermal gradient, and then cooled (normal thermal gradient) before approaching the base of the continental lithosphere. A blob is
assumed to be heated/cooled by conduction only. The following conduction equation is used to model blob thermal history:

\[ C_p \cdot \frac{\partial T}{\partial t} + \nabla \cdot (-k \cdot \nabla T) = 0 \]  

where:

- \( C_p \) - thermal capacity 3.3 (MJ/m\(^3\)°K),
- \( T \) - temperature (°C),
- \( t \) - time (Myr),
- \( k \) - thermal conductivity 3.1 (W/m °K),

The equation is solved numerically inside the blob, with boundary conditions of mantle wedge temperature taken from the thermal modeling. We used 3,000 triangle elements to solve equation (2) inside the spherical blobs. The trajectory of the blobs and the time steps are taken from the dynamic model for the blob tracers described above.

Finally, a tomographic image is obtained using the thermal models and temperature dependence of seismic-wave velocities from (Karato, 1993). The seismic-velocity perturbations are calculated relative to the PREM model of Dziewonski and Anderson (1981). The following equation (Karato, 1993) is used:

\[ \frac{\partial \ln V}{\partial T} = \frac{\partial \ln V_0}{\partial T} - \left( \frac{Q_p^{-1}}{\pi} \cdot \frac{H}{R \cdot T^2} \right) \]  

where:

- \( V \) - velocity (km/s),
- \( V_0 \) - the reference velocity (Dziewonski and Anderson, 1981),
- \( T \) - temperature (°C),
\[
\frac{\partial \ln V_0}{\partial T} = -\alpha \cdot \frac{\partial \ln V_0}{\partial \ln \rho}
\] (3.1)

with:

\( \alpha \) pressure dependence of thermal expansion (Stacey, 1977),
\( \frac{\partial \ln V_0}{\partial \ln \rho} \) from (Chopelas, 1992).

\( Q_{p^{-1}} \) - seismic anelasticity for compressional waves in peridotite (spinel lhezoite) determined by Sato et al., 1988, 1989 as follows:

\[
Q_p = Q_{pm} \cdot e^{[g' \left( \frac{T_m}{T}-a \right)]}
\] (3.2)

where:

\( Q_{pm} \) - is the \( Q_p \) at solidus temperature and it has the following form:

\[
Q_{pm} = Q_1 + \frac{P}{P_0}.
\] (3.3)

with:

\( Q_1 = 3.5 \)
\( P_0 = 73 \) MPa
\( P \) - lithostatic pressure (MPa),
\( g' = 6.75, a = 1.00 \) for \( T_m/T < 1 \),
\( g' = 8.47, a = 1.00 \) for \( 1 < T_m/T < 1.08 \).
g' = 13.3, a = 1.03 for $T_{m}/T > 1.08$,

$T_{m}$ - dry peridotite solidus (Wyllie, 1979),

$H$ - activation enthalpy for olivine: 500 kJ/mol (Karato and Spetzler, 1990),

$R$ - universal gas constant (8.31451 J/mol.K),

$T$ - temperature (°C).
MODELING RESULTS

Thermal models

The thermal models that correspond to isoviscosity and temperature-dependent viscosity are presented in Fig. 4 and Fig. 5. The isoviscous mantle wedge model predicts temperatures of \( \sim 950 \, ^\circ\text{C} \) in the asthenosphere, beneath the volcanic chain indicating that, at least for dry olivine, melting should not occur. The geotherm of the oceanic plate surface does not intersect the solidus (Fig. 6-A) neither for basalt nor for wet sediments (Fig. 6-B) indicating that the oceanic plate and the subducted sediments do not undergo melting. The mantle-wedge velocity field has a rather small back flow of \( \sim 3 \, \text{cm/yr} \) which is responsible for the low temperature in the wedge (Fig. 7-A). A small amount of frictional heating (a pore pressure ratio \( \lambda = 0.96 \) or an effective friction coefficient \( \mu = 0.034 \) and an average shear stress along the thrust fault \( \tau \sim 14 \, \text{MPa} \)) is added at the contact between the subducted slab and the overriding plate. Although the surface of the slab beneath the mantle wedge is heated by more than 150\(^\circ\text{C} \), this is not sufficient to melt the slab surface (Fig. 6-A). It is evident that this simple model with the isoviscous mantle wedge cannot create any source of the volcanic material.

Applying temperature-dependent viscosity in the mantle wedge produces an important increase of the temperature beneath the volcanic front. The maximum temperature then rises to more than 1,300\(^\circ\text{C} \) (Fig. 5). The mantle wedge viscosity, \( \eta_0 \), at a potential temperature, \( T_0 \) (1,450\(^\circ\text{C} \)), might be between \( 10^{17} \, \text{Pa s} \) and \( 10^{20} \, \text{Pa s} \). A benchmark for the numerical scheme applied in this study (Manea et al., 2004-B) shows a very small variation in temperature in the wedge of \( \Delta T < 15\, ^\circ\text{C} \) for \( \Delta \eta_0 = 1,000 \, \text{Pa s} \) (from \( 10^{17} \, \text{Pa s} \) up to \( 10^{20} \, \text{Pa s} \)). Therefore, a unique value for the mantle wedge viscosity at a potential temperature \( T_0 \) of \( \eta_0 = 10^{20} \, \text{Pa s} \) is used, since its effect is negligible on the overall wedge thermal structure.
The velocity field (Fig. 7-B) has back-flow velocities of ~ 7.5 cm/yr, which are responsible for the higher temperature in the tip of the wedge. Without frictional heating, the geotherm of the subducting plate surface does not intersect the solidus neither for basalt nor for sediments (Fig. 6-A) therefore melting of the subducted sediments and basaltic oceanic crust does not occur for this model. The frictional heating fraction corresponding to $\lambda = 0.96$ might cause fluid-saturated melting of subducted sediments at a relatively shallow depth of ~ 35 km (Fig. 6-B). The same amount of frictional heating ($\lambda = 0.96$) is not sufficient to melt the basaltic subducted crust. For greater amounts of frictional heating (i.e. $\lambda = 0.95$) melting of basalt oceanic crust might be possible, but this is inconsistent with the lack adakites volcanism in southern Kamchatka.

**Metamorphic sequences and dehydration within the descending oceanic crust**

The estimated variation of wt% H$_2$O content with depth along the subducting plate for both models, with isoviscosity and temperature-dependent viscosity, with and without frictional heating, are presented in Fig. 8 and Fig. 9. The isoviscous thermal model without frictional heating (Fig. 8-A) shows a fairly simple metamorphic structure: from Lawsonite - Blueschist facies, the oceanic crust enters at a depth of ~ 25 km into the stability field of Jaedite – Lawsonite - Blueschist; from ~ 55 km depth to ~ 85 km depth, the metamorphic facies is represented by Lawsonite - Amphibole - Eclogite. With a small amount of Zoisite - Eclogite from ~ 85 to ~ 100 km depth, the oceanic crust loses completely its hydrous phase entering finally into the Diamond - Eclogite stability field. Intensive dehydration occurs during these phase changes, more than 5 wt% H$_2$O being released into the overlying mantle up to a depth of ~ 100 km (see Fig. 8-A – inset). The same isoviscous thermal model, but with frictional heating included, predicts a more complicated metamorphic structure (Fig. 8-B). Up to the contact between the Moho and the subducted slab, the oceanic crust is represented by Lawsonite - Blueschist
(up to ~ 25 km depth), Epidote - Blueschist (25 - 28 km depth) and Epidote - Amphibolite (28 - 35 km depth) facies. Deeper, the metamorphosed structure is characterized by Eclogite - Amphibole (35 - 40 km depth), Zoisite - Amphibole - Eclogite (40 - 80 km depth), Zoisite - Eclogite (80 - 100 km depth) and from ~ 100 km by Diamond - Eclogite facies. Again, rigorous dehydration occur all the way through these phase changes, more than 5 wt% H\textsubscript{2}O being released into the overlying mantle up to a depth of ~ 100 km (see Fig. 8-B – inset).

The thermal model with temperature-dependent viscosity and without frictional heating reveals a similar metamorphic pattern as the isoviscous model (without frictional heating): Lawsonite - Blueschist, Jaedite – Lawsonite - Blueschist, Jadeite - Epidote - Blueschist, Amphibole - Eclogite, Zoisite - Eclogite and Diamond - Eclogite (see Fig. 9-A – inset). Strong dehydration (more than 5% wt H\textsubscript{2}O released) of wedge peridotite up to ~ 100 km depth is suggested by this model. The last thermal model proposed in this study, with temperature-dependence of viscosity and frictional heating ($\lambda = 0.96$ or $\mu = 0.034$) exposes a slightly different and more complicated metamorphic arrangement along the oceanic crust: Lawsonite - Blueschist, Greenschist, Epidote - Amphibolite, Eclogite - Amphibole, Garnet - Amphibolite, Zoisite - Amphibole - Eclogite, Zoisite - Eclogite, Coesite - Eclogite and Diamond - Eclogite facies (see Fig. 9-B - inset). Hydration of mantle wedge peridotite is likely to occur up to ~ 90 km depth and melting of fluid-saturated oceanic sediments at shallower depth of ~ 40 km is suggested by this last thermal model (Fig. 5-B and Fig. 6-B).

Vertical thermal profiles (A-A') beneath the volcanic front (190 km from the trench) through the mantle wedge (Fig. 4 and Fig. 5) illustrate that for the models with isoviscosity, mantle peridotite melting is not possible, while for both strong temperature-dependent viscosity models melting of wet peridotite is likely to occur (Fig. 6-B).
Diapiric ascent of melted material constrained by geochemical tracers

The thermal model with temperature dependent viscosity and frictional heating confirms the possibility of melting of fluid-saturated sediments and hydrated mantle wedge peridotite (Fig. 6-B and Fig. 6-A). A strong dehydration flux from the slab surface might lower the solidus of mantle wedge allowing the occurrence of partial melt of hydrated peridotite in the vicinity of the slab surface (< 10 km from the slab surface - see Fig. 9). Using the wet peridotite solidus from Mysen & Bottcher (1975) (i.e. 800°C at ~ 80 km) the wet peridotite solidus is even closer to the slab surface (~ 5 km).

Hydrated peridotite with a density lower than the asthenospheric density can develop buoyant plumes (e.g., Gerya and Yuen, 2003). Gerya and Yuen (2003) and Manea et al. (2004-B) suggested that these blobs might be lubricated by a very low wrapping viscosity due to viscous heating against the surrounding mantle wedge. The source of the wrap is melted material coming from the subducting slab, including melted subducted sediments. Wrapping viscosities down to $10^{14}$ Pa s control diapiric ascent in the wedge. The surface geotherm of the subducted slab intersects the dehydration melting solidus at ~ 90 km (Fig. 6-A).

The blobs are not necessarily melts; they might be compositionally instabilities at the slab-wedge interface (Gerya and Yuen, 2003). Actually the blobs are not restricted to be melted, they need only positive buoyancy in order to detach from the slab surface. The origin of this positive buoyancy (low density) might be compositional and/or thermal. Consequently we selected an initial point to calculate blob trajectories at a depth of ~ 110 km on the slab surface, just below the EVF.

The blob trajectories are shown in Fig. 10 for different wrapping viscosities ($10^{14}$ - $>10^{17}$ Pa s) and blob diameters (0.4 - 10.0 km). Very low viscosity is essential for the blob to rise to the base of continental crust. Extremely low viscosity ($10^{14}$ - $10^{17}$ Pa s) is also essential to explain the high pressure and ultra-high pressure metamorphic rocks exhumation from great depths (Burov et al., 2000).
For blobs of 4 km diameter (Fig. 10-A) and wrapping viscosity $\eta > 3 \cdot 10^{16}$ Pa s, the drag force is predominant and the blob cannot rise. For lowered viscosity the drag force is less significant and at the depth of $\sim 175$ km the blob intercepts the mantle wedge back flow, which returns it toward the tip of the wedge. Finally the blob rises and impinges on the continental crust after $\sim 8$ Myr. For lower viscosity ($< 10^{16}$ Pa s) blobs rise faster (Fig. 11) and impact at approximately the same point below the volcanic chain.

The larger the blob size the less time is required to reach the continental crust (Fig. 11). The buoyancy force of large blobs becomes more dominant than the drag force and this yields a substantially upright trajectory (Fig. 10-B). At a wrapping viscosity of $10^{15}$ Pa s blobs with a diameter of less than $\sim 0.8$ never rise up to the continental crust. Blobs smaller than $\sim 1$ km reach the crust at some distance (up to 10 km), depending on the blob size. For a viscosity of $> 10^{17}$ Pa s, blobs with diameters $\geq 8$ km are able to escape from the downward flow in the mantle and accumulate at the base of the continental crust.

An important constraint on the magma transport time through mantle wedge comes from U-series isotope disequilibria, $^{10}$Be and fluid soluble trace elements as Th, Sr and Pb isotope studies.

U-series isotope disequilibria come from the mobility of U in aqueous fluids under oxidizing conditions. On the other hand Th in not mobile, therefore the timescale of such U-Th disequilibria might be used to infer the fluid-transfer and melt generation in the mantle wedge before surface eruptions.

Turner and Hawksworth (1997) proposed a very rapid ascent of magma ($\sim 1000$ years) from the place where might be formed (near the slab surface) toward the earth surface through a channel network. Regelous et al. (1997) infer the magma transport rate from Th, Sr and Pb isotope data. They analyzed data from Tonga-Kermadec arc lavas, the same as Turner and Hawksworth (1997). The final conclusion of Regelous et al. (1997) is that magmas are erupted at the surface $< 350$ kyr after the melts are generated in the mantle wedge. These results are quite different, clearly showing that this issue with U-Th isotope disequilibria and Th, Sr
and Pb isotope data used to infer the magma transport rates across the mantle wedge is an ongoing debate.

The cosmic ray produces $^{10}$Be, which is subducted together with the ocean sediments. Studies of $^{10}$Be show the Be can be transported from the trench through the mantle wedge and finally erupted in surface lavas over a period of ~ 7 Myr (Brown et al. (1982); Morris et al. (1990)). With a subduction velocity rate of 7.8 cm/yr it took ~ 1 Myr for the sediments to arrive at ~ 100 km depth beneath the EVF. Then the remaining ~ 6 Myr represents the residence time for Be in the mantle wedge.

The U series and $^{10}$Be studies show contradictory estimates of the magma transport times through the mantle wedge, suggesting variable transport rates for different elements and/or real differences in transport times probably due to a variable melted volume generated in the wedge.

Whether we consider in our models the results of Regelous et al. (1997), (350 kyr for magma transport), then from Fig. 11 - inset can be seen that a magma transport mechanism as buoyant blobs still represents a reliable mechanism. The blobs with the size of less than 10 km can reach the continental crust in less than 350 kyr for wrapping viscosities less than ~ $10^{16}$ Pa s.

Assuming a residence time of magma of ~ 6 Myr (from $^{10}$Be studies), then buoyant blobs with diameters of 0.4-10 km are allowed to travel through mantle wedge toward the Moho for a very wide range of wrapping viscosities of $10^{14}$ - > $10^{17}$ Pa s.

**Thermal history of the blobs**

Heat transfer accompanies the journey of cold buoyant blobs through the mantle wedge. A temperature-dependent viscosity thermal model was used for a 10-km diameter buoyant blob. The thermal history of this blob, which reaches the Moho in 2.6 Myr, is presented for nine time periods in Fig. 12-A relatively cold blob ($\sim 800^\circ$C) initiates its voyage through the inverted thermal gradient. Its top is
heated up to ~950°C by the surrounding mantle (Fig. 12-A). After ~0.33 Myr, the blob has moved down few km from its initial position, being dragged by the vigorous mantle wedge flow. The top hot region grows in size and temperature (~1,170°C) while the cold core shrinks and becomes warmer (~810°C) (Fig. 12-B). The downward trajectory of the blob continues, and after ~0.66 Myr inner temperatures between 1,000°C and 1,290°C are expected (Fig. 12-C). From this point the blob starts to rise, being close from the maximum wedge temperature after ~1 Myr, when the temperatures inside the blob are from 1,180°C up to 1,330°C. Beyond this point the blob rises in a normal thermal gradient and after ~2.6 Myr reaches the base of the continental crust with a >950°C hot core while the surrounding mantle has ~800°C (Fig. 12-A-i). Assuming that the main composition of the blob is peridotite and using the wet peridotite solidus from Mysen and Boettcher (1975) (e.q. 850°C at 35 km depth) the main blob’s volume is not solidified when touches the Moho due to hot core temperature above 950°C (Fig. 12 and Fig. 13).

Neither viscous heating around the diapir nor thermal convection inside are incorporated in this model. For a blob with \( d = 10 \) km diameter, with a minimum blob viscosity \( \mu = 10^{17} \) Pas and a maximum thermal contrast of \( \Delta T = 350°C \) (see Fig. 12-A-b), the Rayleigh number is

\[
Ra^{\text{max}} = \frac{\rho_b \cdot g \cdot \alpha \cdot \Delta T \cdot d^3}{\mu \cdot k} = 1,050
\]

(other parameters for \( Ra^{\text{max}} \) calculation are: \( \rho_b = 3,000 \) kg/m\(^3\), \( g = 10 \) m/s\(^2\), \( \alpha = 10^{-5} \) °C\(^{-1}\), \( k = 10^{-6} \) m\(^2\)/s). Although \( Ra^{\text{max}} \) exceeds \( Ra^{cr} = 660 \) necessary for convection to begin in fluid layers heated from below (or above) (Turcotte and Schubert, 2002), in the present study we do not incorporate the effect of thermal convection on the blob’s thermal structure. Every parameter in the Rayleigh number is pretty well known except blob’s viscosity, which can vary by orders of magnitude. Future investigations will focus on the effect of viscous heating and thermal convection over the thermal history of buoyant blobs.
**Velocity anomaly estimation from thermal modeling**

The tomographic image computed using the thermal model with temperature dependent viscosity is presented in *Fig. 14-A*. The high temperature in the mantle wedge beneath the volcanic chain (> 1,300°C) produces a strong negative velocity anomaly up to −7% (*Fig. 13*) (relative to PREM). The cold subducting slab produces a positive velocity anomaly up to +4%.

The tomographic image estimation from isoviscous thermal models revealed very low amplitude velocity perturbations in the mantle wedge, because of the thermal structure mainly controlled by the left boundary condition which is consistent with the PREM model.

The procedure applied to estimate the tomography anomaly from a thermal model applying *Karato (1993)* uses the dry solidus for peridotite (*Tm*). As is shown in *Fig. 6-B*, the wedge temperature is well above the wet solidus for peridotite but is below the dry solidus, therefore no partial melting effect is included in this tomography estimation. The thermal models with strong temperature dependence of viscosity show a temperature of ~ 1,300°C at ~ 90 km depth (*see Fig. 6-B*). This corresponds to ~ 90% of the dry peridotite solidus. Tomography anomalies are usually interpreted as indicating a partially molten asthenosphere, but *Sato et al. (1989)* shows that this may reflect instead a hot solid asthenosphere where the temperature approaches 90% of the dry peridotite solidus. Experimentally anelastic properties of peridotite determined by *Sato et al. (1989)* illustrate that the attenuation mechanism of peridotite might be the weakness (or “softening”) of grain boundaries at high temperature below the solidus. This might be the case in southern Kamchatka too, since the wedge temperatures come close to ~ 90% of the peridotite dry solidus.
DISCUSSION AND CONCLUSIONS

Numerical models of steady-state temperature and velocity fields in the mantle wedge of the Kamchatka subduction zone are developed using the numerical scheme of Manea et al. (2004-B). Based on this, a dynamic model of buoyant blob migration in the mantle wedge velocity field is developed. The thermal history of a 10 km blob, which moves up through the mantle wedge thermal field is predicted. Finally, the thermal structure of the mantle wedge is used to estimate the seismic P-wave velocity anomalies (referenced to PREM) associated with subduction of the Pacific plate beneath Kamchatka. The velocity anomalies estimates are compared with a seismic tomography image inferred from P-wave arrivals for the same cross-section (Gorbatov et al., 1999).

Four different models are considered, the first two with isoviscosity in the wedge (with and without frictional heating) and the second two with the temperature-dependent viscosity (with and without frictional heating). Both type of thermal models (isoviscous and with the temperature dependent viscosity), show a velocity inflow-outflow in the mantle wedge beneath Moho and the slab surface. This type of flow might induce elastic anisotropy in the wedge peridotite, which is consistent with the trench normal strike of the inferred anisotropic fast axes in southern Kamchatka (Levin et al. 2002).

The isoviscous thermal models do not predict any melting in the asthenosphere beneath the volcanic chain (Fig. 4). With a temperature > 1,300 °C (Fig. 5), the model with temperature-dependent viscosity in the mantle wedge shows a significant increase in temperature beneath the volcanic arc. Two different sources of melt are possible: sediments and wedge peridotite beneath the volcanic front). A small amount of frictional heating (a pore pressure ratio $\lambda = 0.96$ or an effective friction coefficient $\mu = 0.034$ and an average shear stress along the thrust fault $\tau \sim 14$ MPa) is added at the contact between the subducted slab and the overriding plate (Fig. 6-A). For larger amounts of frictional heating (i.e. $\lambda = 0.95$)
Partial melting of peridotite in subduction zones is initiated by an influx of fluids derived from the metamorphosed slab and sediments (Tatsumi, 1986; Davies and Stevenson, 1992). Major dehydration of the basaltic oceanic crust (> 5 wt% H$_2$O release) occurs just below the volcanic chain up to a depth of ~100 km (Fig. 8 and Fig. 9). Despite the variation in parameters and model input, dehydration of the top of the slab is complete in all models at basically the same depth. This depth estimation is rather robust and does not depend on model parameters. The H$_2$O contents in the phase diagrams for mafics and harzburgite (Hacker et al., 2003) represent the maxima. Whether such H$_2$O contents are reached depends on pre-subduction alteration and fluid flow in the subducted slab. Experimental studies of fluid-saturated peridotite show a solidus as low as ~800 °C at pressures between 2-3 GPa (Mysen and Boettcher, 1975). As a result, just above the subducting slab (~5 km) a layer of melted peridotite might exist, as is the case for the temperature-dependent viscosity thermal models in the present study. The existence of such a melted layer has been suggested by Okada (1979). He deduced a low-velocity layer in the vicinity of the mantle wedge-slab interface from the efficient conversion of ScS and ScSp phases. Gerya and Yuen (2003) demonstrate that Rayleigh-Taylor instabilities can develop and rise up from the top of cold subducting slabs. They also suggest that plumes detached from the slab might be lubricated by partially melted, low-viscosity material from the subducted crust and hydrated mantle.

Modeling of blob motion in the mantle wedge viscous flow field induced by the subducting slab, shows that this simple approach may shed light on the origin of the volcanism beneath south Kamchatka. Two parameters control the trajectories of blobs rising from the slab: the diameter of the blob and the wrapping viscosity. Very low values of wrapping viscosity (10$^{14}$ - 10$^{17}$ Pa s) are necessary to counter out the drag and buoyancy forces, which is critical for the blob to rise.
Blob rise time decreases nonlinearly as its diameter increases and the wrapping viscosity diminishes (Fig. 10). The time required for a blob of 10 km diameter to rise from the slab surface up to the continental crust varies from < 2,000 yr up to 10 Myr for viscosities between \(10^{14}\) Pa s and \(> 10^{17}\) Pa s respectively. For a diameter of \(\geq 8\) km and a viscosity of \(\geq 10^{17}\) Pa s blobs rise upward until they reach the base of the continental lithosphere. This low value of viscosity is in the lowermost range experimentally determined for olivine at the upper mantle pressures and temperatures (Hirth and Kohlstedt, 2003).

Dynamic models of blob trajectories in the mantle wedge velocity field shows that “fast” trajectories terminate at the same location on the base of the continental lithosphere (Fig. 10), while the final points of “slow” trajectories, which are more common for the blobs of smaller size (< 1 km), are dispersed.

An important constraint on the magma transport time through mantle wedge comes from U-series isotope disequilibria, \(^{10}\text{Be}\) and fluid soluble trace elements as Th, Sr and Pb isotope studies. Turner and Hawksworth (1997) proposed a very rapid ascent of magma (~ 1,000 years or ~ 60 m/yr) from the slab surface toward the earth surface through a channel network. On the other hand, Regelous et al. (1997) infer a magma transport rate of < 350 kyr (or ~ 17 cm/yr) from Th, Sr and Pb isotope data. These results are quite different but in a recent study for U-Th-Pa-Ra disequilibria for Kamchatka of Dosseto et al. (2003), is discussed the existence of a dynamic melting model which does not require a high upwelling velocity (i.e. 1 m/yr) within the mantle wedge. A more contradictory conclusion comes from \(^{10}\text{Be}\) studies, which show that residence time for Be in the mantle wedge is ~ 6 Myr before the surface eruption.

The U-Th-Pa-Ra, Th, Sr and Pb isotope disequilibria and \(^{10}\text{Be}\) studies show very contradictory estimates of the magma transport times through the mantle wedge from < 350 kyr up to 6 Myr. Such high variability in magma transport suggests a very variable transport rates for different elements and/or real differences in transport times probably due to a variable melted volume generated in the wedge. Apart of its inherent simplicity, the advantage of a magma transport
model using buoyant blobs proposed in this chapter is its ability to cover the wide range of residence magma time inside the mantle wedge (Fig. 11).

The proposed positively buoyant blobs might have a complex composition of melted saturated peridotite and melted sediments. An H₂O-rich component (likely with sediments) resulting from dehydration penetrates overlying peridotites. Ascending into the hotter mantle, this material passes the wet solidus of peridotite (see Fig. 14). Partial melting starts and the buoyant blobs begin to move through the strong mantle wedge flow. Finally the blobs reach the base of the lithosphere and form a magmatic chamber beneath the volcanic chain.

Levin et al. (2002) proposes a scenario where the anisotropy in the top layer (~ 15 km beneath Moho) of the mantle wedge beneath EVF in southern Kamchatka is generated by melt lenses and/or sheeted diapirs. In this model, the anisotropy might be generated without the presence of a mantle wedge inflow beneath Moho. This model is in good agreement with the flow model presented in Fig. 7-B, where very small inflow velocities are obtained from the thermal model with strong temperature dependence of olivine. Moreover, the magma propagation model with buoyant blobs shows an accumulation area of melted blobs just beneath the volcanic arc at the base of Moho (Fig. 10).

One important control regarding the reality of buoyant blobs comes from a study of deformed peridotite xenolites from Avachinscky volcano (Fig. 1) (Graybill et al., 1999). In this paper the xenolites strains do not indicate a shear induced by a corner flow, rather they seems to belong from individual diapirs traveling through the mantle wedge.

The thermal evolution of the blobs was investigated by applying the heat conduction equation. Thermal convection inside a blob of 10 km diameter is not likely to occur because the Rayleigh number is very small (~ 2.7 for a viscosity of 10^{17} Pa s and a maximum thermal contrast of 300ºC). Therefore, in the present study, the blob is heated/cooled only by conduction. After about 1 Myr, the cold blob (~ 800ºC) moves toward the hotter region of the wedge where a temperature of more than 1,300ºC is estimated (Fig. 12 and Fig. 14). After being heated the
blob moves upwards toward the base of the continental lithosphere where it arrives with a hot core (> 900°C). According to the fluid-saturated peridotite solidus of Mysen and Boettcher (1975), most of the 10 km blob (T > 900°C at 1 GPa) is melted when arrives at the Moho base (Fig. 12-A-i and Fig. 13). The blobs may carry fluid-saturated melts (i.e. melted peridotite) and sediments from the subducted slab to the base of the continental lithosphere and therefore trace elements and the chemical signature finally might reach the earth’s surface through volcanic eruptions. Although Fedotov and Masurenkov (1991) classified the active volcanoes in Kamchatka as calc-alkaline, Tatsumi et al. (1994) interpreted as tholeiitic the southern Kamchatka volcanoes. However, Kelemen (1990) shows that “slow” ascent of a tholeiitic melt through a hot mantle wedge produces calc-alkaline magmatism at the surface.

The thermal model with temperature-dependent viscosity and frictional heating is used to estimate a seismic tomography image below southern Kamchatka. The high temperatures in the mantle wedge beneath the volcanic chain (> 1,300°C) produce a strong negative velocity anomaly of up to −7% (Fig. 14-A) (relative to PREM). On the other hand, the cold subducting slab produces a positive velocity anomaly up to +4%. Since the seismic tomography of Gorbatov et al. (1999) has a small uncertainty of ~10%, our estimation is in good agreement with the velocity anomalies obtained by Gorbatov et al. (1999) for a 2D profile identically located with our 2D cross sections. The shape of our tomographic image inferred from thermal modeling differs from the P-wave seismic tomography image of Gorbatov et al. (1999), especially for the continental lithosphere and the uppermost mantle beneath the volcanic chain (Fig. 14-B). This is likely due to the fact that the thermal models in this study do not consider the magma transport effect toward the surface. However, since the resolution of the seismic tomography is very low (50 km) the position of the low velocity zone beneath the EVF has very large uncertainties. Nevertheless, good agreement of the velocity perturbation beneath the volcanic arc (at least in magnitude) between the tomography image
from P-wave arrivals and our estimation from thermal modeling suggests satisfactory modeling the mantle wedge beneath southern Kamchatka.
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Table 1. Summary of thermal parameters used in the models. (Compilation from Peacock and Wang, 1999; Smith et al., 1979)

<table>
<thead>
<tr>
<th>Geological Unit</th>
<th>Density (g/cm³)</th>
<th>Thermal Conductivity (W/m °K)</th>
<th>Radiogenic heat production (µW/m³)</th>
<th>Thermal Capacity (MJ/m³ °K)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Oceanic sediments</td>
<td>2.20</td>
<td>1.00 – 2.00*</td>
<td>1.00</td>
<td>2.50</td>
</tr>
<tr>
<td>Upper continental crust (0-15 km)</td>
<td>2.70</td>
<td>2.00</td>
<td>1.3</td>
<td>2.50</td>
</tr>
<tr>
<td>Lower continental crust (15-35 km)</td>
<td>2.70</td>
<td>2.00</td>
<td>0.2</td>
<td>2.50</td>
</tr>
<tr>
<td>Mantle wedge and Blob</td>
<td>3.10</td>
<td>3.10</td>
<td>0.01</td>
<td>3.30</td>
</tr>
<tr>
<td>Oceanic lithosphere</td>
<td>3.00</td>
<td>2.90</td>
<td>0.02</td>
<td>3.30</td>
</tr>
</tbody>
</table>

* Increases linearly with distance from the deformation front up to a depth of 10 km.
**FIGURE CAPTIONS**

**Figure 1.** Tectonic settings and position of the modeled cross-section (purple thick line) in Kamchatka. Orange triangles show the location of trench-side and back-arc volcanoes. Open, semi-transparent arrows show convergence velocities between the Pacific and North American plates.

**Figure 2.** A grid with 12,000 triangles with ~2 km mesh resolution was used to solve the numerical thermal models. The green spherical mesh was used to solve the heat transfer equation (2) (see text) inside the blob and contains 3,000 triangles.

**Figure 3.** Boundary condition and parameters used in the modeling. The upper and lower boundaries have constant temperatures of 0°C and 1,450°C, accordingly. The continental plate is fixed. The left (landward) vertical boundary: 18.5°C/km thermal gradient in the continental crust (down to 35 km); between 35 km and 180 km depth the thermal gradient is of 5.5°C/km; underneath 180 km no horizontal conductive heat flow is specified. Zero tractions are considered beneath Moho (35 Km), at the boundary, which belongs to the mantle wedge. The right (seaward) boundary condition is a one-dimensional geotherm for the 70 Myr old oceanic plate. The PAC plate referred to the North American plate has the convergence velocity of 7.8 cm/yr. Volumetric shear heating is imposed along the plate interface up to a maximum depth of 35 km (red dashed line), using the Byerlee's friction law (*Byerlee, 1978*).

**Figure 4.**

(A). Calculated steady-state thermal field for the isoviscous mantle wedge. Horizontal dashed line shows the Moho (35 km depth). Thick solid magenta line denotes the top of the subducting slab. No frictional heating along the thrust zone is included in this model. A-A' is the vertical temperature profile in Fig. 6-B.
(B). The same as A., but frictional heating ($\lambda = 0.96$ or $\mu = 0.034$) along the thrust zone is included in this model.

**Figure 5.**

(A). Steady-state thermal field for strong temperature-dependent viscosity in the mantle wedge. Horizontal dashed line shows the Moho (35 km depth). Thick solid magenta line denotes the top of the subducting slab. No frictional heating along the thrust zone is included in this model. B-B' is the vertical temperature profile in *Fig. 6-B*.

(B). The same as A., but frictional heating ($\lambda = 0.96$ or $\mu = 0.034$) along the thrust zone is included in this model.

**Figure 6.**

(A). Phase diagrams for MORB and maximum $\text{H}_2\text{O}$ contents (*Hacker et al., 2003*). Z - Zeolite (4.6 wt% $\text{H}_2\text{O}$), PP - Prehnite - Pumpellyite (4.5 wt% $\text{H}_2\text{O}$), PA - Pumpellyite - Actinolite (4.4 wt% $\text{H}_2\text{O}$), G - Greenschist (3.3 wt% $\text{H}_2\text{O}$), LB - Lawsonite - Blueschist (5.4 wt% $\text{H}_2\text{O}$), EpB - Epidote - Blueschist (3.1 wt% $\text{H}_2\text{O}$), EpA - Epidote - Amphibolite (2.1 wt% $\text{H}_2\text{O}$), JEpB - Jadeite - Epidote - Blueschist (3.1 wt% $\text{H}_2\text{O}$), EcA - Eclogite - Amphibole (2.4 wt% $\text{H}_2\text{O}$), A - Amphibolite (1.3 wt% $\text{H}_2\text{O}$), GA - Garnet - Amphibolite (1.2 wt% $\text{H}_2\text{O}$), Gr - Granulite (0.5 wt% $\text{H}_2\text{O}$), GGr - Garnet - Granulite (0.0 wt% $\text{H}_2\text{O}$), JLB - Jaedite - Lawsonite - Blueschist (5.4 wt% $\text{H}_2\text{O}$), LAEc - Lawsonite - Amphibole - Eclogite (3.0 wt% $\text{H}_2\text{O}$), JLTS - Jaedite - Lawsonite - Talc - Schist, ZAEc - Zoisite - Amphibole - Eclogite (0.7 wt% $\text{H}_2\text{O}$), AEc - Amphibole - Eclogite (0.6 wt% $\text{H}_2\text{O}$), ZEc - Zoisite - Eclogite (0.3 wt% $\text{H}_2\text{O}$), Ec - Eclogite (0.1 wt% $\text{H}_2\text{O}$), CEc - Coesite - Eclogite (0.1 wt% $\text{H}_2\text{O}$), DEc - Diamond - Eclogite (0.1 wt% $\text{H}_2\text{O}$). Calculated geotherms: dashed blue line - top of subducting oceanic crust for isoviscous mantle wedge and no frictional heating; solid blue line – top of subducting oceanic crust for strong temperature-dependent viscosity and no frictional heating; dashed pink/red line - top of subducting oceanic crust isoviscosity in the mantle wedge and frictional heating ($\lambda = 0.96$ or $\mu = 0.034$);
solid pink/red line - top of subducting oceanic crust for strong temperature-dependent viscosity and frictional heating ($\lambda = 0.96$ or $\mu = 0.034$). The maximum depth of the stable hydrous phases in the oceanic slab is ~ 90 km in case of variable rheology in the wedge (and with frictional heating ($\lambda = 0.96$ or $\mu = 0.034$)).

(B). Phase diagram for harzburgite, and maximum H$_2$O contents (Hacker et al., 2003). A - Serpentine - Chlorite - Brucite (14.6 wt% H$_2$O), B - Serpentine - Chlorite - Phase A (12 wt% H$_2$O), C - Serpentine - Chlorite - Dunite (6.2 wt% H$_2$O), D - Chlorite - Harzburgite (1.4 wt% H$_2$O), E - Talc - Chlorite - Dunite (1.7 wt% H$_2$O), F - Anthigorite - Chlorite - Dunite (1.7 wt% H$_2$O), G - Spinel - Harzburgite (0.0 wt% H$_2$O), H - Garnet - Harzburgite (0.0 wt% H$_2$O).

Continuous thick yellow line indicates the wet solidus for sediments from Nichols et al. (1994). Calculated geotherms are the same as in A. The temperature profiles A-A’ and B-B’ (Fig. 4 and Fig. 5) show that for an isoviscous mantle wedge thermal structure, melting of wet peridotite is not possible, while for temperature-dependent viscosity melting of wet peridotite beneath the volcanic chain is likely to occur. The wet and dry peridotite are taken from Wyllie (1979).

Figure 7.

(A). The velocity field in the isoviscous mantle wedge. The maximum inflow velocity is about 3 cm/yr. The return flow (backflow) is horizontal.

(B). The velocity field with the strong temperature-dependent mantle wedge viscosity. Note that the maximum velocity of the inflow region is about 7.5 cm/yr that is comparable with the subducting slab velocity. The velocity field presents a diagonally upward pattern. Note the very low velocity (< 1 cm/yr) just beneath the Moho.

Figure 8.

(A). Metamorphic facies along the subducting oceanic crust corresponding to the isoviscous model without frictional heating. The inset represents the variation of wt% H$_2$O along the subducting crust as function of metamorphic sequence.
More than 5 wt% H₂O may be released from hydrous phases in the subducting slab through continuous dehydration.

(B). The same as for Fig. 8-A but with frictional heating (λ = 0.96 or μ = 0.034). More than 3 wt% H₂O may be released from hydrous phases in the subducting slab through continuous dehydration.

Figure 9.

(A). The same as Fig. 8-A but for the temperature-dependent viscosity model without frictional heating. More than 5 wt% H₂O may be released from hydrous phases in the subducting slab through continuous dehydration.

(B). The same as Fig. 8-A but for the temperature-dependent viscosity model with frictional heating (λ = 0.96 or μ = 0.034). More than 3 wt% H₂O may be released from hydrous phases in the subducting slab through continuous dehydration.

Figure 10. Blob trajectories in steady mantle wedge flow (Fig. 7-B). The initial point for all trajectories is at a depth of 110 km on the surface of the subducting slab below the trench-side volcanic belt.

(A). Blob diameter is fixed at 4 km. The trajectories, corresponding to different wrapping viscosities of less than 10^{16} Pa s, have the same final point below the volcanic chain.

(B). The wrapping viscosity is fixed at 10^{15} Pa s. The trajectories correspond to different blob size (0.8 - 10.0 km). Blobs with a diameter less than ~ 0.8 never rise to the continental crust.

Figure 11. Blob rise time as a function of wrapping viscosity. The curves annotated with blob diameter show that the rise time decreases for bigger blobs and the lower viscosity. Blobs with a size of less than 10 km can reach the continental crust in less than 350 kyr for wrapping viscosities less than 10^{16} Pa s (see inset).
Figure 12.

(A). The thermal history of a 10-km blob which reaches the Moho in 2.6 Myr at nine time periods. The wrapping viscosity is $10^{17}$ Pa s. The circles represent the blob cross-section.

(B). The trajectory followed by the blob in a thermal structure for temperature-dependent viscosity (Fig. 5-B).

(C). The same as B. but zoomed.

Figure 13. P-T Trajectory of the 10-km diameter blob (see Fig. 12) through the mantle wedge. Colour disks represent the blob at nine time periods from Fig. 12. The number inside the disks shows the rising time. Dark blue dashed line represents wet peridotite solidus from Wyllie (1979). Green dashed line represents wet peridotite solidus from Mysen and Boettcher (1975) (limited to ~ 3 GPa).

Figure 14.

(A). Velocity anomaly estimation below south Kamchatka inferred from thermal models with temperature-dependent viscosity (Fig. 5-B). Red and blue colors reveal the slow and fast velocities according to the vertical scale.

(B). Seismic tomography of south Kamchatka inferred from P-wave arrivals from Gorbattov et al. (1997).
Figure 4

Isoviscous Mantle Wedge
No Frictional heating

Isoviscous Mantle Wedge
Frictional heating: \( \lambda = 0.96 \) or \( \mu = 0.034 \)
Figure 5

Reference viscosity $\eta_r = 10^9 \text{Pa s}$
Activation energy $E_{\text{activation}} = 300 \text{ KJ/mol}$
No Frictional heating

Reference viscosity $\eta_r = 10^9 \text{Pa s}$
Activation energy $E_{\text{activation}} = 300 \text{ KJ/mol}$
Frictional heating $\lambda = 0.96$ or $\mu = 0.934$
Figure 6
Figure 7
Figure 8

A

Isoviscous Mantle Wedge
No Frictional heating

Depth (km)

Distance from Trench (km)

B

Isoviscous Mantle Wedge
Frictional heating:
\[ \lambda = 0.96 \text{ or } \mu = 0.034 \]

Depth (km)

Distance from Trench (km)
Figure 9
Figure 10

A

B

Fixed diameter: 4 km

Fixed wrapping viscosity: $10^{10}$ Pa s

- $\eta > 1 \times 10^{15}$ Pa s
- $\eta = 3 \times 10^{14}$ Pa s
- $\eta = 5 \times 10^{14}$ Pa s
- $\eta = 1 \times 10^{15}$ Pa s
- $\eta = 2 \times 10^{15}$ Pa s
- $\eta = 3 \times 10^{16}$ Pa s

- $d > 4.0$ km
- $d = 2.0$ km
- $d = 1.6$ km
- $d = 1.2$ km
- $d = 1.0$ km
- $d = 0.8$ km

Distance from Trench (km)

Distance from Trench (km)
Figure 13
Figure 14