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Coupled and decoupled regimes of continental collision: Numerical modeling

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ABSTRACT

Useful geodynamic distinction of continental collision zones can be based on the degree of rheological coupling of colliding plates. *Coupled* active collision zones (which can be either retreating or advancing) are characterized by a thick crustal wedge and compressive stresses (i.e. Himalaya and Western Alps), while *decoupled* end-members (which are always retreating) are defined by a thin crustal wedge and bi-modal distribution of stresses (i.e., compressional in the foreland and extensional in the inner part of the orogen, Northern Apennines). In order to understand physical controls defining these different geodynamic regimes we conducted a 2D numerical study based on finite-differences and marker-in-cell techniques. In our experiments we systematically varied several major parameters responsible for the degree of rheological coupling between plates during collision such as convergence rate, crustal rheology and effective velocity of upward propagation of aqueous fluids and melts in the mantle wedge. Low convergence rates and fluids/melts propagation velocities favor continuous coupling and convergence between the plates. *Coupled collision zones* are characterized by continuous accretion of the weak upper continental crust resulting in the development of a thick and broad crustal wedge, by hot temperature in the inner parts of the orogen due to radiogenic heating of the thickened crust, by compressive orogenic stresses and appearance of a double seismogenic (brittle) layer involving upper crust and sub-Moho mantle. In contrast high convergence rates and fluid/melt percolation velocities produce efficient weakening of the mantle wedge and of the subduction channel triggering complete decoupling of two plates, mantle wedging into the crustal wedge and retreating style of collision. The evolution of *fully decoupled collision zones* are characterized by the disruption of the accretionary wedge, formation of an extensional basin in the inner part of the orogen and delamination of the weak portion of the continental crust that is first thrusted toward the foreland and, subsequently, dissected by extensional tectonics. Transition from coupled to decoupled regime occurs always at the early stages of continental collision indicating that insertion of rheologically weak crustal material in the subduction channel is critical for the subsequent evolution of the collision zone. We found good correlations of our numerical results with some of the major collisional orogens. In particular, the decoupled retreating collision regime reproduces what is observed in the Northern Apennines.

1. Introduction

Subduction zones are sites on the Earth's surface where gravitationally unstable plates descend into the mantle. Normally, old negatively buoyant oceanic plates are good candidates for such a behavior, though actively subducting young plates are sometimes observed at trenches (i.e., Cascadia and Nankai subduction zones). As the oceanic lithosphere enters the trench, a small part of its crust (mainly the weak overlying sedimentary cover) is scraped off and accreted at the overriding plate margin. The size of the accretionary prism is variable and depends on the convergence rate and the sediments flux from both the subducting and overriding plates (Clift and Vannucchi, 2004). However, with except of few cases (e.g., the Makran accretionary complex, northern Arabian sea, Kopp et al., 2000), the dimensions are limited by the small flux of weak sediments (normally 10% of the oceanic crust, Schmidt and Poli, 1998) lying atop of the incoming oceanic plate.

An alternative to this scenario is represented by subduction zones where, despite their intrinsic positive buoyancy, continents are colliding and a large amount of gravitationally stable and relatively weak material (the continental crust) is entering the trench causing slow down of the convergence that, eventually, may stop. However, before collision ceases, convergence between the plates can continue actively for tens of millions of years after ocean closure as it is testified by the 50 Ma active collisions in the Western Alps and in the Himalayas.

Active collision zones group in two distinct classes (Royden, 1993a): advancing margins, where the convergence rate induced by far-field stresses exceeds the subduction rate, and retreating

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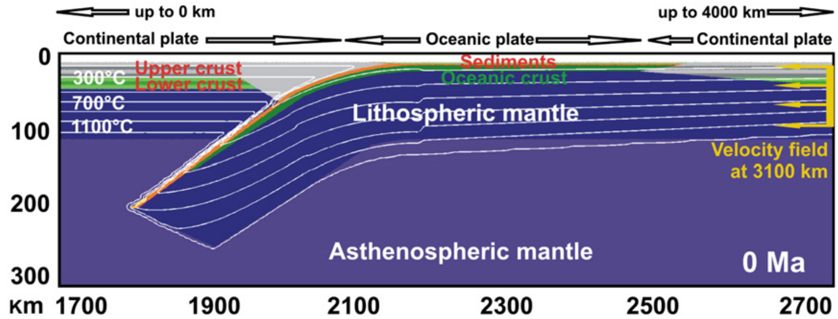

Fig. 1. Initial setup of numerical models: a) enlargement (1000×300 km) of the numerical box (4000×300 km). Free slip is imposed on all boundaries except the lower one that is permeable. Initial geodynamical setting is that of a young (5 Ma) oceanic subduction, with the right plate subducting under the left one. The prescribed plate convergence velocity is imposed at 3100 km. Color scale for different lithologies is described in Fig. 2. White lines are isotherms in °C.

boundaries, where subduction induced by slab pull is faster than plate convergence. Another useful geodynamic distinction of continental collision zones derived from common geological and geophysical data can be based on the degree of rheological coupling (i.e. interaction) of colliding plates. *Coupled* active collision zones are characterized by a thick crustal wedge and compressive stresses (i.e. Himalaya and Western Alps), while *decoupled* end-members are defined by a thin crustal wedge and bi-modal distribution of stresses (i.e., compressional in the foreland and extensional in the inner part of the orogen, Northern Apennines). While fully decoupled collision zones are strictly related to retreat and delamination of the lower plate, the same is not true for coupled collision zones that can form during either an advancing or retreating phase. As an example, recent tectonic reconstruction (Rosenbaum and Lister, 2004) and paleomagnetic data (Maffione et al., 2008) indicate that the Western Alps, a coupled active collision zone, formed in response to the retreat of the eastward subducting Eurasian plate during the Tertiary.

In the last two decades numerical simulations have been widely used to investigate coupled advancing collision (e.g. Beaumont et al., 2001; Burg and Gerya, 2005; Burov et al., 2001; Pfiffner et al., 2000; Pysklywek, 2001) and significant progress has been achieved in our understanding of this major orogenic regime. More recently, then, numerical and analog models focused on dynamics of trench motion (e.g., Di Giuseppe et al., 2008; Faccenna et al., 2007) and tectonic implications arising from a foregoing oceanic subduction (such as slab delamination, slab breakoff and metamorphism) and regarding post-subduction styles of advancing collision (e.g. Faccenda et al., 2008; Regard et al., 2003; Warren et al., 2008).

Decoupled retreating continental subduction was first investigated by Bird (1979), as a mechanism referred to continental delamination. Since then, no detailed numerical study has been done on the evolution of such a system. Beaumont et al. (2006) showed that retreating subduction is promoted by a relatively dense slab, while other studies refer of an increase of plates' decoupling proportional to the imposed viscous strain weakening or related to the presence of a weak subduction channel (De Franco et al., 2006; Warren et al., 2008).

In the present paper we display results of a series of thermo-mechanical models where *rheological weakening* of the plates' boundary induced (i) by propagation of aqueous fluids and melts in the mantle wedge and (ii) by shear heating along the plate interface is acting as the major process affecting degree of plates' coupling and consequently the style of continental collision.

In order to differentiate the development and the structures of both coupled and decoupled collision zones, we further give a description of the major characteristics of the two collisional end-members and, because of the gap left in modeling decoupled retreating collision, we discuss in details such type of model and constrain the results by means of a comparison with the Northern Apennines.

2. Numerical model

2D numerical simulations were performed with I2VIS, a thermo-mechanical code based on a combination of the finite-difference method applied to a staggered grid and with nondiffusive-marker-in-cell technique (Gerya and Yuen, 2003). Realistic visco-plastic rheologies are used for all the lithologies (e.g. Gerya and Stöckhert,

Table 1
Description of numerical experiments

Model	Upper continental crust rheology[a]	Convergence rate (cm/yr)	$v_{y(\mathrm{fluid})}$ (cm/yr)[b]	$v_{y(\mathrm{melt})}$ (cm/yr)[c]	Collision style[d]
1	WQ	2	0.01	0.1	c
2	WQ	3	0.01	0.1	c
3	WQ	5	0.01	0.1	c
4	WQ	2	0.1	0.1	c
5	WQ	3	0.1	0.1	c
6	WQ	5	0.1	0.1	c
7	WQ	2	1	0.1	c
8	WQ	3	1	0.1	c
9	WQ	5	1	0.1	c
10	WQ	2	10	0.1	c
11	WQ	3	10	0.1	d
12	WQ	5	10	0.1	d
13	WQ	2	0.01	1	c
14	WQ	3	0.01	1	c
15	WQ	5	0.01	1	c
16	WQ	2	0.1	1	c
17	WQ	3	0.1	1	c
18	WQ	5	0.1	1	c
19	WQ	2	1	1	c
20	WQ	3	1	1	d
21	WQ	5	1	1	d
22	WQ	2	10	1	d
23	WQ	3	10	1	d
24	WQ	5	10	1	d
25	WQ	2	0.01	10	c
26	WQ	3	0.01	10	c
27	WQ	5	0.01	10	c
28	WQ	2	0.1	10	c
29	WQ	3	0.1	10	c
30	WQ	5	0.1	10	d
31	WQ	2	1	10	c
32	WQ	3	1	10	d
33	WQ	5	1	10	d
34	WQ	2	10	10	d
35	WQ	3	10	10	d
36	WQ	5	10	10	d
37	PLAG	2	0.1	10	d
38	PLAG	2	0.1	0.1	d
39	PLAG	3	0.1	0.1	d
40	PLAG	5	0.1	0.1	d

[a] WQ = Wet Quartzite (Ranalli, 1995); PLAG = Plagioclase An_{75} (Ranalli, 1995).
[b] Effective vertical propagation velocity of the fluid phase.
[c] Effective vertical propagation velocity of the melt phase.
[d] c = coupled collision; d = decoupled collision.

2006; Gorczyk et al., 2007). Since mechanisms for subduction initiation are still poorly known, the starting geometry of the computational domain (4000×300 km, Fig. 1) is that of an early subduction zone where 300 km of the slab has entered in the trench. An 800 km long, spontaneously bending oceanic plate lies between the fixed overriding continental plate (on the left) and the attached incoming continental passive margin (on the right). A prescribed convergence rate is imposed for around 1000 km, after which the subducting plate is allowed to move freely underneath the left continental upper plate.

Continental plates are made of a relatively wet and weak upper crust (20 km thick, Wet Quartzite flow law, Ranalli, 1995) overlying a dry and strong lower crust (15 km thick, Plagioclase An_{75} flow law, Ranalli, 1995), while the oceanic plate has an 8 km thick and relatively strong crust (Plagioclase An_{75} flow law, Ranalli, 1995) with a fixed serpentinization degree (1.5 wt.% water, Schmidt and Poli, 1998). 2 km of sediments (7.6 wt.% water in hydrous minerals and 2 wt.% of porous water) are placed on top of the oceanic crust.

The model accounts for shear heating that has been shown to be an important component of the overall heat budget (Burg and Gerya, 2005; England et al., 1992; Molnar and England, 1990). The degree of rheological coupling between the two plates defined by the effective strength of rocks composing the plate interface is inversely proportional to the amount of heat produced at this shearing interface that, in turn, depends on the convergence velocity (Faccenda et al., 2008).

A thermodynamic database is implemented (Gerya et al., 2006) in the code and accounts for phase transformations, de-hydration reactions and partial melting of the used lithologies. The starting mantle is considered to be formed by anhydrous peridotite, whereas the mantle above the subducting slab can become hydrated or partially molten because of fluids released from the de-hydrating oceanic crust. The models consider decrease in density and rheological strength due to hydration and partial melting of all the lithologies. As

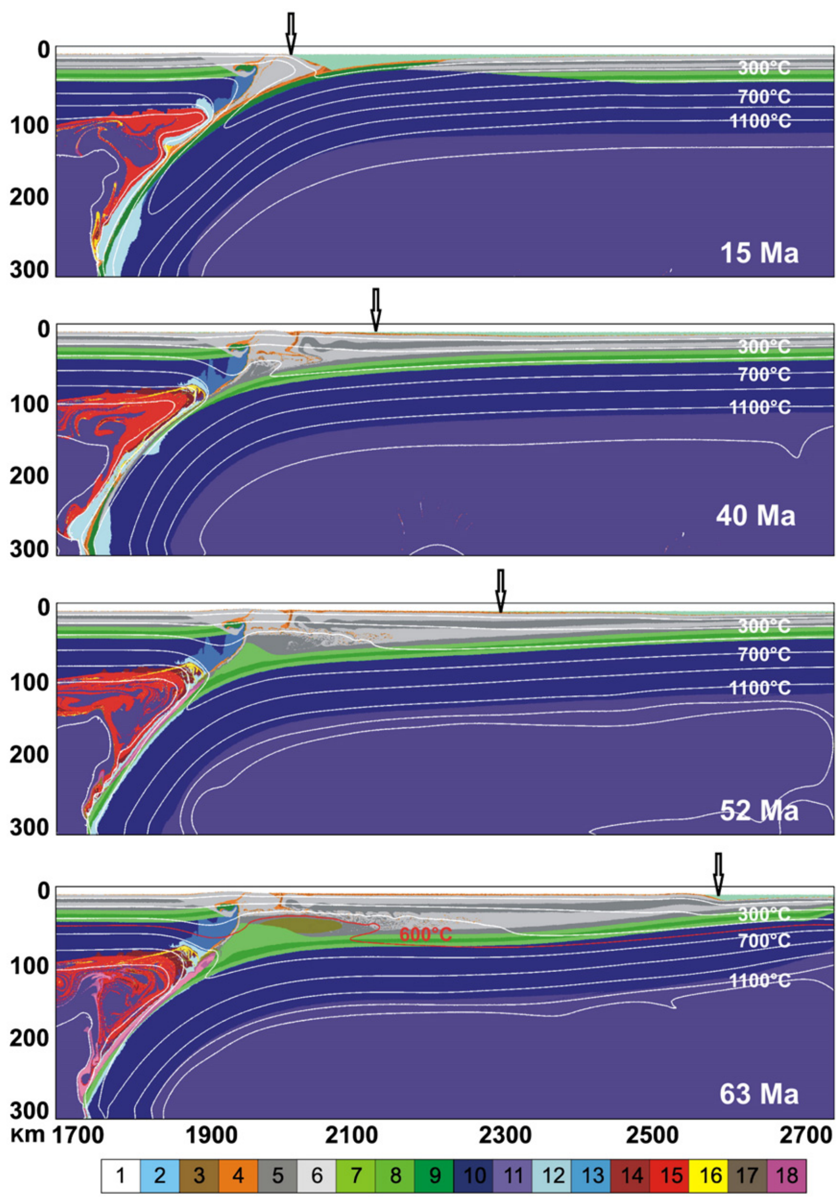

Fig. 2. Development of the coupled collision reference model (Model 19 in Table 1). The white lines are isotherms in °C. The red line at 63 Ma is the 600 °C isotherm. Arrows on top of each snapshot show position of the coastline. Color scale: 1 = air; 2 = sea water; 3,4 = stratified sediments; 5,6 = stratified upper crust; 7,8 = stratified lower crust; 9 = oceanic crust; 10 = lithospheric mantle; 11 = asthenospheric mantle, 12 = hydrated mantle; 13 = serpentinized mantle; 14 = mantle quenched after partial melting; 15 = partially molten mantle; 16,17 = partially molten felsic rocks (sediments, upper crust); 18 = partially molten lower crust.

the subducting slab has a constant serpentinization degree and a fixed length, the amount of hydrated or partially molten (and, thus, less viscous and more buoyant than their anhydrous counterpart) mantle wedge rocks will depend on the rates at which aqueous fluids and melts can propagate (by various mechanisms) through the anhydrous peridotite. Because of uncertainties in our knowledge of physical parameters controlling water transport above slabs, we used prescribed effective upward propagation velocities of aqueous fluids and melts as two independent model parameters and varied them systematically in our numerical experiments.

A detailed description of the modeling and numerical techniques employed in our study is given in the Supplementary Information section.

3. Results

Numerical experiments were designed to cover systematically a range of convergence settings with different convergence velocities and vertical percolation rates of aqueous fluids and melts in the mantle wedge (Table 1). In a few additional cases the strength of the

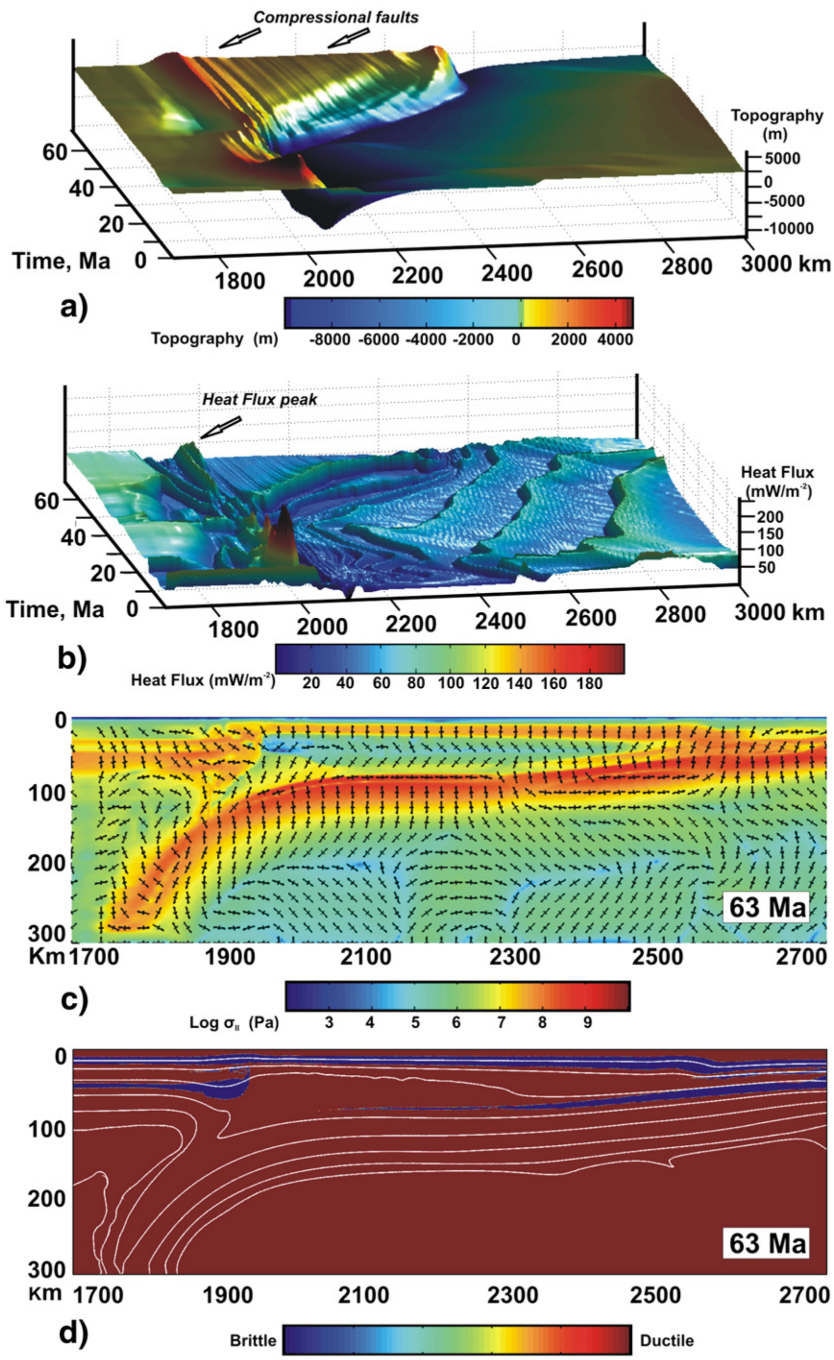

Fig. 3. Analysis of the coupled collision reference model (Model 19, Fig. 2). a) Topographic profile evolution with time. After collision, a wide plateau forms limited by a high mountain chain. b) Heat flux profile evolution with time. The peak of heat flux corresponds to the inner part of the orogen. c) Distribution of the second stress invariant at 63 Ma. Color scale corresponds to the magnitude. Long and short axes indicate orientations of minimum (extension) and maximum (compression) principal stresses, respectively. d) Rheological behavior of rocks at 63 Ma. Red and blue domains correspond to dominating viscous and brittle deformation, respectively.

continental upper crust was varied to constrain the role that this layer plays in determining the style of post-subduction collision.

We describe first the general development of the oceanic subduction phase; subsequently, a description of models with different parameters is given in order to characterize two (i.e., coupled and fully decoupled) post-subduction collisional regimes identified by our study and the role that the physical variables have in controlling these regimes.

3.1. Initial development: oceanic plate subduction

The early phases of the models consist of oceanic plate consumption and ocean closure (Figs. 2, 5, and S1–S3). As the serpentinized oceanic slab subducts into the mantle, the oceanic crust is heated up and de-hydration reactions take place. The amount of water in subducting oceanic slab is the same for all models. However, the size of mantle wedge areas affected by hydration and partial melting (that, in terms of rheology, means a reduction of the strength of the material) varies according to the specific fluid and melt propagation velocities.

In general, models with low fluid/melt velocities have a limited width of hydrated low viscosity mantle wedge, while models with high velocities develop broad hydrated and partially molten regions below the overriding plate and in proximity of the plates contact.

3.2. Continental collision

The collisional regime is greatly affected by the buoyant and rheologically weak materials entering the subduction channel from above (i.e., upper continental crust) and from below (i.e., hydrated and partially molten mantle). Two different post-subduction collisional regimes have been obtained in our numerical experiments: 1) *coupled* collision, characterized by continuous coupling and convergence of the plates and 2) *decoupled* collision, dominated by post-subduction full plates' decoupling and by retreat of the subducting continent at a faster rate than the imposed convergence (Table 1).

Coupled collision occurs for low convergence rates and slow percolation velocities of fluids and melts (e.g. Model 19, Figs. 2–4), while decoupled collision is triggered by high convergence rates and by enhanced fluid and melt percolation velocities that produce large hydrated and partially molten low viscosity areas in the mantle wedge (e.g. Model 34, Figs. 5–7) (Fig. 8). Transition from coupled to decoupled style of convergence always occurs at the early stages of continental collision, indicating that subduction of the upper, weak and relatively buoyant continental crust plays a major role in modifying plates' interaction.

3.2.1. Model 19: reference model of coupled collision (Figs. 2–4)

At the onset of collision (Fig. 2, 40 Ma), partially molten and hydrated rocks of the mantle wedge affect the base of the plates' contact. However, coupling between the two plates is still significant because: 1) low fluid/melt propagation rates of 1 cm/yr create relatively small hydrated mantle portion in the cold tip of the mantle wedge, and 2) the low convergence rate (2 cm/yr) generates minor amounts of shear heating and associated thermal weakening (Fig. S4a). As collision continues (Fig. 2, 52 Ma), the weak (upper) part of the continental crust is accreted at the margin, while large fractions of the stronger lower crust are subducted and recycled into

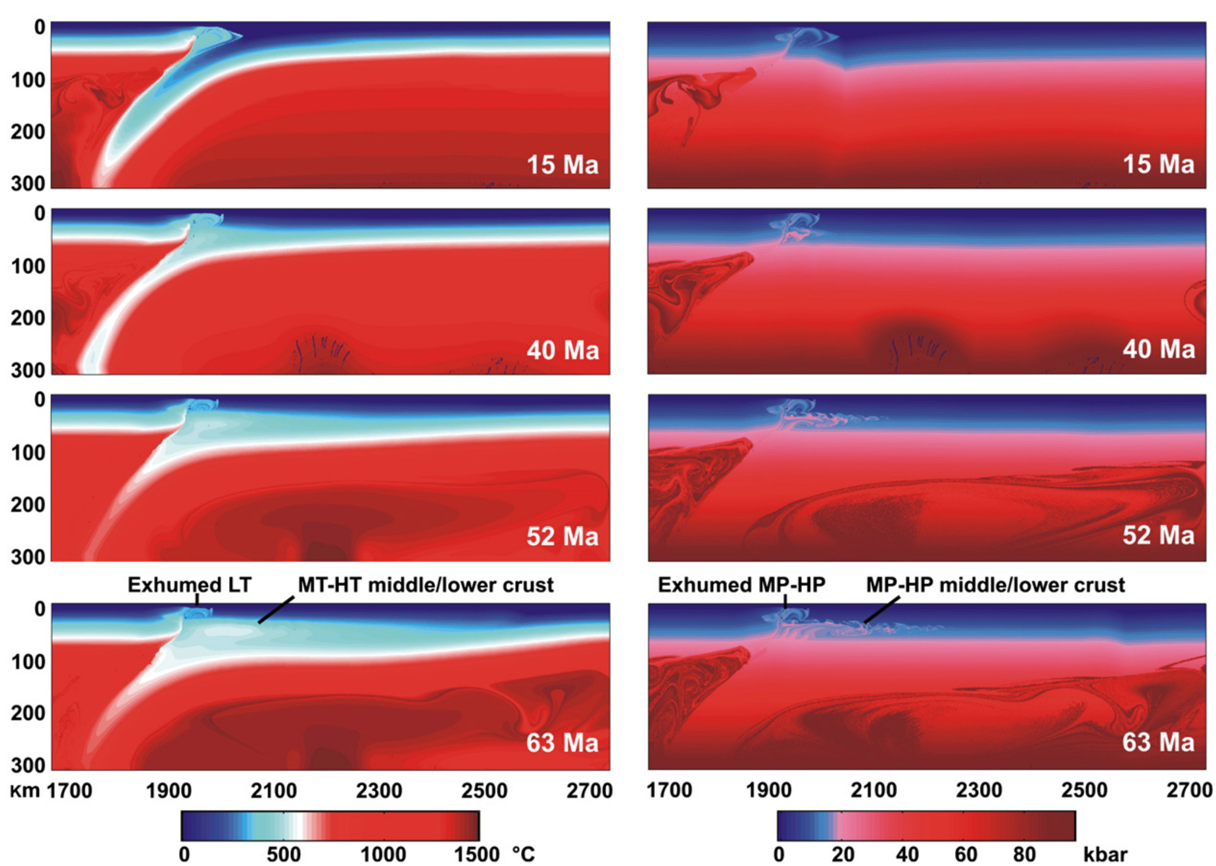

Fig. 4. Evolution of metamorphic zoning for the coupled collision reference model shown in Fig. 2 (Model 19 in Table 1). Left column: distribution of peak temperature conditions. Right column: distribution of peak pressure conditions. Exhumation of LT-HP rocks at the suture zone occurs mainly before collision. Also, the middle/lower crust is composed of a medium-to-high grade rock units derived from different depths.

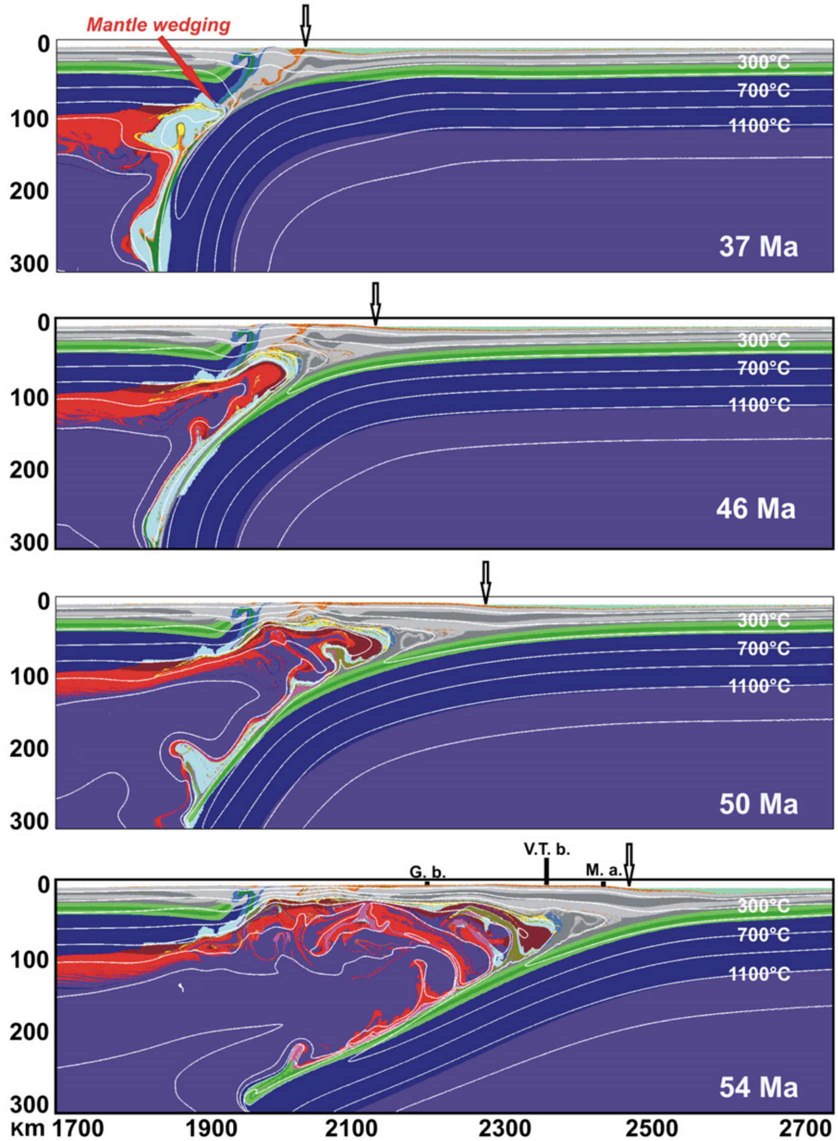

Fig. 5. Development of the decoupled collision reference model (Model 34). Isotherms, color code and arrows are the same as in Fig. 2.

the deep mantle. Crustal stacking produces the progressive rightward migration of the trench and the development of an elevated plateau inward to the frontal mountain range (Fig. 3a).

The final structure of this coupled collisional zone is characterized by a broad and thick crustal wedge overlying a subducting plate with a flat-ramp-flat geometry (Fig. 2, 63 Ma), by stresses that are mainly compressive with minor extension due plate bending (Fig. 3c) and by a double seismogenic (brittle) layer expression of the upper crust and of the sub-Moho mantle (Fig. 3d). Radiogenic heating of the thickened crust increases the temperature above the crustal rocks solidus producing partial melting and high heat flux in the inner parts of the orogen (Fig. 2, 63 Ma). Elsewhere, heat flux measured at the surface is low because of frontal accretion of cold upper crustal slices (Fig. 3b).

Pre- and syn-collisional exhumation of LT-HP rocks occurs at the active margin due to corner-flow-like mass circulation in the accretionary prism (Fig. 4, 15 and 40 Ma). Crustal thickening and heating triggers the formation of a medium to high metamorphic grade crustal channel propagating laterally toward the foreland (Fig. 4, 52 and 63 Ma).

3.2.2. Model 34: reference model of decoupled collision (Figs. 5–7)

This model differs from the previous one (Model 19) by high effective upward fluid/melt propagation rates of 10 cm/yr. This translates in a more efficient hydration and weakening of the mantle wedge above the subducting slab.

During incipient collision (Fig. 5, 37 Ma), a large volume of weak crustal material is interposed between the plates. Chemically buoyant hydrated and partially molten mantle produced during oceanic subduction stage wedges between the plates because of: 1) the easily deformable subducted upper continental crust, sited at the plates' contact, and directly connected with the asthenosphere (Bird, 1979), 2) the loss of coupling due to overriding lithosphere hydration (e.g. Gorczyk et al., 2007) and 3) the negative pressure gradient caused by dense retreating slab that favors sucking of the mantle into the accretionary prism (Dvorkin et al., 1993) (Fig. 5, 46 Ma).

As lithospheric delamination proceeds (Fig. 5, 46 to 54 Ma), the rheologically weak part of the crust is scraped off from the retreating lithosphere by the stronger and denser asthenospheric mantle that acts as a backstop moving in the opposite direction to convergence. The removed crust is first thrusted over the foreland

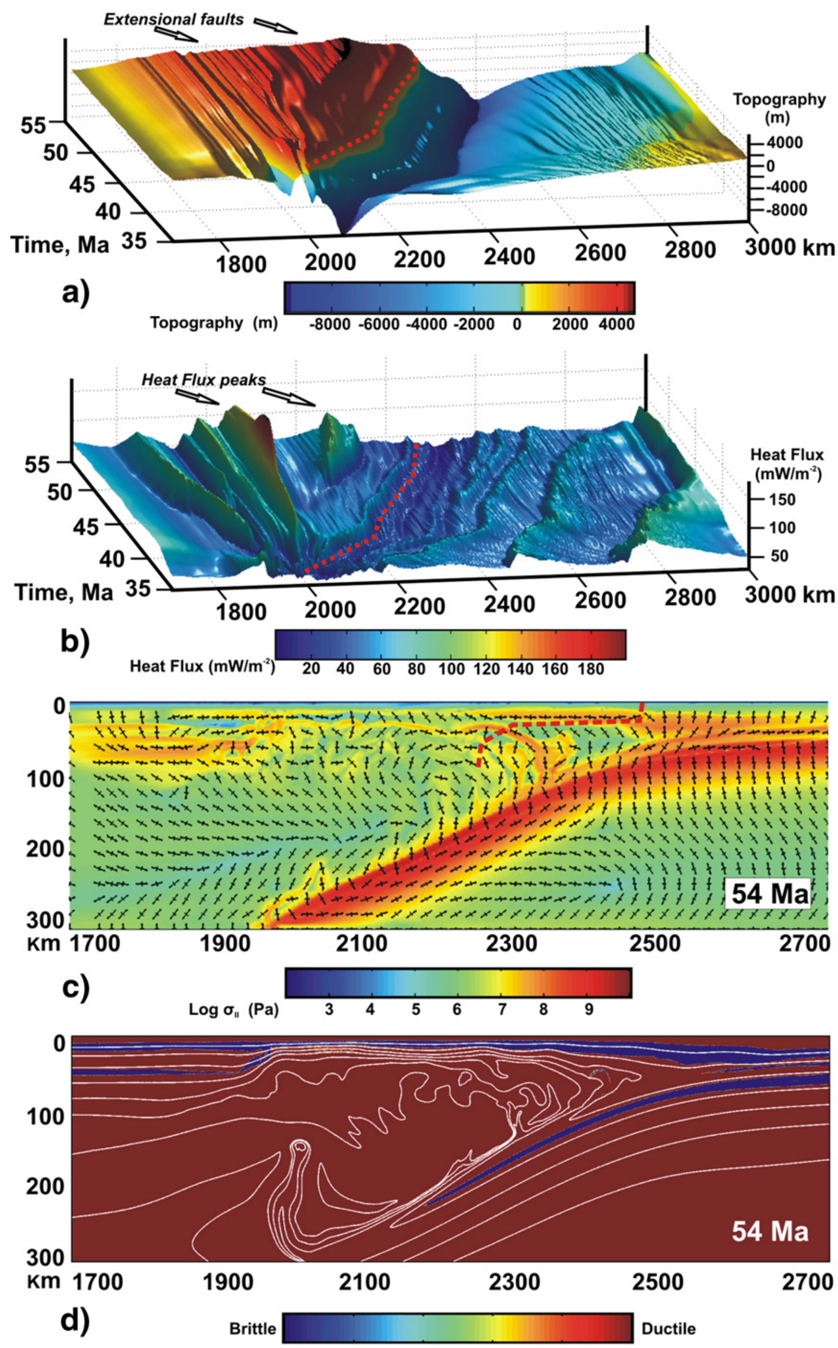

Fig. 6. Analysis of the decoupled collision reference model (Model 34, Fig. 5). a) Topographic profile evolution for the last 20 Myr. Major uplift is caused by emplacement of partially molten rocks at the base of the crust (see Fig. 5, 54 Ma). The red line is the surface expression of the extension/compression boundary evolving with time. b) Heat flux profile evolution for the last 20 Myr. Lateral shift of heat flux peaks can be associated with the eastward migration of volcanic centers as evidenced by the last 20 Ma geological history of the Northern Apennines (see Section 4.3.1). The red line is the surface expression of the extension/compression boundary shifting with time. c) Distribution of the second stress invariant at 54 Ma. Color scale corresponds to the magnitude. Long and short axes indicate orientations of minimum (extension) and maximum (compression) principal stresses, respectively. The red line is the boundary between the extensional and compressional domains. d) Rheological behavior of rocks at 54 Ma. Red and blue domains correspond to dominating viscous and brittle deformation, respectively.

and then, as the subduction zone is progressively shifted, undergoes extensional tectonics in the broadening basinal region (Fig. 6a). Moreover, shallow and partially molten mantle produces uplift and activates a high heat flux regime with peaks that can be associated with localized magmatic centers migrating toward the trench (Fig. 6a–b). As a result, the upper plate is composed of the pre-collisional accretionary prism, removed crustal slices and solidified magmatic bodies. The stress field at 54 Ma (Fig. 6c) is defined by two zones with distinct stress regimes: the foreland is a compressional domain with some extensional areas due to the plate bending, while the inner basin is dominated mainly by extension with some compression due to partial melting advection and intrusion. These two domains are separated by a transitional zone with extension at the surface and compression at depths.

Differently from the coupled collision model, the decoupled collisional orogen has single seismogenic (upper crustal) layer in the

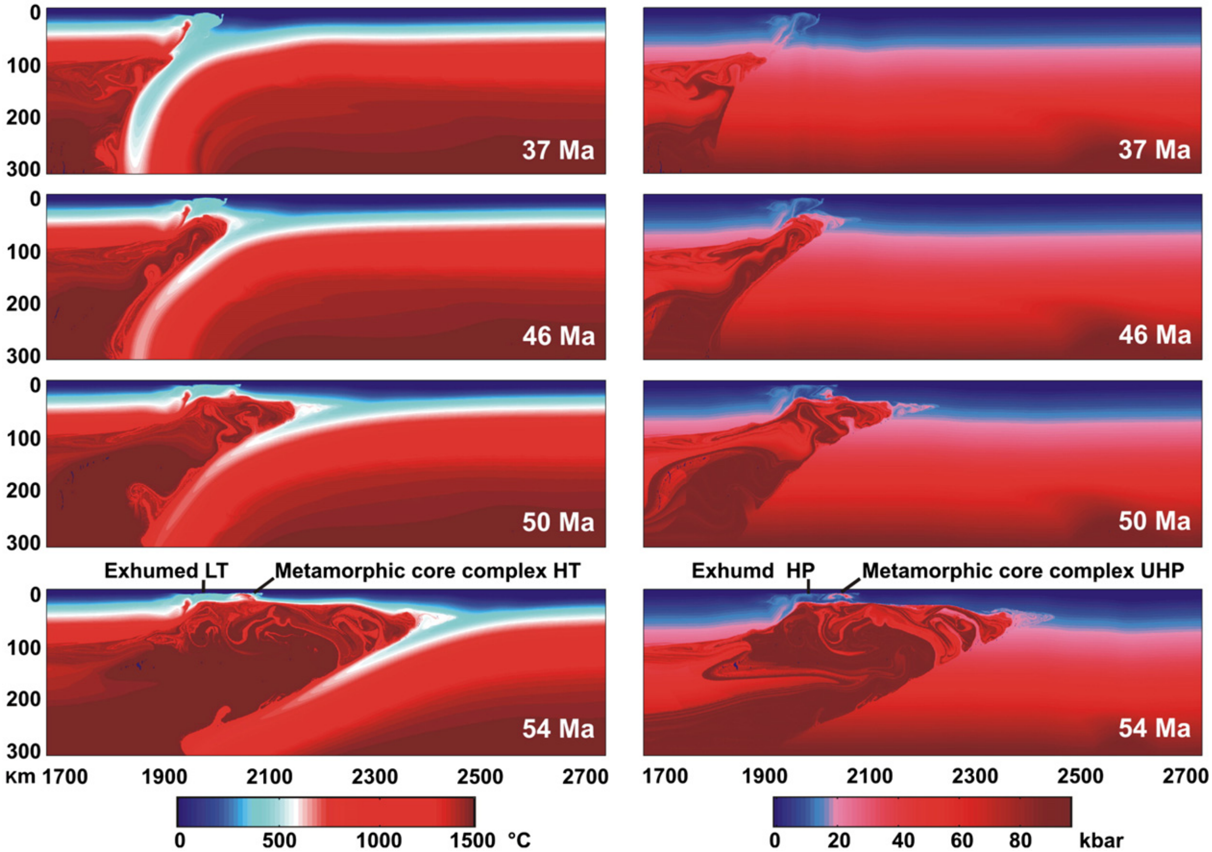

Fig. 7. Evolution of metamorphic zoning for the decoupled collision reference model shown in Fig. 5 (Model 34 in Table 1). Left column: distribution of peak temperature conditions. Right column: distribution of peak pressure conditions.

inner part of the orogen, while in the frontal part a second (sub-Moho) brittle layer is progressively subducted together with retreating plate (Fig. 6d).

Exhumation of LT-HP rocks at the active margin occurs mainly before collision due to corner-flow-like mass circulation in the accretionary prism (Fig. 7), while post-collisional retreating favors exhumation of a high grade metamorphic core complex made of deeply buried upper crustal rocks (Figs. 5 and 7, 37 and 54 Ma).

3.2.3. Model 21 (Fig. S1)

The present model differs from the coupled collision reference model (Model 19) by a higher convergence rate of 5 cm/yr.

Although fluid/melt propagation velocities are set to the same low values of 1 cm/yr, decoupling and retreating of the lower plate is triggered by shear heating and related thermal weakening of the subduction channel due to the higher convergence velocity (Fig. S1, 16 Ma). The amount of shear heat produced in the subduction channel reaches tens of μW/m^3 (Fig. S4b), while in models with lower convergence rates (Fig. S4a, Model 19) shear heating in this area amounts only a few μW/m^3. Resulting modification of the thermal structure of the incipient collision zone is also quite significant (on the order of 100 °C, cf. isotherms in Fig. S4a and b) and is consistent with previous numerical estimates of Burg and Gerya (2005). Because an effective channel viscosity is inversely proportional to the strain rate

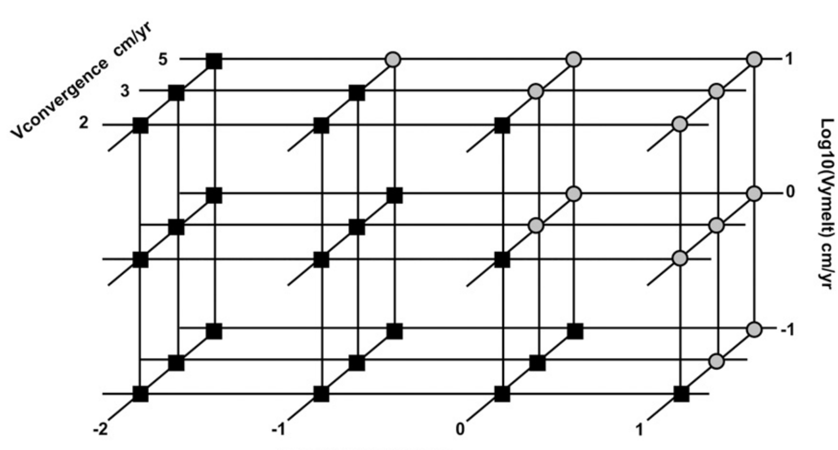

Fig. 8. Styles of post-subduction continental collision as a function of convergence rate and upward propagation velocities of aqueous fluids and melts. Black squares: coupled collision. Grey circles: decoupled collision.

and temperature, enhanced strain rates and shear heating along the plate interface lower viscosity of the channel which in turn triggers wedging of chemically buoyant partially molten hydrated asthenospheric mantle between two plates.

Our results (Fig. S4) indicate that the subduction channel forming atop the slab is notably (by around 100 °C) hotter in models with higher convergence velocity because of shear heating. In contrast, the thermal structure of the subducting plate interior in these models is colder because of the faster down-ward advection of isotherms. Therefore, rapidly subducted upper crust is generally colder and stronger than the crust subducted in slowly converging collision zones. Consequently, this strong upper crust can be subducted much deeper and dehydrate at asthenospheric depths with the consequence that extensive partial melting occurs in the hydrated mantle wedge above the slab (compare Model 34, Fig. 5, 54 Ma with Model 21, Fig. S1, 24 Ma). This could in turn trigger wedging of the hydrated and/or partially molten mantle in the hot and weak subduction channel and further into the crustal wedge.

The extreme case of a very stiff and deeply buried upper crust is presented in Section 3.2.5 (Model 40, Fig. S3).

3.2.4. Model 6 (Fig. S2)

This model differs from the previous one (Model 21) by lower upward fluid/melt propagation velocities of 0.1 cm/yr (Table 1) and from the coupled collision reference model (Model 19) by higher convergence rate of 5 cm/yr.

General development of this model is the same as for Model 21 except that at the onset of collision (Fig. S2, 16 Ma) smaller amounts of hydrated and partially molten material are present in the mantle wedge. As part of the wet continental crust is being subducted, more water is brought into the upper mantle and more partially molten rocks are produced (Fig. S2, 22 Ma). However, the final orogen is that of a coupled collisional regime because weakening of the plate's contact *from below* is limited by the slow percolation velocities of fluids and melts.

3.2.5. Model 40 (Fig. S3)

This model has the same parameters as the last one (Model 6) but differs from all the others by rheologically “stronger” (Plagioclase An_{75} flow law, Ranalli, 1995) upper continental crust.

When the subducting continent goes into the trench, strong upper crust is not accreted to the margin but is subducted to asthenospheric depths releasing water and undergoing extensive partial melting because of the lowered solidus temperature compared to mafic rocks (Fig. S3, 16–20 Ma). Partially molten subducted crust establishes a weak interface between two plates (Fig. S3, 20 Ma). This in turn triggers plate decoupling and onset of retreating collision mode associated with delamination of the buoyant continental crust from subducting plate. We note also that the amount of retreat is less than in “normal” models with rheologically weak upper continental crust.

4. Discussion and comparison of model results with collision zones

3D area diagram in Fig. 8 shows influences of three major model parameters (convergence rate and aqueous fluid/melt propagation velocities) for determining regimes of post-subduction collision zones in convergent margins (Fig. 8). By analyzing this diagram the following important tendencies can be observed.

During oceanic plate subduction, the leading edge of the upper continental plate is slightly up-warped and the friction in the subduction channel is lowered by the expulsion of porous water from the sediments above the oceanic crust. Here we note that upbending of the upper plate margin is minor compared to models without wet sediments (e.g. Faccenda et al., 2008), and that the intensity of deformation depends on the friction between the two plates that, in turn, is a function of the size of the wetted interface. Hence, high porous water contents or thick sedimentary layer above the oceanic crust translate into less friction in the subduction channel and less deformation of the continental plate margin. For instance, aseismic subduction margins are often characterized by a thick sedimentary cover lying on top of the subducting oceanic crust (e.g. Makran, Clift and Vannucchi, 2004; Kopp et al., 2000).

High fluid and melt percolation rates produce extensive rheological weakening at the base of the plates' contact and, thus, favor mantle wedging at the onset of collision. Considering the uncertainties about the efficiency of fluid phase movement and the degree of hydration and melting in the upper mantle, in nature weakening of the mantle wedge is, to a first order, directly proportional to the degree of serpentinization and the length of the subducted oceanic slab and inversely proportional to the efficiency of melt extraction processes. For example, an overall amount of aqueous fluids released in the mantle wedge and subduction channel in a long-lived Andean-type margin will be notably higher compared to our models where two continents are separated by a relatively short slab (1100 km). On the other hand, our simulations do not contemplate melt extraction processes that would cause a decrease of partially molten and rheologically weakened regions. In a mature Andean-type margin, steady-state conditions between continuous dehydration and melt extraction that act with opposite effects in terms of rheological weakening of the mantle wedge will be approximately reached before continental collision. In our experiments, the amount of water released by the oceanic plate in the mantle wedge is the same for all the models because before the slab leaves the computational domain at 300 km of depth all dehydration reactions take place due to the warming of the slab, even in models with fast convergence rates. Hence, the non-steady variations in amount of hydrated and partially molten regions present at the onset of collision in our models are mainly controlled by the imposed fluid/melt percolation velocities. These non-steady variations may, indeed, have similar geodynamic effects as different degrees of steady-state rheological weakening of the mantle wedge in mature collision zones.

High convergence rates also promote decoupled collision regime because of the enhanced strain rates and shear heating lowering effective viscosity in the subduction channel and due to the deep subduction of relatively cold and wet upper continental crust resulting in strong hydration and melting of the mantle wedge. The role of convergence rate in collisional margins characterized by high coupling (no fluids included) was already described by Faccenda et al. (2008), where post-subduction collisional styles are greatly affected by the amount of shear heating produced on the plates' interface. When aqueous fluids are included as in the present study, the effect of convergence velocity is less important because the amount of shear heating is smaller in a weak hydrated compared to a strong (anhydrous) subduction channel. Furthermore, Gorczyk et al. (2007) showed that in Andean-type margins plates' coupling is directly proportional to the convergence rate because of the thinning of continuously formed hydrated layer atop the slab. In contrast, our experiments (for example, see Model 19, Fig. 2, and Model 21, Fig. S1) indicate that in collisional settings, where a thick subduction channel is present, higher rates of convergence promote plates' decoupling and retreating style of collision. The fact that decoupled retreating collision is favored when the convergence rate is high appears counterintuitive. The Himalayan collision zone is characterized not only by high velocity of subducting plate, coupling and advancing collision style, but also by retreat of the overriding plate at half of the subducting plate velocity (in contrast to our models where the overriding plate is fixed), resulting in a net low convergence rate between the plates of 2.5 cm/yr (Guillot et al., 2003). Despite the fact that this observation is consistent with our results, we stress that the final style of post-subduction collision is determined by the overall interaction of the numerous processes acting in the subduction/collision system rather than by a single parameter. In particular, we note that the application of this result to natural systems with slab pull driven subduction, where the amount of mantle wedge hydration is high enough to ensure low friction along the plates contact, should be limited, since the influence of shear heating is significant for more “anhydrous” settings. Also, the structural and geodynamical local variability and complexity of collision zones make comparison between those and our models more difficult.

Finally experimental results evidence that the strength of the subducting continental crust can also affect the orogenic style. In particular, collision with deep subduction and partial melting of a strong buoyant upper continental crust are likely to evolve into decoupled margins (Model 40, Fig. S3). Such situation can possibly be created locally along subducting margins due to the presence of e.g. rheologically strong magmatic intrusions in the upper crust.

4.1. Sensitivity of the model results to the used parameters

Results obtained in this and previous studies regarding modeling of geodynamical processes rely on our limited knowledge of real processes based on geological observations and laboratory experiments. Since the code I2VIS we use simulates the effects of several geological processes (such as shear heating, dehydration of the slab and hydration of the mantle wedge) occurring on the Earth's uppermost mantle, uncertainties on the real properties and rheological behavior of rocks at depth could condition the results we obtained. However, a more dramatic departure from reality would happen if these processes were not taking into account since they are strongly controlling styles and efficiency of terrestrial one-sided subduction (hydration/dehydration, Gerya et al.,

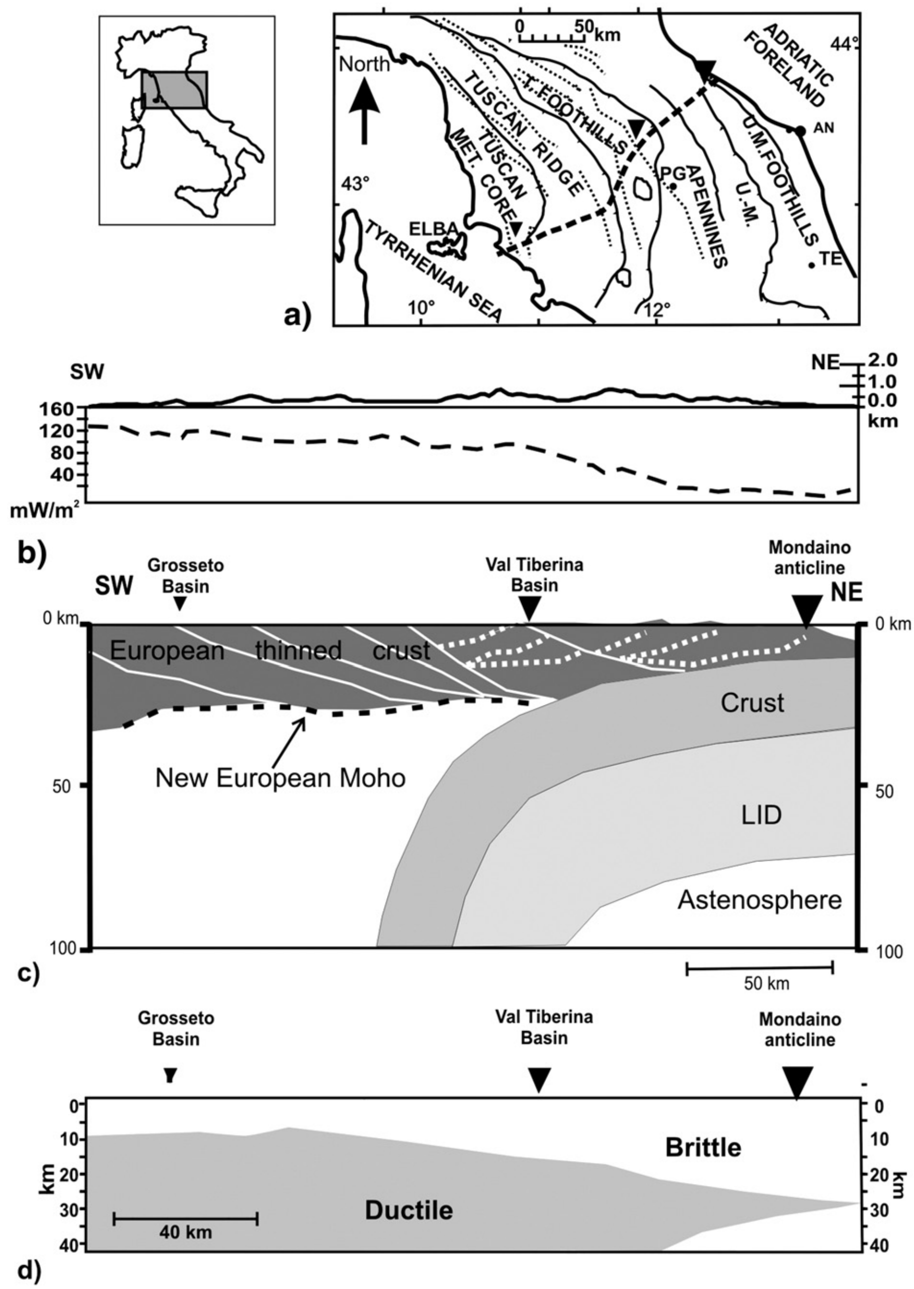

Fig. 9. Location map and geophysical profiles of the study area: (a) Simplified geological map of the Northern Apennines. The dashed thick line indicates the CROP03 NVR seismic line. Barbed lines are the paleo-thrust fronts. Dashed thin lines are normal faults. Reversed triangles indicate localities as in c). b) Topographic (continuous line) and surface heat flow (dashed line) profiles along the CROP03 (Mongelli et al., 1998). (c) Geological cross section of the Northen Apennines along the CROP03 profile. Dark grey is the European thinned crust and the Adria crust being scraped off. Mid grey is the subducting Adria crust. Light grey is the Adria mantle lithosphere. White is the mantle asthenosphere. White thin lines are normal faults. Dashed white lines indicate thrusts. The dashed black line is the new overriding plate Moho. (d) Brittle (light gray) and ductile (dark gray) fields along the CROP03 profile calculated by Pauselli and Federico (2002), using the simplified geological section of Barchi et al. (1998) and the heat flux profile of Mongelli et al. (1998). Modified after Pauselli et al. (2006).

2008) and heat budget in collision orogens (shear heating, Burg and Gerya, 2005). Hence, we believe that accounting for these processes in one way or the other is a necessary requirement for numerical models when we try to understand physical controls of subduction and collision. However, we are aware of large uncertainties of the values of several key parameters involved into our models such as e.g. rheology of rocks and fluid/melt propagation dynamics. Therefore, we concentrate our analyses on the gross scale features and evolution of a subduction/collision zone, rather than on its detailed structure, which is also the case when attempting to compare our models with real orogens. Obviously, when many physical processes are interacting together in the same model it is not always easy to isolate the effect of a single parameter. Therefore, results summarized in Fig. 8 have to be considered as representing *relative tendencies* of influences of tested model parameters.

In spite of these potential pitfalls, we found that, among the various parameters analyzed here together with results obtained in our previous study where only anhydrous cases were taken into account (Faccenda et al., 2008), the most important variable affecting the style of post-subduction continental collision is effective upward propagation velocity of aqueous fluids. Anhydrous models are characterized by continuous coupling between the plates, except for very high convergence rates (10 cm/yr, see Fig. 15 of Faccenda et al., 2008). As soon as porous water and dehydration reactions are included in the models, friction between plates decreases because of the lower viscosity of hydrated and serpentinized rocks forming the subduction channel and located at the base of the plates contact. Indeed, for $v_{y(\mathrm{fluid})}$ and $v_{y(\mathrm{melt})}$ (see Table 1) equal or greater than 1 cm/yr, full decoupling and retreating collisional style occurs, even at convergence rates as low as 2 cm/yr (Fig. 8). Nevertheless, we believe that shear heating also contributes significantly in affecting the thermal structure and the style of collision zones, and it should not be confined as a secondary parameter. It has already been shown that shear heating notably contributes to the overall heat budget in collision zones characterized by Barrovian type metamorphism (Burg and Gerya, 2005), strongly weakens effective rheology of deforming continental lithosphere (Hartz and Podladchikov, 2008; Regenauer-Lieb et al., 2006) and is crucial for localization of the deformation inside the colliding crust (Burg and Schmalholz, 2008) and along mantle detachment faults during continental rifting (Weinberg et al., 2007).

In the next sections our experimental results are compared with geological evolution of real collisional orogens.

4.2. Coupled collisional style

Models with a coupled collision zone are characterized by a thick and broad crustal wedge underlying a wide plateau and limited by a frontal high mountain belt, high heat flux regime in the inner parts of the orogen and low heat flux in outer parts, two seismogenic layers, exhumation of pre-collisional LT-HP metamorphic rocks in proximity of the suture zone and progressive migration toward the foreland of the deformational front (see arrows in Figs. 2–4, and S2).

Two of the most important and active coupled collision zones are the Himalaya and the Western Alps where post-collisional convergence has continued since from Eocene time until present, lasting for a period of nearly 50 Ma (Royden, 1993b). Their geological history is mainly defined by exhumation of pre- and syn-collisional high grade metamorphic rocks in the inner parts of the orogens (Guillot et al., 2007; Compagnoni, 2003) and, after collision, by progressive accretion and stacking of crustal slices and outward migration of the frontal main thrust.

Nowadays, the Himalaya and the Western Alps consist of a broad and thick crustal wedge (Nelson et al., 1996; Schmid and Kissling, 2000), high topography and related high denudation rates, hot crust in the inner parts of the orogen (Nelson et al., 1996; Solarino et al., 1997; Lombardi et al., 2008) and thrust earthquakes that are widespread through all the crust (Monsalve et al., 2006; Schmid and Kissling, 2000). However, in two restricted zones of the southern Tibetan Plateau, earthquakes have a bimodal distribution and hypocenters as deep as 80–90 km have been located in proximity of the Moho and interpreted as nucleating either in the uppermost mantle or in a strong lower crust made of dry granulite (Priestly et al., 2008). The two seismogenic layers in our coupled collision reference model (Model 19, Fig. 3d) can account for the bimodal distribution of the earthquakes and, in particular, for the deeper events. We note here that the presence of the deep brittle layer is closely related to the thermal state of the subducting lithosphere as it is shown by its progressive disappearance in the inner and hotter parts of the orogens.

Pre and syn-collisional metamorphic units form in all models (e.g., Fig. 4); however, only exhumation of the pre-collisional metamorphic rocks occurs because the applied erosion algorithm (see Supplementary Information) does not include focused high denudation rates that control exhumation of syn-collisional medium to high grade metamorphic rocks in the Himalayas (Beaumont et al., 2001) and in the Western Alps (Malusa and Vezzoli, 2006). Nevertheless, the presence of high grade metamorphic rocks in the inner parts of the orogen (Fig. 4) indicates that the gross scale metamorphic pattern of the two collision zone is to a first order reproduced.

4.3. Decoupled collisional style

Models with a decoupled collision zone are characterized by the disruption of the crustal wedge formed during oceanic subduction and the subsequent delamination of the weak part of the continental crust, and by exhumation of LT-HP or higher grade metamorphic core complexes in the inner part of the orogen during the retreating phase (Figs. 5–7, and S1). The collision zone is defined by two distinct domains migrating toward the foreland: the active mountain belt undergoing compression has thick crust, low heat flux and two brittle layers, while the extending basin has thin crust, high heat flux and one brittle layer (Figs. 5–6).

At present day, two of the most studied decoupled collisional zones are the Northern Apennines and the Carpathians. Here we focus on the attempt to make a comparison between the Northern Apennines and our decoupled collision reference model (Model 34).

4.3.1. The Northern Apennines (NA): a case of continental decoupling

The NA are a fold-and-thrust belt that formed in response to the post-collisional (ca. 20 Ma to present-day) roll-back of the Adria microcontinent (e.g. Carminati et al., 2004; D'Offizi et al., 1994; Faccenna et al., 2001; Reutter, 1981).

The geological record indicates that, since the Upper Oligocene–Miocene, the NA underwent two eastward migrating deformational phases that caused early compressional structures (Fig. 3a) being dissected by later extensional deformation due to the opening of the Tyrrhenian Sea (Elter et al., 1975). Extension produced progressive younging toward the east syntectonic extensional basins and associated magmatic activity.

Today the NA consist of two different tectonic domains (Pauselli et al., 2006, and references therein) (Fig. 9): in the Tyrrhenian Domain (TD), shallow mantle is highlighted by a thin crust, high heat flux, recent magmatism and positive Bouguer anomalies. In contrast, a thicker crust, low heat flux, no magmatism and negative Bouguer anomalies in the Adriatic foreland and foredeep Domain (AD) were interpreted as due to crustal doubling. Seismological data (e.g., Ventura et al., 2007) and borehole breakouts (Montone et al., 2004) indicate that the present-day stress distribution in the NA is characterized by extension in the TD and compression and shallow level extension in the AD. Seismic tomography shows evidence of a 650–700 km long slab deepening westward (e.g. Piromallo and Morelli, 2003). Exhumation of LT-HP metamorphic grade rocks (blueschists facies) has been extensively documented in the Tyrrhenian Domain (Jolivet et al., 1998), while rocks outcropping in the Adriatic foreland recorded no metamorphic event.

In comparing the NA with the decoupled collision reference model (Section 3.2.2, Model 34, Figs. 5–7), there is quite good correlation between the geological/geophysical data and the numerical experiment results regarding in particular the well constrained post-collisional evolution and the subdivision of the orogen in two different tectonic areas, the TD and the AD domains (Fig. 9). This could support: 1) no inversion of the subduction polarity in the last 80 Ma (e.g., Faccenna et al., 2001) and 2) the retreating continental plate model as feasible for the last 20 Ma evolution of the NA (e.g. Carminati et al., 2004; D'Offizi et al., 1994; Faccenna et al., 2001; Reutter, 1981).

5. Conclusions

We systematically studied influences of convergence rate and vertical percolation velocities of aqueous fluids and melts in the mantle wedge for establishing of either coupled or decoupled regime of active continental collision. The most important results can be summarized as follows:

1) Upper plate deformation is greatly reduced by the weak and wet sedimentary layer lying on top of the oceanic slab.
2) Low convergence rates combined with low vertical percolation rates of fluid/melt phase translate into continuous plates' coupling and convergence because of the low degree of rheological weakening of the subduction channel and mantle wedge.
3) Coupled collision zones are manifested by the development of a high and broad plateau overlying a thick crustal wedge formed by continuous accretion of crustal slices at the active margin, hot crust in the inner parts of the orogen due to radiogenic heating of the crust, compressive orogenic stresses, presence of two distinct seismogenic (brittle) layers and exhumation of LT-HP metamorphic rocks formed during the previous oceanic subduction.
4) High convergence rates combined with high vertical percolation rates of fluids and melts promote full plates' decoupling, mantle wedging into the crustal wedge, delamination of the upper crust and continental plate retreat.
5) Decoupled collision models are manifested by two distinct and migrating domains. The foreland is dominated by compressional stresses with some extensional areas due to the plate bending, by low heat flux, thick crust and a double seismogenic layer. On the other hand, the basin that develops in the inner part of the orogen is dominated mainly by extension with some compression due to advection and intrusion of partially molten rocks, by high heat flux due to the shallow asthenospheric mantle, thin crust, a single brittle layer and by the exhumation of pre- and syn-collisional blueschists facies metamorphic rocks. These two domains are separated by a transitional zone characterized by extensional stresses at shallow levels and compression at depths.
6) Transition from coupled to decoupled collision occurs always after oceanic subduction have been terminated, indicating that insertion of weak, wet and thick upper crustal layer into the subduction channel plays a major role in triggering plates decoupling and favoring mantle wedging in between the plates.
7) Experiments where deep subduction of a strong continental crust occurs after the early stages of collision due to either the imposed stronger rheology corresponding to "mafic" composition (Plagioclase An_{75}, Model 40) or faster and colder subduction (high convergence rate, Model 21) evolve from coupled to decoupled margins because of dehydration and melting of buoyant upper crust that weakens plate interface and promotes plate decoupling.

Good first-order correspondences between our models and real orogens both in coupled and decoupled collisional settings constrain the validity and reliability of our results and contributes to fill the gap left in modeling retreating (decoupled) collisional margins. In particular, a detailed comparison with the Northern Apennines demonstrates that our decoupled style collision models reproduce main characteristics of this orogen.

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Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.epsl.2008.12.021.

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