Thermo-mechanical model of the mantle wedge in Central Mexican subduction zone and a blob tracing approach for the magma transport

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Abstract

The origin of the Central Mexican Volcanic Belt (CMVB) and the influence of the subducting Cocos plate on the CMVB volcanism are still controversial. In this study, the temperature and mantle wedge flow models for the Mexican subduction zone are developed using the finite element method to investigate the thermal structure below CMVB. The numerical scheme solves a system of 2D Stokes equations and 2D steady state heat transfer equation.

Two models are considered for the mantle wedge: the first one with an isoviscous mantle wedge and the second one with strong temperature-dependent viscosity. The first model reveals a maximum temperature of ~830 °C in the mantle wedge, which is not sufficient for melting of wet peridotite. Also, the geotherm of the subducting plate upper surface does not intersect the dehydration-melting solidus for mafic minerals. The second model predicts temperatures of more than 1200 °C beneath the CMVB for a wide range of rheological parameters (reference viscosity and activation energy). Up to 0.6 wt. % H₂O can be released down to 60 km depth through metamorphic changes in the oceanic crust of the subducting slab. The melting of this oceanic crust apparently occurs in a narrow depth range of 50÷60 km and also melting of the mantle wedge hydrated peridotite is now expected to take place beneath CMVB.

Considering that the melting processes on and in the vicinity of the subducting plate surface generate the most of the volcanic material, a dynamic model for the blob tracers is developed using Stokes flow at infinite Prandtl number. The blobs of 0.2 - 10.0 km in diameter migrate along very different trajectories only at low wrapping viscosities ($\eta_w = 10^{14}÷5\cdot10^{17}$ Pa·s). The modeling results show that the “fast” trajectories terminate at the same focus location at the base of the continental crust, while the arrival points of “slow” trajectories, which are common for the blobs of smaller size (~ 0.4-0.5 km), are scattered away from the average focus location. This observation may give us a hint on a possible mechanism of strato and mono volcanoes genesis. The rise time, which the blob detached from the subducted plate, needs to reach the bottom of the continental crust, is from 0.001 up to 14 million years depending on the blob diameter and surrounding viscosity.

Keywords: Mexican subduction zone, thermal models, mantle wedge flow, blobs.
Introduction

Thermal and flow models in the mantle wedge can give us advance insights on the geodynamic processes in the forearcs as well as beneath the volcanic arcs. There are only a few thermal models of the subduction zone in Mexico (Currie et al., 2002, Manea et al. 2003), but none of them presents a reliable and detailed thermal structure beneath the volcanic arc considering the rheology of the asthenosphere. The present models are further development of the previous study of the Mexican subduction zone in Guerrero (Manea et al., 2003), which plausibly constrained the thermal structure for the forearc area only. The thermo-mechanical modeling of the mantle wedge with temperature and/or stress dependent rheology is in progress (Furukawa, 1993; Conder et al., 2002; Van Keken et al., 2002; Kelemen et al., 2003). Mantle wedge dynamics and thermal structure of shallow flat subduction zones in general have been investigated recently by van Hunen et al. (2002). In this study we explore the thermal structure of the mantle wedge in a specific subduction zone of the Central Mexico (Guerrero), which has an anomalously wide, subhorizontal plate interface and a distant volcanic arc. The models describe a stationary slab-induced convection, in the cases of the constant viscosity (isoviscous mantle) and strong temperature-dependent viscosity of the asthenosphere.

The previous thermal model for the Guerrero subduction zone with a shallow plate interface (Currie et al., 2002) with predefined analytical expression for the mantle corner flow, predicts the temperature of ~900 °C in the asthenosphere beneath the volcanic front (Popocatepetl volcano). In this model the basaltic component of the subducting plate crust does not reach the melting temperature.

Recent geological studies (Luhr, 1997; Marquez et al., 1999a) suggest the existence of a complex mantle beneath the CMVB. However the origin of the OIB-like basalts is still unclear there. Advection of the asthenospheric mantle caused by sinking of the Cocos plate was proposed as a possible source of the OIB-like magmas (Luhr, 1997, Wallace and Carmichael, 1999). There are several mineralogical and geochemical studies of the CMVB (e.g. Marquez and De Ignacio, 2002) suggesting the existence of two different primitive mafic magmas, one with the “asthenospheric” OIB-like component and another with the “lithospheric” component. Although the relationship between the volcanism and the subduction of the Cocos plate is commonly recognized (e.g. Pardo and Suarez, 1995, Wallace and Carmichael, 1999), there are different hypothesis regarding the source of the volcanism in this area: the extension of the Gulf of California fault system (Ferrari et al., 1994), an old wicked cortical zone (Cebull and Schubert, 1987), cortical transpression mechanisms (Shubert and Cebull, 1984; Ferrari et al., 1990) and rifting (Luhr, 1997; Marquez et al., 1999a,b).

The source of fluids that metasomatize the mantle is also unclear. There are three possibilities proposed to explain the metasomatized mantle (Marquez et al., 1999c; Wallace and Carmichael, 1999; Verma, 1999; 2000): a) the mantle wedge affected by the fluids originated from dehydration of the
subducted Cocos plate; b) old enriched lithospheric mantle; c) mantle metasomatized by volatiles from a mantle plume. Direct evidences that the mantle wedge metasomatized by fluids released from the down going oceanic slab is essentially difficult, because the mantle xenoliths are rarely discovered in arc lavas. In the CMVB, Quaternary hornblende andesites erupted near the El Peñon area (see Fig.1) contain xenoliths of 1÷2 cm. These xenoliths are rich in phenocrysts of hornblende and the host andesite is depleted in plagioclase phenocrysts (Blatter and Carmichael, 1998). All these observations suggest an influx of volatiles from the subducting slab into the mantle wedge.

Underneath the CMVB, the magma ascents toward the earth surface producing large stratovolcanic structures and smaller sized monogenetic volcanoes scattered in large areas. The CMVB includes several stratovolcanoes like Popocatepetl, Iztaccihuatl and Nevado de Toluca (Fig.1). The monogenetic volcanoes are basically cinder cones, lava cones, domes and lava flows. While the stratovolcanoes are characterized by a periodicity of magma eruptions, the monogenetic volcanoes present a single one eruptive event, and as a result have a smaller size compared with stratovolcanoes. Fedotov (1981) suggested that the occurrence of one or another type of volcanoes might be related with the magma supply. Other scientists (e.g. Takada, 1989, 1994) propose a dual mechanism related with both the magma supply and regional stress. While the alignment of the monogenetic volcanoes appears to be parallel to a main normal fault system, the arrangement of the large stratovolcanoes seems to be rather orthogonal to the volcanic belt (Alainz-Alvarez et al., 1998).

Using the phase diagrams for mafics (e.g. Hacker et al., 2002) it is possible to investigate whether the mantle wedge is subjected to the hydration by fluids released from the subducting slab. A very thin sedimentary fill at the Mexican trench suggests that about of 95% of these sediments (~200 m) are subducted (Manea et al., 2003). An influx of volatiles from the metamorphosed oceanic crust and sediments might trigger partial melting of the peridotite just above the subducted slab (Tatsumi, 1986; Davies and Stevenson, 1992). Gerya and Yuen (2003) showed that a Rayleigh-Taylor instability developed above the subducting slab generates positively buoyant plumes up to 10 km in diameter that can penetrate the overlying mantle wedge.

Albeit various mechanisms of magma generation have been proposed (e.g. anhydrous decompression melting of peridotite (Klein and Langmuir, 1987; Langmuir et al., 1992); porous flow of hydrated partial melt (Davies and Stevenson, 1992)), in the present study, the magma generation and migration is assumed in a form of partially melted positively buoyant blobs. Regardless of the oversimplification of the blob properties, this model can help to understand the existence of different sources of the volcanism in the area. The buoyant blobs of different size and composition may be generated by melting of the Cocos plate and overlying mantle peridotite, when the pressure and temperature reach the solidus conditions (e.g., Gerya and Yuen, 2003).

It is important to estimate the viscosity range for the reasonable trajectories and average rise times for the blobs reaching the bottom of the continental crust. A recent study by Gerya and Yuen
(2003) revealed that these plume-like blobs may be lubricated by the partially melted material of the subducted crust and hydrated mantle, thus producing a very low viscosity wraps around the blob structures. Burov et al. (2000), apply a similar extremely low viscosity, in order to model the exhumation in the continental lithosphere.

**Modeling procedure**

A system of 2D Stokes equations and 2D steady state heat transfer equation is solved for the Guerrero cross section (Fig. 1) using the finite element solver PDE2D (http://pde2d.com/). The system of equations in an explicit form is:

\[ \begin{align*}
\frac{\partial}{\partial x} \left( -P + 2\eta \frac{\partial u}{\partial x} \right) + \frac{\partial}{\partial y} \left( \eta \left( \frac{\partial u}{\partial y} + \frac{\partial v}{\partial x} \right) \right) &= 0 \\
\frac{\partial}{\partial x} \left( \eta \left( \frac{\partial u}{\partial y} + \frac{\partial v}{\partial x} \right) \right) + \frac{\partial}{\partial y} \left( -P + 2\eta \frac{\partial v}{\partial y} \right) &= -\rho \cdot g - Ra \cdot T \\
C_p \left( u \frac{\partial T}{\partial x} + v \frac{\partial T}{\partial y} \right) &= \frac{\partial}{\partial x} \left( k \frac{\partial T}{\partial x} \right) + \frac{\partial}{\partial y} \left( k \frac{\partial T}{\partial y} \right) + Q + Q_{sh}
\end{align*} \]  

(1)

where \( P \) is pressure (Pa), and

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

is the mantle wedge viscosity (Pa-s).

Other parameters are:

- \( \eta_0 \) – mantle wedge viscosity at the potential temperature \( T_0 \) (reference viscosity) \((10^{17} \div 10^{21}) \text{ Pa-s})\),
- \( T_0 \) - mantle wedge potential temperature (1450ºC),
- \( E_a \) – activation energy for olivine (kJ/mol),
- \( R \) – universal gas constant (8.31451 J/mol.K),
- \( T \) – temperature (ºC),
- \( u \) - horizontal component of the velocity (m/s),
- \( v \) - vertical component of the velocity (m/s),
- \( \rho \) - density (kg/m³),
\( C_p \) - thermal capacity (MJ/m\(^3\)K),
\( k \) - thermal conductivity (MJ/m\(^3\)K),
\( Q \) - radiogenic heating (W/m\(^3\)),
\( Q_{sh} \) – volumetric shear heating (W/m\(^3\)),
\[ Ra = \frac{\rho \cdot \alpha \cdot g \cdot \Delta T \cdot L^3}{\eta_0 \cdot k} \]
- thermal Rayleigh number,
\( \alpha \) - thermal expansion \(3.5 \cdot 10^{-5}\) (1/°C),
\( L \) – length scale (330 km),
\( \Delta T \) – temperature difference between the bottom and top model temperatures (1450 °C),
\( k \) – thermal diffusivity \(10^{-6}\) m/s\(^2\)),
\( g \) - gravitational acceleration (9.81 m/s\(^2\)).

The Stokes equations are solved only for the mantle wedge, while the heat transfer equation is solved for the entire model. The linear system solver used by the present numerical scheme is the frontal method, which represents an out-of-core version of the band solver (uses a reverse Cuthill-McKee ordering). For the model with strong temperature-dependent viscosity, the system of equations becomes strongly nonlinear, therefore Picard iterations are applied, and in order to achieve a convergent solution a cut-off viscosity of \(10^{24}\) Pa·s for the temperature less than 1100 °C is used. For such highly non-linear problems we constructed a one-parameter family of problems using a variable \((V)\), such that for \(V=1\) the problems is easy (e.g. linear) and for \(V > N\) (N is less NSTEPS=20), the problem reduces to the original highly nonlinear problem. The nonlinear terms are multiplied by \(\text{MIN}(1.0,(V-1.0)/N)\). With a wide viscosity range from \(10^{17}\) Pa·s to \(10^{24}\) Pa·s, the fully nonlinear viscosity formulation (e.g., temperature and stress dependence of the viscosity) presented significant numerical instabilities; therefore the strain-rate dependence is neglected in the present study.

In the present numerical scheme, the penalty method formulation is used, \(P\) being replaced by
\[ P = -\alpha' \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right) \],
where \(\alpha'\) is large, on the order of \(\frac{\eta}{\sqrt{\varepsilon}}\) (\(\varepsilon\) is the machine relative precision). In other words, the material is taken to be "almost" incompressible, so that a large pressure results in a small decrease in volume, and the continuity equation \(\left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right) = 0\) is almost satisfied.

The finite elements grid extends from 25 km seaward of the trench up to 600 km landward of it, and consists of 12000 triangular elements with the higher resolution in the tip of the wedge (Fig. 2). The triangles have the height to width ratio equal to 1. A benchmark with different grid resolutions is done to quantify the numerical error introduced by the present numerical scheme. The lower limit of the grid follows the shape of the subducting plate upper surface (Kostoglodov et al., 1996) at 100 km depth distance. The top of the model has a fixed temperature of 0°C. The temperature at the bottom of the model is of 1450°C which represents the mantle temperature at ~100 km depth (Fig. 3).
The continental lithosphere in Guerrero is defined mostly by the crust, which thickness is 40 km in the model. This is consistent with the values inferred from the seismic refraction surveys and gravity modeling (e.g., Valdes et al., 1986, Arzate et al., 1993). The bending geometry of the subducting slab (up to the hinge point: 270 km) is well constrained by gravity modeling (Kostoglodov et al., 1996), seismicity data and recently estimated extension of the coupled plate interface inferred from GPS measurements during the last slow slip event in Guerrero (2001-2002) (Kostoglodov et al., 2003).

A dip of the subducting plate beneath the volcanic arc is poorly constrained because of a very limited number of intraslab earthquakes. The dip of 20° and a hinge point at 270 km from the trench match better the hypocenter locations of the intraslab events (Fig. 5) and as a result, these are enveloped by the modeled “seismicity cut-off” temperature of $T \approx 800°C$ (Gorbatov and Kostoglodov, 1997). The continental crust consists of two layers: the upper crust of 15 km and the lower crust of 25 km. A summary of the thermal parameters used in the models is presented in Table 1 ( compilation from: Peacock and Wang, 1999; Smith et al., 1979; Ziagos et al., 1985; Vacquier et al., 1967; Prol-Ledesma et al., 1989).

The radiogenic heat generation in the continental crust decreases exponentially from 1.3 µW/m$^3$ in the uppermost crust down to 0.2 µW/m$^3$ in its lowermost part (Ziagos et al., 1985). The radiogenic heat production in the models is taken as 50% from the above values to fit the surface heat flow with the observed heat flow data (Fig. 6). This reduction has a minor effect on the thermal structure of the subduction interface and mantle wedge.

A long term sliding between the subducting and the continental plates along the thrust fault should produce frictional heating. We introduced in the models a small degree of volumetric frictional heating using the Byerlee’s friction law (Byerlee, 1978). Frictional heating is ceased at a maximum depth of 40 km, which corresponds to the contact between the oceanic plate and the mantle wedge (Fig.3). The 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; \[\begin{align*}
\tau &= 0.85 \cdot \sigma_n \cdot (1 - \lambda) \quad \text{for} \quad \sigma_n \cdot (1 - \lambda) \leq 200\text{MPa} \\
\tau &= 50 + 0.6 \cdot \sigma_n \cdot (1 - \lambda) \quad \text{for} \quad \sigma_n \cdot (1 - \lambda) > 200\text{MPa}
\end{align*}\]

$\sigma_n$ – lithostatic pressure (MPa),

$\lambda$ - pore pressure ratio, (the ratio between the hydrostatic and lithostatic pressures. $\lambda=0.98$ in the present study. The maximum value, $\lambda=1$, means no frictional heating),

$v$ – convergence velocity (5.5 cm/year),

$w$ – thickness of the oceanic crust involved in friction (200 m).
The right landward vertical boundary condition is defined by 20 °C/km thermal gradient in the continental crust. This value is in agreement with the back arc thermal gradient of 17.8 - 20.2 °C/km reported by Ziagos et al. (1985). Underneath Moho (40 km), the right boundary condition is represented by 10 °C/km thermal gradient down to the depth of 100 km. Below 100 km, no horizontal conductive heat flow is specified. Beneath Moho (40 km depth), for the right boundary corresponding to the mantle wedge, the boundary conditions are:

\[
\begin{align*}
\left(-P + 2 \cdot \eta \cdot \frac{\partial u}{\partial x}\right) \cdot n_x + \eta \cdot \left(\frac{\partial u}{\partial y} + \frac{\partial v}{\partial x}\right) \cdot n_y &= GB_1, \\
\eta \left(\frac{\partial u}{\partial y} + \frac{\partial v}{\partial x}\right) \cdot n_x + \left(-P + 2 \cdot \eta \cdot \frac{\partial v}{\partial y}\right) \cdot n_y &= GB_2,
\end{align*}
\]

which are obtained by balancing the internal (stress induced) forces against the external boundary forces, called tractions (\(GB_1\) and \(GB_2\)). Therefore, beneath Moho, where there is no "external" force applied, \(GB_1=GB_2=0\).

At the intersection between the subducted slab and the right boundary, the velocity of the subducting slab is used.

The left 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 200 m (Moore et al., 1982) at the trench (Fig. 3). The sedimentation history and the porosity-depth profile are used only to calculate the oceanic geotherm and are not included in the modeling procedure.

In terms of displacements, the velocity of the oceanic plate is considered with reference to the continental plate. Thus the convergence rate of 5.5 cm/year between the Cocos and North American plates is used for the Guerrero subduction zone (DeMets et al., 1994). The velocities in the subducting Cocos slab are set of 5.5 cm/yr; therefore the interface with the mantle wedge is predefined. The boundary between the mantle wedge and overlying lithosphere is considered fixed.

The Cocos plate age at the trench is 13.7 Ma according to the interpretation of Pacific-Cocos seafloor spreading magnetic anomaly lineations (Klitgord and Mammerickx, 1982, Kostoglodov and Bandy, 1995). The forearc thermal model constraints are described in detail by Manea et al., (2004). The present study is focusing mostly on the mantle wedge thermal and velocity structure.

Based on the velocity field obtained in the case of temperature dependent viscosity, a dynamic model for the blob tracers is developed using Stokes flow at infinite Prandtl number (Turcotte and Schubert, 1982). The blob moves under the action of drag, mass, and buoyancy forces in the mantle wedge stationary velocity field generated in the previous model (1). To investigate the mere dynamic
effect of the mantle wedge convection on the hypothetical blobs rising from the subducted plate up to the base of the continental lithosphere, the following main assumptions are done:

1. The motion of the asthenosphere is described by the above 2D Stokes equations (1);
2. The blobs are spherical and the drag force is assumed to be similar to that of non-deforming spheres;
3. The velocity field is steady state and the liquid acceleration is negligible because the related forces have small magnitude compared to the steady drag force;
4. The influence of the blob motion on the mantle wedge circulation is irrelevant and there is not interaction between individual blobs.

According to these assumptions the total force acting on the blob is:

\[
\vec{F} = \vec{F}_g + \vec{F}_A + \vec{F}_D,
\]

where:

\[
\vec{F} = \frac{4}{3} \pi r^3 \rho_b \frac{d v_b}{dt} - \text{resultant force},
\]

\[
\vec{F}_g = \frac{4}{3} \pi r^3 \rho_e g - \text{gravitational force},
\]

\[
\vec{F}_A = -\frac{4}{3} \pi r^3 \rho_e g - \text{buoyancy force},
\]

\[
\vec{F}_D = \frac{1}{2} \pi r^3 \rho_e C_D(Re) \cdot |v_a - v_b| \cdot \frac{v_a - v_b}{|v_a - v_b|} - \text{steady drag force},
\]

\[
C_D(Re) = \frac{24}{R_e},
\]

\[
R_e = \frac{2 \cdot r \cdot |v_a - v_b|}{\eta_u / \rho_b} - \text{Reynolds number},
\]

\[
v_b - \text{blob velocity},
\]

\[
v_a - \text{mantle wedge velocity},
\]

\[
r - \text{blob radius (0.1 – 5.0 km)},
\]

\[
\rho_b - \text{blob density (3000 kg/m}^3\text{)},
\]
\( \rho_a \) - mantle wedge density (3200 kg/m\(^3\)),

\( g \) - gravitational acceleration (9.81 m/s\(^2\)),

\( \eta_w \) - the blob’s wrapping viscosity (Pa\( \cdot \)s).

Using the expressions for forces, the equation of the blob motion has the following form:

\[
\frac{d}{dt} v_b = g - \frac{\rho_a}{\rho_b} g + \frac{3 \rho_a C_D(\text{Re})}{8 r \rho_b} (v_a - v_b) \cdot (v_a - v_b) .
\]  

(2)

The equations of the blob motion can be written as a system of four ordinary differential equations, two equations for the coordinates and two for the velocity. Then the system is solved using PDE2D.

The time step size will be chosen adaptively, between an upper limit of \( DT_{\text{MAX}} = TF/N_{\text{STEPS}} \) and a lower limit of 0.0001\( \cdot DT_{\text{MAX}} \). \( N_{\text{STEPS}} \) represents the minimum number of steps (\( N_{\text{STEPS}}=5 \)) and \( TF \) represents the time necessary for a blob to touch the base of the continental crust. Each time step, two steps of size \( DT/2 \) are taken, and that solution is compared with the result when one step of size \( DT \) is taken. If the maximum difference between the two answers is less than the relative tolerance (0.001), the time step \( DT \) is accepted. Then, the next step \( DT \) is doubled, if the agreement is "too" good; otherwise \( DT \) is halved and the process is repeated. As the tolerance is decreased, the global error decreases correspondingly. The Crank-Nicolson scheme is used to discretize the time. The steady mantle wedge velocity \( (v_a) \) field is obtained previously, for the model with temperature dependent viscosity (1).

The trajectories of blobs with the diameters varying between 0.2 and 10.0 km are calculated for different values of \( \eta_w \). The total rise times that the blobs require to reach the base of the continental crust are also estimated.

**Modeling results**

The thermal models which, correspond to the isoviscous mantle wedge and to the temperature-dependent viscosity, are presented in Fig.5, Fig.7 and Fig.13. The isoviscous mantle wedge model predicts temperatures of \( \sim 830 \) °C in the asthenosphere (Fig.5), beneath the Popocatepetl volcano, indicating that melting should not occur at least for dry olivine. The geotherm of the oceanic plate surface does not intersect the *solidus* for basalt (Fig.10A), suggesting that the oceanic plate does not suffer melting as well. The temperature at the base of the continental crust is also low, reaching values
of ~800 °C, which is not sufficient to produce melting. The back flow velocity field from the mantle wedge (the inflow into the mantle wedge) is relatively small, ~1.5 cm/yr. It is evident that simple model with the isoviscous mantle wedge cannot create any source of the volcanic material beneath the CMVB.

A series of benchmark tests with different grid resolutions have been done to verify the accuracy of numerical scheme realized in the present study. The slab-mantle wedge interface is the area where the isotherms have a very small incidence angle with this interface, thus significant errors might appear along this boundary. The benchmark tests on this boundary quantify the numerical errors as a function of mesh resolution. Grids with a systematic increase in element number from 4000 to 12000 elements in steps of 2000 elements are used for this benchmark.

The benchmark results are presented in Fig. 4. Large temperature fluctuations (up to 22 °C) close to the tip of the wedge occur for grids with 4000 and 6000 elements. Increasing the mesh resolution (8000 and 10000 triangles), this fluctuation are diminished and the maximum temperature fluctuation along the slab-wedge interface is of ~7 °C. Increasing the grid resolution to 12000 triangles, the numerical error is less then 5 °C. We stop the benchmark at this point because the computing time is increasing exponentially with the grid resolution (from half of hour for 4000 triangles to more than 10 hours for a model with 12000 elements, running on a Pentium 4 PC at 3 Ghz and 2 Gb RAM memory). To obtain final results we used the 12000 elements grid, thus our thermal models have the numerical error <5 °C.

The temperature dependent viscosity case has been investigated for a systematic variation of the rheological parameters \( \eta_0 \) and \( E_a \). Hirth and Kohlstedt (2003) showed that the viscosities experimentally estimated for olivine at upper mantle pressures and temperatures are in the range of \( 10^{17} \div 10^{21} \text{ Pa} \cdot \text{s} \). Consequently, the reference viscosity, \( \eta_0 \), has been varied in our model from \( 10^{17} \text{ Pa} \cdot \text{s} \) up to \( 10^{21} \text{ Pa} \cdot \text{s} \). The modeling results are presented in Fig. 7. The activation energy for diffusion creep in olivine is fixed at 300 kJ/mol (Karato and Wu, 1993). For the viscosity, \( \eta_0 \), in the range from \( 10^{17} \text{ Pa} \cdot \text{s} \) to \( 10^{20} \text{ Pa} \cdot \text{s} \), only a small increase of temperature (<15 °C) is observed beneath Popocatepetl (Fig. 9). The maximum temperature underneath the Popocatepetl is around 1260 °C. On the other hand, for the reference viscosity of \( \eta_0=10^{21} \text{ Pa} \cdot \text{s} \), a significant decrease of temperature (~200 °C) occurs. For this case a maximum temperature below Popocatepetl is ~1170 °C. The viscosity distributions in the mantle wedge for \( \eta_0=(10^{17} \div 10^{21}) \text{ Pa} \cdot \text{s} \) are presented in Fig. 8.

Experimentally obtained values of the activation energy for diffusion creep in olivine are of 300 kJ/mol (Karato and Wu, 1993) and of 315 kJ/mol (Hirth and Kohlstedt, 1995). The thermal models with variable activation energy from 150 kJ/mol up to 350 kJ/mol are presented in Fig. 13. Since the variation of the reference viscosity has a little effect on the overall thermal distribution (except for \( \eta_0 =10^{21} \text{ Pa} \cdot \text{s} \)), the modeling is done for constant \( \eta_0=10^{20} \text{ Pa} \cdot \text{s} \). Increase of the activation energy from 150
kJ/mol up to 300 kJ/mol results in a relatively small increase of temperature (<25 ºC). For bigger activation energies (~350 kJ/mol) an important increase in temperature (up to 70 ºC) is observed, and the maximum temperature below the Popocatepetl volcano reaches ~1330 ºC (Fig.12). The mantle wedge viscosity distributions for $E_a=(150\div350 \text{ kJ/mol})$ are presented in Fig. 14.

The geotherms along the slab surface are presented in the phase diagram for mafics and harzburgite (Hacker et al., 2002) for a variable reference viscosity, $\eta_0$, in Fig. 10, and for variable activation energy, $E_a$, in Fig. 15. The slab geotherm intersects the dehydration melting solidus for basalt at 50÷60 km for the reference viscosities $\eta_0=(10^{17}\div10^{20} \text{ Pa·s})$ and the activation energy $E_a=(150\div300 \text{ kJ/mol})$ (Fig. 10A and Fig. 15A).

The effect of the mantle wedge flow on the blob’s trajectory is an interesting aspect of the volcanic magma source problem. We selected an initial point for the blob trajectories at the depth of 70 km on the surface of the subducted slab. This initial point corresponds to the geometric vertical projection of the Popocatepetl volcano on the subducting Cocos plate. Estimated blob trajectories are presented in Fig. 16 for different wrapping viscosities ($10^{14} \div 5\cdot10^{17} \text{ Pa·s}$) and blob diameters ($0.2\div10.0 \text{ km}$).

For a wide range of activation energies ($150\div350 \text{ kJ/mol}$) and reference viscosities ($10^{17}\div10^{20} \text{ Pa·s}$), the blob trajectories and rising times are practically indistinguishable (Fig. 17). Only for the reference viscosity of $10^{21} \text{ Pa·s}$, an ~1 My increase in rising time is obtained due to lower velocities in the mantle wedge. A very low wrapping viscosity is essential to let the blob to rise up to the continental crust. Such low viscosity could result from the lubricating wrap around the blob. The source of the wrapping is apparently the melted material coming from the subducting slab, including the melted subducted sediments. Indeed, the water saturated sediments are likely to melt at a depths of 50÷58 km for $\eta_0=(10^{17}\div10^{21} \text{ Pa·s})$ (Fig. 11A), and of 45÷50 km for $E_a=(150\div350 \text{ kJ/mol})$ (Fig. 11B).

For the blobs of 2 km size (Fig. 16C) and the wrapping viscosity, $\eta_w>2\cdot10^{16} \text{ Pa·s}$, the drag force is predominant and the blob cannot rise. Decreasing the viscosity the drag force is less significant and at the depth of ~110 km the blob intercepts the mantle wedge back flow, which returns it toward the tip of the wedge. Finally the blob rises up and touches the continental crust after ~8 My. For the lower viscosity (<$9\cdot10^{15} \text{ Pa·s}$) the blobs are rising faster (Fig. 17) and pop up at approximately the same point below the Popocatepetl volcano (~350 km from the trench). The larger is the blob’s size the less time is necessary to reach the continental crust. Any blob of the size >0.6 km will touch the bottom of the continental crust in less then 4 My when the wrapping viscosity is of $1\cdot10^{15} \text{ Pa·s}$ (Fig. 16A). For the blob with the diameter of ~2 km, ~1 My is necessary to pop up if the $\eta_w<3\cdot10^{15} \text{ Pa·s}$. The buoyancy force of large size blobs becomes more dominant than the drag force and this yields a substantially upright trajectory. A 10 km blob touches the Moho in almost the same location for $\eta_w<10^{17} \text{ Pa·s}$ (Fig. 16B). On the other hand, 10 km blobs with the $\eta_w>5\cdot10^{17} \text{ Pa·s}$ would never rise up to the continental
crust (Fig. 16D). For a \( \eta_w \) fixed at \( 10^{17} \) Pa·s, blobs with diameters greater than 4 km will be able to traverse the mantle wedge flow and to go up toward the surface.

**Discussion and Conclusions**

Numerical models of temperature and velocity fields in the mantle wedge can contribute to our understanding of the magma generation, dynamics and volcanic sources in very atypical subduction zone of Central Mexico. In particular, the models, allowing for the ascending of buoyant material melted from the subducted slab and mantle in the form of plumes or blobs, can reveal the mechanisms of magma transport to the bottom of the continental plate.

Two basic thermo-mechanical models for the mantle wedge in the Central Mexico are considered in this study. The first model is restricted with the isoviscous mantle wedge whereas the second one is advanced with the temperature-dependent mantle viscosity. The first model (Fig.5) does not predict any melting conditions in the asthenosphere, beneath the CMVB, at the base of the continental crust and on the surface of the subducting slab. The slab surface geotherm intersect neither the dehydration melting solidus for basalt (Fig.10A) nor the solidus for H\(_2\)O saturated sediments (Fig.11A). The vertical temperature profile just beneath the Popocatepetl volcano through the mantle wedge does not encounter the wet peridotite solidus (Fig.10B), thus the melting of the hydrated peridotite should not occur.

The temperature and velocity fields in the second model depend on the rheological parameters, the reference viscosity, \( \eta_0 \), and the activation energy, \( E_a \). Variation of the reference viscosity, \( \eta_0 \), from \( 10^{17} \) Pa·s to \( 10^{20} \) Pa·s causes a slight temperature increase (<15 °C) close to the tip of the wedge (Fig.9). Increasing the reference viscosity up to \( \eta_0=10^{21} \) Pa·s provokes a significant drop of the tip temperature of about 200 °C. The phase diagram for mafic minerals shows (Fig.10A) that the melting of basaltic oceanic crust might take place at the depths of ~58 km for \( \eta_0=(10^{17}÷10^{20} \) Pa·s), and at ~80 km for \( \eta_0=10^{21} \) Pa·s. The water saturated sediments start to melt at shallower depths, between 50 km and 55 km (Fig.11A). The vertical temperature profile beneath the Popocatepetl volcano achieves the temperature which is high enough to melt the wet peridotite (Fig.10B).

The thermal models show that basalt melting initiates at ~60 km depth for the activation energy, \( E_a=150÷300 \) kJ/mol, and at ~50 km depth for \( E_a=350 \) kJ/mol (Fig.15A). In the range of reference viscosity values (\( 10^{17}÷10^{21} \) Pa·s), the saturated sediments may melt at the depth between 45 km and 50 km (Fig.15B). The temperature below the Popocatepetl volcano is well beyond the wet peridotite solidus for the whole range of activation energies applied in this study (150÷350 kJ/mol).

The modeling results ascertain that the melting of the oceanic crust is likely to occur in a narrow depth range of 50÷60 km (except for a reference viscosity of \( 10^{21} \) Pa·s which predicts the
melting at ~80 km depth). The subducted sediments begin to melt at shallower depths of 45÷55 km for the entire range of \( \eta_0 \) and \( Ea \).

As the oceanic crust melting starts, at first, the dacitic-rhyolitic magma is formed; afterward as it ascends and interacts with the mantle wedge, the adakitic magma might be formed. The oceanic slab contribution to the volcanism in Eastern Mexican Volcanic Belt has been reported by Gomez-Tuena et al. (2003). Magmatic rocks with the adakitical signature have been found recently in the Quaternary series in CMVB. The stratovolcano Nevado de Toluca shows evidences of the same adakitic mark as well (Gomez-Tuena - personal communication).

The temperature of the hydrated peridotite below the CMVB is beyond wet peridotite solidus, but it is still lower the dry solidus. This suggests that the hydration of the mantle wedge by fluids released from the subducted Cocos plate is a necessary condition for the partial melting of the mantle. Indeed, the metamorphic dehydration predicted by our thermal models might occur down to 80 km depth. The estimated variation of wt. %H\(_2\)O content with the depth along the subducting plate is presented in Fig.10A-inset. By the transformation of zoisite and amphibole into eclogite, ~0.6 wt. % H\(_2\)O may be released into the mantle wedge from the hydrous phases in the subducting slab through a dehydration at the depths between 40 km and 60 km.

Mantle xenolites (oxidized peridotite) have been found in Mexico near El Peñon (Fig.1), suggesting an important flux of volatiles from the subducting Cocos slab (Blatter and Carmichael, 1998). This dehydration would drop down the mantle viscosity in the tip of the wedge. This effect is not included in the present thermal models. The amount of serpentine in the tip of the wedge is much smaller for the model with temperature dependent viscosity (Fig.10B-upper left inset) than for the model with the isoviscous mantle wedge (Fig.10B-upper right inset). Furthermore, the amount of serpentine decreases as the activation energy increases (Fig.15B-upper insets). The serpentinized tip of the mantle wedge may have bearing on the down dip limit of the forarc coupled plate interface that controls an extension of aseismic slow slip transients (Kostoglodov et al., 2003; Manea et al., 2003).

The calcalkaline rocks are the most common series in the CMVB. This series includes the igneous rocks from basalts up to rhyolites, and is characterized by depletion of Fe. The depletion probably occurs because of the Fe and Ti oxides crystallization, which initially might be facilitated by the presence of fluids in the magma. The influx of volatiles from the metamorphosed oceanic crust and sediments triggers partial melting of peridotite above the subducted slab (Tatsumi, 1986; Davies and Stevenson, 1992). The majority of the calcalkalines in the CMVB is represented by the rocks with high content of K\(_2\)O and Na\(_2\)O, which originally might represent low degrees of partial melting in the mantle wedge. In fact, the wet solidus conditions for peridotite (Wyllie, 1979) are developed close to the slab surface (see Fig.13A).
Felsic magma formations were found in the monogenetic volcanic field of Chichinautzin (Marquez and De Ignacio, 2002). The models with the temperature-dependent viscosity show that at the base of the continental crust, the temperature exceeds 1100 °C (Fig. 7 and Fig. 13). That may create felsic magma sources in Sierra Chichinautzin by partial melting of the basaltic lower crust under low water fugacity conditions.

Recent paper of Gerya and Yuen (2003) demonstrates that Rayleigh-Taylor instabilities can develop and rise up from the surface of the cold subducting slab. They also suggest that the plumes detached from the slab might be lubricated by partially melted, low viscosity material of the subducted crust and hydrated mantle.

The modeling of the blob motion in the mantle wedge viscous flow induced by the subducting slab shows that this simple approach may help to understand the origin of the volcanism in the CMVB. The plume like blobs might origin at the slab-mantle wedge interface as a consequence of the thermal instability. It appears that the detachment area along the subducted slab from which the blobs might emerge is not very large. Within the wide range of values of the rheological parameters, the sediment and oceanic crust melting, and metamorphic dehydration is expected to occur in the present models at depths of 45–60 km.

A certain distance along the slab surface is necessary for the melted material to accumulate and to form the blob. Nevertheless, the same blob detachment point in the present model is selected on the subducted slab surface, just below the main volcanic structure, Popocatepetl volcano (70 km depth). In reality, the spatial (and temporal) variation of the composition of the material transferred from the subducted Cocos plate into the overlying mantle wedge is expected. Therefore the volcanic arc lavas may be enriched with incompatible elements and volatiles, and highly variable composition of the subducted-related lavas is probable. The proposed positively buoyant blobs might have a complex composition of melted H₂O saturated peridotite, melted sediments and oceanic crust.

Two parameters control the trajectories of the blob structures rising from the slab: the diameter of the blob and the wrapping viscosity. Very low values of the wrapping viscosity (10^{14} ± 5·10^{17} Pa·s) are necessary to reduce the drag force, which is critical for the blob to pop up. The lowest dynamic viscosity reported for dry mantle is ~10^{17} Pa·s (Moore et al., 1998). On the other hand, the viscosity of hydrated, partially molten blobs might be as low as 10^{14} Pa·s (Gerya and Yuen, 2003). Since the viscosity of the surrounding dry mantle controls the propagation of the blob, a certain mechanism is responsible for the occurrence of the low wrapping viscosity.

As the blob ascents through the mantle wedge, porous flow (Davies and Stevenson, 1992) of the melt and fluid might penetrate the mantle around the blob, dropping there the viscosity down to 10^{14} Pa·s. The rate of this penetration by porous flow has to be at least the same as the ascending rate of the blob. Recent laboratory experiments by Hall and Kincaid (2001) show that the positively buoyant blobs
might pass through the low-viscosity, low-density paths created by previous blobs. The mechanism that produces the low wrapping viscosities may be a combination of the porous flow and a low viscosity conduit left by previous blobs.

Since the rigid sphere approximation is used to estimate the drag force applied to weak partially molten blobs, the estimates of wrapping viscosity in the present study should be considered as rather approximate. The blobs of 10 km diameter reach the base of the continental crust in ~6 My when the wrapping viscosity is of \(3 \times 10^{17}\) Pa·s. Since this viscosity value is in the lower range reported for dry mantle (Moore et al., 1998), no wrapping viscosity mechanism is needed in this case. The blobs of such big size can affect locally the mantle wedge flow pattern and the effective viscosity distribution (Gerya and Yuen, 2003), especially for the temperature and strain-rate dependent viscosity (Ranalli, 1995). These effects are not included in the present models.

The time required for the blob of 1 km diameter to rise from the slab surface up to the continental crust varies between 0.001 My and 14 My for the viscosity between \(10^{14}\) Pa·s and \(5 \times 10^{17}\) Pa·s respectively. In general, the blob rising time decreases nonlinearly as its diameter increases and the wrapping viscosity is diminishing (Fig.17). The reference viscosity, \(\eta_0\), in the range up to \(10^{20}\) Pa·s, has a negligible effect on the blob trajectory. The effect can be noticed for the higher values of the reference viscosity (\(10^{21}\) Pa·s) when the raising time is more than ~2 My (Fig.17).

The blob tracing dynamic model in the mantle wedge velocity field shows that the “fast” trajectories end at the same focus location (below the Popocatpetl volcano, ~350 km from trench) on the base of the continental crust (Fig.16). This result may be interpreted as a possible condition for the development of the stratovolcanoes. The ending points of “slow” trajectories, which are common for the blobs of smaller size (~0.4÷0.6 km), are scattered from the focus location (Fig.16A). This observation may give us a hint on a possible mechanism of monovolcanic volcanism.

Recent studies (Lucia Capra - personal communication) revealed at least two magmatic pulses with a time span of ~1 My on the stratovolcano Nevado de Toluca. The maximum volume of each of these magmatic events is of 3.5 km³ (equivalent blob diameter is of ~2 km). It is interesting that both pulses started with a magmatic signature of melted sediments (Ce anomaly). These new studies will provide an important constrains on the blob size, rising time and the blob composition. From our model (Fig.16), a blob of 2 km in diameter reaches the base of the continental crust in 1 My, if the wrapping viscosity is of \(2 \times 10^{15}\) Pa·s. The low viscosity is essential for the smaller size blobs to rise to the base of the continental crust.

The average volume of a monogenic cinder cone in the CMVB is less than 1 km³ (Hasenaka, 1994), which corresponds to blob diameters of ~1.3 km. If the origin of monogenic cones are described by the blob tracing model, then the wrapping viscosity should be of \(\eta > 5 \times 10^{15}\) Pa·s to produce the
“slow” trajectories. We need further model enhancement to verify the relation between the blob (or plume) hypothesis and the origin of the strato and monogenic volcanism in the CMVB.

Acknowledgments

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References


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Table 1. Summary of the thermal parameters used in the models. (Compilation from: Peacock and Wang, 1999; Smith et al., 1979; Ziagos et al., 1985; Vacquier et al., 1967; Prol-Ledesma

<table>
<thead>
<tr>
<th>Geological Unit</th>
<th>Density (kg/m³)</th>
<th>Thermal Conductivity (W/mK)</th>
<th>Radiogenic heat production (µW/m³)</th>
<th>Thermal Capacity (MJ/m³K)</th>
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<td>Lower continental crust</td>
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<td>2.90</td>
<td>0.02</td>
<td>3.30</td>
</tr>
</tbody>
</table>

* Increase linearly with distance from the deformation front up to a depth of 10 Km.
Figure Captions

Figure 1. Tectonic setting and position of the modeled cross-section (straight line), in Guerrero. Triangles show the location of active volcanoes in Mexico. Dashed ellipse is the CMVB – Central Mexican Volcanic Belt. Light blue rectangle represents the El Peñon area, where mantle xenolites have been found. Arrows show convergence velocities between the Cocos and North American plates (DeMets et al., 1994).

Figure 2. The grid with 12000 elements used to solve the numerical models. This particular shape of the grid is designed to better resolve the velocity field nearby the tip of the mantle wedge.

Figure 3. Boundary conditions and parameters used in the modeling. The upper and lower boundaries have constant temperatures of 0 ºC and 1450 ºC, accordingly. The continental plate is fixed. The right (landward) vertical boundary: 20 ºC/km thermal gradient in the continental crust (down to 40 km); between 40 km and 100 km depth the thermal gradient is of 10 ºC/km; no horizontal conductive heat flow is specified beneath 100 km depth. Zero traction is considered beneath Moho (40 Km), at the boundary, which belongs to the mantle wedge. The convergence velocity is specified for the intersection between the right boundary and the subducting slab. The left (seaward) boundary condition is a one-dimensional geotherm for the oceanic plate. The Cocos plate motion is referred to the North American plate with the convergence velocity of 5.5 cm/yr. Volumetric shear heating is imposed along the plate interface up to a maximum depth of 40 km, using the Byerlee's friction law (Byerlee, 1978).

Figure 4. The results of benchmark test with different grid resolutions: 4000, 6000, 8000, 1000 and 12000 triangles. The temperature differences between two subsequent models are calculated along the slab-wedge interface. Note the temperature fluctuation up to 22 ºC for low grid resolution (blue curve). Less than 5 ºC numerical error results when the grid with 12000 triangles is used (red curve).

Figure 5. Calculated steady-state thermal field for the isoviscous mantle wedge. Horizontal black dashed line shows the Moho (40 km depth). The hinge point is at 270 km from the trench. Thick violet solid line denotes the top of the subducting slab. Orange triangles are the two principal stratovolcanoes in CMVB: Nevado de Toluca and Popocatepetl. The maximum temperature beneath Popocatepetl volcano is about of 830 ºC. The lower left inset is the magnified thermal structure close to the tip of the mantle wedge. Thin dark blue arrows in the mantle wedge represent the velocity field. The intraslab
earthquakes with magnitudes, $M_w \geq 5.5$, are represented by the focal mechanisms. The two brown clouds of hypocenters beneath the coast denote the smaller magnitude seismicity associated probably with the bending of the subducted plate.

**Figure 6.** Variation of the surface heat flow along the Guerrero profile. The dots with vertical error bars are heat flow data reported by Ziagos et al. (1985).

**A.** Surface heat flow for steady-state thermal models reference viscosities $\eta_0 = 10^{17} \div 10^{21}$ Pa·s and $E_a = 300$ kJ/mol.

**B.** Surface heat flow for steady-state thermal models activation energies $E_a = 150 \div 350$ kJ/mol and $\eta_0 = 10^{20}$ Pa·s.

**Figure 7.** Steady-state thermal models with strong temperature-dependent viscosity in the mantle wedge. The reference viscosity, $\eta_0$, varies from $10^{17}$ Pa·s (A.) up to $10^{21}$ Pa·s (E.). The activation energy for olivine of 300 kJ/mol is used. Note an important increase of the temperature close to the tip of the wedge (A.+D.). NP and P represent the Nevado de Toluca and Popocatepetl volcanoes. The maximum temperature beneath the Popocatepetl volcano is about of 1230 °C. Other notations are the same as in *Fig. 5*.

**Figure 8.** Distribution of the mantle wedge viscosity for steady-state thermal (*Fig. 5*) models with strong temperature-dependent viscosity. The reference viscosity, $\eta_0$, is from $10^{17}$ Pa·s (A.) up to $10^{21}$ Pa·s (E.).

**Figure 9.** Vertical temperature profile (A-A’) for the mantle wedge thermal models with the reference viscosities of $10^{17} \div 10^{21}$ Pa·s. The activation energy is fixed at 300 kJ/mol. The temperature profile for the model with the isoviscous mantle is also shown as a dashed line.

**Figure 10. A.** Phase diagrams for the MORB (Hacker et al., 2002). 1 – Zeolite (4.6 wt%H$_2$O), 2 - Prehnite-Pumpellyite (4.5 wt%H$_2$O), 3 - Pumpellyite-Actinolite (4.4 wt%H$_2$O), 4 – Greenschist (3.3 wt%H$_2$O), 5 – Lawsonite-Blueschist (5.4 wt%H$_2$O), 6 – Epidote-Blueschist (3.1 wt%H$_2$O), 7 – Epidote-Amphibolite (2.1 wt%H$_2$O), 8 - Jadeite-Epidote-Blueschist (3.1 wt%H$_2$O), 9 - Eclogite-Amphibole (2.4 wt%H$_2$O), 10 – Amphibolite (1.3 wt%H$_2$O), 11 – Garnet-Amphibolite (1.2 wt%H$_2$O), 12 – Granulite (0.5 wt%H$_2$O), 13 – Garnet-Granulite (0.0 wt%H$_2$O), 14 – Jaedite-Lawsonite-Blueschist (5.4 wt%H$_2$O), 15 – Lawsonite-Amphibole-Eclogite (3.0 wt%H$_2$O), 16 – Jaedite-Lawsonite-Talcschist, 17 – Zoisite-Amphibole-Eclogite (0.7 wt%H$_2$O), 18 – Amphibole-Eclogite (0.6 wt%H$_2$O), 19 – Zoisite-Eclogite (0.3 wt%H$_2$O), 20 – Eclogite (0.1 wt%H$_2$O), 21 – Coesite-Eclogite (0.1 wt%H$_2$O), 22
Diamond-Eclogite (0.1 wt%H$_2$O). Slab surface geotherms are calculated for the reference viscosity range: $10^{17} \div 10^{21}$ Pa.s (see inset).

**B.** Phase diagram for harzburgite (*Hacker et al., 2002*). 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). Calculated geotherms for the slab surface are the same as in A. The vertical temperature profile beneath the Popocatepetl volcano are the same as in *Fig.9*. The amount of serpentine (green-yellow hatched insets) in the tip of the mantle wedge is noticeably smaller for the model with temperature dependent viscosity (upper left inset) than for the model with the isoviscous mantle wedge (upper right inset).

**Figure 11.** A. Calculated slab surface geotherms for models with reference viscosities in the range of $10^{17} \div 10^{21}$ Pa.s (see inset), and fluid saturated sediment *solidus* from (*Nichols et al., 1994*). The horizontal dashed arrows mark the depths where the sediment meting might occur. Dashed line represents the geotherm for the isoviscous mantle.

B. Calculated slab surface geotherms for models with activation energy range: 150÷350 kJ/mol (see inset) and fluid saturated sediment *solidus* from *Nichols et al., 1994*. The horizontal dashed arrows mark the depths where sediment meting might occur.

**Figure 12.** Vertical temperature profile (A-A’) through the mantle wedge. The thermal modeling is done for activation energies of 150÷350 kJ/mol. The reference viscosity is fixed at $10^{20}$ Pa·s. Dashed line shows, for a comparison, the temperature profile for the model with the isoviscous mantle wedge.

**Figure 13.** Steady-state thermal models with strong temperature-dependent viscosity in the mantle wedge for activation energies, $E_a$, from 150 kJ/mol (A.) to 350 kJ/mol (E.). The reference viscosity of $10^{20}$ Pa·s is used. The maximum temperature beneath the Popocatepetl volcano is about of $\sim 1250$ °C (A.÷D.) and more than 1300 °C (E.). Other notations are the same as in *Fig.5*.

**Figure 14.** Distribution of mantle wedge viscosity with strong temperature-dependence for steady-state thermal models (*Fig.12*). Activation energies, $E_a$, are from 150 kJ/mol up to 350 kJ/mol.

**Figure 15.** A. Phase diagrams for the MORB and maximum H$_2$O contents (*Hacker et al., 2002*). Modeled slab surface geotherms are plotted for the activation energy range: 150÷350 kJ/mol (see inset). The other notations are the same as in *Fig.10*.
B. Phase diagram for harzburgite, and maximum H$_2$O contents (Hacker et al., 2002). The calculated geotherms are the same as in A, as well as those from the vertical profile beneath Popocatepetl (see Fig.14). The amount of serpentine (green-yellow hatch upper insets) in the tip of the wedge decreases as the activation energy increases.

**Figure 16.** Blob trajectories in the steady mantle wedge flow Initial points for all trajectories are selected on the surface of the slab, right below the Popocatepetl volcano at 70 km depth.

A. The wrapping viscosity is fixed at $10^{15}$ Pa·s. The blobs with the diameter less than 0.4 would never rise up to the continental crust.

B. The blob diameter is fixed at 10 km. The trajectories corresponding to different wrapping viscosities of less than $10^{17}$ Pa·s have the same final point below the Popocatepetl volcano. The blobs would never rise up to the continental crust if with the wrapping viscosities is $>5\cdot10^{17}$ Pa·s.

C. The blob’s diameter is fixed at 2 km. The trajectories corresponding to different wrapping viscosities of less than $9\cdot10^{15}$ Pa·s, have the same final point below the Popocatepetl volcano. The blobs with the wrapping viscosities $>2\cdot10^{16}$ Pa·s would never rise up to the continental crust.

D. The wrapping viscosity is fixed at $10^{17}$ Pa·s. The trajectories correspond to different blob size (4÷10 km). The blobs with the diameter less than 4 km would never rise up to the continental crust.

**Figure 17.** Blob rising time as a function of the wrapping viscosity. For a wide range of rheological parameters ($\eta=10^{17}÷10^{20}$ Pa·s and $Ea =150÷350$ kJ/mol) the blob trajectories and rising times are nearly identical (blue curves). For the reference viscosity of $10^{21}$ Pa·s, the rising time becomes longer after ~1 My, for a given wrapping viscosity (dashed green curves). The curves annotated with the blob’s diameter show that the rising time is decreasing for the bigger blobs and the lower viscosity. The blobs with the size of ~ 2 km can reach the continental crust in less than 1 My (black dashed line) at a wide range of low wrapping viscosities.
Figures

Figure 1.
Figure 3.
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Figure 13.
Figure 14.

A. Reference viscosity $\eta_0 = 10^{20}$ Pa.s
Activation energy $E_a = 150$ KJ/mol

B. Reference viscosity $\eta_0 = 10^{20}$ Pa.s
Activation energy $E_a = 200$ KJ/mol

C. Reference viscosity $\eta_0 = 10^{20}$ Pa.s
Activation energy $E_a = 250$ KJ/mol

D. Reference viscosity $\eta_0 = 10^{20}$ Pa.s
Activation energy $E_a = 250$ KJ/mol

E. Reference viscosity $\eta_0 = 10^{20}$ Pa.s
Activation energy $E_a = 350$ KJ/mol
Figure 15.
Figure 16.
Figure 17.