# Subduction with Variations in Slab Buoyancy: Models and Application to the Banda and Apennine Systems

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**Abstract** Temporal variations in the buoyancy of subducting lithosphere exert a first-order control on subduction rate, slab dip and the position of the associated volcanic arc. We use a semi-analytic, three-dimensional subduction model to simulate "unforced" subduction, in which trench motion is driven solely by slab buoyancy. Model rates of subduction and model slab dip respond almost immediately to changes in the buoyancy of the subducting lithosphere entering the trench; as more buoyant slab segments correlate with slower subduction rates and steeper slab dip. The results are largely consistent with observations from the Banda and southern Apennine subduction systems, where subduction slowed and ended shortly after the entry of continental lithosphere into the trench. Over a 2 m.y. period, model subduction rates decrease from ~70 mm/year to ~30 mm/year for the Banda Arc, and from ~40 mm/year to ~20 mm/year for the Apennine Arc. Increases in model slab dip and decreases in arc-trench distance are likewise consistent with hypocenter locations and volcanic arc position along the Banda and Sunda arcs. In contrast, a time period of ~10 m.y. is needed for model subduction rates to slow to near zero, much longer than the ~3 m.y. upper bound on the observed slowing and cessation of trench motion in the Apennine and Banda systems. One possible explanation is that slab break-off, or the formation of large slab windows, occurred during the last stages of subduction, eliminating toroidal flow around the slab and allowing the slab to steepen rapidly into its final position.

## 1 Introduction

Slab buoyancy provides the primary driving force for subduction (e.g., Forsyth and Uyeda, 1975; Chapple and Tullis 1977; Conrad and Lithgow-Bertelloni, 2004). However, several studies have revealed a poor correlation between observed slab buoyancy and subduction kinematics (Jarrard, 1986; Doglioni et al., 1999; Lallemand et al., 2005). This may reflect the importance of other factors that affect subduction rate and has led to a number of studies that explore a wide range of factors that might affect subduction (e.g., Kincaid and Olson, 1987; Funiciello et al., 2003; Schellart, 2004, Bellahsen et al., 2005; Royden and Husson, 2006; Capitanio et al., 2007; Billen and Hirth, 2007, Stegman et al., 2006). Alternatively, it may be that temporal variations in slab buoyancy exert a short-term effect on subduction that is as important as the mean buoyancy.

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Temporal variations in slab buoyancy occur commonly during subduction, for example when oceanic plateaus, seamounts or fragments of continental crust enter a subduction system. Thus by examining the relationship between slab density, slab geometry and subduction rate from theoretical models, and comparing these results to the time-history of subduction and magmatism in natural systems, we can begin to understand how variations in slab density may control a host of geological phenomenon. Yet, despite the extensive literature on subduction process, only a handful of studies have examined the role of variable slab buoyancy in the subduction process. These have generally aimed at examining the role of slab buoyancy in controlling slab shape (i.e., van Hunen et al., 2004) and arc curvature (i.e., Morra et al., 2006). In contrast, Martinod et al. (2005) use an analog approach to show that buoyant slab material entering the subduction system both steepens the slab angle and reduces the velocity of the trench. In this paper we investigate more systematically the length and timescales over which variations in slab buoyancy control on subduction velocity, slab geometry and subduction dynamics, with a brief comparison to two natural subduction systems.

## 2 Subduction Model

We use a semi-analytical model updated from Royden and Husson (2006) that includes a Newtonian viscous mantle and a slab of finite width, as measured parallel to the trench. Figure 1 shows the geometry and primary sources of stress on the slab. This model uses approximate solutions for toroidal flow around the slab and for flow above and beneath the slab. This provide a reasonable approximation of stress on the slab at distances of more than ~100 km from its side boundaries (Royden and Husson, 2006). We have neglected possible curvature of the trench and slab along strike, which also affects stress on the slab near the side boundaries.

A uniform viscosity is assigned to the slab, making up a viscously competent zone that is 50 km thick and localized within the upper (cold) portion of the slab (Table 1). Subduction is assumed to be "unforced," in that the rate of subduction and trench migration are determined only by the negative buoyancy of the subducted lithosphere and viscous stresses induced in the slab and surrounding mantle. In the model, far-field motions of the plates do not contribute to the subduction process nor play a role in trench migration. Unless otherwise specified, the foreland (unsubducted lithosphere in front of the trench) is assumed to be stationary relative to the underlying lower mantle; therefore the rate of trench motion is equal to the rate of subduction (see later portions of this paper and Royden and Husson, 2006, for alternative assumptions). This is equivalent to the analog models that fix the end of the slab to the sidewall of the tank (e.g., Kincaid and Olson, 1987).

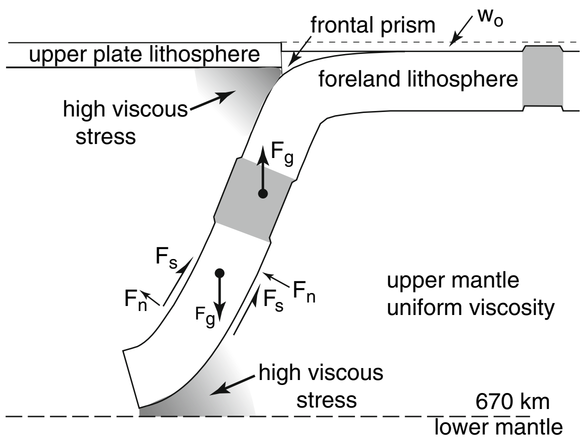



**Fig. 1** Schematic diagram of the model subduction zone. *Shaded areas* within the slab denote low-density (high buoyancy) slab segments. $F_g$, $F_n$ and $F_s$ are gravitational force, and normal and tangential components of viscous forces, respectively. $w_0$ is pre-subduction water depth of the foreland. *Shaded areas* in the upper mantle outside of the slab denote areas of high viscous stress on the slab due to the flux of mantle in or out of this constricted region

**Table 1** Model parameters for Figs. 2 and 3

| | | |
|---|---|---|
| $w_a$ | Mid-ocean ridge depth | 2.5 km |
| $l$ | Thickness of slab | 100 km |
| | Trench length | 600 km |
| | Thickness of upper plate | 50 km |
| $\rho_a$ | Density of the asthenosphere | 3,300 kg m$^{-3}$ |
| $\rho_c$ | Density of the crust/frontal prism | 2,800 kg m$^{-3}$ |
| $\rho_w$ | Density of water | 1,000 kg m$^{-3}$ |
| | Viscosity of the upper mantle (excluding slab and lithosphere) | $2.5 \times 10^{20}$ Pa |
| | Viscosity of slab "core" | $10^{23}$ Pa s |
| | Thickness of slab "core" | 50 km |

Subduction is "mature" in that the end (or "tail") of the slab extends to the top of the lower mantle, treated in this paper as a rigid boundary. This condition prevents mantle flow beneath the slab tail (Garfunkel et al., 1986). Note, however, that the position of the slab tail is determined only by viscous forces within the slab and upper mantle. The slab position is not *a priori* fixed with respect to the top of the lower mantle. This differs from tank models where the deep slab becomes fixed onto the base of the tank (see Funiciello et al., 2003). Because the process of slab deformation within the transition zone is poorly known, it is not clear which, if either, of these assumptions is more correct. There should be little difference between them when the foreland lithosphere, and thus the slab tail, is relatively stationary with respect to the top of the lower mantle. However, differences in predicted behavior could become large when there is rapid relative motion between them.

Prior to subduction, slab buoyancy can, to first order, be linearly equated with the bathymetry of the soon-to-be-subducted lithosphere (its "pre-subduction" water depth, $w_o$). For a slab thickness of 100 km and an asthenospheric density of 3300 kg/m$^3$, each kilometer change in pre-subduction water depth translates to a change in mean slab density of ~23 kg/m$^3$. Slab buoyancy is neutral if $w_o$ equals ~2.5 km, very negative if $w_o$ is ~6.5 km (e.g., old oceanic lithosphere), and very positive if $w_o$ is ~0 km (e.g., typical continental lithosphere). (See Royden and Husson, 2006, for further discussion of buoyancy and bathymetry.) Note, however, that slab buoyancy is a function only of the lithospheric material that is actually subducted and the density of that material as it is subducted. If sediments or crust are stripped from the slab as it enters the subduction zone, or if phase changes occur during subduction, the slab buoyancy changes accordingly. In this paper we ignore both of these effects, although they could easily be included with appropriate changes in slab density with depth.

## 3 Subduction of a Model Foreland

In Figs. 2 and 3 we show results for subduction of a model foreland consisting of a deep oceanic domain ($w_0$ = 6.5 km) embedded with several contrasting buoyancy domains, including a continental island (250 km wide, $w_0$ = 0 km), an oceanic plateau (800 km wide, $w_0$ = 3.5 km), and a continental margin ($w_0$ = 0 km) (Fig. 2a). Slab buoyancy is varied in a direction normal to the trench while all other parameters are held constant.

After subduction of oceanic lithosphere has achieved a steady-state geometry and rate (at ~70 mm/year), the contrasting buoyancy domains reach the trench and are subducted sequentially. When the continental island

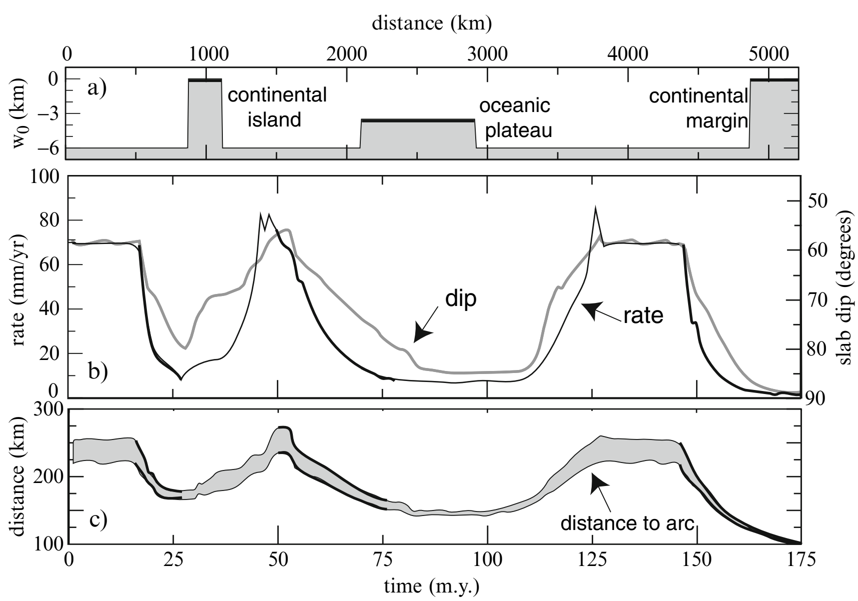
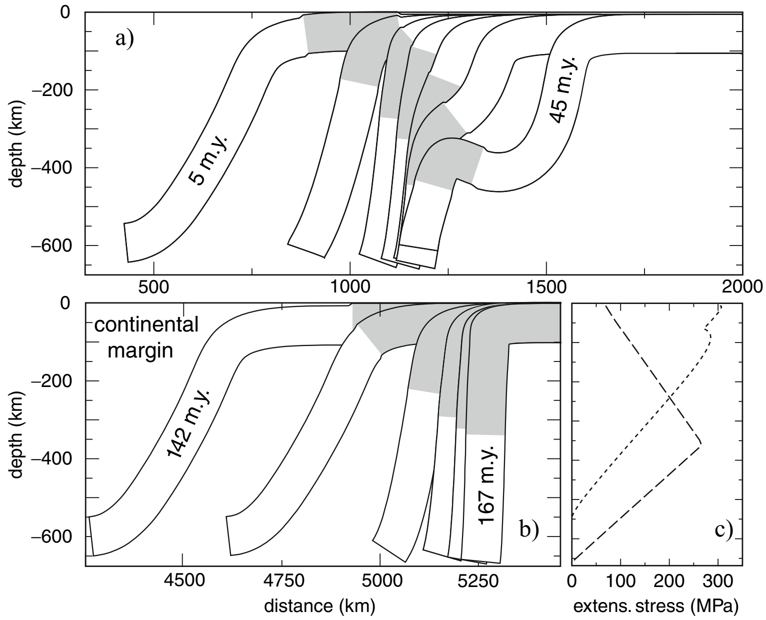



**Fig. 2** (**a**) Pre-subduction water depth for the model foreland to be subducted, as a function of horizontal distance. Large water depths are equated with large negative slab buoyancy, see text for details. *Thick solid lines* represent more buoyant segments of the foreland, consistent in all panels. (**b**) Subduction rate (*black line*) and slab dip at 100 km depth (*gray line*) as a function of time. (**c**) Distance from the trench to the volcanic arc, taken as the horizontal distance over which slab deflection increases from 2 to 100 km (lower value) or to 150 km (higher value)

**Fig. 3** Slab geometries corresponding to two subduction events of Fig. 2: (**a**) subduction of the continental island and (**b**) subduction the continental margin. *Shaded slab* units denote continental lithosphere. The time interval between slab positions is 5 m.y. with beginning and ending times as indicated. (**c**) Extensional stress along the subducted slab for steady-state subduction of oceanic lithosphere and after subduction of a continental margin




enters the trench and reaches the uppermost asthenosphere (~65 km depth), the rate of subduction drops abruptly and trench retreat slows to 5–10 mm/year (Fig. 2b). Subsequently, as the leading edge of the next oceanic segment enters the trench and reaches the uppermost asthenosphere, the subduction rate increases again to ~70 mm/year. A similar pattern of decreasing and increasing rate occurs with subduction of the oceanic plateau. Finally, rates slow and subduction is terminated when the continental margin enters the system; its leading edge is ultimately subducted to ~300 km depth km (Fig. 3, see also Ranalli et al., 2000; Regard et al., 2003).

The time-scale over which subduction rates respond to changes in slab buoyancy is generally on the order of just a few million years. When subduction rates change in response to slab buoyancy, the rates of change can be rapid only when the subduction rate is rapid. This is because each new lithospheric segment, with new buoyancy, enters the system at the rate of subduction, and the subduction system adjusts to the new slab buoyancy accordingly. Thus rapid changes in subduction rate can only occur when the subduction velocity is high; when subduction velocity is low, changes in subduction rate must occur slowly, but the rate of change can become rapid as the subduction rate speeds up.

Model steady-state subduction of the dense oceanic slab occurs at an average dip of ~60–65° (Figs. 2b and 3). When continental or plateau lithosphere enter the system, the slab steepens to more than 80° because the buoyant part of the slab at shallow depth resists sinking and slows the rate of trench retreat; meanwhile the deeper, denser part of the slab continues to sink at a faster rate. Flattening of the shallow slab occurs when dense lithosphere follows buoyant lithosphere into the trench because the dense slab at shallow depth sinks more rapidly than the buoyant material at greater depth. Thus there is a strong correlation between slab dip and trench migration rate (Fig. 2b). For narrow, highly buoyant slab segments, the dip of the slab may become inverted at depth, with a local depth-minimum coinciding with the buoyant slab segment (Fig. 3a).

Equating the distance from the subduction boundary to the volcanic arc with the horizontal distance over which slab deflection increases from 2 to 100 and 150 km, the distance from the subduction boundary to the volcanic arc is ~220–250 km during steady-state oceanic subduction for the parameters used in this paper (Fig. 2c). This decreases to ~150–170 km during subduction of the continental island and oceanic plateau areas. The greatest change in the location of the volcanic arc occurs when subduction terminates against the continental margin and the distance from the thrust belt to the volcanic arc is decreased to ~100 km. The precise distance to the arc is highly sensitive to factors like the density of material in the frontal prism (Fig. 1), but the significant decrease in arc-trench distance during slowing of subduction is a robust feature of the model results.

# 4 Case Studies: Banda and Apennine Systems

The subduction model used here supposes that subduction occurs solely as a result of negative slab buoyancy, so the model results are not applicable to orogenic belts like the Himalaya, which accommodate rapid inter-plate convergence driven largely by far-field forces (e.g., Conrad and Lithgow-Bertelloni, 2004). It is best applied to unforced subduction systems that display upper-plate extension concurrent with subduction, such as those of the Mediterranean region (e.g., Malinverno and Ryan, 1986; Wortel and Spakman, 2000; Royden, 1993a, b; Jolivet and Faccenna, 2000; Faccenna et al., 2003), and the region extending eastward from Indonesia to the Pacific plate (Hall, 1996; Charlton, 2000; Daly et al., 1991; Rangin et al., 1990; Hinschberger et al., 2000). The Apennine and Banda Sea systems offer excellent opportunities to compare model results with observations in systems where subduction has terminated following entry of continental lithosphere into the subduction zone. Conversely, there are also subduction systems where an increase in subduction rate can be correlated with entry of denser lithosphere into the trench, but the amount of geologic information needed to document rates and timing renders such comparisons beyond the scope of this paper.

## 4.1 Geological Settings

Subduction of Australian lithosphere northward beneath the Banda Arc occurred prior to Pliocene time at a rate of ~70 mm/year (Fig. 4a, after Hinschberger et al., 2000). The pre-subduction depth, $w_o$, of the subducted oceanic lithosphere was probably similar to that observed today in the adjacent Indian Ocean, ~5.7 km. During Pliocene time, the eastern end of the subduction system encountered the continental margin of northwestern Australia and arc-continent collision occurred diachronously from west to east. During subduction of the margin, $w_o$ decreased to its current value of ~0.05 km. The rate of subduction dropped to zero by 0.5–1.0 Ma while to the west (Sunda arc), subduction of oceanic lithosphere continues at ~70 mm/year (Kreemer et al., 2000; Richardson and Blundell, 1996; Hughes et al., 1996). Thus the Banda Sea offers an excellent opportunity to observe the reaction of subduction system to the entry of continental material into the trench, and at the same time to observe steady-state subduction of oceanic lithosphere (Sunda arc) that is probably similar to that of the Banda arc prior to collision.

The Late Cenozoic Apennine system consumed mixed oceanic and continental lithosphere along a west-dipping subduction system (Fig. 4b). Reconstruction of the southern Apennines suggests a Late Miocene subduction rate of ~30–50 mm/year or perhaps higher (e.g., Patacca et al., 1990; Fig. 5b, detailed rates taken from the unpublished work of P. Scandone, pers. com., 2006, average rate from Faccenna et al., 1997). The apparent oscillations in rate are probably artifacts of an incomplete geological record, and the maximum rates are probably the more significant. In any case, we choose a rate of 40 mm/year as the pre-collisional rate of subduction for the Apennine system. In Pliocene time, the southern Apennine trench encountered the continental Adriatic lithosphere and thrusting and subduction had

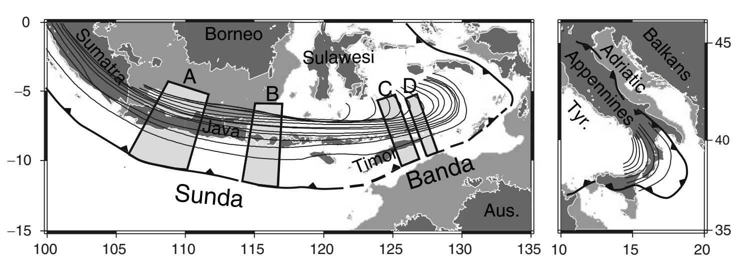



**Fig. 4** Location maps for the Banda Arc and Apennine subduction systems. *Boxes* show locations of swath profiles for earthquake hypocenters in Fig. 6. Slab contours are 50 km isodepths, adapted from Sambridge and Gudmundsson (1998). Light gray domains are shallow water domains. *Solid and dashed bold curves* indicate active and fossil trench locations

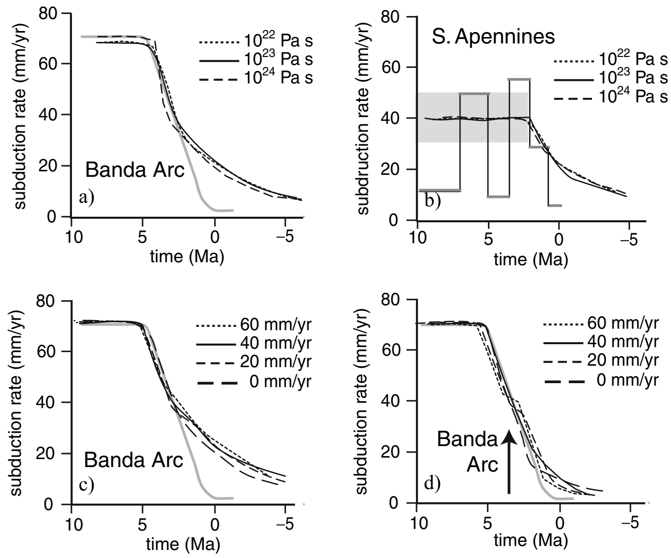



**Fig. 5** Observed (*gray curves*) and model (*black curves*) rates of subduction for the Banda arc the Southern Apennines. Parameter values are given in Table 2. (**a**), (**b**) Subduction rate versus time following entry of continental lithosphere into the trench at 5 Ma. Foreland velocity relative to the top of the lower mantle is zero. *Labeled curves* correspond to different values of slab viscosity. (**c**) Subduction rate versus time for the Banda Arc following entry of continental lithosphere into the trench at ~5 Ma. *Labeled curves* correspond to different imposed values foreland velocity (from foreland toward trench) relative to the top of the lower mantle. (**d**) Same as (**c**) except that slab detachment occurs at 470 km depth at the time indicated by the *vertical arrow*

**Table 2** Parameters for Figs. 5 and 7

| | Pre-subduction water depth | Northward foreland velocity (relative to lower mantle) | Slab viscosity | Mantle viscosity |
|---|---|---|---|---|
| Figure 5a (Banda) | 5.7 km | 0 mm/year | $10^{22}$ Pa s | $1.9 \times 10^{20}$ Pa s |
| | 5.7 km | 0 mm/year | $10^{23}$ Pa s | $1.8 \times 10^{20}$ Pa s |
| | 5.7 km | 0 mm/year | $10^{24}$ Pa s | $1.5 \times 10^{20}$ Pa s |
| Figure 5b (Apennines) | 5.0 km | 0 mm/year | $10^{22}$ Pa s | $2.4 \times 10^{20}$ Pa s |
| | 5.0 km | 0 mm/year | $10^{23}$ Pa s | $2.2 \times 10^{20}$ Pa s |
| | 5.0 km | 0 mm/year | $10^{24}$ Pa s | $1.8 \times 10^{20}$ Pa s |
| Figure 5c,d (Banda) | 5.7 km | 0 mm/year | $10^{22}$ Pa s | $1.9 \times 10^{20}$ Pa s |
| | 5.7 km | 20 mm/year | $10^{22}$ Pa s | $2.1 \times 10^{20}$ Pa s |
| | 5.7 km | 40 mm/year | $10^{22}$ Pa s | $2.4 \times 10^{20}$ Pa s |
| | 5.7 km | 60 mm/year | $10^{22}$ Pa s | $2.7 \times 10^{20}$ Pa s |
| Figure 7 | 5.7 km | 0 mm/year | $10^{23}$ Pa s | $1.8 \times 10^{20}$ Pa s |
| | 5.7 km | 0 mm/year | $10^{24}$ Pa s | $1.5 \times 10^{20}$ Pa s |

ended by Quaternary time (see e.g., Patacca et al., 1990; Faccenna et al., 2001; Jolivet and Faccenna, 2000).

## 4.2 Model Parameters

The transition from oceanic to continental foreland was modeled for these two systems using the same parameters as in Fig. 2 and Table 1, except as noted in Table 2. In particular, the pre-subduction water depth of oceanic lithosphere subducted in the Banda region was estimated from the depth of the adjacent ocean basin (5.7 km for the Banda Arc). The pre-subduction water depth for the deepwater portions of the Apennine slab is unknown, so we chose a moderate depth of 5 km. In both cases the pre-subduction water depth of continental portion of the lithosphere is set to sea-level. A range of slab viscosities, from $10^{22}$ to $10^{24}$ Pa s, were used in the model runs, with the viscosity of the surrounding mantle chosen in each case to provide a subduction velocity of ~70 mm/year for the Banda system and ~40 mm/year for the Apennine system.

## 4.3 Subduction Rates

Model rates of subduction are in excellent agreement with the rapid termination of subduction within the Apennine and Banda systems for the initial period of slow-down, with velocities slowing down from ~70 to ~30 mm/year over a 2 m.y. period for the Banda Sea model, and from ~40 to ~20 mm/year over a 2 m.y. period for the Apennine system (Fig. 5a and b). This indicates that the timescale over which subduction initially responds to changes in slab buoyancy is similarly rapid in the natural systems and in the model, and is largely independent of slab viscosity. However, for all slab viscosities, the latter stages of subduction termination are significantly faster in the observed systems than in the model results. In particular, the slowing of model subduction from 20 to 0 mm/year occurs over at least 10 m.y., with a gradual slowing of rate, while subduction in the Apennine and Banda systems terminates abruptly.

One possible source of the mismatch between observed and model results is the imposed velocity of the foreland relative to the top of the lower mantle (see e.g., Schellart, 2005). The foreland lithosphere for the Apennine subduction system is approximately stationary, or at least not moving very rapidly, in the hot spot reference frame (e.g., Gordon and Jurdy, 1986). Its motion with respect to the top of the lower mantle is, presumably, similar. However, the Indian-Australian plate (the foreland for the Banda subduction system) is moving northwards in the hotspot reference frame at ~60 mm/year, bringing into question the importance of the absolute velocity of the foreland with respect to the top of the lower mantle.

Figure 5c shows the model decrease in subduction rate expected for imposed foreland velocities, relative to the top of the lower mantle, ranging from 0 to 60 mm/year, in a direction orthogonal to and toward the trench. In order to match the initial subduction velocity of ~70 mm/year and to keep the initial slab buoyancy consistent with a pre-subduction water depth of 5.7 km, the viscosity of the mantle had to be adjusted for each individual case (Table 2). With these constraints, there is little predicted variation in the rate at which subduction slows as a function of imposed foreland velocity, indicating that this is probably not the cause of the misfit between observations and model in the latter stages of subduction termination. This result is consistent with the Apennine and Banda systems exhibiting the same rapid termination in subduction regardless of foreland velocity. It appears that some process, not included in our subduction model, must be responsible for the rapid termination of subduction in the latter stages of continental margin subduction, as we discuss further below.

## 4.4 Slab Dip and Migration of the Volcanic Arc

A robust result of the model is that the slab dip steepens as buoyant material enters and descends into the subduction system. For example, Fig. 3 shows the model slab dip steepening from about 60–65° during steady-state subduction of oceanic lithosphere to 85° or more after buoyant lithosphere enters the trench. This compares favorably to the slab geometries indicated by earthquake hypocenters from the Sunda and Banda arcs (Fig. 6); the average slab dip in the Sunda arc, where oceanic lithosphere is being subducted in steady-state is ~55–60° between 200 and 400 km depth. The slab dip along the Banda arc, where subduction has ceased, is ~75–80°. Both model and observations show an increase in slab dip of about 20° during entry of the continental margin into the subduction system. The modern slab dip beneath the Apennines is also very steep, perhaps 80° or more (Lucente et al., 1999; Piromallo and Morelli, 2002).

Another robust result from the model is the migration of the volcanic arc (equated in the model with a depth of 100–150 km at the top of the subducted slab) toward the trench during subduction of buoyant lithosphere. Thus there is a strong correlation between subduction rate and the location of the volcanic arc. The modern position of the volcanic arc within the Southern Apennines is ~100 km from the thrust front, similar to

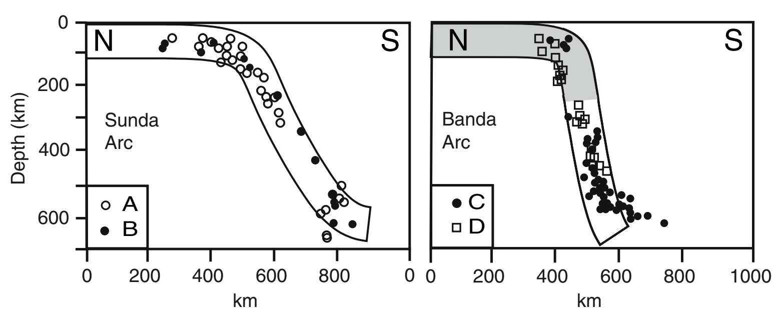



**Fig. 6** Observed earthquake hypocenters (*dots*) from the Banda and Sunda arcs as given by Schoffel and Das (1999) and Das (2004). Swath locations shown in Fig. 4. *Solid lines* show model slab profiles from Fig. 3, for steady-state subduction of oceanic lithosphere (*left panel*) and following entry of the continental lithosphere (*shaded region*) into the subduction boundary (*right panel*, geometry corresponding to 152 m.y. in Fig. 2)

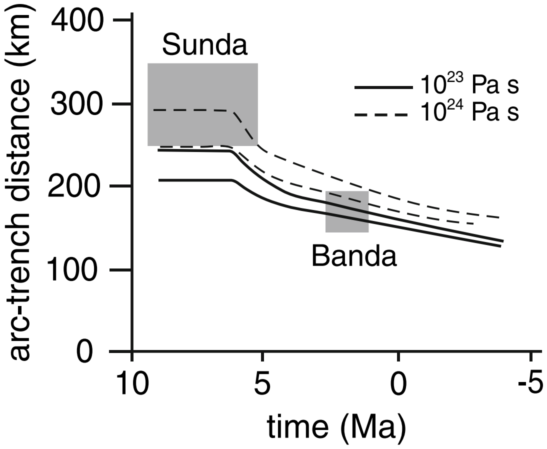



**Fig. 7** Observed distance from the Banda and Sunda volcanic arcs to the adjacent trench or frontal thrust belt (*shaded squares*) as a function of time following entry of the continental lithosphere into the subduction boundary at ~5 Ma. *Lines* show model distances from trench to the front and back of the volcanic arc for the viscosities shown. Front and back of the model volcanic arc are equated with the positions where the upper slab boundary descends below 100 and 150 km depth, respectively

the model results of Fig. 2c after subduction of the continental margin. Within the Banda-Sunda arc system, the volcanic arc is ~250–350 km from the Sunda trench but only ~150–200 km from the inactive Banda thrust front. Using the Sunda arc as a proxy for the Banda arc prior to its interaction with the Australian continental margin, there appears to have been a 100 km decrease in the distance from the thrust front to the volcanic arc that is associated with subduction of the margin (Fig. 7). While we cannot rule out that some of the decrease in arc-trench distance might be due to shortening in the fore-arc region after the eruption of the last volcanic arc rocks at ~3 Ma, the fact that the slab dip is steeper slab the Banda region than in the Sunda region is also consistent with a decrease in the arc-trench distance during slowing and cessation of subduction.

## 5 Slab Detachment?

A number of authors (e.g., Buiter et al., 2002; Wortel and Spakman, 2000) have suggested that during the entry of continental lithosphere into a subduction zone, the lower part of the slab detaches from the upper part, in a process commonly referred to as "slab break-off" or "slab detachment." This process has been specifically suggested for both the Apennine (Buiter et al., 1998) and Banda arc (Elburg et al., 2004) systems. While a detailed investigation of this process is beyond the scope of this paper, we can make a preliminary assessment of whether this process could lead to a sufficiently rapid termination of the last stages of subduction, as indicated in Fig. 5.

Figure 5d shows the results for a simulated slab break-off at the time indicated by the vertical arrow. Although our model does not really allow for proper slab break-off, we simulate instantaneous slab break-off by truncating the slab at 470 km depth and setting the viscous stresses on the bottom of the slab to zero. This somewhat *ad hoc* modification of the forces on the slab eliminates the high stress area beneath the lower part of the slab (Fig. 1) and also eliminates the viscous stresses that result from toroidal flow around the slab (because mantle can flow through the region formerly occupied by the deep part of the slab). The results indicate a very rapid transition from slow subduction to no subduction, in good agreement with observations from the Banda Sea (Fig. 5d) and the Apennines (not shown but similar to Fig. 5d). Although not a precise simulation of slab break-off, these results suggest that slab break-off may provide a mechanism whereby the last stages of subduction occur very quickly.

Importantly, the reduction in weight of the deep slab is not the most important factor in causing rapid cessation of subduction. Rather, it is the creation of a shortcut for mantle flow that is critical in enabling rapid steepening of the slab into its final, stationary position. In particular, the creation of gaps in the slab enables rapid flow out of the geometrically confined, high-stress region beneath the deep slab (Fig. 1), thereby allowing for rapid motion of the deep slab.

## 6 Discussion and Conclusions

In general, rates of subduction and trench motion respond rapidly to changes in slab buoyancy. Subduction rates respond more rapidly to changes in slab buoyancy when the subduction rate is fast, and less rapidly when subduction rates are slow. Because subduction rates are faster for more negatively buoyant slabs, there is a discrepancy between the amount of time that various buoyancy domains spend in the shallow part of the subduction system and their original extent within the foreland. In Fig. 2, continental lithosphere and oceanic plateau regions initially make up ~25% of the foreland, but account for ~55% of the subduction time. Thus the duration of events within a thrust belt cannot be used as a direct measure of the original extent of various buoyancy or facies domains because the low buoyancy regions will be over-represented in the duration of thrusting.

The dip of subducted slabs at shallow depth (~100 km) tracks closely with the rate of trench migration. For the parameters used in this paper, very dense slabs (old ocean) subduct at angles of ~60–65° whereas more buoyant lithospheres, like the oceanic plateau modeled in Fig. 2, subduct at steeper angles. (The precise slab dip depends on a variety of parameters and may vary from system to system even if slab buoyancy is identical). When all other parameters are held constant, the correlation between subduction rate and slab dip at 100 km depth is surprisingly good. This is counter to the common misconception that high-density slabs dip more steeply than low-density slabs (e.g., Luyendyk, 1970). The latter is correct if the trench retreat rate is held constant while slab buoyancy is varied, but not when both subduction rate and trench-migration rate are controlled by the negative buoyancy of the slab. This agrees with the fact that little correlation is found between observed slab dip and slab density (e.g., Jarrard, 1986; Heuret and Lallemand, 2005).

Because arc-magmas are generated above subducted slabs at depths of 100–150 km, there is also a correlation between the slab buoyancy, slab dip and location of the volcanic arc. As slab dips steepen, the volcanic arc should move closer to the trench (or front of the thrust belt). This occurs most markedly during subduction of a continental margin or other large buoyant object and is illustrated by the Banda/Sunda subduction system (Fig. 2). Equating the distance from the trench to the volcanic arc with the horizontal distance over which the slab depth increases from 10 to 100 km, the model distance from the trench to the volcanic arc decreases by ~100 km during subduction of the continental margin.

When subduction terminates, the viscous stresses on the slab disappear and the stresses on the slab arise only from its buoyancy. If subduction is terminated by subduction of a continental margin or other buoyant object, the depth to which the leading edge of the buoyant lithosphere can be subducted depends on its buoyancy and the buoyancy of deeper, denser portions of the slab. This is reflected in the extensional stresses transmitted along the slab (Fig. 3c). For a uniformly buoyant slab subducting in steady-state, the extensional stress along the slab increases toward the surface. (Note that extensional stress is not necessarily zero at the surface; only the vertical component of stress, integrated along the slab, must be zero.) When subduction terminates after subducting a continental margin, the extensional stress within the subducted slab is greatest at the former ocean-continent transition. This result is intuitive because at that stage there are no longer stresses on the slab from viscous flow in the surrounding mantle. Thus the extensional stress on the slab increases upwards to the point where the density of the slab is less that the density of the surrounding mantle. This suggests slab attenuation and possible slab detachment may be most likely to occur at this site (see Fig. 3c and Wortel and Spakman, 2000; Regard et al., 2003).

Our results apply largely to unforced subduction systems, where trench motion is controlled by slab buoyancy and not by far-field constraints on relative plate velocities. Modeling studies (e.g., Hassani et al., 1997) and observations (e.g., on the Nazca subduction zone, Gutscher et al., 1999) indicate that velocities imposed on the upper plate can play a large role in determining slab geometry and kinematics. For this reason, we have chosen natural examples where the upper plate lithosphere experienced extension during the subduction process, suggesting relatively little stress coupling between the over-riding and down-going plates.

Nevertheless, some of the general concepts developed here may be extrapolated more broadly. We suggest that, for large plates like the Pacific, local and short-term variations in upper plate deformation, such as in the Japan or Kurile regions, may be more closely related to temporal and spatial variations in slab buoyancy than to the mean buoyancy of the slab. Even small variations in slab buoyancy result in speeding or slowing of trench motion. Although the change in rate may be but a minor fraction of the total trench velocity, even a change in rate of 5 mm/year over 10 m.y. can result in significant stretching or shortening of the upper plate. If this interpretation is correct, it helps to explain the apparent contradiction that slab buoyancy is the most important driver of subduction, yet there appears to be a lack of correlation between local slab buoyancy and associated upper plate deformation.

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# Continental Collision and the STEP-wise Evolution of Convergent Plate Boundaries: From Structure to Dynamics

Rinus Wortel, Rob Govers and Wim Spakman

**Abstract** Particularly interesting stages in the evolution of subduction zones are the two main transient stages: initiation and termination. In this contribution the focus is on the second of these: terminal stage subduction, often triggered by continental collision or arc-continent collision. The landlocked basin setting of the Mediterranean region, in particular the western-central Mediterranean, provides unique opportunities to study terminal stage subduction and its consequences.

We use seismic tomography results on lithosphere and upper mantle structure as a source of information on plate boundary structure, and concentrate on the lithospheric scale aspects. Combining this structural information with process-oriented numerical modelling studies and regional observations, we present a 3D model for convergent plate boundary evolution after collision, in which slab detachment and the formation of tear or STEP (Subduction-Transform-Edge-Propagator) faults are key elements. A STEP fault laterally decouples subducting lithosphere from non-subducting lithosphere in a scissor type of fashion. It enhances the ability of a slab to retreat through the mantle flow around the edge of the subducted slab. In this way collision and back-arc extension may occur in close proximity. In our study area this specifically pertains to collision along the north African margin, STEP formation in easterly direction, CCW rotation of the southern Apennines slab and the opening of the Tyrrhenian Sea. Vertical tearing of subducted lithosphere may play an important role as well, but is probably not crucial. On the basis of the good agreement between the Mediterranean-based model and the evolution of the Tonga-Fiji region we expect that the model may shed light on other complex convergent plate boundary regions, as well.

In summary: Upon continental (or arc-continent) collision, along-trench variations in lithospheric properties of the subducting lithosphere may lead to disruption and segmentation of the subduction system. Following slab detachment along limited segments of a convergent plate boundary, the development of STEP faults is expected. These faults contribute to an increase in arc curvature within plate boundary segments. This contributes to the sinuous geometry of long subduction systems such as in the western and southwest Pacific.

**Keywords** Subduction • Seismic tomography • Continental collision • Slab detachment • STEP fault • Roll-back

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## 1 Introduction

Among the stages that may be distinguished in subduction zone evolution, the most intriguing ones probably are the transient stages of initiation and termination. Transient stages typically result in a wider spectrum of expressions than approximately steady state conditions. In this contribution the focus is on terminal stage subduction, often triggered by continent-continent or arc-continent collision (both hereafter referred to as continental collision). In this terminal stage the topic of back-arc extension deserves special attention, in view of Royden's (1993) finding of a peculiar combination of collisional events and escape type of arc migration and back-arc extension.

Since Karig (1971), the relation between subduction and back-arc extension (in combination with migration of the arc involved) is well-known. Identifying hinge migration as resulting from the gravitational instability of dense oceanic lithosphere (Elsasser, 1971) was a major step in understanding the physical cause of the connection. The process became commonly referred to as (slab) roll-back or slab retreat. Subsequent progress was hampered by the intrinsic short-comings of 2D modelling. This led Dvorkin et al. (1993) to investigate the significance of 3D aspects of subduction zones, particularly a finite width slab allowing mantle flow around the slab edges. Several studies expanding on this topic and other 3D aspects of subduction have further elucidated their role in subduction zone evolution and back-arc extension (Funiciello et al., 2003; Schellart, 2004; Bellahsen et al., 2005; Stegman et al., 2006; Schellart et al., 2007).

Because continental collision plays a key role in terminating subduction we choose the Mediterranean region, the evolution of which is characterized by continental collision and terminal stage subduction (Le Pichon, 1982; Jolivet and Faccenna, 2000; Wortel and Spakman, 2000), as our study area. We start from the regional geodynamic setting. Deploying information on lithosphere and upper mantle structure as obtained from seismic tomography we concentrate on the geodynamic process causing and allowing roll-back. Adopting a lithospheric scale approach, we combine the structural information with physical guidance from modelling studies to arrive at an evolutionary model for terminal stage subduction and accompanying back-arc extension.

## 2 Terminal Stage Subduction in Mediterranean Region: North Africa and the Sicily-Tyrrhenian Region

Collision has been part of geodynamic analyses of the Mediterranean for many decades (e.g., Argand, 1924), albeit originally predominantly in the context of the Africa-Europe collision in the Alps. In a plate tectonic framework, the role of collision in the evolution of the wider Mediterranean region has been highlighted in different ways by Le Pichon (1982) and Royden (1993).

Le Pichon (1982) focused on the role of Africa-Europe collision in the westernmost Mediterranean (Gibraltar arc region) and the Arabia-Anatolia collision in the east (Bitlis region) in establishing a landlocked basin setting. In such a setting the gravitational instability of the African lithosphere of the Mediterranean promoted roll-back of subduction zones and associated back-arc extension. Jolivet and Faccenna (2000) further advanced our understanding of the regional evolution by pointing out the contemporaneity of the start of extension in the Mediterranean and the role of the decrease in absolute velocity of the African plate in causing an acceleration in slab retreat. On a somewhat smaller scale, Royden (1993) concentrated on the kinematic relationship between collisional processes in parts of the Alpine-Mediterranean region and escape type of arc migrations in adjacent regions, such as the Carpathians, the Hellenic and Apennine arcs and the Betics-Rif thrust belt.

In this paper we use tomographic results to delineate 3D aspects of subduction zones and process-based understanding to investigate the physical processes giving rise to a kinematic pattern very similar to the type described by Royden (1993).

### 2.1 Deep Structure and Regional Evolution

Seismic tomography has played a major role in producing new information concerning the deep structure of subduction zones, in particular where the subducted lithosphere's geometry is not or only partially outlined by earthquake hypcentres. This has very clearly been the case in the Mediterranean region where indeed earthquake hypocentres only provided very limited windows on the deep structure of the convergent plate boundary zone (Giardini and Velonà, 1991; Spakman et al., 1993; Wortel and Spakman, 1992, 2000; Spakman and Wortel, 2004; Piromallo and Morelli, 2003; Lucente et al., 1999). Our early studies of the geodynamic implications of seismic tomography results (Spakman et al., 1988; Wortel and Spakman, 1992) showed the significance of the new source of information on structure of the lithosphere and upper mantle.

The Mediterranean region shows arc migration and back-arc extension in both the western Mediterranean (Liguro-Provencal Basin and Tyrrhenian Sea) and the Aegean region. Since in the Aegean case back-arc extension is fully active (McClusky et al., 2000), terminal stage subduction can be studied best in the western Mediterranean. We refer to Spakman and Wortel (2004) for a detailed overview of the tomographic results available for the western-central Mediterranean, in combination with resolution tests, a discussion of other tomographic results and a regional scale interpretation.

There is general consensus on the first order kinematic evolution of the western Mediterranean. Characteristic feature is the Late Oligocene to Recent arc migration starting from a location along the eastern margin of Iberia and southern France (e.g., Dewey et al., 1989; Auzende et al., 1973; Rehault et al., 1984; Doglioni et al., 1997; Faccenna et al., 2001a; Schettino and Turco, 2006). The migrating trench system made contact with the north African continental margin in the Langhian (~15 Ma). In a second episode of extension starting at about 10 Ma, the extension continued with an eastward shift of activity (Fig. 1), towards the Tyrrhenian realm (Carminati et al., 1998; Faccenna et al., 2001b). Malinverno and Ryan (1986) illustrated the relevance of Elsasser's (1971) gravitational sinking and arc migration model for the Tyrrhenian region.

The present-day plate boundary between Africa/Adria and Eurasia is often drawn from the Apennines, through the highly curved Calabrian Arc towards the Maghrebides in northern Africa. As shown by Carminati et al. (1998) on the basis of seismic tomography results, however, this apparently continuous arc does not correspond to an equally continuous subduction zone. Here we display the regional seismological structure in a series of horizontal slices through the BS2000 model of Bijwaard and Spakman (2000) (Fig. 2). The sections show a high velocity anomaly of the Calabrian slab and clearly indicate the absence of a subducted slab underneath the plate boundary segment in and to the west of western Sicily.

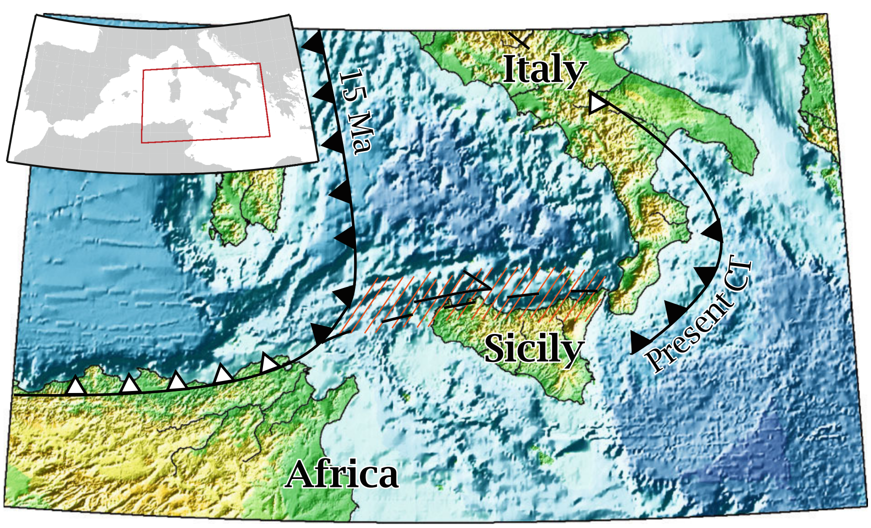


**Fig. 1** Regional topographic/bathymetric map of the Sicily-Tyrrhenian Sea region (see inset for location of the main figure within the larger western-central Mediterranean region). It shows the location of the convergent plate boundary in the western-central Mediterranean in the Langhian (~15 Ma) and at Present. *CT* = Calabria Trench. White dents indicate detachment of subducted slab, black dents indicate continuous slabs. The exact location of the transition from detached slab towards continuous slab in Southern Italy is uncertain. The STEP fault is shown as a deformation zone, with an overall dextral sense of shear

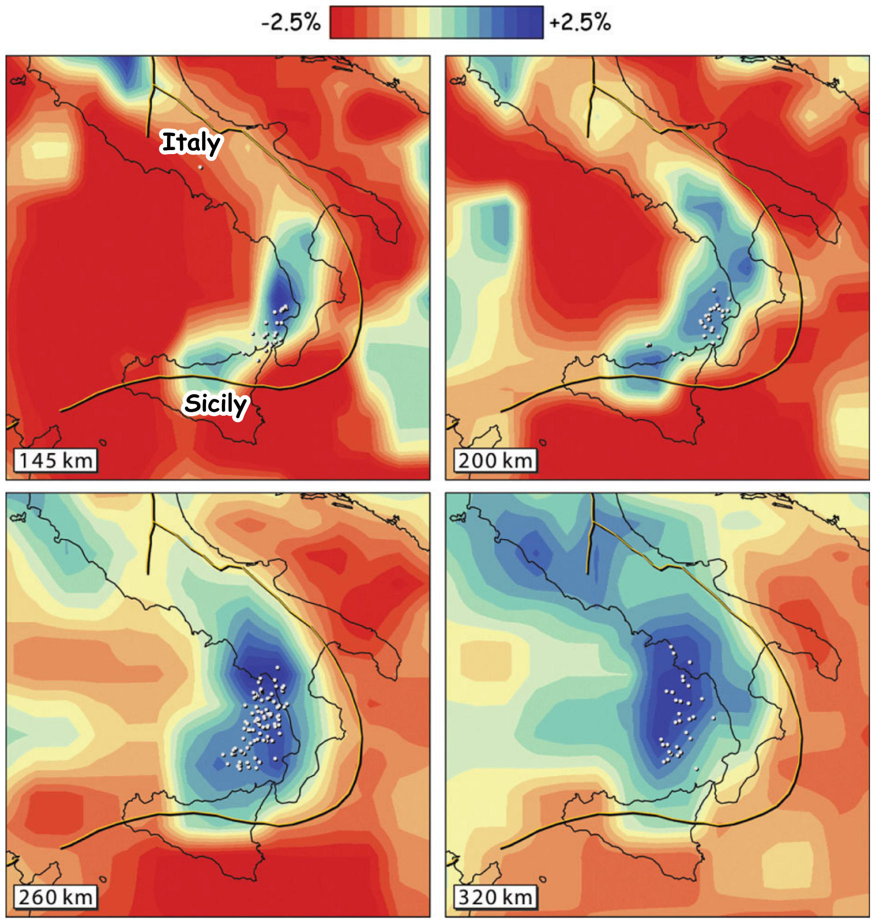


**Fig. 2** Horizontal sections through the BS2000 seismic tomography model of Bijwaard and Spakman (2000) for the Sicily-Tyrrhenian region, at the depths indicated. P-wave velocity anomalies are given in percentages relative to the ambient depth-dependent velocity of the one-dimensional reference model ak135 (Kennett et al., 1995). White dots indicate earthquake hypocentres

## 2.2 Plate Boundary Evolution

The seismological structure in Fig. 2 pertains to the present-day situation. In order to develop a model for the evolution of the plate boundary we use results from modelling studies and follow the plate boundary evolution in a process-oriented way. We start from the stage of collision between the outwardly migrating (in S, SW and E direction) trench system and the north African continental margin (Fig. 1). In doing so we summarize earlier work and put it in a regional perspective.

### 2.2.1 Continental Collision and Slab Detachment

The migrating trench in the western Mediterranean reached the North African margins at ~15 Ma. Van de Zedde and Wortel (2001) numerically investigated the evolution of stress and strength in subducting lithosphere in case of transition from oceanic to continental lithosphere subduction. They showed that continental collision (in particular the soft type of collision, i.e., upon slab retreat) can lead to very shallow slab detachment (breakoff) of the subducted slab. This is schematically shown in Fig. 3a.

The seismic velocity distribution (Spakman and Wortel, 2004; Wortel and Spakman, 2000) shows evidence for the absence of high velocity slab material, and hence for slab detachment, below the north African margin in the ~100–300 km depth range, almost along the margin's entire length. It does not show clear evidence for a distinct pattern of lateral migration of the slab detachment process, neither to the west nor to the east. In analyses of convergent margin evolution it is important that slab detachment occurs with a considerable delay time (several Myr) relative to the time of

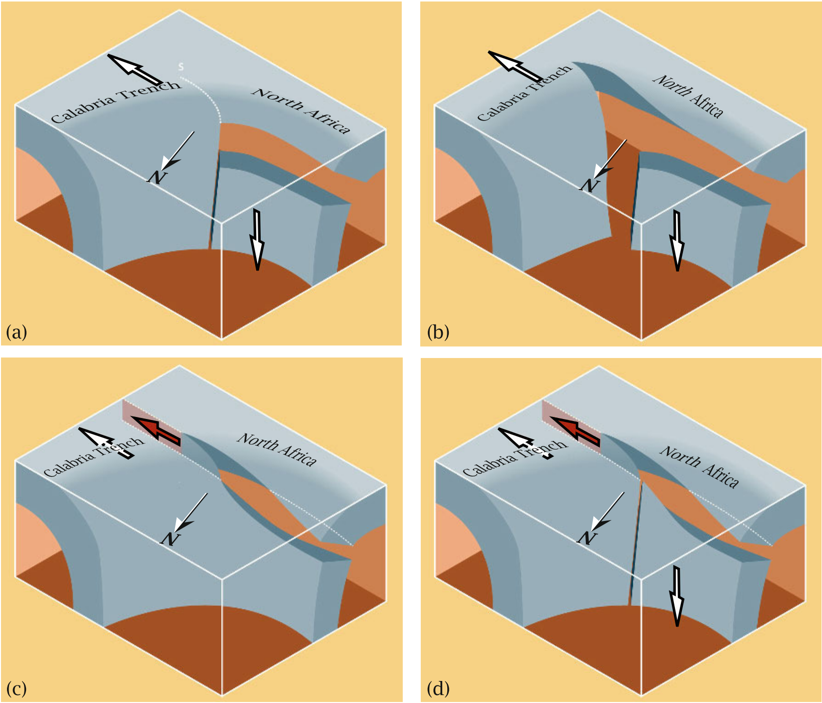


**Fig. 3** Schematic representation of the subduction zone geometry in the Sicily-Tyrrhenian region. The region approximately corresponds with that in Fig. 1. The vertical scale of the boxes is 300 km. (**a**) Detached slab after collision of the African continent with the migrating trench system in the western-central Mediterranean. A vertical tear separating the detached part from the continuous part is indicated. Dotted line, labelled S, indicates future STEP fault. (**b**) Formation of a STEP fault and subsequent trench retreat (roll-back). (**c**) and (**d**) An alternative scenario for STEP fault generation: The detachment fault in the subducted slab may shoal and rotate towards an approximately vertical orientation. The detached slab segment (in c) may tear away from the continuous segment (see **d**). In this case the configuration closely resembles that in (**a**)

first arrival of the continental lithosphere at the trench (van de Zedde and Wortel, 2001).

The lateral variation in contact properties along the trench system in the western-central Mediterranean implied that slab detachment was probably confined to the north African segment, whereas the Calabrian segment, in which the oceanic Ionian lithosphere was consumed, remained continuous. This leads to the next phase.

### 2.2.2 Formation of a Vertical Slab Tear

A detached slab along a major part of the plate boundary gives rise to in-plate bending stresses (Yoshioka and Wortel, 1995), tending to tear away the detached part from the continuous part. We assume a vertical tear to develop as in Fig. 3a (this is not a critical assumption; an alternative is presented below and in Fig. 3c and d). The vertical tear leads to the creation of a free slab edge. This is an important evolutionary step in the process, since both analogue and numerical experiments have convincingly shown the prominence of transport of asthenospheric and deeper mantle material around the edge of the slab, allowing the slab to retreat further and faster (see references given in Introduction).

### 2.2.3 Formation of a STEP Fault

Carminati et al. (1998) and Carminati and Wortel (2000) proposed a tear fault as the connection between the north African and the Calabrian plate boundary segments (Fig. 1). The physical need for this type of fault was recognized from the beginning of plate tectonic theory (Isacks et al., 1968: "scissors type of faulting"; also regularly referred to as tear fault or transfer fault). Govers and Wortel (2005) drew attention to the non-rigid body nature of the relative motions along such a fault, and named it STEP fault (**S**ubduction-**T**ransform-**E**dge-**P**ropagator). They investigated the potential for propagation and the stress field and vertical motions associated with its activity. The motion along a STEP fault, in general, and the dextral motion indicated in Fig. 1, in particular, are to be understood as the overall sense of motion of a possibly broad deformation zone, in which also block rotations can be involved. The STEP is formed by the action of the slab pull acting on the continuous slab, around the edge of which the mantle material flows and accommodates the roll back. STEP faulting and roll-back are intrinsically coupled aspects; a STEP is a crucial element in the 3D evolution of arc systems exhibiting roll-back.

We note that vertical tearing as described above (Fig. 3a) is not required for the initiation of STEP faulting. In view of the possibly very shallow level of slab breakoff (~35 km; van de Zedde and Wortel, 2001) we envisage another possibility, namely shoaling and rotation of the detachment fault in the subducted slab towards an approximately vertical orientation (Fig. 3c) to form a STEP fault. Shoaling may be dependent on the curvature of the trench. Whether the detached slab remains linked to the continuous segment (Fig. 3c) or not (Fig. 3d) is not critical for the further evolution. The configuration in Fig. 3d is virtually the same as in Fig. 3a, with a vertical tear.

If the detached part of the slab (in Fig. 3: below North Africa) remains laterally connected to non-detached subducted lithosphere (in Fig. 3c: below Calabria) the slab pull acting on the non-subducted lithosphere at the surface, is concentrated in the trench segment where the subducting slab is still continuous in the downdip direction (in Fig. 3c: along the Calabria Trench segment). STEP fault formation, indicated by the red arrow, is to be expected in a very similar way as in Fig. 3a and b.

#### 2.2.4 Slab Retreat with or without Back-Arc Extension

Once a STEP fault has been initiated the adjacent continuous slab segment can roll-back. Depending on the regional tectonic setting, the slab retreat can be accompanied by either advective motion as a rigid body or deformation of the overriding plate (Fig. 4; see also Govers and Wortel, 2005). Figure 4 illustrates this by two situations with different boundary conditions on the left side of the model: (a) no "east-west" motion between left and right side of the model, and (b) east-west motion is possible. For the Tyrrhenian situation eastward rigid body type advection (relative to the African lithosphere) of the overriding lithosphere was/is not possible (situation as in Fig. 4a) and deformation developed, giving rise to the back-arc extension in the Tyrrhenian Sea, of which the Marsili and Vavilov basins with their oceanic basement are the most prominent expressions. The Caribbean plate is an example of an overriding plate which largely follows the eastward motion of the Lesser Antilles trench through rigid body motion (similar to Fig 4b).

In summary, we connected a series of processes into an evolutionary model for plate boundary evolution after continental collision. The components are (Fig. 3): (1) Continental collision ("soft", in the north African case); (2) Slab detachment; (3) Formation of a vertical tear which laterally separates the detached part of the slab from the adjacent still continuous part of the subducted slab. This step is not a crucial one; (4) Formation of a STEP fault and retreat of the continuous segment of the slab; (5) Back-arc extension, unless advection of the overriding follows the retreating trench. Basically, the entire process is governed by the transient distribution of slab pull forces acting on the subducting slab.

The above analysis was aimed at abstracting an essential aspect of collision and subsequent plate boundary evolution, which – as was realized by Royden (1993) – can be studied very well in the Mediterranean context. For other analyses of back-arc extension and arc migration in the western-central Mediterranean we refer to Faccenna et al. (2001a, b, 2004), Wortel and Spakman (2000) and Spakman and Wortel (2004), and other references given in Sect. 2.1.

## 3 Test of Evolutionary Model

### 3.1 *Magmatism*

Although the scenario developed for the Sicily-Tyrrhenian region seems physically sound, we would like to subject the above model scenario to new tests, preferably against data which have not been used in the development of the scenario. This could be done by using observations from the same region but of a different type. Magmatic data are valuable in this context, since they bear information on transient process, including their timing. Maury et al. (2000) and Coulon et al. (2002) concluded that alkaline magmatism in northern Africa is accounted for by slab breakoff, as was proposed by Carminati et al. (1998). In the Tyrrhenian realm, data on spatial and temporal varia-

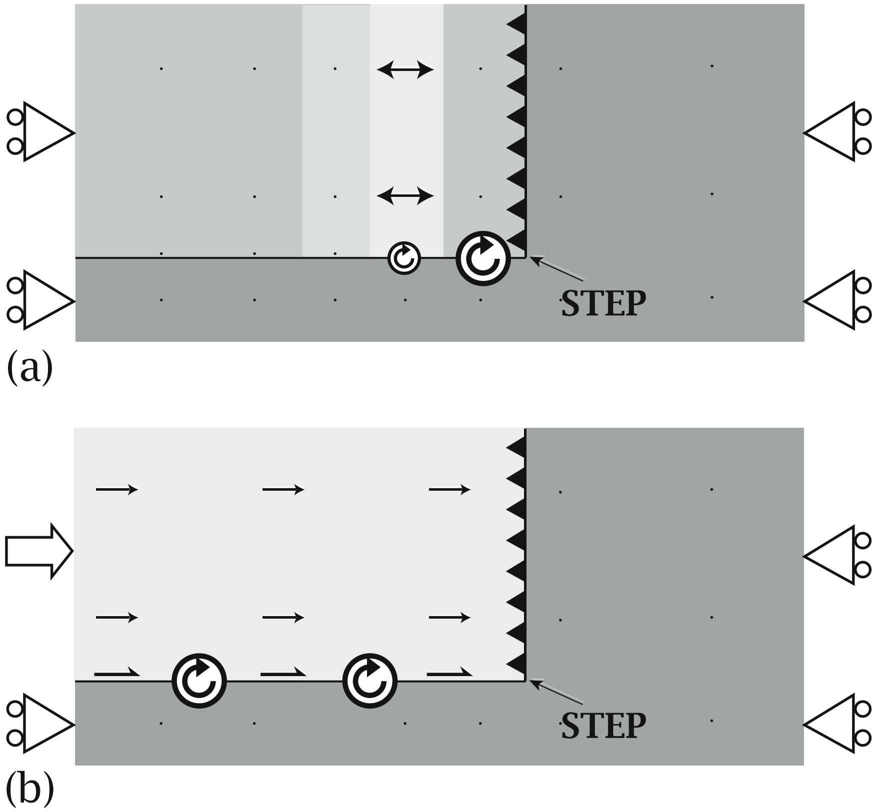


**Fig. 4** Schematic representation (map view) of two situations showing a STEP fault near the edge of a subducting slab, in combination with the motion or deformation of the overriding plate. Subduction and subduction direction are indicated by black dents. Left and right are taken to correspond with west and east, respectively. The NW part represents the overriding plate. (**a**) The left side boundary condition is: no "eastward" motion (indicated by rollers). Subduction of (oceanic) lithosphere is accompanied by eastward trench retreat (roll-back) and back-arc extension (light areas in overriding plate). The region of extension may migrate with the retreating trench, towards the right/east. The STEP fault shows dextral motion, not uniform along the STEP fault, possibly in combination with CW rotations. (**b**) At the left boundary a velocity boundary condition applies, implying that the overriding plate moves relative to the subducting plate. If the relative motion equals the trench retreat rate, no intraplate deformation in the overriding plate occurs. This is rigid body advection of the overriding plate, and the STEP fault acts like a regular transform fault, with dextral sense of motion

tions in magma composition (Peccerillo, 2005) support the scenario developed. However, we consider this as not yet convincing in view of the lack of spatial resolution.

### *3.2 Tonga-Vitiaz Trench Region*

Alternatively, we can test the scenario in a completely different tectonic environment, with a similar starting situation. Collisional (arc-continent) events must have occurred in many regions in the geological history. Preferably, such a test is to be carried out in a region which experienced collision recently enough to show observational evidence concerning the evolution of the process. There are several reasons for selecting the Tonga-Fiji region as an excellent candidate for such a test: among these are (1) The starting situation is very similar to the one in the western-central Mediterranean; a long uninterrupted trench system, north-northeast of Australia (Hall, 2002; Schellart et al., 2006); (2) Magnetic anomaly data allow reconstruction of the complicated recent history of back-arc extension, and last but not least, (3) In this region the presence of STEP faults is demonstrated in a uniquely convincing way by a detailed seismicity and focal mechanism study of Millen and

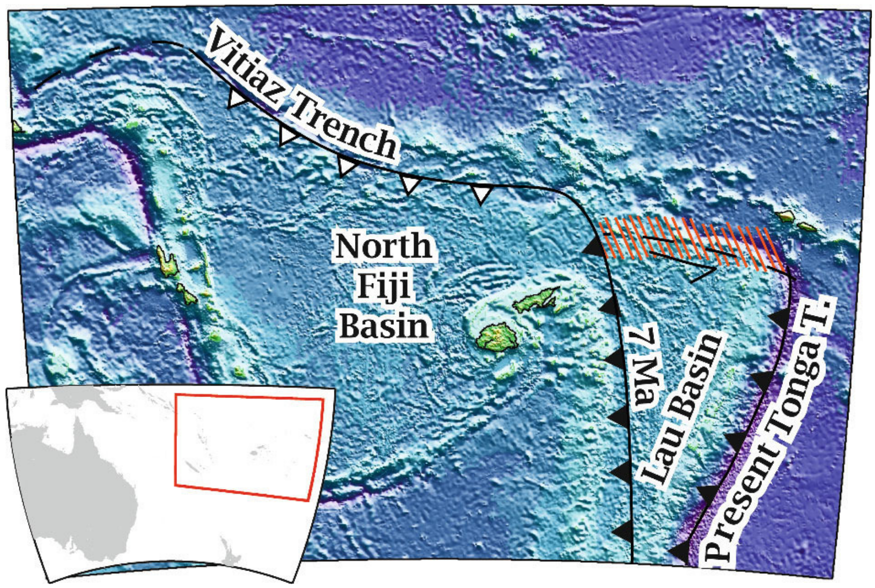


**Fig. 5** Regional topographic/bathymetric map for the Tonga-Vitiaz Trench region (see inset for location of the main figure within the larger southwest Pacific region). It shows the approximate location of the convergent plate boundary in this part of the SW Pacific in the Middle–Late Miocene and the present location of the Tonga Trench. White dents indicate presumed detachment of subducted slab, black dents indicate continuous slabs. The STEP fault is shown as a deformation zone, with an overall sinistral sense of shear

Hamburger (1998): A narrow band (~40 × 80 km) of normal faulting events ("hinge faulting", along vertical fault planes) with focal depths in the range 18–88 km provides direct evidence for tearing of the Pacific plate.

The region shown in Fig. 5 encompassing the Tonga-Fiji, the New Hebrides and the Vitiaz trench is a complicated segment of the plate boundary between the Pacific plate and the (Indo)-Australian plate. The sum of the back-arc spreading rates and the motion of the Pacific plate produces the world's highest subduction rate (~24 cm/year, Bevis et al., 1995; Doglioni et al., 2007), in the Tonga trench.

Figure 6 shows horizontal sections from the seismic tomography model BS2000 (Bijwaard and Spakman, 2000). In view of the high subduction velocity at the Tonga Trench effects of recent slab segmentation are to be found at greater depths than in the Calabrian subduction zone. A distinct slab edge is visible at the northern limit of the Tonga trench. Note the difference in orientation between the slab anomaly at 870 km depth and the Tonga slab anomalies in the upper mantle at 420 and 330 km depth. Another conspicuous feature is the distribution of deep earthquake hypocentres, not only in the very active Tonga subduction zone, but also outside the Tonga slab, in the northwestern part of the region displayed (see Fig. 6, section at 630 km depth).

In a reconstruction of the regional evolution Hall (2002) shows a long trench system, the Melanesian Arc, which was subject to major reorganization, probably in Miocene time. Hall (2002) identified the collision of continental type of lithosphere (the Ontong Java Plateau) with the Melanesian Arc as the likely cause of the plate boundary reorganization. Whether this collision affected the region of the present-day Vitiaz trench (Fig. 5) or rather the Solomon Islands region more to the west is uncertain. A similar analysis was made by Schellart et al. (2006) who distinguished a smaller continental type of block to the east of the Ontong-Java Plateau, the Melanesian Borderland, which in his reconstruction did collide with the trench system at the location of the present-day Vitiaz trench, in the Middle Miocene (~10 Ma).

In both reconstructions active motion of the Pacific plate brought a continental block (Melanesian Borderland, Ontong Java Plateau) into collisional contact with part of the trench system. Presumably, continent-arc collision of this type led to termination of the convergence and subsequent slab detachment in that particular part of the trench, now called Vitiaz Trench (Fig. 7a). Several studies have proposed a fossil subduction origin for the bathymetric feature of the Vitiaz trench (Pelletier and Auzende, 1996, and references therein). Subsequent development of a

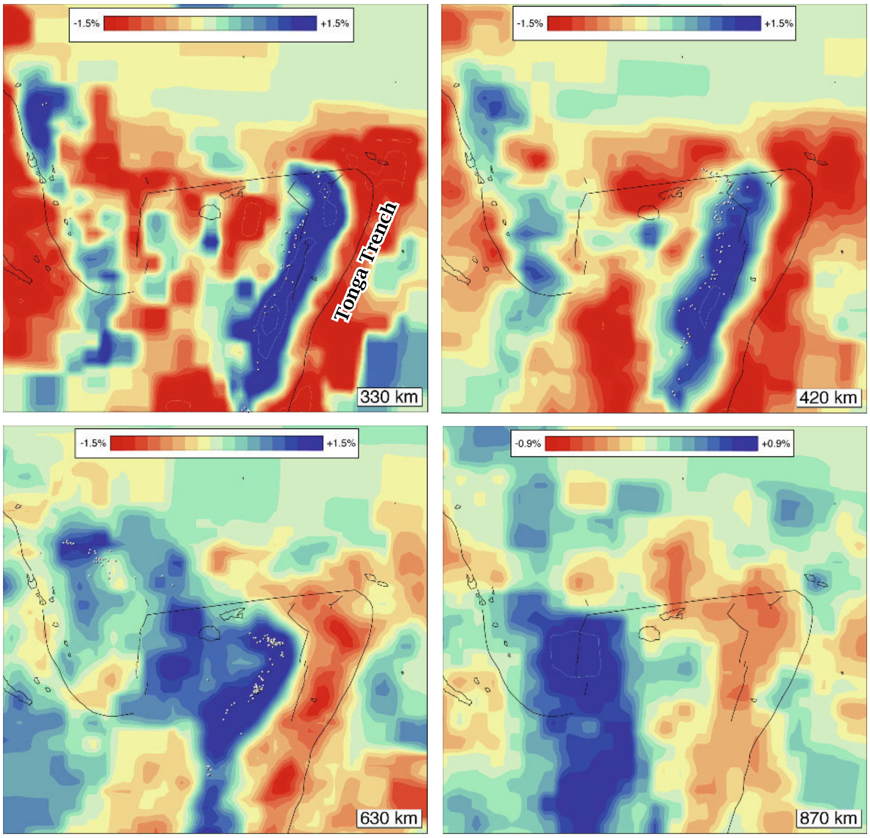


**Fig. 6** Horizontal sections through the BS2000 seismic tomography model of Bijwaard and Spakman (2000) for the Tonga-Vitiaz Trench region, at the depths indicated. P-wave velocity anomalies are given in percentages relative to the ambient depth-dependent velocity of the one-dimensional reference model ak135 (Kennett et al., 1995). White dots indicate earthquake hypocentres

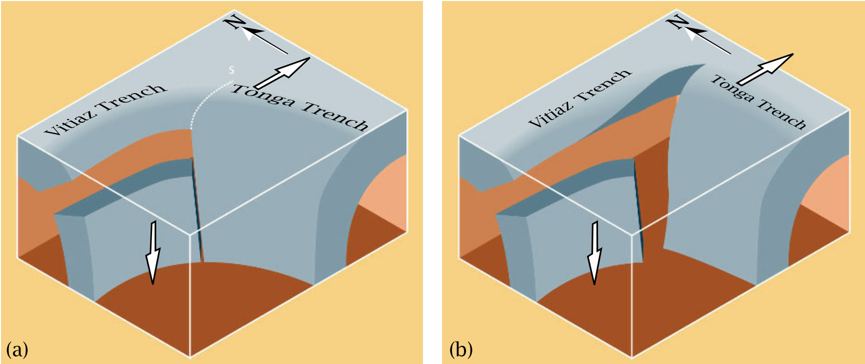


**Fig. 7** Schematic representation of the subduction zone geometry in the Tonga-Vitiaz Trench region. The model applies to the region approximately corresponding with that in Fig. 5. Vertical scale of the box is 300 km. a Detached slab after collision of continental type of plateau (Melanesian Borderland, Schellart et al., 2006) with trench system in the SW Pacific (the present-day Vitiaz Trench). A vertical tear separating the detached part from the continuous part is indicated. *Dotted line*, labelled S, indicates future STEP fault. b Formation of a STEP fault and subsequent trench retreat (roll-back) of the Tonga trench

free slab edge set the stage for STEP faulting to be initiated. This in turn allowed for the rapid eastward retreat of the Tonga slab (Fig. 7b), and the opening of the Lau Basin (see Fig. 5 for location). Recognizing that the resolution in the northern part of the 330 km depth section (Fig. 6) is not sufficient to draw direct conclusions from the tomographic images it appears that there is no anomaly in the uppermost mantle corresponding to a slab consumed in the Vitiaz Trench. The earthquake hypocentres at the base of the upper mantle (630 km depth section), however, are considered to occur in the remnants of lithosphere once consumed in the Vitiaz Trench (Chen and Brudzinski, 2001; Brudzinski and Chen, 2003). In combination this is indicative of detachment in the Vitiaz slab segment. The absence of high velocity anomalies to the west of the northern Tonga slab edge is in agreement with the presence of the STEP fault zone indicated in Fig. 5 The above noted change in orientation of the slab anomaly (Fig. 6) testifies to the rotation Tonga Trench. In combination with the Millen and Hamburger (1998) results on tear faulting in the STEP fault zone and the young age of the Lau Basin (start of spreading at about 5.5 Ma, with rifting starting a few Myr earlier [Parson and Hawkins, 1994; Schellart et al., 2006]), these inferences strongly support the evolutionary model in Fig. 7 which is identical to the one derived for the Sicily-Tyrrhenian region. In fact, the Fig. 7a and b displaying the Vitiaz-Tonga slab structure are exact mirror images of the ones in Fig. 3a and b for the Sicily-Tyrrhenian region.

We conclude that the Middle Miocene to Recent evolution of the Tonga-Vitiaz Trench region is in very good agreement with the evolutionary model derived from the Sicily-Tyrrhenian region. This is taken as an indication that the model is of more than only regional (western-central Mediterranean) validity.

## 4 Discussion

### 4.1 Arc Straightening and/or Increasing Curvature in Subduction Roll-Back

The free slab edge allows for sideways migration of asthenospheric and deeper mantle material around it, and consequently higher roll-back velocities relative to adjacent segments of the same trench system. In case there is only one STEP fault in the system the retreat of the slab will lead to a rotation of the plate boundary segment. If the starting configuration was arc-shaped, such a rotation leads to straightening of the convergent arc segment involved. On the other hand, if we consider the combination of the convergent segment and the STEP segment(s) of the evolving arc system it will appear to lead to an increase in arc curvature. This aspect is particularly clear in case of the Apennines, but it can also be recognized in the Tonga Arc (Fig. 7).

The Apennines-Calabrian arc evolution bears similarities with aspects of the Hellenic arc evolution in the eastern Mediterranean, where the African plate subducts beneath the Crete and the Aegean region. The southwestern arc segment is convergent and relatively straight, and the eastern part (from south of central Crete to Rhodes) represents the expression of a STEP type segment of the arc (Huchon et al., 1982; Govers and Wortel, 2005).

If a second STEP develops, the above straightening does not apply and an approximately translational type of motion of the arc segment between the two STEPS may occur, as proposed for the Gibraltar arc (Lonergan and White, 1997; Spakman and Wortel, 2004), the recent development of the Calabrian arc and the relative motion of the Caribbean plate relative to the South American plate (Govers and Wortel, 2005). For this 2-STEP case, arc curvature may further be increased by mantle flow around both slab edges (Schellart, 2004; Morra et al., 2006).

### 4.2 Other Regions

In Sect. 3.2 we have seen that STEP fault are not regional features of Mediterranean geodynamics only. Govers and Wortel (2005) have compiled a series of identified or proposed STEP faults in several other regions. Among these are the northern boundary of the Sandwich plate in the southern Atlantic, the northern and southern boundaries of the Caribbean plate, the northern and southern segments of the Carpathian arc, the eastern Hellenic arc, and the western edge of the north Sulawesi trench in Indonesia. In addition, Wortel et al. (2006) identified a STEP fault along the eastern part of the Cyprus arc.

In the Pacific realm, apart from the Tonga region, Schellart et al. (2003) identified roll-back in combination with STEP type of faulting in the Kuriles arc and back-arc region and investigated the STEP's near-surface expression by analogue modelling. In fact, the western Pacific Ocean provides a setting where STEP formation is to be expected because a basic condition for slab roll-back, gravitational instability of the oceanic lithosphere involved in subduction, is amply fulfilled in view of the predominantly Mesozoic lithospheric age. In this light, the tear in the subducting Izu-Bonin slab identified by Miller et al. (2004) may lead to a new arc segmentation phase in the region.

We expect STEP faults to be intrinsic tectonic elements in highly curved arcs, and anticipate that the number of identified STEP faults will further increase.

### 4.3 Arc Segmentation as a Repetitive Process?

Both in the Western Mediterranean and in the Tonga-Vitiaz Trench region we started from long continuous arcs, as postulated in regional studies. One may wonder whether perceptions of that type are the consequence of incomplete knowledge about early stages of arc development and subduction zone evolution. The postulated long arcs and/or the associated subducted slabs were possibly already segmented in early stages of the regional evolution, of which direct evidence was lost. Along these lines we hypothesize that the above evolutionary model applies to long arc segments which, upon growing by outward arc migration, become prone to collision and segmentation along the lines of the model considered. The new segments may again grow in length (as expressed at the surface) till new continental collisional events cause a repetition of the segmentation process. This may happen to a long trench system such as the Tonga-Kermadec trench, in the future, by the interaction with continental plateaus or ridges. The process we have considered in this paper, then leads to a new complication added to the earlier arc structure, and as such it may be a modular process within a repeating scheme.

### 4.4 Other Aspects of Terminal Stage Subduction Involving Slab Detachment

In this contribution we concentrated on continental collision and arc migration and back-arc extension. If continental collision results in slab detachment other expressions or accompanying processes are to be expected, notably in the field of vertical motions and magmatism. We briefly mention some pertinent studies.

Buiter et al. (2002) and Gerya et al. (2004) studied vertical motions by using an elastic rheology and a viscous rheology for the descending plate, respectively. In the former study the total magnitude of vertical motion was found to possibly amount to as much as 2–6 km. A striking result in the latter study was the long time span of the uplift process after the occurrence of slab detachment (~20 Myr).

Wortel and Spakman (1992) and in particular Davies and von Blanckenburg (1995) drew attention to the implications of slab breakoff for magmatic processes, with emphasis on the generation of alkaline magmas. For several other regions slab detachment has been proposed to account for special types of (post-collisional), usually K-rich magmatism, for example, for the Himalayas by Kohn and Parkinson (2002), for Central-Southern Italy by Wortel et al. (2003), and for Mexico by Ferrari (2004). De Boorder et al. (1998) have highlighted the association of mineralization zones with regions where slab detachment has been proposed. A crucial element in both magmatism and mineralization is the possible rise of asthenospheric material to unusually shallow depths after shallow slab breakoff.

## 5 Conclusions

We have identified a series of sequential process components for the evolution of a plate boundary in part of which continental collision occurs and combined them into a three-dimensional evolutionary model. Lateral (i.e., along strike of the trench) variation in nature of the lithosphere arriving at the trench, from continental lithosphere to oceanic lithosphere, sets the stage for the entire process. Continental collision triggers the process by causing slab detachment, and STEP fault formation is a key element in the further evolution. The various components represent stages in a continuous process, governed by the transient distribution in slab pull forces. As such, it is an evolutionary model for arc segmentation, which leads to strongly varying patterns of back-arc spreading.

Whereas we have derived the model for the Sicily-Tyrrhenian area, the Tonga-Vitiaz Trench region shows a strikingly similar evolution, which testifies to its more general validity. Hence, we expect that it will shed light on other convergent plate boundary regions with a history involving collision.

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