Shallow slab detachment as a transient source of heat at midlithospheric depths

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Shallow slab detachment as a transient source of heat at midlithospheric depths

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Abstract. Slab detachment (or breakoff) has been proposed as a cause of temperature-related processes associated with subduction, such as postcollisional magmatism, mineralization, and metamorphism. In this study, we quantitatively investigate the breakoff process and the subsequent thermal evolution of a plate boundary region involving continental collision after a prolonged period of oceanic lithosphere subduction. Our two-dimensional time-dependent thermomechanical modeling shows that the dense, oceanic part of the slab can become detached at depths as shallow as 35 km. The detached part of the slab sinks into the mantle, creating a gap in the lithospheric system which is filled with upwelling hot asthenosphere. The resulting temperature increase in the overlying material can be more than 500°C. It allows for partial melting of the asthenosphere and the overriding metasomatized lithosphere for a timespan of a few millions of years. Crustal anatexis and related magmatism and mineralization can proceed over a considerably longer period. The quantification of the conditions required for shallow slab detachment will contribute to warranted assessments concerning the role of slab detachment (relative to other proposed heat sources, such as tectonically accreted radioactive material) in the geodynamical evolution of former convergent plate boundaries.

1. Introduction

By their very nature, active continental margins and convergent plate margins in general have played, and continue to play, a primary role in plate tectonics studies. As a result, much insight has been gained into the relation between plate convergence and several closely associated geodynamic processes of widely different nature. Although some of these processes can be understood on the basis of stationary mode subduction, many subduction-related processes point to distinctly nonstationary aspects of the subduction process. In this study we address a subgroup of the latter category, i.e., those with a strong thermal component, such as magmatism, mineralization, and metamorphism.

Several types of transient processes may be envisaged as underlying causes of this nonstationarity, e.g., ridge subduction [*Atwater,* 1970; *Vlaar,* 1983; *Haeussler et al.,* 1995] and slab window formation [*Furlong et al.,* 1989], and removal of thickened lithosphere [*Platt and Vissers,* 1989; *Platt et al.,* 1998]. Here we focus on another transient aspect of subduction: slab detachment, also referred to as slab breakoff. Progress in the investigation of this process was prompted by developments in seismic tomography resulting in a significant increase in information concerning the deep structure of convergent plate margins. Seismic tomographic studies provided evidence for the breaking off of parts of subducted plates in a number of subduction zones, among which those in the Mediterranean region were subject of detailed analysis [*Spakman et al.,* 1988, 1993; *Wortel and Spakman,* 1992]. By now, considerable attention has been given to the study of slab detachment and to assess its role in, for example, temporal variations in state of stress, vertical and horizontal motions, and arc migration [*Wortel and Spakman,* 1992; *Meijer and Wortel,* 1996; *van der Meulen et al.,* 1998, 1999, 2000; *Carminati et al.,* 1998a, 1998b]. The combined results and the underlying tomographic studies provide considerable evidence in support of the (past) occurrence of slab detachment in the convergence zones involved (see *Wortel and Spakman* [2000] for a review).

Whereas the above mentioned investigations predominantly addressed the mechanical response of the lithosphere to slab detachment, *Davies and von Blanckenburg* [1995] drew attention to the thermal aspects of the process and proposed that slab detachment could also account for a series of (features of) temperature-related processes, such as metamorphism and magmatism. The key issue in this context is the premise that slab breakoff may occur at a quite shallow depth (~50 km) allowing hot asthenospheric material to quickly rise to the corresponding depth, thus providing an advective type of heat source. *Davies and von Blanckenburg* [1995, 1998] corroborated this premise by results of simple steady state and one-dimensional modeling and presented data on composition of volcanics in the Alpine region strongly supporting this mechanism [*von Blanckenburg and Davies,* 1995; see also *von Blanckenburg et al.,* 1998]. *De Boorder et al.* [1998] noted that a spatial relationship exists between regions where slab detachment is inferred to have taken place and zones of mineralization. These zones could be expressions of the above mentioned rise of asthenospheric material into the gap created by slab detachment, and the subsequent melt generation. In addition, evidence on reheating phases, during exhumation, at depths of 30 to 40 km,

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with temperature increases of up to 200°C, is inferred from metamorphic petrology studies of, for example, the Ronda peridotite in southern Spain [*van der Wal and Vissers*, 1993; *Vissers et al.*, 1995]. Slab detachment has been suggested as the underlying cause of this temperature increase by *Blanco and Spakman* [1993] and *Zeck* [1996]. Similar PTt results were obtained for the Alpi Arami peridotite body in the central Alps [*Brouwer et al.,* 1999].

These results and considerations lead us to investigate the thermal evolution of an active margin following a slab detachment event. Acknowledging the transient nature of the slab detachment process is the basis of our approach. The stage for our analysis was set by *Wong A Ton and Wortel* [1997], who modeled, in 2-D, the time-dependent dynamics (force, stress, and strength distribution) of a subduction zone involving continental collision after a prolonged period of oceanic lithosphere subduction. Continental collision zones were chosen because they are prime candidates for slab detachment to occur. We extend their thermomechanical analysis of the timing and depth of slab detachment to quantitatively study, in an internally consistent manner, the effect of breakoff on the thermal evolution of both the overriding and downgoing plate. Next, the results are analyzed in the light of the first-order question: Can slab detachment be the source of reheating, melt generation, and magmatism? With respect to PTt paths of exhumed rocks, this means that we investigate the mechanism underlying the reheating, and not the actual exhumation process itself. Similarly, in the context of postcollisional magmatism and mineralization we are concerned with the source of heat which generates melting, as a prerequisite for the generation of both magma and fluids which subsequently migrate upward to erupt or to form a mineralized body, respectively.

2. Model

2.1. Geometry of the Model

The geometry of the downgoing plate, especially the top of the slab, is modeled as a circular arc [*Van den Beukel and Wortel*, 1988]. The descending slab can consist of oceanic and/or continental lithosphere (Figure 1). In general, a combination of the two is used, whereby oceanic lithosphere subduction precedes continental lithosphere subduction. The oceanic lithospheric thickness is age dependent and about 100 km for a 100 Ma oceanic slab. The accompanying crust has a thickness of 10 km. The thickness of the subducting continental lithosphere is 125 km. Its crust is divided into an upper and lower part, each 15 km thick. A subducting continental margin is modeled by a transition zone between the oceanic and continental lithospheric slab. This transition zone, in which the thicknesses change gradually, has a variable width. The continental lithosphere of the overriding plate is 110 km thick with a 25-km-thick crust. A trench system is modeled by a 50-km-wide zone in which the topography increases from a depth of 6.5 km at the trench to zero depths in the overriding plate.

2.2. Thermal Model

The temperature distribution changes because of the subduction of relatively cold material. The variation of temperature with time is calculated by solving the heat transfer equation numerically through finite difference methods. The advection velocity is equal to the convergence velocity of the two plates involved. For material in the overriding plate the advection velocity is assumed to be zero, indicating no internal deformation. The boundary conditions are as specified in Figure 1.

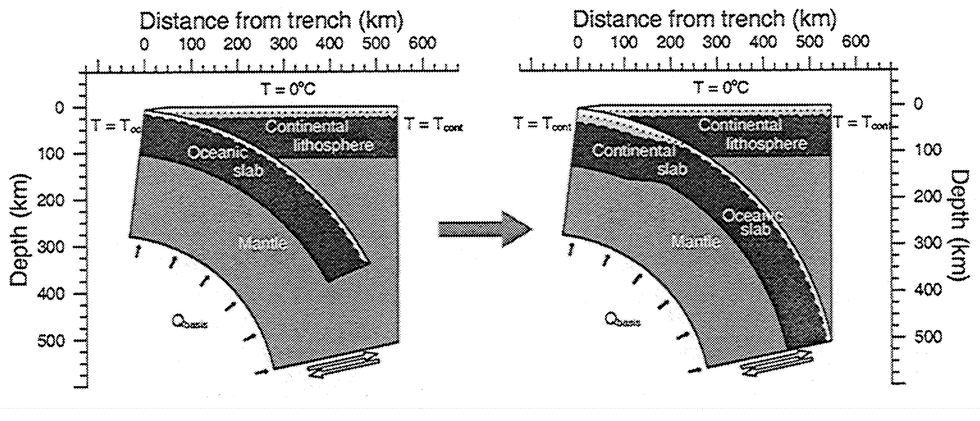


Figure 1. Geometry of the time-dependent model with the various boundary conditions. The slab can consist of both oceanic and continental lithosphere with a transition zone of variable width in between. The boundary conditions are a zero temperature at the surface, an oceanic or continental geotherm at the left side of the model (depending on the nature of the subducting lithosphere), a constant heat flux at the lower left boundary, and a continental geotherm at the right boundary. At the lower right boundary, heat transfers conductively only parallel to that boundary.

The initial thermal state of oceanic lithosphere is determined by the boundary layer model with a uniform constant basal heat flux [*Crough,* 1975]. The initial temperatures in a continental lithosphere are calculated with an analytical function inferred from the heat transfer equation by assuming no advection and a steady state situation, and by neglecting lateral changes [e.g., *Chapman,* 1986]. This temperature distribution is parameterized by the continental surface heat flow and is adjusted in such a way that the temperature at the base of the lithosphere, T_m, is 1325°C. For the starting temperatures in the whole model, a surface heat flow of 60 mW m^{-2} is used. The initial mantle temperatures are computed from a temperature gradient due to adiabatic compression and depend on the value of T_m. The specific heat has a value of 1050 J kg^{-1} $°C^{-1}$. The crustal thermal conductivity remains constant during modeling, namely, 2.5 W m^{-1} $°C^{-1}$. The conductivity of the subcrustal region changes according to the specific temperature dependence of this parameter obtained by *Schatz and Simmons* [1972]. The densities of the upper and lower continental crust and the mantle are 2700, 3000, and 3300 kg m^{-3}, respectively. The density of the oceanic crust is also 2700 kg m^{-3}. Radiogenic heat production within the oceanic crust is evenly distributed (0.5 μW m^{-3}), whereas it exponentially decreases with depth in the continental crust. The heat production at the surface (A_0) is computed with the assumption that 60% of the surface heat flow stems from subcrustal regions [*Pollack and Chapman,* 1977].

The amount of frictional heat generated at the plate contact linearly depends on the convergence velocity and the shear stresses, and it is inversely proportional to the thickness of the friction layer [*van den Beukel and Wortel,* 1988]. The shear stress depends on temperature and depth since pressure-dependent and temperature-dependent rheologies are used. The inverse proportionality to the thickness of the friction layer results in a maximum frictional heat for very thin layers.

2.3. Strengths and Forces

The maximum differential stresses in a cross section of the slab are integrated over its thickness to give the integrated strength of the slab as a function of the downdip distance [*Wong A Ton and Wortel,* 1997]. Pressure-dependent brittle deformation and temperature-dependent ductile deformation are used to calculate the stresses in both the oceanic and the continental descending plate. Byerlee's law is used to obtain the maximum principal stress differences for extension and compression, whereby one of the principal stresses is oriented vertically and equal to the lithostatic pressure [*Brace and Kohlstedt,* 1980; *Kohlstedt et al.,* 1995]. Near-hydrostatic pore fluid pressure is assumed. Ductile stresses for the mantle part of the slab are obtained from a dry dunite power law creep [*Chopra and Paterson,* 1984] and an olivine Dorn creep flow law [*Goetze,* 1978], for differential stresses below and above 200 MPa, respectively. Power law rheologies of wet quartzite [*Koch et al.,* 1980] and wet diorite [*Hansen and Carter,* 1982] are used to describe the stresses for ductile deformation in the upper and lower continental crust. The strain rate is taken to be 10^{-15} s^{-1}.

The forces acting on a subducting slab are formulated in terms of the slab pull force F_{sp} and resistive forces F_R. Following *Davies* [1980], they are calculated accumulatively as a function of downdip distance along the slab's upper surface; this implies integration over the slab's thickness and over the distance from the trench to a particular downdip distance. Downdip-oriented forces are defined as positive, and updip forces are defined as negative. The total accumulative downdip force is the sum of the driving and resistive forces acting on the slab. Positive and negative values of the total accumulative force correspond with downdip compression and downdip tension, respectively [*Davies,* 1980].

The accumulative slab pull force is calculated through density differences between the relatively cold, dense slab and the surrounding mantle. The total resistive force consists of compositional buoyancy of the subducting crust, a shallow resistance in the plate contact region (interplate friction) and a deep resistance (including viscous shear), arising from the slab plunging into the viscous mantle. Its distribution is related to the convergence velocity by the terminal velocity of the downgoing plate. A slab descending at rates higher than its terminal velocity experiences a large resistance from the mantle, which results in a minimum tensional force at the trench [*Forsyth and Uyeda,* 1975]. On the other hand, a nonmoving slab does not give rise to a resistive force at the lower end of the slab. Instead, the resistive forces are concentrated at the top, in the trench region; the maximum (upward) resistive force is then taken to be equal to the strength of the slab at the trench. A linear trade-off relation can be used in between these two end-member situations.

Note that the requirement of dynamical equilibrium implies that the total accumulative force acting on the slab should be zero at the leading edge of the slab (Figure 2a). A contribution to the dynamical force balance of the subducting slab may result from stress transfer (in the trench region) from the horizontal part of the plate. This contribution can be incorporated in the shallow resistance. Forces associated with major phase changes, in particular those near 410 and 660 km depth [see *Jackson and Rigden,* 1998], contribute to the force balance, but their effect is not relevant in the present context of (shallow) slab breakoff. A more detailed description of the calculation of the forces was given by *Wong A Ton and Wortel* [1997].

Strengths and forces are monitored in time. At first, only oceanic material subducts. A steady state situation is established within a few millions of years (Figure 2a). The strength increases from the trench downward, mostly due to the pressure dependence of the brittle deformation. The tensional strength of the slab starts to decrease when duc-

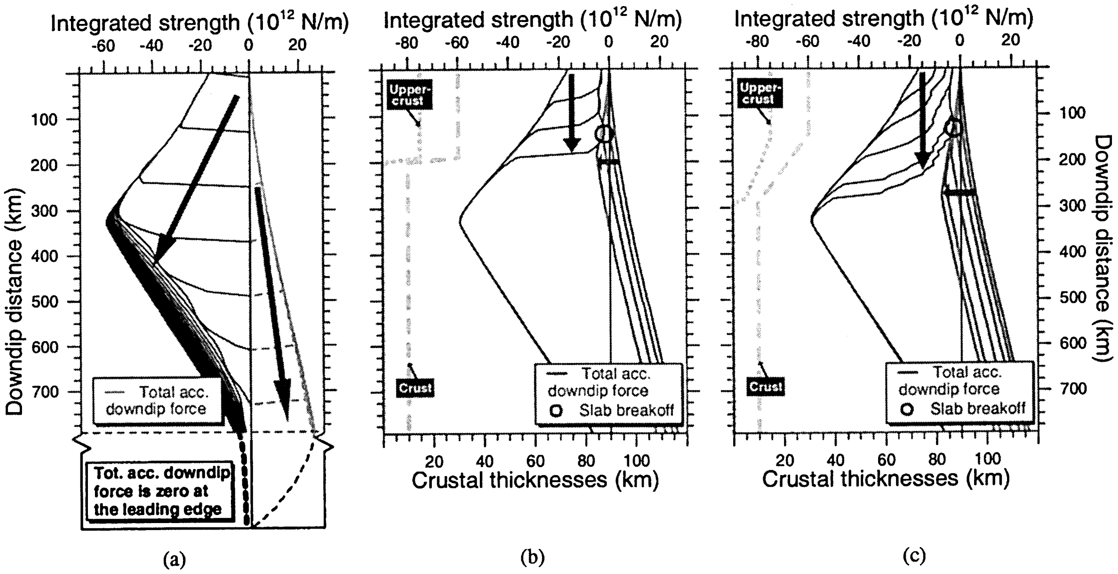


Figure 2. Integrated strength versus downdip distance from the trench and the total accumulative downdip forces. (a) A 100 Ma oceanic lithosphere is subducted with a velocity of 6 cm yr^{-1} for 50 Myr. The arrows denote the change of the strength and force curves in time. Because the slab must at all times be dynamically balanced, the total accumulative downdip force must be zero at the leading edge of the slab (schematically indicated at the bottom of Figure 2a). (b) The subduction of continental lithosphere with a surface heat flow of 80 mW m^{-2} causes a dramatic weakening of the slab and a change in the sign of the forces. The forces exceed the strength of the slab at a downdip distance of 140 km (at 3 Myr after the onset of continental subduction). There is a sharp transition between the oceanic and continental part of the slab. (c) A transition zone of 150 km width is used to model the subducting continental margin. Notice the change in thickness of the (upper)crust (light gray). The strength drops more gradually after the onset of continental lithosphere subduction, resulting in a breakoff depth of 130 km (after 4.5 Myr).

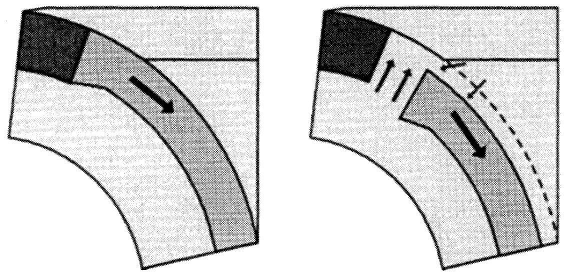
tile deformation produces the smallest maximum differential stresses. The total accumulative downdip force is positive along the entire slab and increases with depth.

The strength of the slab decreases rapidly if continental lithosphere starts to subduct, while the forces become upwardly directed due to the compositional buoyancy of the continental crust (Figure 2b). Once the forces acting on the slab exceed its tensional strength, the slab is assumed to break off. This appears to take place in or near the transition zone between the oceanic and continental lithospheric slab. A small transition zone results in a very rapid decrease in the strength of the slab due to the thicker weak (lower)crust. If the subducting continental margin is wider, it takes longer before continental crust of normal thickness starts to subduct. Since this is the weakest part of the slab it is most probable that breakoff occurs in this part of the slab (as indicated by our calculations). The width of the zone has a negligible effect on the breakoff depth (compare Figures 2b and 2c).

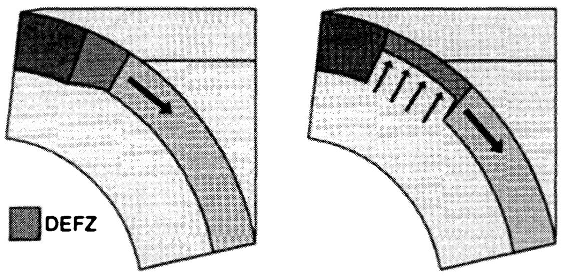
In order to arrive at a conservative estimate (i.e., an overestimate) of the breakoff depth, we neglected the shallow resistance, including the stress transfer from the horizontal part of the subducting plate. Incorporating this shallow resistance would normally result in a reduction of the resulting breakoff depth. The case of a relatively large compressive stress transfer, in which slab breakoff occurs either at greater depths or not at all, is not considered in this study. The breakoff results are represented by the breakoff depth and the breakoff time, i.e., the time elapsed since the start of continental subduction. Note that the breakoff depth is in terms of downdip distance from the trench.

2.4. Breaking and Sinking Models

2.4.1. FreeSink model. Within realistic limits the broken-off part sinks in a particular direction. The rate of descent is controlled by two velocities. The first ($V_{s//}$), referred to as the parallel sinking velocity or as the velocity at which the gap opens, is directed parallel to the top of the slab. The second, $V_{s\perp}$, is directed perpendicular to the first and allows the slab to detach immediately from the overriding lithosphere. Mantle material from above and beneath the slab flows into the gap as soon as the slab is assumed to break off, causing instantaneous temperature increases in the overriding plate (Figure 3a). A special case of the FreeSink-model features a zero $V_{s\perp}$. The slab sinks into the mantle along the circular arc with the parallel sinking velocity $V_{s//}$. For points in the overriding plate situated much deeper than the breakoff depth, the temperatures continue to decrease until the broken-off piece has passed entirely before rising due to the upwelling hot asthenospheric material.



(a) FreeSink model



(b) Stretching model

Figure 3. Two different models for the breaking and sinking of the slab and the filling of the gap. (a) FreeSink model; The broken-off piece of the slab moves into a particular direction. The gap is filled with asthenospheric material from beneath and above the lithospheric slab. (b) Stretching model. A deformation zone (DEFZ) is predefined in which stretching takes place. The slab sinks along the circular arc, and the gap is filled from below.

2.4.2. Stretching model. *McKenzie* [1978] proposed a, by now classical, model for the development and evolution of sedimentary basins involving stretching and thinning of continental lithosphere. An analogous model is used to represent the breaking of the slab. A deformation zone is defined in which the stretching takes place (DEFZ in Figure 3b). As the gap widens, the zone becomes thinner, allowing upwelling hot asthenosphere to fill the gap below the stretched zone. The temperature distribution in the extended area is adjusted in such a way that the temperatures are preserved at the top and the basis of the thinned lithosphere. Stretching of the zone and of the inflown mantle material continues until the zone is very thin. The broken-off piece of the slab sinks into the mantle along the arc geometry.

The FreeSink model can be regarded as an upper bound for the heat input at the plate contact. The Stretching model gives a lower bound, provided that the combination of the width of the deformation zone and $V_{s//}$ gives relatively slow thinning of the slab.

3. Slab Breakoff and Thermal Evolution of the Overriding Plate

The magnitude of the temperature increase following slab breakoff is directly related to the difference between the temperature in the overriding plate just before breakoff and the temperature of the hot mantle material filling the gap. Since the temperature of the inflowing asthenospheric mantle is always about the temperature T_m, the temperature distribution just before breakoff in the overlying continent, $T_0(r)$, is the key factor. A large difference between T_0 and T_m will result in a large temperature jump.

Apart from the effect of the breaking mechanism and the sinking velocity, we can subdivide the various parameters affecting the temperature increase into three different classes: (1) factors which only influence the breakoff position, e.g., the rheology of the continental slab, (2) parameters controlling both breakoff and T_0, e.g., the convergence velocity during continental subduction, and (3) variables which only affect T_0, such as the duration of oceanic subduction before continental subduction.

3.1. Reference Model

To assess the effects of various parameters on the thermal evolution of the convergent plate boundary, we first choose a set of reference parameter values (Table 1). The temporal distributions of the strengths and forces calculated with this reference set are shown in Figures 2a and 2b. The slab detaches at a downdip distance of 140 km after 3 Myr of continental lithosphere subduction. The resulting temperature distributions at specific points in time are displayed in Plate 1. Plate 1b shows the familiar pattern of depressed geotherms in a subduction zone. The temperature distribution just before the slab detaches is depicted in Plate 1c. After breakoff, the filling of the gap with hot asthenospheric material causes the temperatures in the overriding continent to increase. The area near the plate contact is affected most. While the detached part of the slab sinks further into the mantle, a new thermal equilibrium is formed in the upper part of the model. Convergence is assumed to stop after slab detachment occurs.

Hereinafter, we will display the thermal evolution of the overriding material by using stationary monitoring points

Table 1. Reference Parameter Values

Variable	Reference Value
Convergence velocity (V_c) (until breakoff)	6 cm yr^{-1}
Duration of oceanic subduction	50 Myr
Age of the oceanic lithosphere	100 Ma
Surface heat flow of the slab (Q_{slab})	80 mW m^{-2}
Surface heat flow of overriding plate (Q_{cont})	60 mW m^{-2}
Rheology of the upper crust	wet quartzite
Rheology of the lower crust	wet diorite
Rheology of the mantle	dry dunite
Strain rate ($\dot{\varepsilon}$)	10^{-15} s^{-1}
Thickness of the friction layer (Z_{fr})	10 km
$V_{s//}$	6 cm yr^{-1}
$V_{s\perp}$	1 cm yr^{-1}

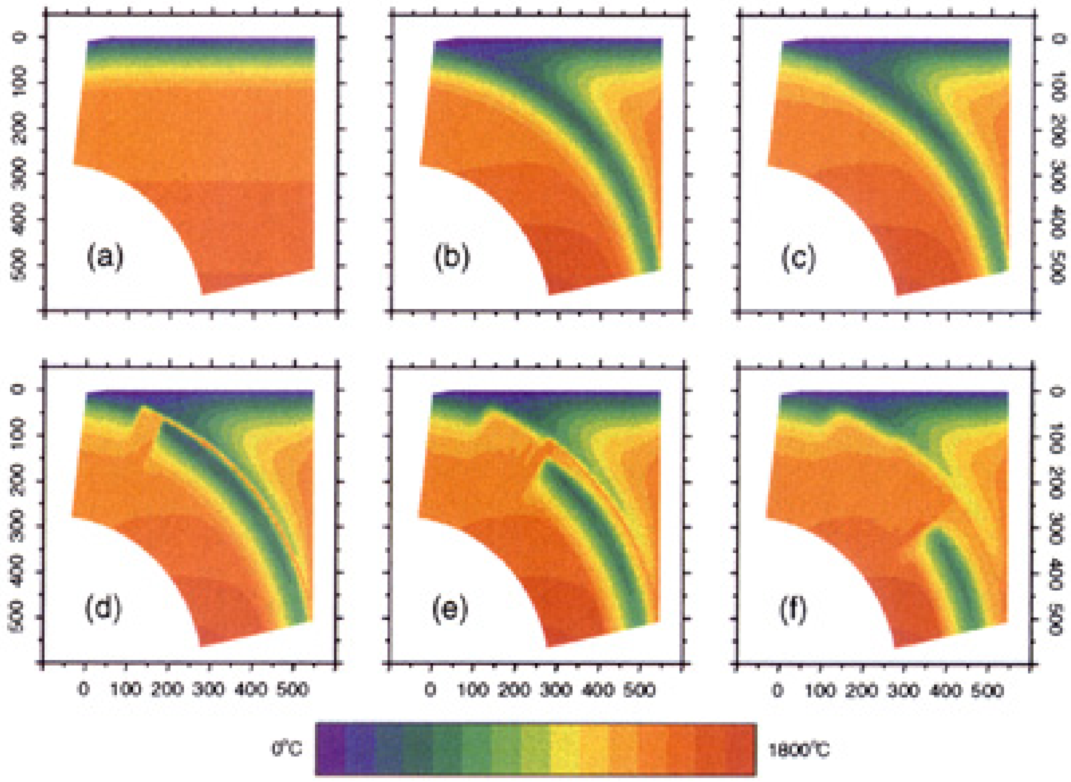


Plate 1. Six stages in the thermal evolution of the model: (a) Initial situation, (b) 50 Myr of oceanic lithosphere subduction, (c) 3 Myr of continental lithosphere subduction, (d) 1 Myr after slab breakoff, (e) 3 Myr after breakoff, (f) 6 Myr after breakoff. The FreeSink model with a $V_{s//}$ of 6 cm yr^{-1} (same as convergence velocity) and a $V_{s\perp}$ of 1 cm yr^{-1} is used. See Table 1 for the other parameter values.

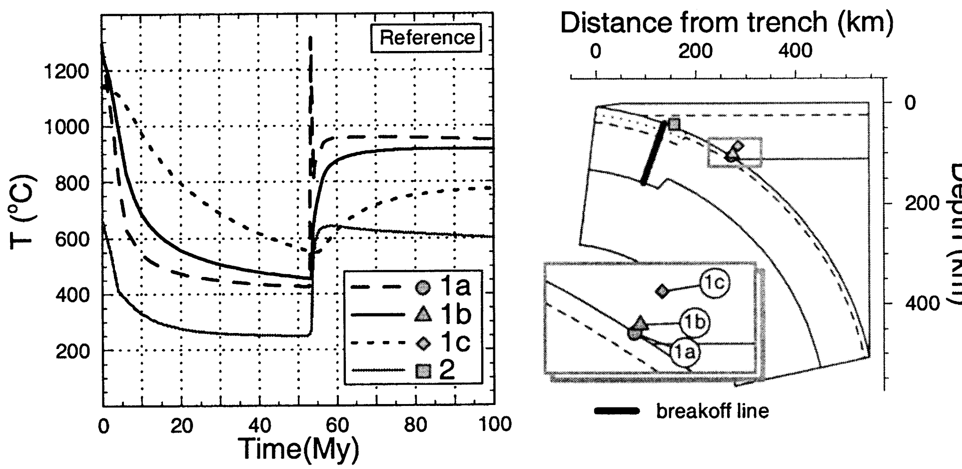


Figure 4. Temperature distribution in the overriding plate for the reference model (Table 1). Shown are temperature-time paths of four points which are situated at the locations indicated on the right. Points 1a, 1b, and 1c are positioned at different distances from the plate contact, i.e., at 0, 5, and 25 km, respectively. Point 2 is emplaced at a much shallower depth and at 5 km distance of the plate contact. Temperatures decrease in the period of oceanic lithosphere subduction. After 3 Myr of rapid continental lithosphere subduction the slab detaches at a downdip depth of 140 km, and hot mantle material fills the newly formed gap (FreeSink model as in Plate 1).

which record the temperature in time. Figure 4 shows the temperature-time paths of four points for the reference set of parameter values. The first three points (1a, 1b, and 1c) are situated near the bottom of the overriding lithosphere at depths of ~100 km, i.e., at a downdip distance from the trench of ~300 km. Point 2 is placed at a depth of 40 km, at 170 km downdip distance from the trench. The effects of subduction and slab breakoff are clearly visible in all points, though the influences at the points near the plate contact are much more pronounced, as expected. During oceanic lithosphere subduction, the temperature at point 1b (5 km from the plate contact) decreases with 806°C in 50 Myr. After slab detachment it increases with 419°C in only 7 Myr. Directly at the plate contact, i.e., point 1a, the temperature jumps to the temperature of the inflowing asthenospheric material, after which it falls to temperatures comparable with those obtained for point 1b. At point 1c, at 25 km from the plate contact, the temperature reduction is 575°C in the first 50 Myr. The temperature increase after breakoff is also much more gradual and not as large, ~110°C in 10 Myr. Hereinafter, we focus on points 1b and 2 to investigate the influence of varying parameter values on both the breakoff depth and the temperature increase after slab detachment.

3.2. Factors That Only Affect Breakoff

3.2.1. Rheology. *Wong A Ton and Wortel* [1997] showed the importance of temperature and rheology in the context of slab detachment processes. They control the breakoff time and depth to a large extent. In the present study, the breakoff depth in particular is important for the distribution of the temperature increase after detachment. If a point is situated deeper than or close to the breakoff position, it experiences a rapid temperature rise (Figure 5). Points situated at much shallower levels than the breakoff depth only experience gradual temperature increases.

Of all parameters associated with rheology, the rheologies of the upper and lower crust have the greatest effect on the strength of the slab. Various combinations of relatively weak wet quartzite [*Koch et al.,* 1980], wet diorite [*Hansen and Carter,* 1982], and strong dry diabase rheologies [*Caristan,* 1982] were used to model the rheology of the crust. The thickness of this crustal part of the descending continent is of importance since it defines the size of the region of relative weakness. If this region is smaller, e.g., 20 km, the integrated strength of the slab will be larger, leading to deeper breakoff. Other rheology-related parameters like the pore fluid pressure and the rheology of the mantle part of the lithosphere have smaller effects on breakoff.

3.2.2. Initial temperature distribution of the slab. The temperatures within the continental part of the slab are controlled by the temperature distribution just before subduction, which is parameterized by the surface heat flow at the trench. A lower value corresponds to a colder, stronger slab, resulting in a later and deeper detachment. For example, a value of 60 mW m^{-2} results in breakoff after 5.5 Myr continental subduction at a downdip depth of 330 km (Figure 5). This is later and, more important, deeper compared with the 3 Myr and 140 km computed for a surface heat flow of the slab of 80 mW m^{-2} (reference model). The thermal structure of the continental slab also affects the temperatures at the plate contact directly, but only to the depth to which the continental material has been subducted. Since detachment is likely to occur in or near the transition zone to the oceanic slab, the temperatures within the continental slab have no

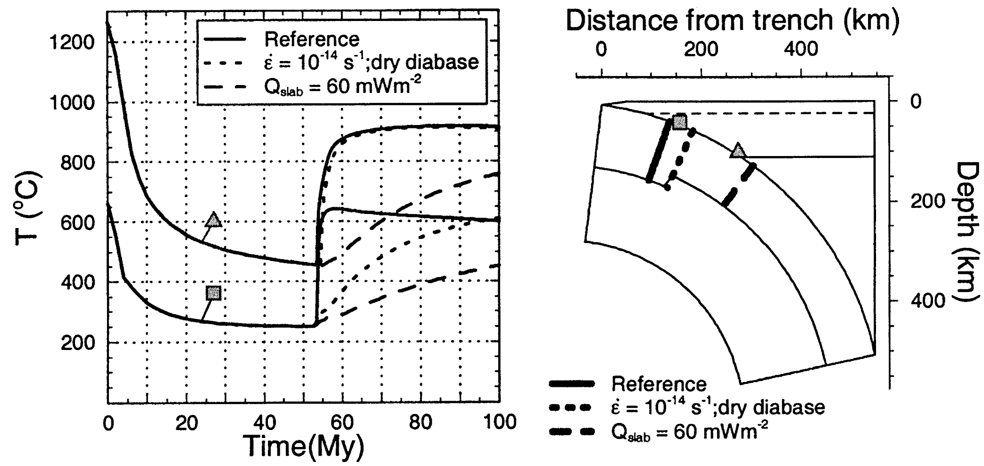


Figure 5. Influence of the rheology and the thermal structure before subduction on the breakoff depth and the resulting thermal evolution of the overriding plate. A strain rate of 10^{-14} s^{-1} and a lower crust rheology of dry diabase give slab breakoff after 4 Myr of continental subduction at a downdip depth of 190 km. A lower surface heat flow of 60 mW m^{-2} results in a later and deeper breakoff, namely, after 5.5 Myr at a breakoff depth of 330 km, which is not shallow enough to even increase the temperature at the point just above the base of the overlying plate (triangle) very quickly. Note that the bottom of the lithospheric slab drawn on the right only represents the structure at the time of detachment in case of the reference parameter values.

direct effect on thermal evolution at depths greater than the breakoff position.

3.2.3. Thickness of the subducting lithosphere. For temperatures higher than a specific critical temperature the differential stresses for ductile deformation become too low to contribute to the integrated strength of the slab. This critical temperature depends on the rheology used. For the mantle part of the slab it is ~1000°C for dry dunite as well as wet Aheim and wet Anita Bay dunite [*Chopra and Paterson,* 1984]. The depth at which the temperatures exceed the critical value is partially controlled by the total thickness of the lithosphere. A very thick continent has lower subcrustal temperatures, pushing the 1000°C isotherm to a deeper position. A greater part of the slab contributes to its strength, resulting in a stronger slab which detaches at a later point in time at a greater depth. The effect is most noticed for relatively low surface heat flows since they give a low temperature at Moho depths, leaving a larger subcrustal zone to contribute to the strength.

3.3. Factors That Affect Both Breakoff and T_0

3.3.1. Convergence velocity during continental lithosphere subduction. The thermal structure of the continental slab is also strongly controlled by the velocity at which it subducts. Note that what we discuss here is the convergence velocity during continental lithosphere subduction only. During the first 40 Myr of oceanic lithosphere subduction the convergence velocity is 6 cm yr^{-1}. In the last 10 Myr of this period, V_c is linearly decreased in time to 1 cm yr^{-1} if a low convergence velocity during continental lithosphere subduction is considered. In that case, the descending plate has more time to warm up than a faster subducting plate, which results in detachment at a shallower depth but at a much later point in time. However, the effect on the depth of detachment can be quite small, namely, 130 km (Figure 6) compared to 140 km downdip distance (Figure 4). This is due to the high initial surface heat flow of the continental slab of 80 mW m^{-2}. For colder slabs the situation is much more clear (compare Figures 5 and 6 where a Q_{slab} of 60 mW m^{-2} is used). A slab descending at a rate of 6 cm yr^{-1} breaks off at a level deeper than the bottom of the overlying lithosphere, whereas the slowly subducting plate detaches at a much shallower depth, namely, at 250 km from the trench. The effect of the convergence velocity during continental lithosphere subduction on the breakoff depth is thus largest for initially cold subducting continents.

The effect of a low convergence velocity on T_0 is twofold. Firstly, a slowly descending plate has higher steady state temperatures than a rapidly subducting slab. This means that if a long period of fast subduction is followed by slower subduction, the temperatures in the plate contact region might increase again (Figure 6). In addition to this, a slowly descending plate detaches at a later point in time, which allows the slab to be warmed up even more. In general, a lower convergence velocity during continental subduction results in a higher $T_0(r)$. The convergence velocity during oceanic lithosphere subduction, which is discussed below, has a much greater influence on the temperature just before breakoff.

3.3.2. Frictional heat. A slab subducting with a relatively low V_c or with a thick friction layer will produce very little frictional heat. Although the direct influence on T_0 and

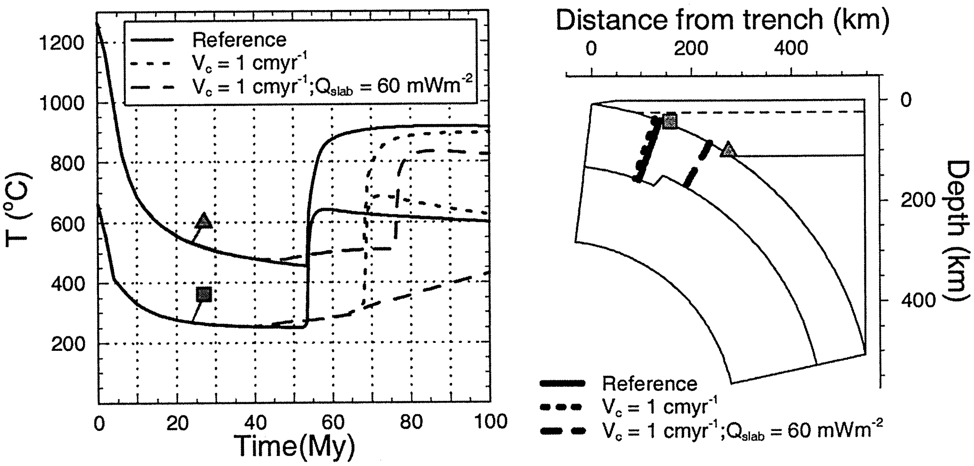


Figure 6. Effect of the convergence velocity during continental lithosphere subduction and the initial thermal structure of the continental slab on both the breakoff depth and the temperature just before breakoff ($T_0(r)$). A convergence velocity of 1 cm yr^{-1} during continental subduction gives breakoff after 17 Myr at 130 km downdip depth. Both points experience an instantaneous temperature increase, only later. A surface heat flow of 60 mW m^{-2} in combination with a V_c of 1 cm yr^{-1} results in slab detachment at 250 km depth after 25 Myr.

the subsequent temperature increase can not be disregarded, the effect on the strength of the slab is negligible.

3.3.3. Age of the oceanic lithosphere. Breakoff is often calculated to occur near the zone where the transition from oceanic to continental material takes place. The strength in this region is therefore very important and is partly controlled by the age of the oceanic lithosphere. A younger lithosphere is warmer which also weakens the continental material near the oceanic slab. The part of the transition zone just next to the oceanic lithosphere can become very weak which results in slab detachment exactly at the position where oceanic lithosphere first changes into continental lithosphere. The temperature before breakoff is affected because of the age-dependent temperature distribution of the oceanic lithosphere. However, the maximum difference in T_0 is only ~25°C. The radioactive heat generation in and the thickness of the oceanic crust are found to have negligible influences on both breakoff and T_0.

3.3.4. Mantle temperature. T_m controls the depth at which the critical temperature is reached (with respect to ductile deformation, see previous section), though its effect on the strength of the slab remains small. However, T_m does influence the temperature increase because it defines the temperature of the hot asthenospheric material which fills the gap.

3.4. Factors That Only Affect T_0

The period of oceanic lithosphere subduction prior to continental lithosphere subduction has an important influence on the magnitude of the temperature increase following breakoff because it strongly affects $T_0(r)$. For fast descending slabs the effect will not be very large since a steady state situation is reached within a few millions of years of subduction. However, for low convergence velocities the difference between a relatively long and short period of oceanic subduction can be considerable, i.e., up to a few hundred degrees C. The interplay between the duration of oceanic lithosphere subduction and the convergence velocity during this period thus controls T_0 to a great extent (Figure 7a). The initial thermal distribution in the overriding continent, parameterized by the surface heat flow and the thickness of the lithosphere, can also affect T_0, but only if the period of subduction prior to detachment is very short. If the period of subduction is sufficiently long for the temperatures to reach a steady state situation, the effect is negligible.

3.5. Breaking Mechanism and Sinking Velocity

3.5.1. FreeSink model. In the FreeSink model, the temperatures at the whole plate contact start to rise after slab breakoff (Figures 3a and 4). The effect of a lower sinking velocity $V_{s//}$ on the temperature increase is not very large, except for points situated near the bottom of the overlying lithosphere. These points experience a larger T jump after breakoff because of the ongoing flow of warm mantle material below the overriding plate, while the slab sinks into the mantle (Figure 3a). The sinking velocity in the direction perpendicular to the slab top also has a modest effect on the magnitude of the temperature increase. A higher $V_{s\perp}$ results in a somewhat larger ΔT which is reached within a shorter period.

In the case of the FreeSink model with a zero $V_{s\perp}$, the temperature increase strongly depends on $V_{s//}$. An instantaneous thermal impulse is only present exactly above the gap. So, there is no immediate effect on the whole plate contact if the sinking velocity is relatively low. A point situated much deeper than the breakoff depth will therefore ex-

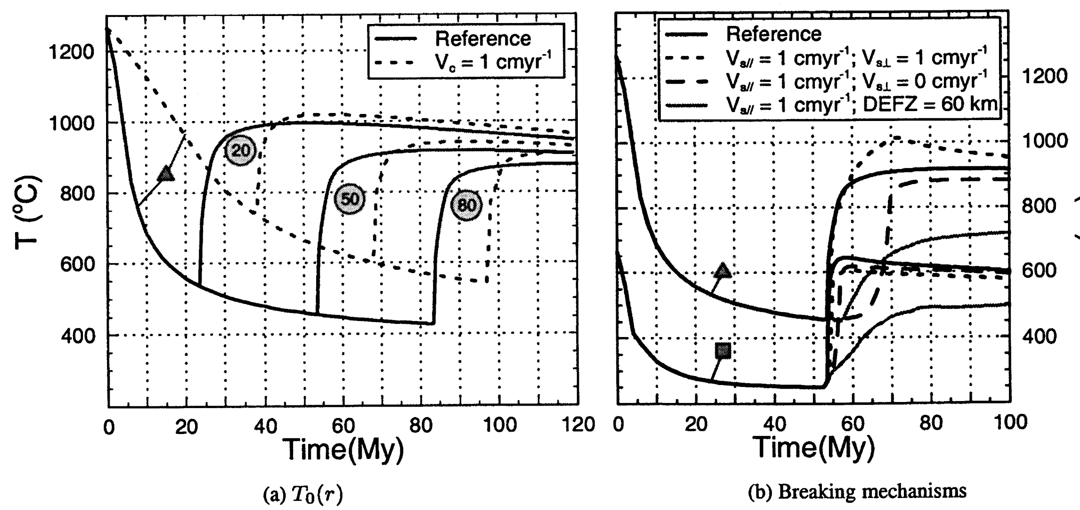


Figure 7. (a) Influence of $T_0(r)$ on the temperature increase after slab detachment, shown by varying the period of oceanic lithosphere subduction prior to continental lithosphere subduction (the circled numbers denote this period in Myr). Temperature-time paths of point 1b (as in Figure 4) are shown for constant convergence velocities of 1 and 6 cm yr^{-1} (reference model). For a V_c of 1 cm yr^{-1}, ΔT (in 6 Myr) ranges from 249°-339°C. The temperature increase in 6 Myr is more than 400°C for a V_c of 6 cm yr^{-1}. (b) Effect of the breaking mechanism on the temperature jump after breakoff. Tt paths of points 1b and 2 (see Figure 4) show the influence of the sinking velocities and the breaking mechanism on ΔT.

perience ongoing cooling until the entire slab sinks below this point (Figure 7b). The temperature increase then highly depends on the temperature just before the head of the slab passes (instead of depending on T_0).

3.5.2. Stretching model. In the Stretching model, heating of the overriding plate after slab detachment strongly depends on the width of the deformation zone and the sinking velocity. If the zone is very small and the velocity relatively high, the model resembles the zero-$V_{s\perp}$ FreeSink model. A combination of a wide stretching area and a low $V_{s//}$ will produce smaller and more gradual temperature increases in the entire region below the breakoff position (Figure 7b).

3.6. Summary of Modeling Results

The breakoff depth is controlled by the thermal structure of the continental part of the slab. A minimum breakoff depth of 130 km downdip distance is obtained for an initially warm continental slab subducting at a low rate of 1 cm yr^{-1}. This corresponds to 35 km depth in the standard vertical sense. Note that the breakoff depth reduces if resistive forces at the trench are included. Higher values of V_c, stronger rheologies and/or lower values of Q_{slab} result in breakoff at deeper levels (Figure 8). If detachment occurs deeper than the bottom of the overriding lithosphere, i.e. for continental lithosphere with a $Q_{\text{slab}} \leq 60$ mW m^{-2}, the effect of the inflow of hot asthenospheric material on the thermal structure of the overlying lithosphere is negligible. The breakoff times associated with shallow slab detachment vary from 2 to 25 Myr, measured from the onset of the subduction of continental lithosphere.

4. Analysis and Discussion

4.1. Modeling Aspects

By evaluating the effects of various parameters on temperatures at points near the plate contact an upper bound for the temperature increase following shallow slab detachment was obtained. If a monitoring point were to be placed directly at the plate contact (see Figure 4), the temperature increase would be even higher, i.e., it would rise to the temperature of the inflowing asthenospheric material. Subsequently, the temperature would drop to values comparable to those obtained for points situated at a few kilometers from the plate contact (such as point 1b). Points situated much further away will not experience such a pronounced temperature jump. The same holds for points at shallower depths than the breakoff depth (Figure 5). Note that a distinction must be made between heating due to the ceasing subduction, i.e., less cooling, and actual heating by conduction. The presence of the relatively cold, remaining part of the slab counteracts this heating to a certain extent. In our modeling, we assume that convergence stops at the time of breakoff. This is a reasonable assumption in view of the vanishing slab pull force F_{sp}. However, if underthrusting of cold material continues after slab breakoff, the timespan in which the heat pulse is active in the overriding plate is reduced to only a few millions of years (depending on the subduction velocity after breakoff).

Return flow of warm mantle material into the wedge is not incorporated in the model. This corner flow would counteract cooling of the area in the mantle wedge, and the bottom

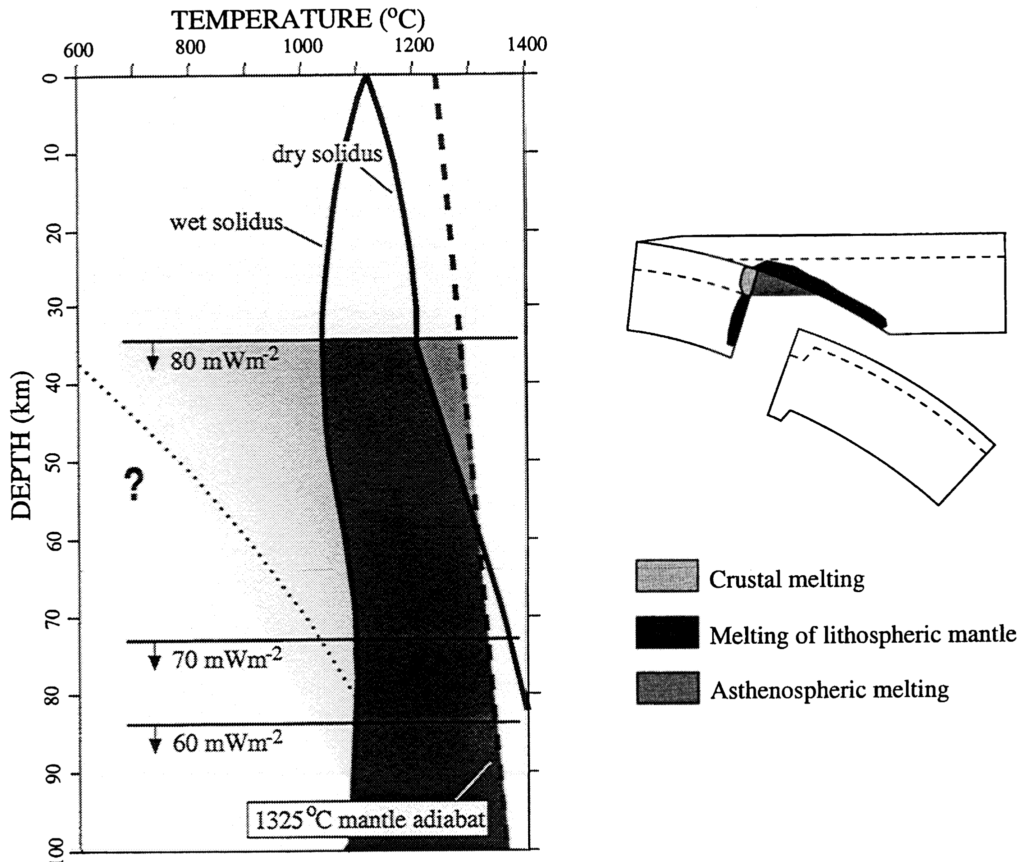


Figure 8. Summary of breakoff modeling results in the context of melt generation. The temperature-depth diagram displays a normal geotherm with a surface heat flow of 60 mW m^{-2} (dotted), the 1325°C mantle adiabat (dashed), and both wet and dry melting curves of lherzolite (solid). The horizontal lines indicate the minimum breakoff depth for a value of Q_{slab} with a convergence velocity during continental lithosphere subduction of 1 cm yr^{-1} and a weak rheology. Higher values of V_c or stronger rheologies, or both, give deeper breakoff (denoted by the arrows). Shading corresponds to the types of melt sources indicated on the right.

of the overriding plate and the upper part of the slab, during (fast) subduction [*Kincaid and Sacks,* 1997]. The temperature before breakoff of points located near the bottom of the overriding continental lithosphere can therefore be higher, resulting in a smaller temperature increase. However, the effects of return flow in the mantle wedge on the strength of and the forces acting on the slab are negligible since the areas influenced by corner flow are of relative unimportance to shallow slab detachment. Shallow breakoff is mostly concerned with the strength of the continental material which is too far located from the wedge to become affected. Model runs in which we prevented the temperatures in the mantle wedge from dropping below 1200°C confirm that cornerflow has no influence on the breakoff times and depths.

Figure 7b clearly demonstrates the ability of the two sinking mechanisms to produce considerable temperature increases after breakoff. However, are these mechanisms physically feasible? It is obvious that the slab will not detach exactly as presented in the FreeSink model, i.e., with a sharp, instantaneous cut. The immediate flow of hot asthenospheric material to the former plate contact is also a gross simplification, though the inflow of asthenosphere is likely since it is the least viscous material near the deforming zone [*Davies and von Blanckenburg,* 1998]. The Stretching model resembles the mechanism of narrow rifting as suggested by *Davies and von Blanckenburg* [1995].

4.2. Comparison With Other Mechanisms

The slab breakoff model shows a number of similarities with models involving convective removal [*Platt and England,* 1994; *Turner et al.,* 1999] or delamination [*Bird,* 1979]. Hot asthenospheric material can rise up to the base of the crust, causing uplift, magmatism, and metamorphism. However, the differences between the various mechanisms give rise to, for example, different spatial distributions of the vertical motions or states of stress. Also, a subduction zone setting, as in the case of slab detachment, provides a number of possibilities to exhume metamorphic rocks from greater depths. Another model which is associated with anomalously high subcrustal temperatures beneath orogens is viscous heating whereby the subcontinental lithosphere is not removed and replaced [*Kincaid and Silver,* 1996].

Jamieson et al. [1998] recognized the problem of producing high peak metamorphic temperatures at crustal depths. They used a thermomechanical model to investigate the thermal state of convergent orogens and concluded that the temperatures are dominated by the interplay between cooling due to subduction and radioactive heating. Instead of using an average value for the crustal heat production, they introduced a heat-producing wedge, referred to as TARM (tectonically accreted radioactive material), to counteract the cooling by subduction.

We inserted a similar heat-producing wedge in the overriding plate whereby we used a depth-independent distribution of the heat production to assess the effect on the entire upper 50 km (Figure 9b). The radioactive wedge is instantaneously emplaced as soon as modeling starts. These simplifications clearly overestimate the effect of the TARM on the temperatures since the accretion of radioactive material is actually a rather time-consuming process. During oceanic lithosphere subduction the overriding plate experiences less cooling. The same holds for the subduction of continental material which starts after 50 Myr of oceanic lithosphere subduction. When continental material first passes a monitoring point, the temperatures in that point increase because of the higher heat production within the subducting continental crust. TARM model temperatures are relatively high for a subduction zone. However, these high temperatures are only representative for the TARM wedge itself (Figure 10a).

Another TARM model represents the situation in which subduction is assumed to stop at 67 Myr (the time at which subduction ceases in the case of slab breakoff with the same set of parameter values). Temperatures start to rise immediately but in a much more gradual fashion than in the case of slab detachment. The total increase in temperature after a specific timespan depends on the amount of heat-producing material present and on the overall thermal structure of the continental part of the slab (compare Figures 10b and 10c).

In conclusion, the TARM model is capable of producing higher temperatures in a continental collision zone. However, fast in situ temperature increases are not obtained. Note that a distinction must be made between in situ temperature variations and temperature changes as experienced by an exhuming rock volume. The latter could experience a temperature increase by moving through the locally raised temperature in the TARM wedge. The rate of the ΔT is then controlled by the exhumation velocity. The implications of the accretion of radioactive material for, for example, crustal anatexis is discussed in section 4.3.

4.3. Implications for Melt Generation and Exhumation

Figure 8 shows a summary of the slab breakoff modeling results in the context of melt generation. The possible occurrence of various types of melting is indicated: asthenospheric melting, partial melting of the mantle part of the lithosphere, and crustal melting. *Davies and von Blanckenburg* [1995] considered melting of the upwelling asthenosphere itself doubtful since breakoff at depths $\leq$ 50 km ($\leq$ 170 km downdip distance from the trench in our model-

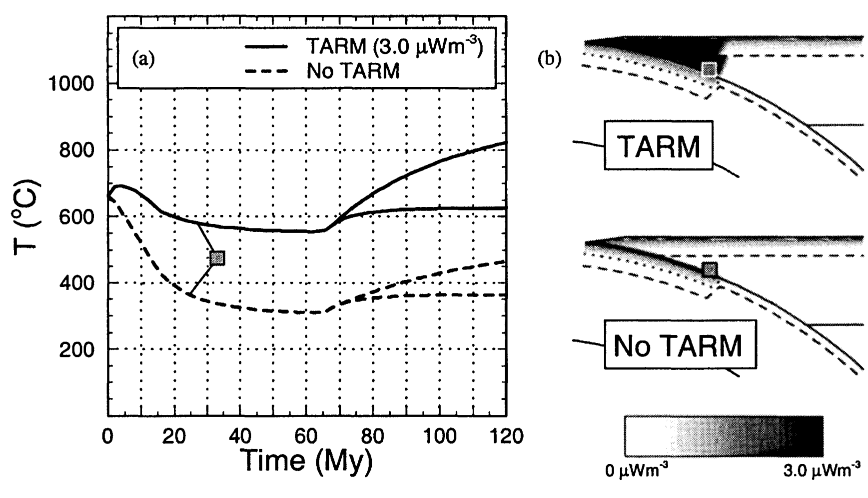


Figure 9. Effect of TARM on temperatures in the overriding plate. All parameters are as stated in Table 1, except for the convergence velocity which is 1 cm yr^{-1}. (a) Temperatures remain higher during subduction because of the heat-producing wedge. Subduction either continues or is assumed to stop at 67 Myr (the time at which subduction ceases in the case of slab breakoff with the same set of parameter values). In both cases the temperatures increase after that time because of the presence of newly subducted warm continental crust which generates more radiogenic heat (Figure 9b). They continue to increase if subduction is assumed to stop. (b) Distribution of heat producing material at 67 Myr for models with and without TARM.

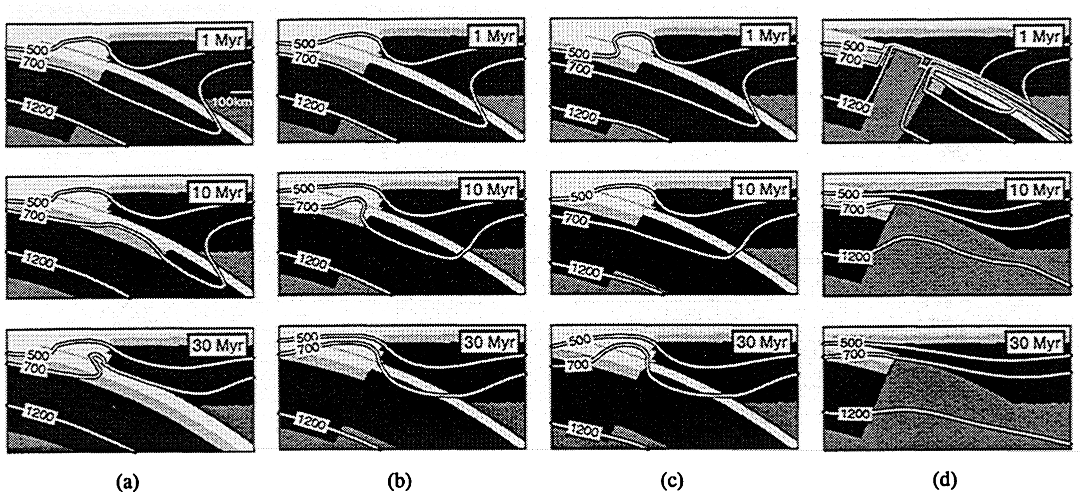


Figure 10. Thermal evolution of the subduction zone in case of a tectonically accreted radioactive wedge (TARM) and in case of slab detachment. Shown are isotherms of 500°, 700°, and 1200°C at 1, 10, and 30 Myr after 67 Myr, the time at which slab breakoff occurs for slow subduction of initially warm continental material. Values of V_c and Q_{slab} are 1 cm yr^{-1} and 80 mW m^{-2}, respectively, except for Figure 10c in which Q_{slab} is 60 mW m^{-2}. Shading indicates (upper/lower) crust and lithospheric mantle and asthenosphere, respectively. Note the scale in Figure 10a, top. (a) TARM model with continuing subduction. Temperatures are only locally raised in the tectonically accreted radioactive material. (b) TARM model with subduction assumed to stop at 67 Myr. Temperatures increase gradually in and near the TARM wedge. (c) TARM model with subduction of an initially colder (Q_{slab}=60 mW m^{-2}) continental slab, stopping at 67 Myr. (d) Slab detachment at 17 Myr after the onset of continental lithosphere subduction (see also Figure 6). Isotherms are elevated along the entire plate contact.

ing) is required. Our results, however, show that breakoff at such shallow depths is possible for warm, weak slabs (Q_{slab} $\geq$ 80 mW m^{-2}). At those depths, partial melting of asthenospheric material can be associated with temperatures of $\sim$1200°C (Figure 8). Figure 10d clearly shows that asthenospheric melting associated with shallow slab detachment has a distinctive transient character.

This transient behavior also holds for partial melting of the mantle part of the continental lithosphere which is also associated with high temperatures [*McKenzie and Bickle,* 1988]. However, the overriding continent is metasomatized during subduction whereby the presence of fluids lowers the solidus considerably. This can lead to partial melting of the mantle part of the enriched overriding lithosphere at temperatures of 1000° to 1100°C (Figure 8). Such temperatures are restricted to an area very close to the plate contact or to deeply situated rocks (near the base of the overriding lithosphere) and are bound to timescales of only a few millions of years (Figure 10d).

A precise solidus for crustal material is difficult to construct because of the (petrological) complexity of the crust. A reasonable lower bound for partial melting of crustal material is 700°C [e.g., *Wyllie,* 1977; *England and Thompson,* 1986] (Figure 8). The calculated 700°C isotherms after shallow slab detachment are plotted in Figure 10d. Anatexis of the overriding crust directly after breakoff is possible, especially when a thickened crust is considered. The crustal parts of the remaining upper part of the slab might be an even larger source for melt generation. Alternatively, migration of lithospheric melts into the overlying crust can also cause local magmatism.

In the case of a TARM wedge (Figures 10a-10c), temperatures associated with partial melting of both the ascending asthenosphere and dry lithospheric mantle remain restricted to greater depths. This holds for the situation in which subduction continues as well as for ceasing subduction at 67 Myr. In the latter case, it is possible for crustal anatexis to occur (Figure 10b), just as it is with slab detachment. However, the timescales related with partial melting of the crust in the TARM model are considerably different from those related to slab breakoff. The 700°C isotherm only starts to impinge on crustal material at 10 Myr after the stop of subduction. Thereafter, this isotherm is locally raised at the position of the TARM wedge, and a new steady state situation will set in. Note that a relatively high value of heat production is used, i.e., 3.0 μW m^{-3}. A TARM wedge containing less heat generating material does not produce any crustal anatexis.

The origin of partial melts can be determined from major and trace elements and isotopic geochemical data. The result can be used not only to discriminate between tectonic processes [*Wilson and Bianchini,* 1999], but also to give a clue on the possible depth at which a slab detached [*von Blanckenburg and Davies,* 1995]. It is important to recognize that the anatexis of the lithospheric mantle and the upwelling asthenosphere are transient processes, i.e., only for timespans in the order of a few millions of years. As a result, this distinctive transient behavior is expected to be present in the episodes of volcanism, for example, associated with the lateral migration of slab detachment beneath Italy [*van der Meulen et al.*, 1998]. Note, however, that the timespans of these temperature-related processes can be longer if convective processes are also considered. In summary, melt generation due to the ascent of hot asthenosphere after slab detachment is possible, which is important for mineralization [*de Boorder et al.*, 1998] and volcanism [*Wortel and Spakman*, 1992].

The model does not encompass vertical motions in a sense that it can simulate the exhumation of the magmatic and/or high-temperature metamorphic rocks. The complicated problem of getting these rocks to the surface is subject to much debate. Recent analog modeling of continental subduction [e.g., *Chemenda et al.*, 1996, 2000] shows the possibility to emplace rocks from the subducting continental crust into the overriding crust. In the process of continental subduction, slivers of crustal material can become detached from the downgoing lithosphere and pile up as thrust sheets in front of the subducting plate [*van den Beukel*, 1992]. Next, they cool off during ongoing subduction. If the continental slab then detaches at a shallow level, the subsequent heat pulse due to the upwelling hot asthenosphere can heat these scraps of continental crust considerably. Next, the plate contact can act as a channel in the process of exhumation.

The presence of these slivers in the overriding plate does not necessarily affect the overall geometry of the collisional zone [*van den Beukel and Wortel*, 1992]. It can still be represented as one plate subducting beneath the other. However, since the thrust sheets are relatively cold, the actual temperatures near the plate contact could be lower than our modeling indicates. This would give a lower temperature just before breakoff and, subsequently, result in a higher temperature increase. In this light, it means that our calculated temperature jumps could be an underestimation of the actual temperature increase.

5. Conclusions

From our 2-D time-dependent modeling we conclude that shallow slab detachment provides the opportunity for asthenospheric material to rise to depths as shallow as 35 km. Our results indicate under which conditions this occurs: In general, the arrival of relatively warm continental lithosphere at a trench system, and subduction at a low rate, favors shallow slab detachment. The associated breakoff times, measured from the onset of continental lithosphere subduction, range from 2 to 25 Myr after the onset of continental subduction.

The rise of asthenospheric material constitutes a first-order (advective) heat source which may account for temperature increases leading to postcollisional melting of both mantle and crustal material and subsequent magmatism and mineralization. In view of the transient nature of the thermal anomaly the resulting postcollisional magmatic (and mineralization) activity is predicted to be transient, as well. Provided slab detachment is a viable process in the region studied, it should be taken into account in the analysis of the thermal evolution of exhumed rocks, in particular those involving a temperature increase experienced during exhumation.

The quantification of the conditions required for shallow slab detachment will contribute to warranted assessments concerning the role of slab detachment (relative to other proposed heat sources, such as TARM) in the geodynamical evolution of former convergent plate boundaries.

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References

Atwater, T., Implications of plate tectonics for the Cenozoic evolution of western North America, *Geol. Soc. Am. Bull., 81*, 3513-3535, 1970.

Bird, P., Continental delamination and the Colorado Plateau, *J. Geophys. Res., 84*, 7561-7571, 1979.

Blanco, M.J., and W. Spakman, The *P*-wave velocity structure of the mantle below the Iberian Peninsula: Evidence for subducted lithosphere below southern Spain, *Tectonophysics, 221*, 13-34, 1993.

Brace, W.F., and D.L. Kohlstedt, Limits on lithospheric stress imposed by laboratory experiments, *J. Geophys. Res., 85*, 6248-6252, 1980.

Brouwer, F.M., R.L.M. Vissers, M.J.R. Wortel, and W.M. Lamb, Testing geodynamical models for exhumation of high-pressure metamorphic rocks in the Alps, paper presented at Metamorphic Studies Group Meeting, University of Rennes, France, Aug. 31 to Sept. 2, 1999.

Caristan, Y., The transition from high temperature creep to fracture in Maryland diabase, *J. Geophys. Res., 87*, 6781-6790, 1982.

Carminati, E., M.J.R. Wortel, W. Spakman, and R. Sabadini, The role of slab detachment processes in the opening of the western-central Mediterranean basins: Some geological and geophysical evidence, *Earth Planet. Sci. Lett., 160*, 651-665, 1998a.

Carminati, E., M.J.R. Wortel, P.Th. Meijer, and R. Sabadini, The two-stage opening of the western-central Mediterranean basins: A forward modeling test to a new evolutionary model, *Earth Planet. Sci. Lett., 160*, 667-679, 1998b.

Chapman, D.S , Thermal gradients in the continental crust, in *The Nature of the Lower Continental Crust*, edited by J.B. Dawson et al., *Geol. Soc. Spec. Publ., 24*, 63-70, 1986.

Chemenda, A.I., M. Mattauer, and A.N. Bokun, Continental subduction and a mechanism for exhumation of high-pressure metamorphic rocks: New modelling and field data from Oman, *Earth Planet. Sci. Lett., 143*, 173-182, 1996.

Chemenda, A.I., J.P. Burg, and M. Mattauer, Evolutionary model of the Himalaya-Tibet system: Geopoem based on new modelling, geological and geophysical data, *Earth Planet. Sci. Lett., 174*, 397-409, 2000.

Chopra, P.N., and M.S. Paterson, The role of water in the deformation of dunite, *J. Geophys. Res., 89*, 7861-7876, 1984.

Crough, S.T., Thermal model of oceanic lithosphere, *Nature, 256*, 388-390, 1975.

Davies, G.F., Mechanics of subducted lithosphere, *J. Geophys. Res., 85,* 6304-6318, 1980.

Davies, J.H., and F. von Blanckenburg, Slab breakoff: A model of lithosphere detachment and its test in the magmatism and deformation of collisional orogens, *Earth Planet. Sci. Lett., 129,* 85-102, 1995.

Davies, J.H., and F. von Blanckenburg, Thermal controls on slab breakoff and the rise of high-pressure rocks during continental collisions, in *When Continents Collide: Geodynamics and Geochemistry of Ultrahigh-Pressure Rocks,* edited by B.R. Hacker and J.G. Liou, pp.95-115, *Kluwer Acad.,* Norwell, Mass., 1998.

de Boorder, H., W. Spakman, S.H. White, and M.J.R. Wortel, Late Cenozoic mineralization, orogenic collapse and slab detachment in the European Alpine Belt, *Earth Planet. Sci. Lett., 164,* 569-575, 1998.

England, P.C., and A. Thompson, Some thermal and tectonic models for crustal melting in continental collision zones, in *Collision Tectonics,* edited by M.P. Coward and A.C. Ries, *Geol. Soc. Spec. Publ., 19,* 83-94, 1986.

Forsyth, D., and S. Uyeda, On the relative importance of the driving forces of plate motion, *Geophys. J. R. Astron. Soc., 43,* 163-200, 1975.

Furlong, K.P., W.D. Hugo, and G. Zandt, Geometry and evolution of the San Andreas fault zone in Northern California, *J. Geophys. Res., 94,* 3100-3110, 1989.

Goetze, C., The mechanisms of creep in olivine, *Philos. Trans. R. Soc. London, Ser. A., 288,* 99-119, 1978.

Haeussler, P.J., D. Bradley, L. Snee, and C. Taylor, Link between ridge subduction and gold mineralization in southern Alaska, *Geology 23,* 995-998, 1995.

Hansen, F.D., and N.L. Carter, Creep of selected crustal rocks at 1000 MPa (abstract), *Eos Trans. AGU, 63,* 437, 1982.

Jackson, I., and S.M. Rigden Composition and temperature of the Earth's mantle: Seismological models interpreted through experimental studies of Earth materials, in *The Earth's Mantle: Composition, Structure and Evolution,* edited by I. Jackson, pp.405-460, *Cambridge Univ. Press,* New York, 1998.

Jamieson, R.A., C. Beaumont, P. Fullsack, and B. Lee, Barrovian regional metamorphism: Where's the heat?, in *What Drives Metamorphism and Metamorphic Reactions?,* edited by P.J. Treloar and P.J. O'Brien, *Geol. Soc. Spec. Publ., 138,* 23-51, 1998.

Kincaid, C., and I.S. Sacks, Thermal and dynamical evolution of the upper mantle in subduction zones, *J. Geophys. Res., 102,* 12295-12315, 1997.

Kincaid, C., and P. Silver, The role of viscous dissipation in the orogenic process, *Earth Planet. Sci. Lett., 142,* 271-288, 1996.

Koch, P.S., J.M. Christie, and R.P. George, Flow law of "wet" quartzite in the α-quartz field (abstract), *Eos Trans. AGU, 61,* 376, 1980.

Kohlstedt, D.L., B. Evans, and S.J. Mackwell, Strength of the lithosphere: Constraints imposed by laboratory experiments, *J. Geophys. Res., 100,* 17587-17602, 1995.

McKenzie, D., Some remarks on the development of sedimentary basins, *Earth Planet. Sci. Lett., 40,* 25-32, 1978.

McKenzie, D.P., and M.J. Bickle, The volume of composition of melt generated by extension of the lithosphere, *J. Petrol., 29,* 625-679, 1988.

Meijer, P.Th., and M.J.R. Wortel, Temporal variation in the stress field of the Aegean region, *Geophys. Res. Lett., 23,* 439-442, 1996.

Platt, J.P., and P.C. England, Convective removal of lithosphere beneath mountain belts: Thermal and mechanical consequences, *Am. J. Sci., 294,* 307-336, 1994.

Platt, J.P., and R.L.M. Vissers, Extensional collapse of thickened continental lithosphere: A working hypothesis for the Alboran Sea and Gibraltar arc, *Geology, 17,* 540-543, 1989.

Platt, J.P., J.-I. Soto, M.J. Whitehouse, A.J. Hurford, and S.P. Kelley, Thermal evolution, rate of exhumation, and tectonic significance of metamorphic rocks from the floor of the Alboran extensional basin, western Mediterranean, *Tectonics, 17,* 671-689, 1998.

Pollack, H.N., and D.S. Chapman, On the regional variation of heat flow, geotherms and lithosphere thickness, *Tectonophysics, 38,* 279-296, 1977.

Schatz, J.F., and G. Simmons, Thermal conductivity of Earth materials at high temperatures, *J. Geophys. Res., 77,* 6966-6983, 1972.

Spakman, W., M.J.R. Wortel, and N.J. Vlaar, The Hellenic subduction zone: A tomographic image and its geodynamic implications, *Geophys. Res. Lett., 15,* 60-63, 1988.

Spakman, W., S. Van der Lee, and R.D. Van der Hilst, Travel-time tomography of the European-Mediterranean mantle down to 1400 km, in Recent Advances in Geosciences, pp.3-74, *Elsevier Sci.,* New York, 1993.

Turner, S.P., J.P. Platt, R.M.M. George, S.P. Kelley, D.G. Pearson, and G.M. Nowell, Magmatism associated with orogenic collapse of the Betic-Alboran Domain, SE Spain, *J. Petrol., 40,* 1011-1036, 1999.

van den Beukel, J., Some thermomechanical aspects of the subduction of continental lithosphere, *Tectonics, 11,* 316-329, 1992.

van den Beukel, J., and R. Wortel, Thermomechanical modelling of arc-trench regions, *Tectonophysics, 154,* 177-193, 1988.

van den Beukel, J., and R. Wortel, Ridge-trench interaction: A possible mechanism for ophiolite emplacement, *Ofioliti, 17,* 141-154, 1992.

van der Meulen, M.J., J.E. Meulenkamp, and M.J.R. Wortel, Lateral shifts of Appenninic foredeep depocentres reflecting detachment of subducted lithosphere, *Earth Planet. Sci. Lett., 154,* 203-219, 1998.

van der Meulen, M.J., T.J. Kouwenhoven, G.J. van der Zwaan, J.E. Meulenkamp, and M.J.R. Wortel, Late Miocene rebound in the Romagnan Apennines and detachment of subducted lithosphere, *Tectonophysics, 315,* 319-335, 1999.

van der Meulen, M.J., S.J.H. Buiter, J.E. Meulenkamp, and M.J.R. Wortel, An Early Pliocene uplift of the central Apenninic foredeep, and its geodynamic significance, *Tectonics, 19,* 300-313, 2000.

van der Wal, D., and R.L.M. Vissers, Uplift and emplacement of upper mantle rocks in the western Mediterranean, *Geology, 21,* 1119-1122, 1993.

Vissers, R.L.M., J.P. Platt, and D. van der Wal, Late orogenic extension of the Betic Corderilla and the Alboran Domain: A lithospheric view, *Tectonics, 14,* 786-803, 1995.

Vlaar, N.J., Thermal anomalies and magmatism due to lithospheric doubling and shifting, *Earth Planet. Sci. Lett., 65,* 322-330, 1983.

von Blanckenburg, F., and J.H. Davies, Slab breakoff: A model for syncollisional magmatism and tectonics in the Alps, *Tectonics, 14,* 120-131, 1995.

von Blanckenburg, F., H. Kagami, A. Deutsch, M. Wiedenbeck, F. Oberli, M. Meier, S. Barth, and H. Fischer, The origin of Alpine plutons along the Periadriatic Lineament, *Schweiz. Mineral. Petrogr. Mitt., 78,* 57-68, 1998.

Wilson, M. and G. Bianchini, Tertiary-Quaternary magmatism within the Mediterranean and surrounding regions, in *The Mediterranean Basin: Tertiary Extension Within the Alpine Orogen,* edited by B. Durand et al., *Geol. Soc. Spec. Publ., 156,* 141-168, 1999.

Wong A Ton, S.Y.M., and M.J.R. Wortel, Slab detachment in continental collision zones: An analysis of controlling parameters, *Geophys. Res. Lett., 24,* 2095-2098, 1997.

Wortel, M.J.R., and W. Spakman, Structure and dynamics of subducted lithosphere in the Mediterranean region, *Proc. K. Ned. Akad. Wet., 95,* 325-347, 1992.

Wortel, M.J.R., and W. Spakman, Subduction and slab detachment in the Mediterranean-Carpathian region, *Science, 290,* 1910-1917, 2000. (Correction, *Science 291,* 437, 2001.)

Wyllie, P.J., Crustal anatexis: an experimental view, *Tectonophysics, 43,* 41-71, 1977.

Zeck, H.P., Betic-Rif orogeny: Subduction of Mesozoic Tethys lithosphere under eastward drifting Iberia, slab detachment shortly before 22 Ma, and subsequent uplift and extensional tectonics, *Tectonophysics, 254,* 1-16, 1996.

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