



# From oceanic plateaus to allochthonous terranes: Numerical modelling

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## ARTICLE INFO

### Article history:

Received 29 April 2012

Received in revised form 25 August 2012

Accepted 4 November 2012

Available online 23 November 2012

### Keywords:

Crustal growth

Subduction

Terrane accretion

Oceanic plateau

## ABSTRACT

Large segments of the continental crust are known to have formed through the amalgamation of oceanic plateaus and continental fragments. However, mechanisms responsible for terrane accretion remain poorly understood. We have therefore analysed the interactions of oceanic plateaus with the leading edge of the continental margin using a thermomechanical–petrological model of an oceanic–continental subduction zone with spontaneously moving plates. This model includes partial melting of crustal and mantle lithologies and accounts for complex rheological behaviour including viscous creep and plastic yielding. Our results indicate that oceanic plateaus may either be lost by subduction or accreted onto continental margins. Complete subduction of oceanic plateaus is common in models with old (> 40 Ma) oceanic lithosphere whereas models with younger lithosphere often result in terrane accretion. Three distinct modes of terrane accretion were identified depending on the rheological structure of the lower crust and oceanic cooling age: frontal plateau accretion, basal plateau accretion and underplating plateaus.

*Complete plateau subduction* is associated with a sharp uplift of the forearc region and the formation of a basin further landward, followed by topographic relaxation. All crustal material is lost by subduction and crustal growth is solely attributed to partial melting of the mantle.

*Frontal plateau accretion* leads to crustal thickening and the formation of thrust and fold belts, since oceanic plateaus are docked onto the continental margin. Strong deformation leads to slab break off, which eventually terminates subduction, shortly after the collisional stage has been reached. Crustal parts that have been sheared off during detachment melt at depth and modify the composition of the overlying continental crust. *Basal plateau accretion* scrapes oceanic plateaus off the downgoing slab, enabling the outward migration of the subduction zone. New incoming oceanic crust underthrusts the fractured terrane and forms a new subduction zone behind the accreted terrane. Subsequently, hot asthenosphere rises into the newly formed subduction zone and allows for extensive partial melting of crustal rocks, located at the slab interface, and only minor parts of the former oceanic plateau remain unmodified.

Oceanic plateaus may also *underplate* the continental crust after being subducted to mantle depth. (U)HP terranes are formed with peak metamorphic temperatures of 400–700 °C prior to slab break off and subsequent exhumation. Rapid and coherent exhumation through the mantle along the former subduction zone at rates comparable to plate tectonic velocities is followed by somewhat slower rates at crustal levels, accompanied by crustal flow, structural reworking and syndeformational partial melting. Exhumation of these large crustal volumes leads to a sharp surface uplift.

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## 1. Introduction

The oceanic crust (which covers 60% of the Earth surface) is not homogenous, but contains significantly thicker crust than norm. About 10% of the present day's ocean floor is covered by anomalous thick crust typified by a high bathymetric relief, low upper crustal velocities, lack of clear magnetic lineations and steep margins (Nur and Ben-Avraham, 1982; Schubert and Sandwell, 1989). Although the origin of these oceanic rises remains controversial most of them are thought to represent extinct arcs or spreading ridges, detached

continental fragments, volcanic piles or oceanic swells (e.g. Stein and Ben-Avraham, 2007 and references therein).

Regardless of their origin, these oceanic features may collide with continental margins to form collisional orogens and accreted terranes in places where oceanic lithosphere is recycled back into the mantle (Taylor, 1966; Ben-Avraham et al., 1981; Taylor and McLennan, 1985; Schubert and Sandwell, 1989). Hence, it has been argued that the rapid growth of some major segments of the continental crust is related to accretionary processes by which new material is added to the continental crust (e.g. Coney et al., 1980; Jones et al., 1982; Reymer and Schubert, 1986; Stein and Goldstein, 1996; Dobretsov et al., 2004). Large areas in western North America (Jones et al., 1977; Coney et al., 1980; Monger et al., 1982), Alaska (Jones and Silberling, 1986), and

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the Caribbean (Kerr et al., 1997; Kerr and Tarney, 2005) are believed to have formed through extensive accretion along its active margin. Schubert and Sandwell (1989) have estimated an upper bound to the continental crust addition rate by the accretion of all oceanic plateaus to be  $3.7 \text{ km}^3/\text{year}$ , which over a time span of 100 Ma would account for 5% of the total crustal volume of the continental crust.

The Ontong-Java Plateau in the southwestern Pacific is a present day example of an oceanic plateau that resists subduction and thus modifies subduction between the Pacific and Indian plate (Hughes and Turner, 1977; Mann and Taira, 2004). Despite the broad evidence that some plateaus may resist subduction to be accreted in form of collisional terranes others may be lost by subduction (Cloos, 1993), such as in the circum-Pacific where several oceanic plateaus are currently being consumed along with oceanic lithosphere (Rosenbaum and Mo, 2011). Among those are the Nazca (Pilger, 1981) and Juan Fernandez Ridges (von Huene et al., 1997) that are presently being subducted beneath South America. Geological and geochemical observations (Hilton et al., 1992) as well as analogue (Boutelier et al., 2003; Boutelier and Chemanda, 2011) and numerical experiments (Ranalli et al., 2000; van Hunen et al., 2002; Gerya et al., 2009) have supported the idea of deep subduction of crustal material. Thus significant amounts of continental crust may be recycled back into the mantle or be incorporated into active arcs with geochemical and tectonic implications that still need to be explored (e.g. McGeary et al., 1985; Hilton et al., 1992; Chopin, 2003; Rosenbaum and Mo, 2011).

Despite its implications to crustal growth and/or loss, mechanisms responsible for terrane accretion or its deep subduction remain poorly understood. Previous analytical studies have concentrated on the buoyancy of the oceanic lithosphere and bathymetric rises (Cloos, 1993), while analogue (e.g. Boutelier et al., 2003) and early numerical studies have focused on the rheological strength of these features (Ellis et al., 1999). Ellis et al. (1999) have shown that continental fragments of low crustal strength may be deformed and folded within the subduction channel as they approach the continental margin. However, these models concentrated on the upper crustal section and did not take the sublithospheric mantle into account that will significantly affect the dynamics involved in terrane accretion. Subsequent studies have mainly focused on the consequences of plateau subduction/accretion upon the slab and overriding plate. It has been demonstrated in terms of geodynamic models, that oceanic plateaus might alter trench behaviour leading to flat subduction, slab break off, trench advance and trench retreat (van Hunen et al., 2002; Gerya et al., 2009; Mason et al., 2010). Three-dimensional numerical experiments on the influence of a buoyant oceanic plateau on subduction zones show that oceanic plateaus may spread laterally along the trench during collision, if the plateau itself has a sufficiently low density (Mason et al., 2010). Most recently, Tetreault and Buiter (2012) have presented a detailed numerical study on accretion of various crustal units. Their study emphasizes that lithospheric buoyancy alone does not prevent subduction during constant convergence and that a weak detachment layer is necessary in order to accrete crustal units onto the overriding plate. The depth of this detachment layer controls the amount of accreted crust and may lead to crustal underplating or collisional accretion. However, this recent study has employed a constant prescribed subduction velocity and has moreover neglected slab dehydration and melting processes. According to recent results on subduction zones of Sizova et al. (2012) a prescribed convergence velocity and the neglect of fluid- and melt-related weakening effects may inhibit the development of several important collisional processes, such as slab breakoff, vertical crustal extrusion, large scale stacking, shallow crustal delamination and relamination, and exhumation of the continental plate. Geodynamic models on collisional orogens that have employed spontaneously moving plates demonstrate that crustal delamination and accretion processes are critically controlled by the rheology of the lower crust and age of the subducting slab (Duretz et al., 2011, 2012; Sizova et al., 2012; Ueda et al., 2012). The latter

parameter has not yet been explored in relation to terrane accretion processes.

In this present work we aim to extend previous terrane accretion models and explore geodynamic regimes with implications to magmatic activity using spontaneously moving plates. We have undertaken a detailed study of 2-D petrological–thermomechanical numerical experiments to (i) characterise the variability of accretion processes, and (ii) investigate possible effects of melting of subducted crustal units upon magmatic addition rates associated with terrane accretion. Our parametric study is primary focussed on influences of two major parameters which control crustal accretion in collisional zones: (1) the age of the subducting oceanic plate and (2) the rheology of the lower continental crust.

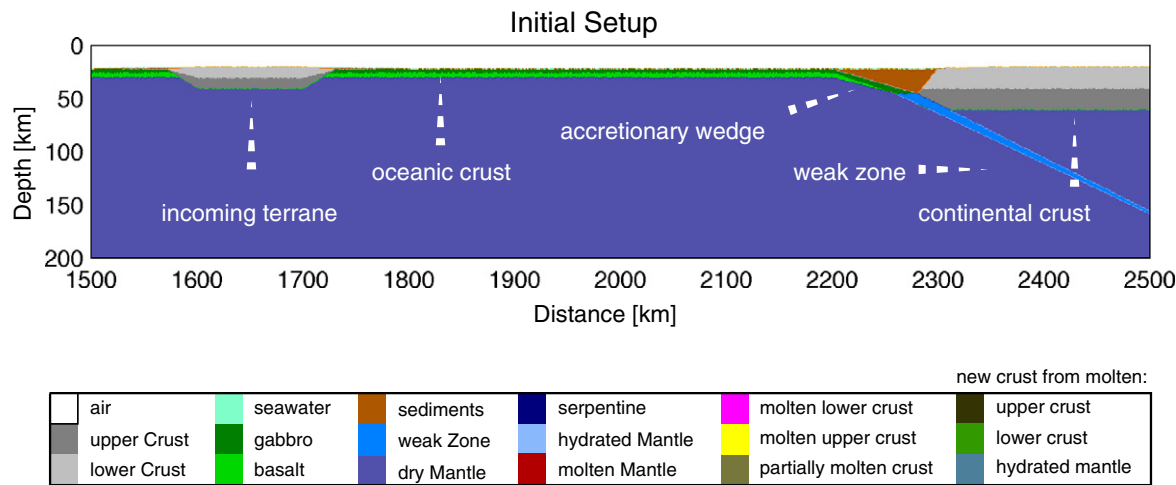
## 2. Model Setup

The numerical model simulates forced subduction of an oceanic plate beneath a continental margin on a lithospheric to upper mantle cross-section (4000 km by 1400 km) Fig. 1. The rectangular grid with  $1361 \times 351$  nodal points is non-uniform and contains a (1000 km wide) high-resolution area of  $1 \text{ km} \times 1 \text{ km}$  in the centre of the domain. The rest of the model remains at a lower resolution ( $10 \times 10 \text{ km}$ ).

The oceanic crust contains an oceanic plateau that moves with the oceanic lithosphere as it migrates towards a fixed continent, fated to collide with the continental margin. The oceanic crust is composed of 2 km of hydrothermally altered basalt, underlain by 5 km of gabbroic rocks that cover 2500 km horizontally. The continental crust is felsic and has a total thickness of 30 km, composed of 15 km upper and 15 km lower crust that extend over 1500 km. The total thickness of the continental crust corresponds to extended continental crust of Western Europe and Western North America and was adopted according to Christensen and Mooney (1995). Since the composition and thickness of oceanic plateaus are not known in detail we have chosen a simplified description and assume the crustal structure to be similar to continental crust. Schubert and Sandwell (1989) have calculated the average crustal thickness of oceanic and continental plateaus, to vary between ~10 and 20 km based on global topographic data analysis. Sandwell and MacKenzie (1989) on the other hand, estimated that continental plateaus with a relief greater than 4.2 km have roots that extend to 25–35 km depths, while oceanic plateaus have a lower relief and thus shallower roots (15–25 km). For the sake of simplicity we have chosen a total crustal thickness of 20 km, subdivided into upper and lower crust of 10 km each, which cover 100 km horizontally. Both the asthenosphere and the upper mantle are composed of anhydrous peridotite and are defined by the temperature profile. The rheological parameters used in the experiments are summarized in Table S1 (Supplement). All mechanical boundary conditions are free slip only the lower boundary is permeable satisfying an external free slip boundary condition (Gorczyk et al., 2007; Ueda et al., 2008). To allow for topographic build up of the lithosphere, the top surface of the lithosphere is treated as an internal free surface (Schmeling et al., 2008) by using a top layer (of 20–22 km thickness) with low viscosity ( $10^{18} \text{ Pas}$ ) and low density ( $1 \text{ kg/m}^3$  for air,  $1000 \text{ kg/m}^3$  for sea water).

The initial temperature field of the oceanic plate is defined by its oceanic geotherm (Turcotte and Schubert, 2002) for a specific lithospheric cooling age that was varied from 20 Ma to 80 Ma. Embedded into oceanic crust, the oceanic plateau is assumed to have the same thermal structure as the oceanic lithosphere. The initial temperature field of the continental plate increases linearly from  $0^\circ \text{C}$  at the surface to  $1344^\circ \text{C}$  at the lithosphere asthenosphere boundary (140 km depth). For the asthenospheric mantle ( $> 140 \text{ km}$ ) a thermal gradient of  $0.5^\circ \text{C km}^{-1}$  is used.

An internally prescribed velocity field within the convergence condition region enables spontaneous slab bending of the oceanic crust. During the first 6 Ma the oceanic plate is pushed toward a fixed continental plate with a constant velocity, reproducing an active



**Fig. 1.** Initial setup of the numerical model (see Section 2 for details). Staggered grid resolution is 1361 × 351 nodal points, with more than 10 million randomly distributed markers. The grid resolution is 1 × 1 km in the subduction zone area (1500–2500 km) and 10 × 10 km outside of this area. Colours indicate materials (i.e. rock type or melt), which appear in subsequent figures. For rheologies and melting conditions used in the experiments see Table S1 and S3 (Supplement).

continental margin. After 6 Ma of forced subduction, subduction and subsequent collision are only driven by the slab pull. A rheologically weak shear zone at the bottom of the oceanic-continental suture zone prescribes initialization of subduction.

A detailed description of the governing equations and the numerical procedure employed in this study is given in the Supplementary material section provided with this article.

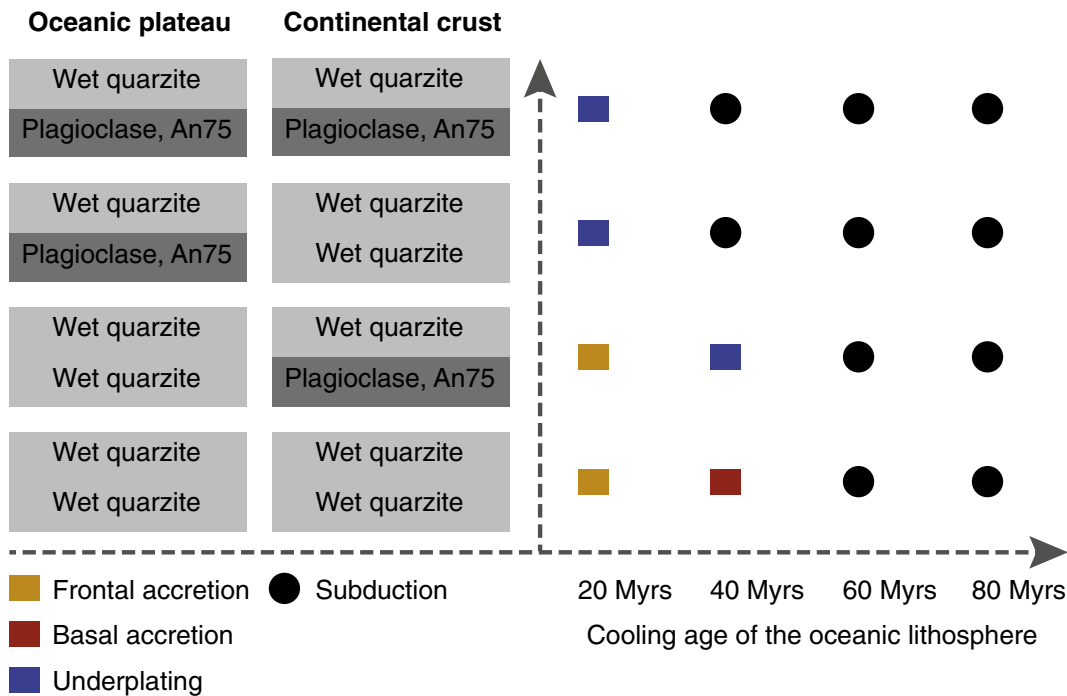
3. Results

Mechanisms responsible for terrane accretion or its deep subduction have been studied in detail, by varying the cooling age of the downgoing slab (20 Ma, 40 Ma, 60 Ma, 80 Ma) and the rheological structure, of the incoming terrane and overriding plate (see Table

S1 for details). Our results indicate that oceanic plateaus associated with the subduction of old (> 40 Ma) and hence dense oceanic lithosphere, are likely to be lost by subduction. In contrast, oceanic plateaus embedded in young oceanic lithosphere (≤40 Ma) are less prone to subduction and may accrete onto continental crust by either (i) frontal or (ii) basal accretion or else may (iii) underplate the continental crust. The results of this study are displayed in Fig. 2.

3.1. Complete plateau subduction

Terrane subduction is likely to occur where old (> 40 Ma) oceanic lithosphere is consumed, but oceanic plateaus that have a strong lower crust (plagioclase, An75) will also subduct if embedded in moderately old crust (40 Myr) (Fig. 2).

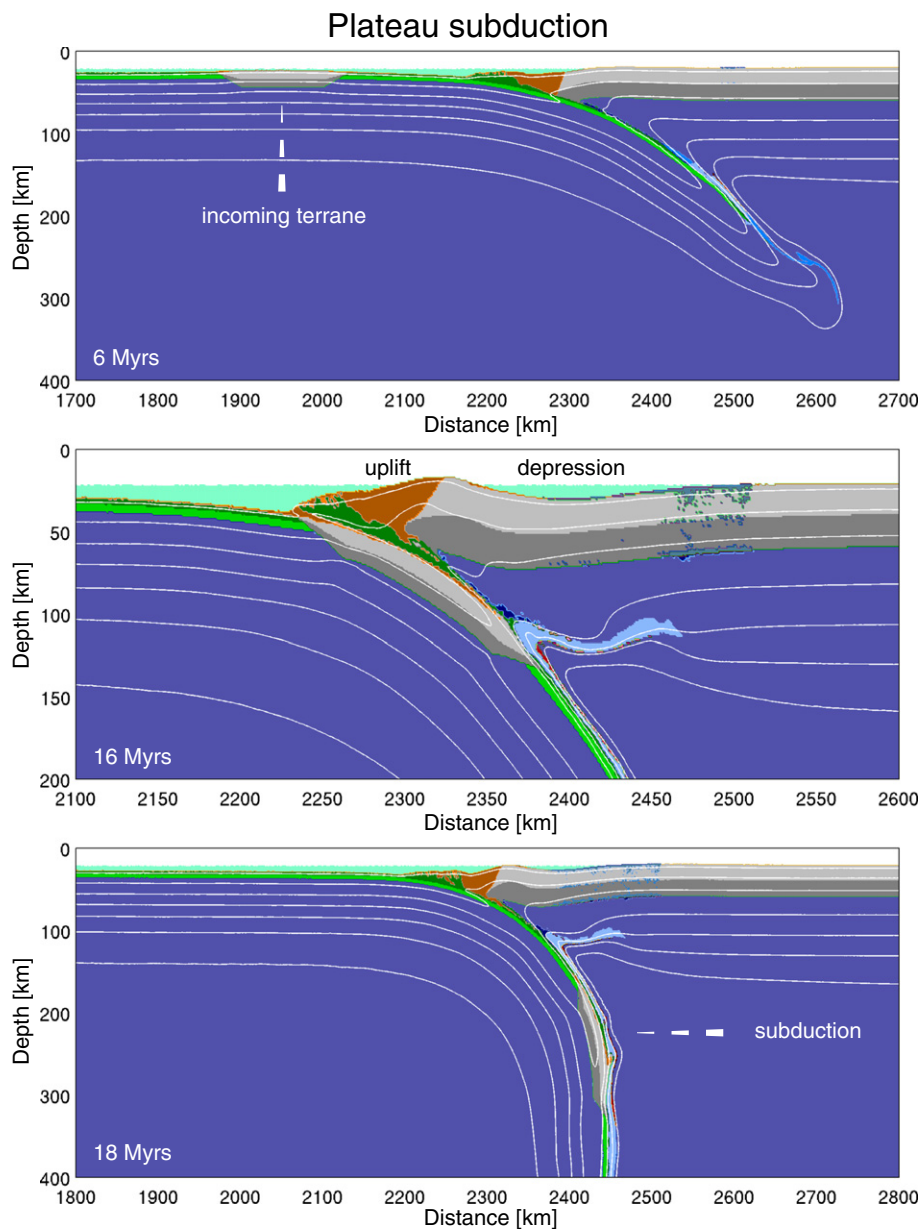


**Fig. 2.** Parameter space showing the explored range of slab ages and crustal rheologies of the upper and lower crust. Wet quartzite and Plagioclase rheologies are used consistently with experimentally determined flow laws (see Supplement for details). Four distinct endmembers are classified: (i) subduction, (ii) frontal plateau accretion (iii) basal plateau accretion and (iv) underplating plateaus.

Shortly after subduction initiation ( $>5$  Ma), prior to collision, water is released from the downgoing slab due to compaction or mineral dehydration reactions. Fluids percolating from the subducting slab into the overlying mantle may form a serpentinized channel at shallow slab interfaces ( $<130$  km) or generate subsolidus metasomatism at greater depth (Fig. 3). Where such fluids encounter the wet mantle solidus they induce partial melting, which enables basaltic melt production. In areas where the extent of melting is sufficient to allow for extraction, melts are emplaced at crustal levels forming extrusive volcanics and intrusive plutons. Discrete addition of magmatic products (Fig. 4a) leads to gradual growth of the magmatic crust ( $20\text{--}30\text{ km}^3/\text{km}/\text{Ma}$ ; see Vogt et al., 2012 for discussion). The depth at which such igneous bodies accumulate varies between lower and mid crustal levels according to the rheological structure of the overriding plate. Arrival of anomalously thick oceanic crust at the subduction zone enhances the magmatic productivity ( $50\text{--}60\text{ km}^3/\text{km}/\text{Ma}$ ), but

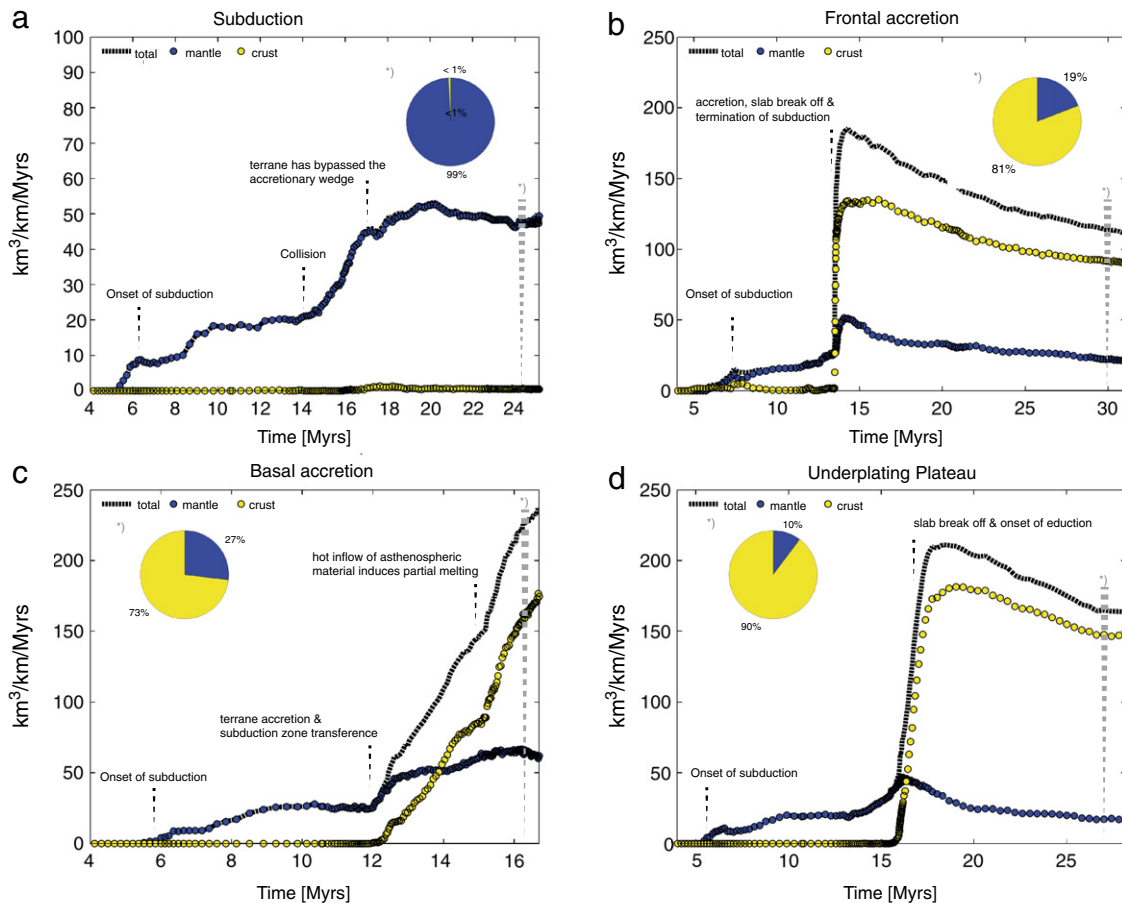
has no effect on the composition of newly formed crust (Fig. 4a). The continental crust grows simply by the addition of basaltic material, derived from partially molten mantle.

Nevertheless, subduction of high bathymetric reliefs becomes all the more evident in the tectonic response of the overriding plate. While the oceanic plateau is dragged down into the subduction zone, material of the forearc is uplifted and pushed towards the continent to form a thickened topographic high (Figs. 3 and 5a). The slab dip steepens significantly and the upper plate is exposed to strong deformation. Localized shear zones of intense plastic deformation are formed that cut through the basement rocks of the overriding plate (Fig. 6a). Driven by the rapid uplift, farther landward the surface subsides and forms a shallow topographic depression or basin (Fig. 5a). Subsequently, the oceanic plateau bypasses the overriding plate and sinks into the deep mantle. While the topography becomes smoother with time, localized shear zones remain visible and may be reactivated during future events.



**Fig. 3.** Tectonic evolution of subducting plateaus (slab age: 40 Ma; lower crust rheology: oceanic plateau: plagioclase, continental crust: plagioclase). Subduction of oceanic plateaus results in a temporary uplift of the forearc and subsidence further landward, followed by topographic relaxation. Water released from the downgoing slab lowers the melting temperature of the overlying mantle and allows for partial melting of the mantle.





**Fig. 4.** Magmatic addition rates ( $\text{km}^3/\text{km}/\text{Ma}$ ) during a) plateau subduction b) frontal plateau accretion c) basal plateau accretion and d) underplating plateau. The inset pie charts illustrate the magmatic composition (crustal (yellow) versus mantle (blue) components) at the time \* marked by the grey dashed line in the time evolution diagrams. a) Plateau subduction: In the course of collision magmatic addition rates increase from 10–20  $\text{km}^3/\text{km}/\text{Ma}$  to typical values of 50–60  $\text{km}^3/\text{km}/\text{Ma}$ . Water released from the subducting plate lowers the melting temperature of the overlying mantle and crustal growth is solely attributed to partial melting of the mantle. b) Frontal plateau accretion: Magmatic addition rates rise from 20–30  $\text{km}^3/\text{km}/\text{Ma}$  to 200  $\text{km}^3/\text{km}/\text{Ma}$  in the course of collision and subsequent accretion. Asthenospheric inflow of hot material to subcrustal levels leads to partial melting of mainly crustal components with minor contribution of the hydrated mantle. c) Basal plateau accretion: Magmatic addition rates increase in the course of collision from 20–30  $\text{km}^3/\text{km}/\text{Ma}$  to 150  $\text{km}^3/\text{km}/\text{Ma}$  and partial melting of crustal lithologies becomes predominant. The outward migration of the subduction zone at 15 Ma is clearly marked in the magmatic record. d) Underplating plateaus: Magmatic addition rates increase from 20–30  $\text{km}^3/\text{km}/\text{Ma}$  to more than 200  $\text{km}^3/\text{km}/\text{Ma}$  following accretion. Large parts of the former oceanic plateau melt at depth and partial melting of crustal lithologies outweighs basaltic melt production from partially molten mantle.

### 3.2. Frontal plateau accretion

Subduction of young oceanic lithosphere (20 Ma) results in frontal plateau accretion if the oceanic plateau contains a weak lower crust (wet quartzite) (Fig. 2).

Collision of the oceanic plateau with the leading edge of the continental margin causes strong deformation in the overriding plate and even more so in the downgoing slab (Fig. 6b). Soon after the collisional stage is reached (1–2 Ma), highly localized fracture zones are formed that trigger tear propagation along the slab-plateau interface followed by segmentation of the downgoing slab, which eventually leads to slab break off and terminates subduction (Fig. 7). The depth at which the oceanic plateau detaches from the downgoing slab may be as shallow as 40 km. While most of the former oceanic plateau is accreted horizontally, minor parts are sheared off and pulled down into the mantle. Following slab detachment, asthenospheric inflow of hot material to subcrustal levels leads to partial melting of mainly crustal components with minor contribution of the hydrated mantle (Fig. 4b). Where such melts exceed the melt extraction threshold they rise to the surface or emplace in predefined fracture zones that have evolved since the onset of collision (Fig. 7). Melt emplacement weakens the continental lithosphere (according to the

adopted  $\lambda_{\text{melt}}$  pressure factor), which in some cases may lead to extension and subsequent decompression melting of dry peridotite.

Tectonic responses to the collision of the oceanic plateau with the continental margin (Fig. 5b) include a slight uplift of the forearc region followed by transpression and transtension in the course of collision. Other characteristics include crustal thickening and the formation of thrust and fold belts in the back arc.

### 3.3. Basal plateau accretion

Subduction of oceanic lithosphere of moderate age (40 Ma) leads to basal accretion if both the incoming oceanic plateau and the continental crust have a weak lower crust (wet quartzite) (Fig. 2).

As the incoming plateau approaches the subduction zone it is for the most part scraped of the downgoing plate. Smaller crustal units can bypass the accretionary wedge and be lost by subduction (Fig. 8). Large stresses that operate during collision lead to strong deformation in the overriding plate and downgoing slab, enabling slab segmentation and slab break off (Fig. 6c). In contrast to frontal accretion, slab break off occurs at somewhat greater depth (~60 km) and does not cease subduction, but allows for the outward migration of the subduction zone (Fig. 8). New incoming oceanic crust underthrusts the fractured terrane

and forms a new subduction zone behind the accreted terrane. Following slab break off, hot asthenosphere rises into the newly formed subduction zone and leads to extensive partial melting of large parts of the former oceanic plateau (Fig. 4c). Rocks that have remained unmodified are exposed to sedimentation and are thus fully incorporated into the accretionary wedge.

The topographic response includes crustal thickening and the formation of thrust and fold belts, similar to frontal accretion (Fig. 5c).

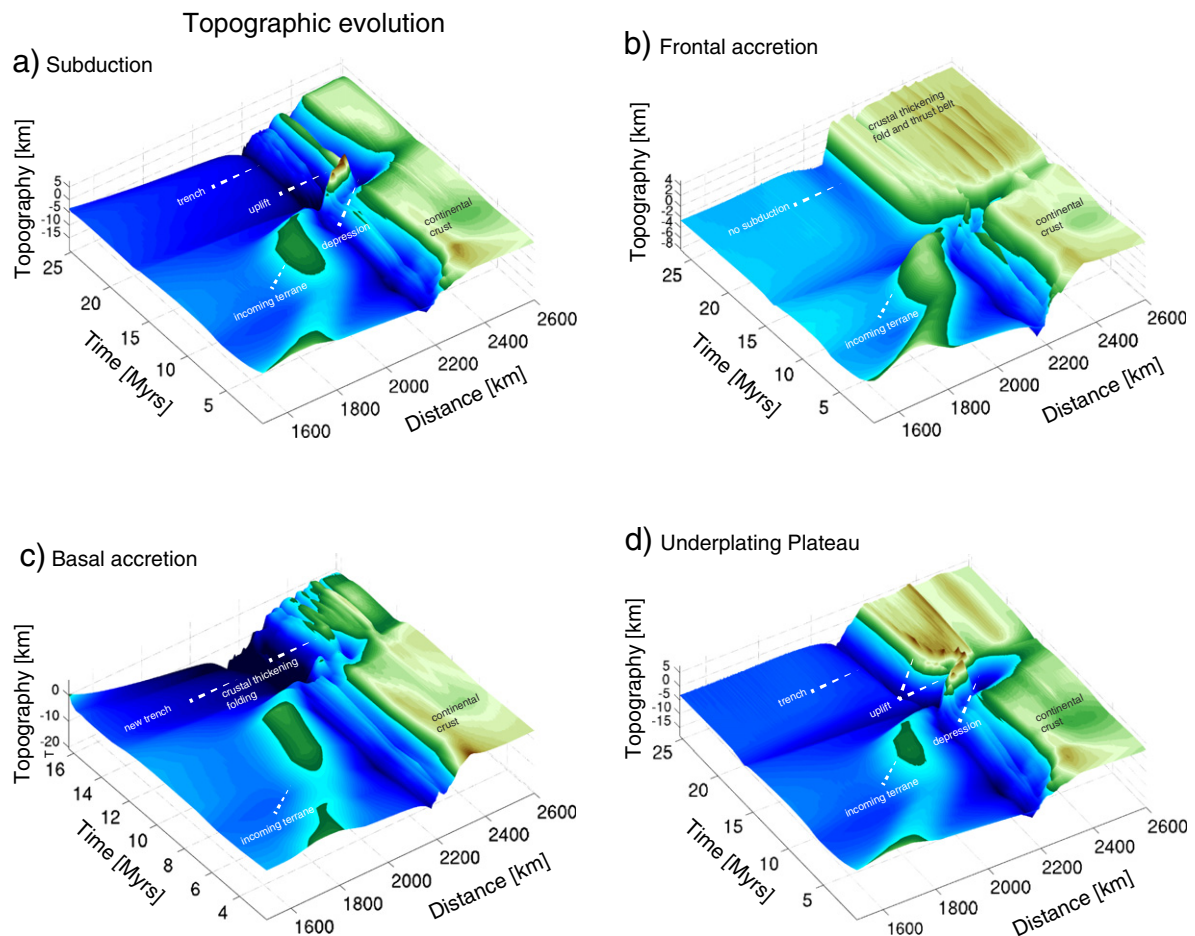
### 3.4. Underplating terranes

Underplating plateaus are associated with a strong lower crust (plagioclase, An75) and young oceanic lithosphere (20 Ma). Subduction of slightly older crust (40 Ma) requires oceanic plateaus to have a weak lower crust (wet quartzite) and continental crust to have a strong lower crust (plagioclase, An75) in order to form underplating plateaus (Fig. 2).

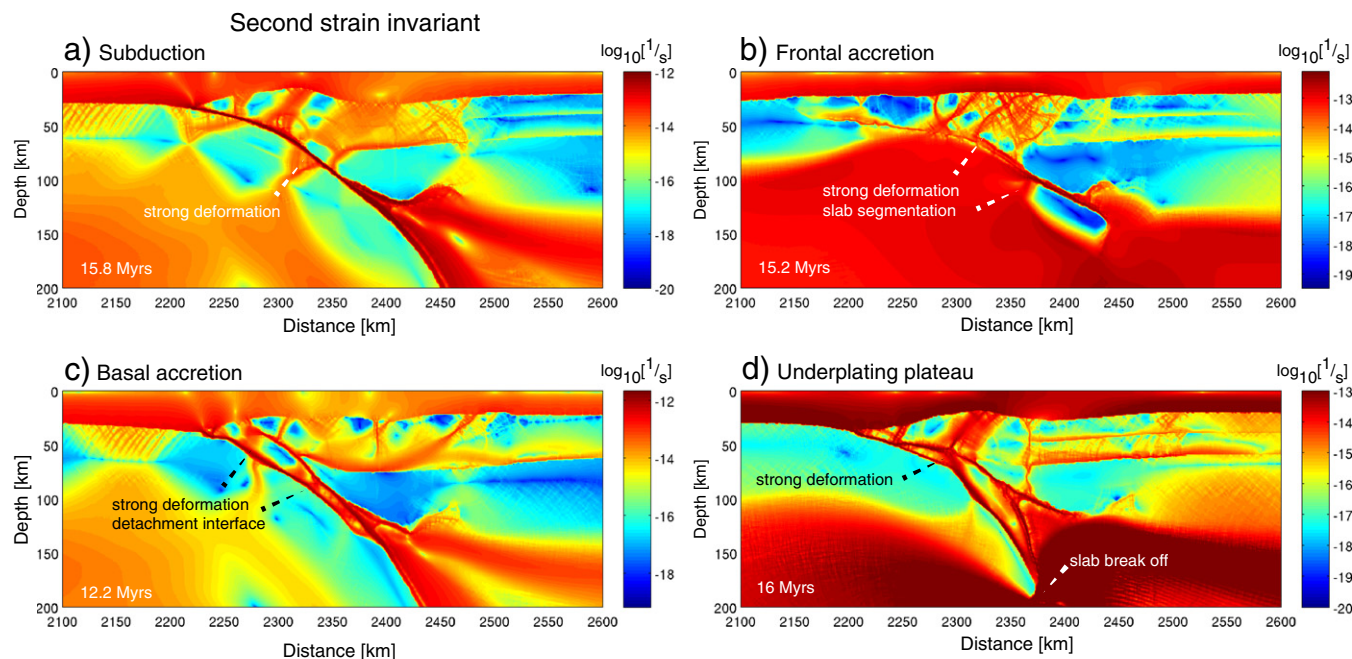
In the course of subduction the oceanic plateau is dragged down to sublithospheric depths (up to 160 km), where it remains for 1–2 Ma (Fig. 9). Large stresses that operate during collision, lead to strong deformation along the plateau–slab interface (Fig. 6d), which eventually leads to slab break off and subsequent slab exhumation along the former subduction zone. In places where the oceanic plateau has encountered greater depth slab exhumation is accompanied by the buoyant rise of

crustal rocks (from the oceanic plateau), which are less dense than the surrounding mantle (Fig. 9). While some of these crustal rocks are brought back to the surface or mid crustal levels, others melt during the inflow of hot asthenosphere and only minor parts of the former oceanic plateau are lost to the mantle. The pressure temperature pathways of rocks that have been traced during collision and subsequent exhumation clearly mark the onset of slab break off by the rapid, nearly isothermal decrease in pressure (Fig. 10a). Rocks that have been exhumed from mantle depth to subcrustal levels reveal elevated exhumation rates on the order of 6–9 cm/year, followed by somewhat slower rates 0.08–0.22 cm/year that mark the final emplacement at mid to upper crustal levels (Fig. 10b). Although some of these rocks remain unmodified, others get incorporated into crustal melting zones and form migmatites or partial melts with complex thermal histories and diverse cooling rates (10–173 °C/Ma; Fig. 10c). Magmatic addition rates (200 km<sup>3</sup>/km/Ma) are for the most part related to partial melting of crustal components rather than to partial melting of the mantle (Fig. 4d).

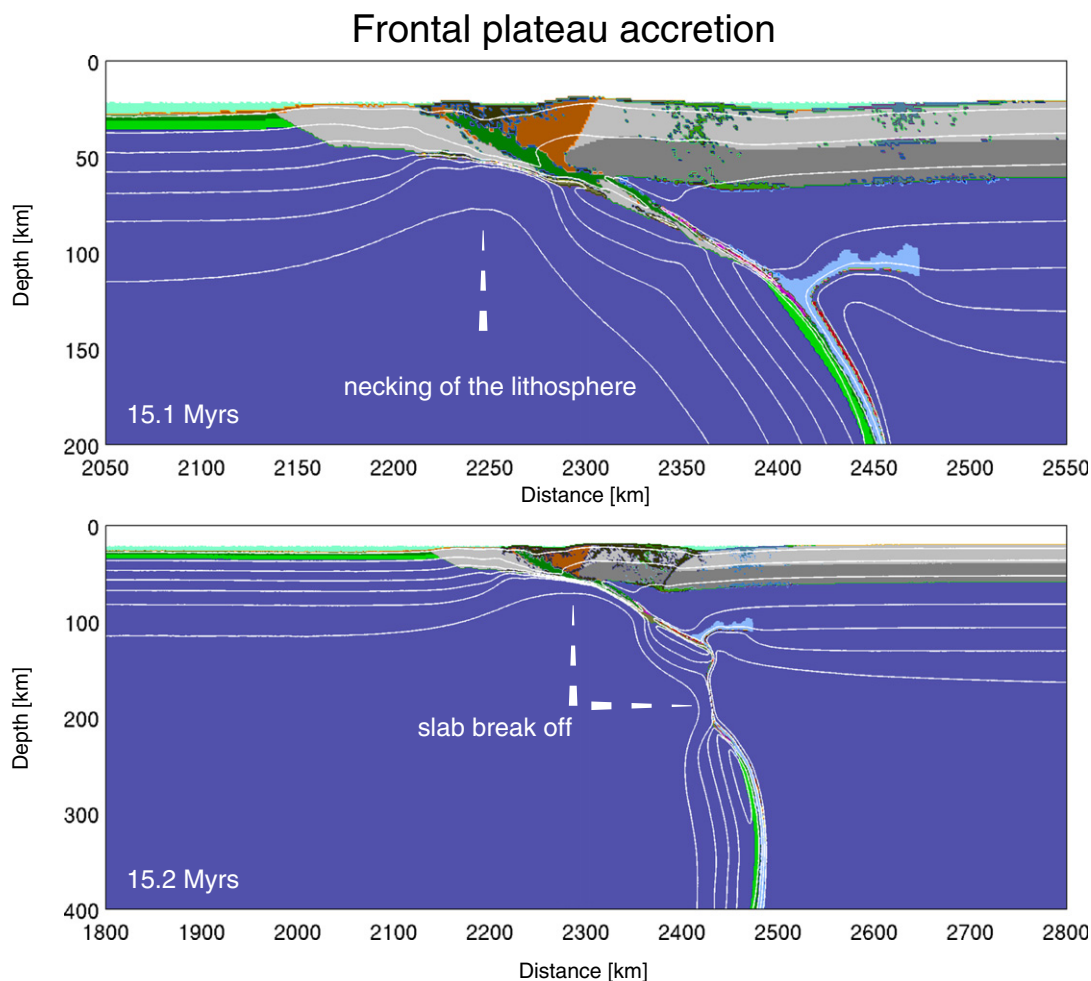
The topographic evolution is at first similar to that of subducting plateaus and is characterized by a sharp uplift of the forearc region and the formation of a depression further landward. However, following slab break off, exhumation leads to strong surface uplift rather than topographic relaxation (Fig. 5d).



**Fig. 5.** Topographic response to a) plateau subduction b) frontal plateau accretion c) basal plateau accretion and d) underplating plateaus. a) Plateau subduction: The most dominant tectonic feature to the subduction of oceanic plateaus is a sharp uplift of the forearc region and the formation of a basin further landward. Once the oceanic plateau has bypassed the accretionary wedge the topography becomes smoother. b) Frontal plateau accretion: Frontal plateau accretion leads to crustal thickening and the formation of fold and thrust belts in the back-arc region. Subduction ceases, following slab break off. c) Basal plateau accretion: Collision of oceanic plateaus with the continental crust leads to crustal thickening and folding of the upper plate. Subduction does not cease, but continues behind the accreted terrane, forming a new subduction zone (subduction zone jump). d) Underplating plateaus: Tectonic responses to the accretion of underplating plateaus include a sharp uplift of the forearc region and the formation of a depression further landward. Following slab break off, crustal material that underplates the continental crust leads to strong surface uplift.



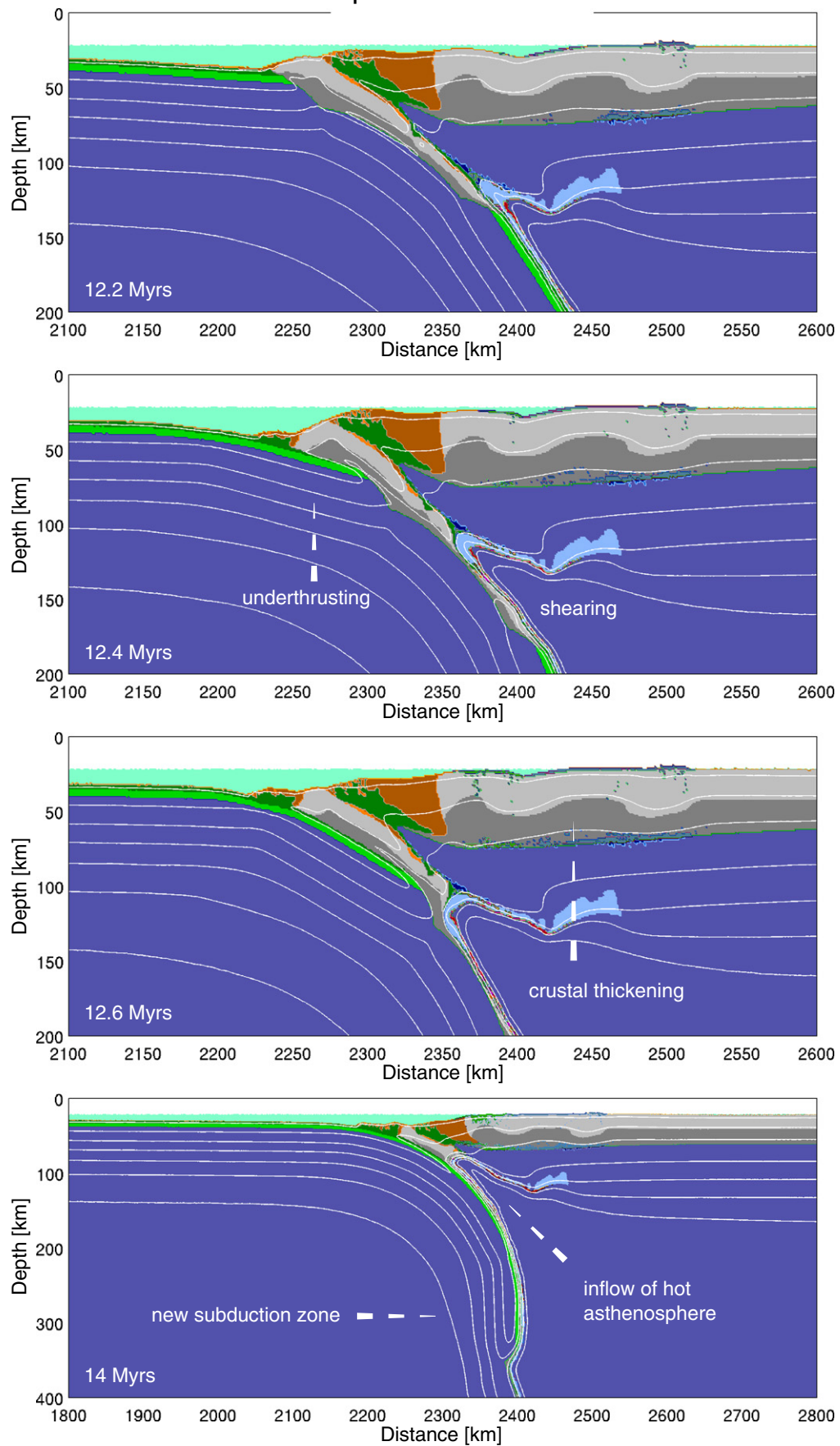
**Fig. 6.** Second strain rate invariant ( $\text{Log}_{10}[1/\text{s}]$ ) of a) Plateau subduction b) Frontal plateau accretion c) Basal plateau accretion and d) Underplating plateaus. Collision of oceanic plateaus with the leading edge of the continental margin leads to strong deformation. Localized shear zones are formed, which may cut through the basement rocks of the upper plate. (b,d) Enhanced plastic failure of the downgoing slab leads to break off.



**Fig. 7.** Tectonic evolution during frontal plateau accretion (slab age: 20 Ma; lower crust rheology: oceanic plateau: wet quartzite, continental crust: plagioclase). In the course of collision, crustal material is docked onto the continental margin, with minor crustal loss. Rocks that have bypassed the accretionary wedge melt at depth. Shortly after the collisional stage has been reached, strong deformation leads to slab break off and terminates subduction. Hot inflow of asthenospheric material leads to partial melting of crustal lithologies.



## Basal plateau accretion





## 4. Discussion

### 4.1. Mode of collision - subduction versus accretion

The negative buoyancy of the downgoing plate is one of the major driving forces in plate tectonics (e.g. Forsyth and Uyeda, 1975). Its magnitude is strongly dependent on the thermal structure of the oceanic lithosphere and increases with time as the lithosphere cools down (Oxburgh and Parmentier, 1977). Old, dense lithosphere readily sinks, while young, buoyant lithosphere may resist subduction (Stern, 2002). While normal oceanic crust is invariably subductable, anomalously thick crust may form collisional orogens and cease subduction, if sufficiently buoyant material is introduced into the subduction zone.

Our results indicate that the mode of accretion is a strong function of the oceanic plate age, which controls both, the negative buoyancy of the plate (and hence the slab pull) and the rheological strength of the oceanic plateau (and therefore the possibility of its separation from the subducting plate). According to the thermodynamic database used in this study (Supplement), an increase in the lithospheric cooling age from 20 Ma to 80 Ma results in an average slab density increase of up to 60 kg/m<sup>3</sup>. This strongly enhances slab pull and thus the possibility for plateau subduction. On the other hand, the rheological strength of rocks decreases exponentially with increasing temperature. Crustal rocks embedded in young and hot oceanic lithosphere have therefore a significantly lower strength compared to rocks embedded in old and cold oceanic lithosphere. Delamination and accretion of oceanic plateaus are therefore favoured for young oceanic plates. Fig. 11 shows the effective crustal strength of the oceanic plateau and underlying mantle for different crustal rheologies and slab ages. Consequently, the general tendency is that oceanic plateaus embedded in old oceanic lithosphere readily sink into the mantle, while oceanic plateaus surrounded by younger oceanic lithosphere may accrete onto the continental margin.

According to our models frontal accretion (Fig. 7), by which most of the oceanic plateau is accreted laterally to the continental margin without crustal loss might be a rare scenario that involves subduction of very young lithosphere ( $\leq 20$  Ma).

Basal accretion (Fig. 8) on the other hand involves detachment of crustal material along the lower surface of the oceanic plateau, whereby some of the material is lost by subduction. This requires a weak lower crust. Based on geological observation of the Caribbean–Colombian igneous provinces Kerr et al. (1997) have postulated that accretion of oceanic plateaus may occur along two detachment layers, hydrothermally altered rocks of the upper crust and weak rocks at the base of the Moho. Hydrothermal circulation may form a rheologically weak zone of altered rocks that is underlain by fresh basement rocks, while temperature dependent weakening affects the base of the crust. Similarly, Schubert and Sandwell (1989) have suggested that in places where the crustal thickness exceeds about 15 km, a low viscosity, ductile layer develops above the crust mantle boundary that acts as a weakness for detaching the crust from the mantle. A recent numerical study on subduction, accretion and collision of various crustal units has revealed that a weak ductile layer is of great importance for the ability of an island arc, oceanic plateau or continental fragment to accrete onto continental crust (Tetreault and Buitert, 2012). The depth of this detachment layer crucially controls the amount of accreted crust (Tetreault and Buitert, 2012).

Underplating plateaus (Fig. 9) on the other hand continue to subduct as long as they remain denser than the surrounding mantle. Deep crustal burial to depth of about 100–160 km precedes slab

detachment and subsequent exhumation to crustal levels. Natural examples of currently underplating plateaus may be found in the forearc region below the North-Western US (Trehu et al., 1994; Gao et al., 2011) and the Bering Sea shelf (McGeary and Ben-Avraham, 1981).

In contrast, oceanic plateaus surrounded by older crust ( $> 40$  Ma) are likely to be lost by subduction (Fig. 3). Buoyancy analysis, based on the contrasts in lithospheric bulk density of 80 Ma old lithosphere have indicated that only bodies of continental and intra-oceanic arc crust greater than 15 km thick may cause collisional orogens and that oceanic plateaus must have crust greater than 30 km thick to have a similar effect (Cloos, 1993), which is in agreement with our experiments. Mann and Taira (2004) have compiled 11 examples of oceanic plateaus and hotspot tracks presently subducting at the circum-Pacific or circum Caribbean plate boundaries. They have concluded that only one, namely the Ontong Java Plateau is being actively accreting, while the remaining 10 examples are subducting without any significant accretion of the uppermost crust.

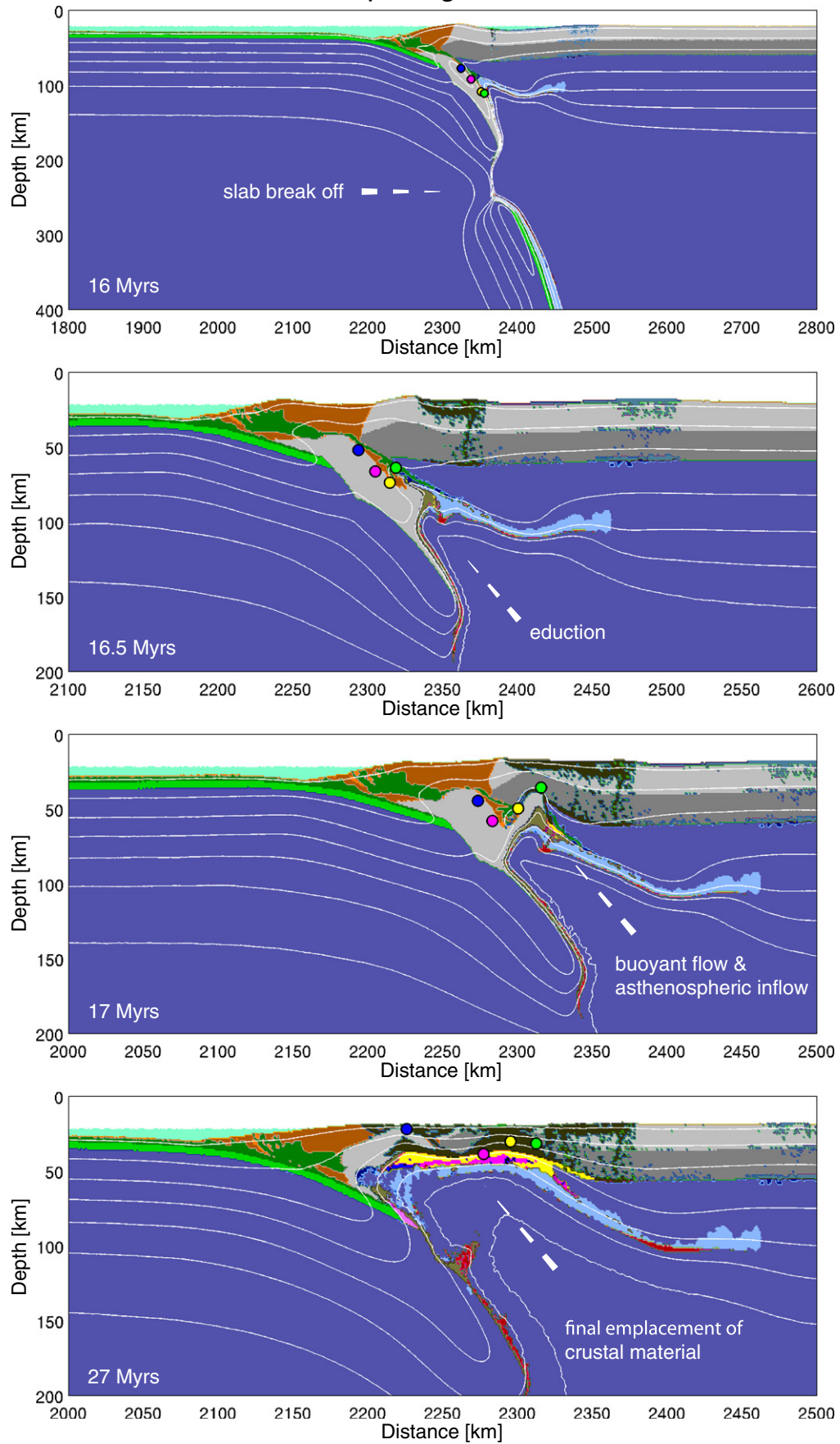
One of the main conclusions of our study is that oceanic plateaus embedded in old and cool oceanic lithosphere ( $> 40$  Ma) will be lost by subduction. However, it is difficult to compare these results with natural geological settings, since accreted terranes have distinct, but diverse stratigraphies and complex tectonic histories that in some cases may involve large scale displacement (e.g. Jones et al., 1977). Also, it is evident that increasing the size of the terrane itself or lowering its strength will lead to accretion, even though the terrane is surrounded by older crust. Fluid entry into the base of the imbricating plateau is known to form pegmatites or hydrous melt (Kerr et al., 1997) that could weaken the crust (Crawford et al., 1987). Recent numerical studies confirm that modifications of fluid and melt related weakening (that is taken constant in this study) modify orogenesis (Faccenda et al., 2009) and hence may significantly affect the dynamics of terrane accretion. In addition, for many present-day oceanic plateaus the age of the surrounding ocean floor is greater than the age of the plateau itself (Schubert and Sandwell, 1989). Crustal accretion can therefore be facilitated, if the oceanic plateau contains some residual formational heat (i.e. in the Caribbean: Kerr and Tarney, 2005). Finally it should be noted, that the third dimension is crucial for evaluating the total effect of buoyant crust on a subduction zone (Stern, 2002). End-on subduction of a linear tract of crust may locally disrupt a subduction zone but not cause it to fail, whereas delivery of similarly buoyant crust parallel to the trench is more likely to lead to subduction zone failure (Stern, 2002). To verify some of these major controls on accretion dynamics further investigations from both numerical studies and geological observation are necessary.

### 4.2. High pressure and ultrahigh pressure terranes

Where underplating plateaus form, crustal material is dragged down to mantle depth, where it might undergo prograde metamorphism, prior to slab break off and subsequent exhumation. High Pressure (HP) to Ultrahigh Pressure (UHP) terranes are formed with peak metamorphic temperatures of 400–700 °C. Exhumation of such (U) HP terranes occurs by coherent exhumation along the former subduction zone or by buoyant flow of crustal rocks, leading to structural reworking and retrograde metamorphism. Geological observations and numerical simulations both support this idea (Andersen et al., 1991; Kylander-Clark et al., 2008; Duretz et al., 2011; Lexa et al., 2011; Little et al., 2011). For example, exhumation of The Western Gneiss region in Western Norway is believed to have been coherent throughout its exhumation history (Kylander-Clark et al., 2008),

**Fig. 8.** Tectonic evolution during basal plateau accretion (slab age: 40 Ma; lower crust rheology: oceanic plateau: wet quartzite, continental crust: wet quartzite). Crustal material of the oceanic plateau is scraped off the downgoing slab and accreted onto the continental margin. Strong deformation leads to slab detachment and the outward migration of the subduction zone. Oceanic crust underthrusts the accreted terrane and forms a new subduction zone (subduction zone jump). Hot asthenosphere rises into the newly formed subduction zone and leads to extensive partial melting of crustal rocks.

## Underplating Plateau



while crustal flow has been inferred to be a major driving force for the exhumation of high-pressure terranes in the Bohemian Massif (Štípská et al., 2004; Lexa et al., 2011) and in the D'Entrecasteaux islands of Papua New Guinea (Little et al., 2011). The rate by which these U(HP) rocks are brought back to crustal levels in our models changes with time, suggesting a multistage history of transport. Rapid exhumation through the mantle at speeds (6–9 cm/year) comparable to plate tectonic velocities, is followed by slower exhumation at the base of the crust (0.08–0.22 cm/year). This is in accordance with several high-pressure/ultrahigh pressure occurrences worldwide for instance in the Alps (Rubatto and Hermann, 2001) or the Himalaya (Parrish et al., 2006) that are known to have undergone such rapid exhumation, followed by slower exhumation at crustal levels. During the first stage of exhumation in our models, deformation is mainly localized at the bottom and top of the oceanic plateau. The later stage involves a more complex deformation pattern, which includes localized deformation of rocks of the upper and lower plate.

Although some of these rocks reach the surface others melt or merge with magma that has been emplaced at crustal levels, revealing differing thermal histories and cooling rates (10–173 °C/Ma). Ultrahigh pressure rocks that have reached peak metamorphic temperatures of around 700 °C are accompanied by rapid cooling (173 °C/Ma) during the first stage of exhumation (~1 Ma), followed by significantly slower rates at crustal levels 10 °C/Ma. Nevertheless, rocks of the same plateau may also encounter lower peak metamorphic temperatures (400 °C) and single stage thermal histories with moderate cooling rates 24 °C/Ma.

#### 4.3. Tectonic responses

Arrival of buoyant material at the subduction zone causes strong surface uplift and considerable structural damage in the upper and lower plates (Figs. 5 and 6). The rheology of the lower continental crust is thereby crucial to the ability to support topography (Clark et al., 2005; Duretz et al., 2011). Significant uplift rates are consistent with a strong lower continental crust (i.e. plagioclase rheology, see Table S1 for details), while crustal thickening is characteristic for a rheologically weaker crust (wet quartzite rheology). Regardless of the mechanisms that lead to crustal accretion or terrane subduction, large stresses that operate during collision form localized shear zones of intense plastic deformation that in some places may cut through the basement rocks of the overriding plate (Fig. 6).

Tectonic features that are solely attributed to the accretion of crustal material onto the continental margin include the formation of a suture zone composed of basalts and sediments of the former oceanic crust and accretionary wedge. Such suture zones have been identified in various locations associated with terrane accretion through geological (e.g. Coney et al., 1980; Monger et al., 1982) and geophysical field studies (Brennan et al., 2011). These may contain thick highly deformed sequences of flysch, deformed ophiolite or high-pressure mineral assemblages of the blueschist facies that render a complex structural history. If subjected to post accretionary collision old suture zones may be reactivated throughout intra-plate deformation (Coney et al., 1980) and could play an important role in the assembly and growth of the continental crust (Brennan et al., 2011 and references therein).

Another tectonic feature of accretionary tectonics is strong deformation of the downgoing slab, which eventually leads to slab break off. Slab detachment has been widely discussed in the literature and although mostly related to continental collision settings (e.g. Davies

and von Blanckenburg, 1995; Wortel and Spakman, 2000; van Hunen and Allen, 2011) slab detachment has also been reported to be an important mechanism associated with the consumption of oceanic lithosphere (Buiter et al., 2002; Haschke et al., 2002; Levin et al., 2002; Rogers et al., 2002). Consistent with our experiments of frontal and basal plateau accretion, van Zedde and Wortel (2001) have shown that slab detachment may occur at depth as shallow as 35 km. While slab detachment related to frontal plateau accretion ceases subduction, subduction zone failure associated with basal plateau accretion allows for the outward migration of the subduction zone. New incoming oceanic crust underthrusts the strongly fractured terrane and forms a new subduction behind the accreted terrane. Both termination of subduction and subduction zone transference/migration have been discussed in the literature as possible consequences to terrane accretion (e.g. Dewey and Bird, 1970; Saunders et al., 1996; Stern, 2004; Cawood et al., 2009).

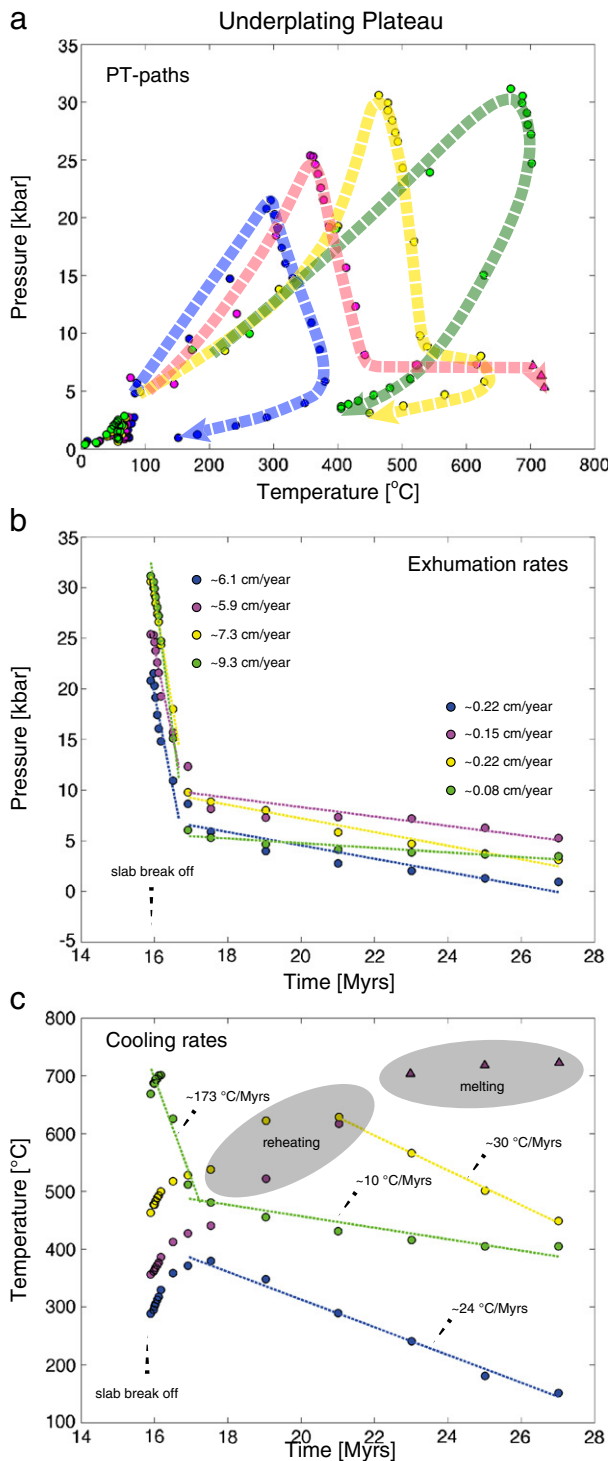
Slab detachment related to underplating plateaus occurs at significantly greater depth. This is in accordance with petrological (e.g. Chopin, 1984, 2003) and numerical (e.g., Gerya et al., 2004; Duretz et al., 2011; van Hunen and Allen, 2011) considerations that suggest deep crustal burial, but it should be noted that these studies have been conducted in relation to continent–continent collision zones and conditions of terrane accretion might be somewhat different and need to be explored in greater detail. Irrespective of the depth of detachment, slab break off causes significant surface uplift (Fig. 6). Elasto-plastic (Buiter et al., 2002) and viscous-plastic models (Gerya et al., 2004; Andrews and Billen, 2009; Duretz et al., 2011) have been used to obtain surface uplifts on the order of 1–6 km.

Natural examples of tectonic responses and consequences to the subduction of high bathymetric relief have recently been reviewed by Rosenbaum and Mo (2011). These include among others thickening and uplift of the forearc region accompanied by thick-skinned deformation and reactivation of basement thrusts, consistent with our observations. Places where surface uplift has been recorded include the Solomon Islands, New Hebrides, Costa Rica and Peru (Rosenbaum and Mo, 2011 and references therein). Subduction of the Nazca Ridge off South America has not only been preserved in the geomorphology and sedimentary facies of the forearc (Hampel, 2002; Clift et al., 2003), but has also been shown to influence far field structure in Amazonian foreland basin, 750 km away from the area of subduction (Espurt et al., 2007). Subduction of aseismic ridges is moreover believed to result in episodic transitions from thin-skinned to thick-skinned deformation and exhumation of basement rocks (Rosenbaum and Mo, 2011), which has been observed in the Andes (Kley et al., 1999).

Although some of these fractures may generate earthquakes, it has been argued that the events tend to be small because the evolving fracture system will limit rupture size (Wang and Bilek, 2011) and act as a seismic barrier to rupture propagation (Kodaira et al., 2000; Wang and Bilek, 2011) consistent with seismic gaps that have been commonly observed in such places (Kelleher and McCann, 1976; McCann et al., 1979; McGeary et al., 1985). Nevertheless it has also been argued that subduction of bathymetric highs enhances seismic coupling at the subduction interface, which causes large magnitude seismicity (Christensen and Lay, 1988; Cloos, 1992; Scholz and Small, 1997). However, it should be noted that subduction of bathymetric highs is not a necessary condition for large subduction earthquakes (Scholz and Campos, 1995) and there are examples of ridge subduction that do not correlate with large earthquakes epicentres (Rosenbaum and Mo, 2011).

**Fig. 9.** Tectonic evolution of underplating plateaus (slab age: 40 Ma; lower crust rheology: oceanic plateau: wet quartzite, continental crust: plagioclase). The oceanic plateau is fully subducted prior to slab break off and subsequent exhumation. The coherent exhumation along the former subduction zone is accompanied by buoyant flow of crustal rocks of the former oceanic plateau. Some of these rocks are brought back to the surface (blue coloured rock), while others accumulate at mid-crustal levels (yellow, pink and green coloured rocks). The pressure temperature paths, exhumation and cooling rates of these rocks (marked by the coloured circles) are displayed in Fig. 10.





**Fig. 10.** Pressure temperature paths, exhumation and cooling rates of crustal rocks of the underplating plateau shown in Fig. 9. a) Pressure–temperature paths of rocks that have been traced during the simulation. Peak metamorphic conditions of 20–30 kbar and 300–700 °C are reached prior to slab break off and subsequent exhumation. Exhumation to sub-crustal levels occurs at nearly isothermal conditions. b) Exhumation rates of crustal rocks. Rapid exhumation through the mantle at speeds (6–9 cm/year) comparable to plate tectonic velocities, is followed by slower exhumation at crustal levels (0.08–0.22 cm/year) showing a two stage history of exhumation. c) Cooling rates. Cooling rates of exhumed crustal rocks vary in detail according to their thermal history. Rocks that have experienced UHP metamorphism reveal a two-stage history of cooling with rapid cooling followed by significantly slower cooling rates (green coloured rock). Other rocks melt (pink coloured rock) or merge (yellow coloured rock) with magma at mid-crustal levels, which modifies their thermal history. Nevertheless, some rocks reveal a simple cooling history with moderate cooling rates (blue coloured rock).

#### 4.4. Crustal composition

Collision and subsequent accretion of oceanic plateaus are preceded by the subduction of oceanic lithosphere. Water released from the subducting plate lowers the melting temperature of the overlying mantle, allowing for flux melting of the hydrated mantle (e.g. [Stolper and Newman, 1994](#); [Tatsumi and Eggins, 1995](#); [Iwamori, 1998](#); [Schmidt and Poli, 1998](#)). Hence, juvenile material of basaltic composition is added to the continental crust by which most arc magmas are believed to have formed. The continental crust, however, has an andesitic bulk composition, which cannot be derived by the basaltic magmatism that dominates present-day crustal growth ([Rudnick, 1995](#)). The andesite composition of the continental crust led Taylor to propose that continents may have formed by the accretion of island arcs of andesitic composition ([Taylor, 1966](#); [Taylor and McLennan, 1985](#)). Later on it has been argued that at least some major segments of the continental crust might have formed through amalgamation of oceanic plateaus and continental fragments ([Coney et al., 1980](#); [Ben-Avraham et al., 1981](#); [Jones et al., 1982](#); [Reymer and Schubert, 1986](#)). However, most island arcs and oceanic plateaus are believed to have basaltic rather than andesitic compositions (e.g. [Stein and Hofmann, 1994](#); [Abbott and Mooney, 1995](#); [Rudnick, 1995](#)). Nonetheless it should be noted that the formation of oceanic plateaus is still not fully understood and while some are considered to have formed by plume related magmatism ([Stein and Hofmann, 1994](#); [Abbott and Mooney, 1995](#)), others are believed to represent rifted continental fragments of crustal structure ([Nur and Ben-Avraham, 1977](#); [Ben-Avraham et al., 1981](#)). Irrespective of their origin, accretion of such structures onto continental crust is of fundamental importance to the growth of continental crust, because even oceanic plateaus of initially basaltic composition may be modified with time to develop a typical upper continental structure ([Stein and Ben-Avraham, 2007](#)). Although some of these plateaus have begun to evolve into continental crust before accretion, most terranes probably began this evolution at the time of accretion ([Condie, 2001](#)).

This is because newly accreted terranes might be rapidly exposed to subduction zone magmatism, which chemically modifies the newly formed crust ([Abbott and Mooney, 1995](#)). On the other hand thick continental crust that is underlain by oceanic plateaus may be hot enough to allow for partial melting or else such melts may mingle and rise with mantle derived basalts along shear zones to form calc-alkaline plutons ([Crawford et al., 1987](#); [Hollister and Andronicos, 2006](#)). Both scenarios are consistent with our results. In places where oceanic plateaus are sheared off the downgoing plate a new subduction zone is formed following slab break off and basalts derived from partially molten mantle may rise and modify the composition of the accreted terrane. Some rocks of the former oceanic plateau may also bypass the accretionary wedge, melt and further modify the composition of the accreted terrane or otherwise add to the growth of the continental crust. [Davies and von Blanckenburg \(1995\)](#) have argued that slab detachment at shallow slab interfaces (<50 km) may cause large thermal perturbations that lead to partial melting of the metasomatized overriding plate, producing basaltic magmatism that may form granitic magmatism in the crust. Underplating plateaus on the other hand are associated with deep crustal burial to depth of about 100–160 km. While some of these rocks may be exhumed to form high-pressure terranes on the surface, extensive partial melting of crustal lithologies may significantly affect arc magmatism and therefore the composition of newly produced crust. High-grade metamorphic and granitic belts that are believed to have formed subsequent to collision of large composite terranes (e.g. [Monger et al., 1982](#); [Crawford et al., 1987](#)) confirm the above-mentioned observations.

#### 4.5. Magmatic addition rates

All our experiments reveal magmatic addition rates on the order of 20–30 km<sup>3</sup>/km/Ma prior to collision that are solely attributed to



partial melting of the mantle. Aqueous fluids that are released from the downgoing slab lower the melting temperature of the overlying mantle allowing for basaltic melt production, which adds to the growth of the continental crust. This is in good agreement with estimates based on natural observations along the circum Pacific. Here, magmatic addition is known to vary between 20 and 95 km<sup>3</sup>/km/Ma among varied locations along the Pacific (Reymer and Schubert, 1984; Tiara et al., 1998; Holbrook et al., 1999; Dimalanta et al., 2002). Less data is available for continental margins, but growth rates that have been measured along the Cordilleran orogenic system are somewhat similar 20–90 km<sup>3</sup>/km/Ma (DeCelles et al., 2009). Following collision, subduction of bathymetric highs is considered to form volcanic gaps, where volcanic activity is absent (Nur and Ben-Avraham, 1982; McGeary et al., 1985; Rosenbaum and Mo, 2011). However, we do not observe such behaviour. Although little magmatism is observed in comparison to accretionary margins, magmatism does not cease but increase (50–60 km<sup>3</sup>/km/Ma) as a response to the deep subduction of oceanic plateaus (Fig. 4). Arrival of buoyant material steepens the slab dip and arc magmas (that are solely attributed to partial melting of the mantle) move closer to the trench. Accretionary margins on the other hand reveal elevated magmatic addition rates on the order of 150–200 km<sup>3</sup>/km/Ma that are mostly related to partial melting of crustal material (Fig. 4). Condie (2007) has estimated average accretion rates in accretionary arcs to be 70–150 km<sup>3</sup>/km/Ma in Phanerozoic orogens and 100–200 km<sup>3</sup>/km/Ma in Precambrian orogens. Production rates of juvenile crust are believed to be typically 10–30% lower than total accretion rates (Condie, 2007).

## 5. Conclusion

We have analysed the dynamics of terrane accretion or its deep subduction along active continental margins, where oceanic crust is

recycled back into the mantle. In addition to terrane subduction three distinct modes of terrane accretion were identified: frontal plateau accretion, basal plateau accretion and underplating plateaus.

### 5.1. Complete plateau subduction

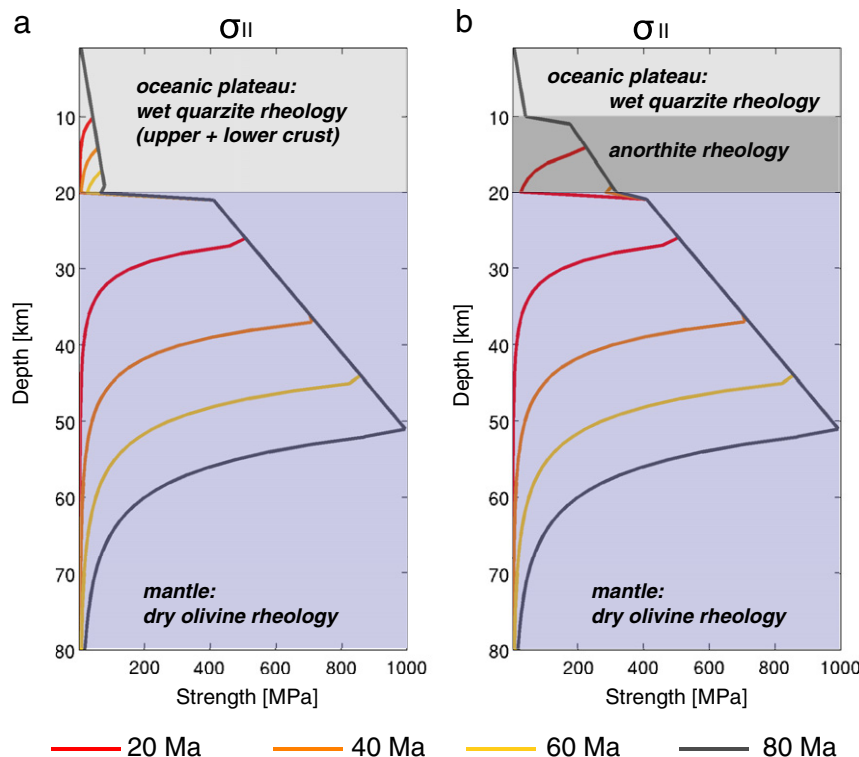
The most dominant tectonic response to the subduction of oceanic plateaus is a temporary uplift of the forearc region and the formation of a depression further landward, followed by subsequent subsidence. Other tectonic features include, steepening of the slab dip and the formation of localized shear zones that in some places may cut through basement rocks of the overriding plate.

### 5.2. Frontal plateau accretion

Oceanic plateaus subjected to frontal plateau accretion are docked onto the continental margin. This leads to crustal thickening. Hence both the downgoing slab and the overriding plate are exposed to intense plastic deformation, which generates deep-seated shear zones. Consequently, the slab detaches at shallow depths and ceases subduction shortly after the collisional stage has been reached. Crustal material that has bypassed the accretionary wedge gets incorporated into arcs that have formed above the upper plate.

### 5.3. Basal plateau accretion

During basal accretion, oceanic plateaus are scraped off the downgoing slab to be accreted along the leading edge of the continental margin. Similar to frontal accretion, strong deformation forms localized shear zones of intense plastic failure. However, slab break off occurs at somewhat greater depth and does not cease subduction, but results in the outward migration of the subduction zone (subduction zone



**Fig. 11.** Strength profile of the oceanic plateau and underlying (lithospheric) mantle as a function of the oceanic cooling age (20 Ma, 40 Ma, 60 Ma, 80 Ma Turcotte and Schubert, 2002) at constant strain rate  $\dot{\epsilon} = 1 \times 10^{-14}$ . The ductile strength of the oceanic plateau and underlying mantle increases with increasing cooling age. a) A wet quartzite rheology has been applied to the upper and lower crust of the oceanic plateau. The rheology of underlying mantle is dry olivine. b) A wet quartzite rheology has been applied to the upper and a plagioclase rheology to the lower crust of the oceanic plateau. The rheology of underlying mantle is dry olivine. A detailed description of the rheological parameters used, is given in the Supplementary material.

jump/ transference). The large thermal contrast between the newly formed subduction zone and the overlying mantle promotes extensive partial melting of crustal lithologies located at the slab interface.

#### 5.4. Underplating plateaus

In contrast to frontal or basal accretion underplating plateaus may be subducted to mantle depth before they underplate the continental crust, following slab break off and subsequent exhumation. The coherent eduction of deep buried material is accompanied by the buoyant flow of crustal rocks, structural reworking and retrograde metamorphism. Rocks that have been brought back to crustal levels form (U) HP terranes or else melt or merge with magma at mid crustal levels. The rate by which these rocks are exhumed changes, whereupon rapid exhumation through the mantle is followed by slower exhumation at crustal levels.

#### Acknowledgements

The manuscript has been improved by comments from Susanne Buiter and Jeroen van Hunen. This work was supported by ETH Research Grants ETH-0807-2, ETH-0807-3, and ETH-0609-2, SNF Research Grants 200020-126832 and 200020-129487, Crystal2Plate program, SNF ProDoc program 4-D-Adamello and TopoEurope Program.

#### Appendix A. Supplementary data

Supplementary data to this article can be found online at <http://dx.doi.org/10.1016/j.gr.2012.11.002>.

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