Continental plate collision: Unstable vs. stable slab dynamics

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ABSTRACT

We study possible scenarios for the evolution of continental collision zones by using a dynamic thermomechanical model that includes brittle-elastic-ductile rheology, surface erosion, and explicit metamorphic changes. This paper focuses primarily on the influence of four key parameters: (1) geotherm or thermotectonic age (which controls the rheological profile), (2) lower-crustal composition (weak or strong rheology), (3) convergence rate, and (4) metamorphic changes in the downgoing crust. The experiments suggest that, depending on these parameters, plate convergence is accommodated by four distinct mechanisms: stable subduction, shortening by pure-shear thickening or folding, and Rayleigh-Taylor instabilities. It appears that stable, oceanic-type subduction can only occur in the case of cold lithospheres (Moho temperature, $T_{Moho} < 550$ °C), and basically needs high convergence rates (>4–5 cm/yr). Depending on the lower-crustal rheology (strong or weak), either the whole (upper and lower) crust or only the lower crust can be involved in subduction. It appears that in the case of weak metamorphic rheologies, phase changes only slightly improve chances for stable subduction. Lithospheric shortening becomes a dominant mechanism when $T_{Moho} > 550$ °C or convergence rates are <4–5 cm/yr. Pure-shear thickening becomes important in all cases of hot lithospheres ($T_{Moho} > 650$ °C). Large-scale folding is favored in the case of $T_{Moho} = 500$–650 °C and is more effective in the case of mechanical coupling between crust and mantle (e.g., strong lower crust). Gravitational (Rayleigh-Taylor) instabilities overcome other mechanisms for very high values of T_{Moho} (>800 °C) and may lead to development of subvertical cold spots.

Keywords: continental collision, subduction, orogens, modeling, rheology.

INTRODUCTION

The possibility of continental subduction (simple shear) provokes lively discussions. In the case of oceanic lithosphere, subduction is a natural consequence of negative buoyancy of aging lithosphere additionally pushed and pulled by far-field forces. The oceanic subduction is fast (5–15 cm/yr), and as a consequence, the slab may remain cold and strong at great depth. It can be pushed and pulled down as a single unit because it is rigid enough to transmit far-field stresses over distances largely exceeding its thickness. Development of continental subduction is more problematic, because the continental plates are positively buoyant owing to their light, thick crust, and because continental convergence rates are small (3–10 times smaller than in the oceans). As a result of the latter, the continental slab–asthenosphere interactions may be dominated by conductive, not advective, heat transfer. In this case, hot temperatures propagate into the slab by thermal conduction from the asthenosphere as fast as the cold temperatures are advected from the surface of the slab. Such a slab may undergo thermal readjustment leading to mechanical weakening before it can descend to any significant depth. In this case, far-field push or pull forces will result in internal shortening or extension of the plate rather than its subduction as a rigid unit.

The alternative collision scenarios are related to accommodation of lithospheric shortening by various mechanisms: (1) pure-shear thickening, (2) folding (Burg and Podladchikov, 2000; Cloetingh et al., 1999), and (3) gravitational (Raleigh-Taylor, RT) instabilities in thickened, negatively buoyant lithosphere. The RT instabilities lead to sinking of subvertical sections of lithosphere into the asthenosphere (e.g., Houseman and Molnar, 1997), which we call "unstable subduction." Superimposed scenarios are also possible: e.g., megabuckles created by lithospheric folding (Burg and Podladchikov, 2000) can localize and evolve into subduction-like zones or result in development of RT instabilities. RT instabilities can also occur in subducting lithosphere (Pysklywec et al., 2000).

Despite these complexities, a number of geologic and geophysical observations point to the possibility of stable continental subduction (e.g., Chopin, 1984). However, this requires additional explanatory models. For example, the problems related to the positive buoyancy of the lithosphere can be circumvented if the light crust separates early from the mantle (Cloos, 1993) or if it undergoes metamorphic changes to become dense and strong (Austrheim, 1991; Le Pichon et al., 1992; Burov et al., 2001). Geodynamic data suggest that during the first million years of transition from fast ocean-continent subduction to slowing continent-continent collision, convergence rates are considerably higher than at later stages (Patriat and Achache, 1984). If true, then the continental slab may remain cold and strong enough to maintain subduction for a few million years. This minimal condition is usually defined by the Péclet number, *Pe*, equal to the ratio of the advection rate to the diffusion rate. The *Pe* must be >20–30 for subduction (Turcotte and Schubert, 1982), which suggests that subduction is unlikely for convergence rates below 1–1.5 cm/yr ($Pe < 20$), but is not impossible for convergence rates >1.5–2.5 cm/yr ($Pe = 30$–70). Yet, additional conditions must be satisfied to enable stable subduction: the crust should be either decoupled from the mantle or metamorphosed; and other modes of deformation such as RT instabilities, folding, and pure-shear thickening should develop slower than the simple-shear mode (subduction).

The multitude of factors influencing continental collision justifies a modeling approach (e.g., Doin and Henry, 2001; Pysklywec et al., 2000; Sobouti and Arkani-Hamed, 2002; Chemenda et al., 1995). However, most existing models are not satisfactory. The analogue models are imperfect because of rheological simplifications, poor thermal coupling, and the absence of phase changes. The numerical models are often limited by simplified visco-plastic rheologies (e.g., Pysklywec et al., 2000) and by use of a fixed-displacement (or velocity) upper-boundary condition, instead of a natural free-surface boundary condition. The use of fixed upper-boundary condition forces stable subduction (Doin and Henry, 2001; Sobouti and Arkani-Hamed, 2002), attenuates pure shear, cancels folding, and does not allow for realistic topography evolution. Many studies also simply predefine subduction (e.g., Beaumont et al., 1996). Consequently, it is crucial to introduce a model that allows for all modes of deformation and accounts for realistic rheology and thermal evolution. The goal of the work presented in this paper has been to study various factors controlling continental collision by using an adequate thermomechanical model. For that, we suggest a new thermomechanical model that combines (1) a free surface boundary condition and surface processes (true topography); (2) realistic brittle-elastic-ductile rheology and structure; and (3) full phase changes (density and rheology). In

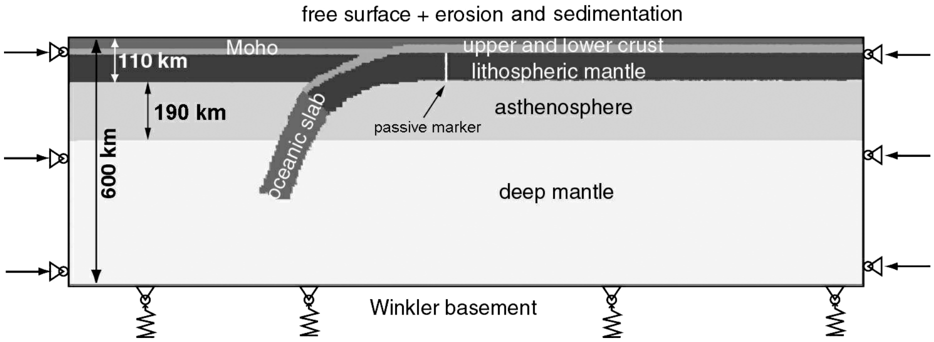

Figure 1. Model setup. Numerical grid is composed of 5 × 5 km quadrilaterals. Upper boundary condition is free surface, erosion, and sedimentation. Bottom boundary condition is pliable Winkler basement. Lateral boundary conditions are velocities. Brittle-elastic-ductile rheology is different for upper crust, lower crust, mantle lithosphere, slab, sediments, asthenosphere, and deep mantle (see Table 1; Fig. 2).

TABLE 1. PHYSICAL PARAMETERS

All rocks	
Conductivity (crust)	2.5 Wm^{-1} $°C^{-1}$
Conductivity (mantle)	3.5 Wm^{-1} $°C^{-1}$
Surface heat production	9.5×10^{-10} W kg^{-1}
Thermal expansion	3.0×10^{-5} $°C^{-1}$
Lamé constants $\lambda = G$	30 GPa
Friction angle	30°
Cohesion	20 MPa
Specific upper and weak lower-crust properties	
ρ (upper crust)	2800 kg m^{-3}
ρ (lower crust)	2900 kg m^{-3}
N	2.4
A	6.7×10^{-6} $MPa^{-n} \cdot s^{-1}$
Q	1.56×10^{5} KJ $\cdot$ mol^{-1}
Specific strong lower-crust properties	
ρ	2980 kg m^{-3}
N	3.4
A	2×10^{-4} $MPa^{-n} \cdot s^{-1}$
Q	2.6×10^{5} KJ $\cdot$ mol^{-1}
Specific eclogite properties	
ρ	3340 kg m^{-3}
Power-law exponent *n*	2.4
A	6.7×10^{-6} $MPa^{-n} \cdot s^{-1}$
Q	1.56×10^{5} KJ $\cdot$ mol^{-1}
Specific mantle properties	
ρ (lithosphere)	3330 kg m^{-3}
ρ (oceanic slab)	3350 kg m^{-3}
ρ (asthenosphere)	3310 kg m^{-3}
N	3
A	1×10^{4} $MPa^{-n} \cdot s^{-1}$
Q	5.2×10^{5} KJ $\cdot$ mol^{-1}

Note: Compilation by Burov et al. (2001). ρ is density; *Q*, *n*, *A* are parameters of the ductile flow law: activation energy, material constant, and power exponent, respectively.

this preliminary study we search for optimistic bounds on the parameters controlling the continental subduction. We consider the early stages of continental plate collision when subduction is assisted by high convergence rates and the downward pull of the subducted oceanic slab.

NUMERICAL MODEL

We further extended the Parovoz code (Poliakov et al., 1993), based on the well-known FLAC (Fast Lagrangian Analysis of Continua) algorithm by Cundall (1989). This ''2.5 D'' explicit time-marching, large-strain Lagrangian algorithm locally solves Newtonian equations of motion in continuum mechanics approximation and updates them in large-strain mode. The solution of these equations is coupled with solutions of constitutive and heat-transfer equations. Our version of Parovoz is fully thermally coupled, handles explicit elastic-ductile-plastic rheologies, a free-surface boundary condition, full metamorphic changes, and surface processes (erosion and sedimentation).

PROBLEM SETUP

We assume an optimistic collision geometry (Fig. 1) that is initially compatible with stable subduction: (1) the oceanic slab first entrains the cold continental slab to a depth equal to one plate thickness, and (2) the convergence rate is high (two-sided closing rate of 2 × 3 cm/yr), as at the beginning of collision. This rate is close to the smallest oceanic subduction rates and is smaller than the initial rate of India-Asia collision (8–10 cm/yr during the first 10 m.y.; Patriat and Achache, 1984).

For physical and rheological parameters of the crust and mantle, we use commonly inferred values derived from rock mechanics (Table 1; Burov et al., 2001). The upper boundary condition is free surface, the lateral boundary conditions are cinematic, and the Winkler (hydrostatic) forces are used as the bottom boundary condition. As in nature, the topography growth is limited by surface erosion, which is modeled by using common diffusion erosion (Burov et al., 2001). The initial geotherm is derived using a common half-space model (e.g., Burov et al., 2001).

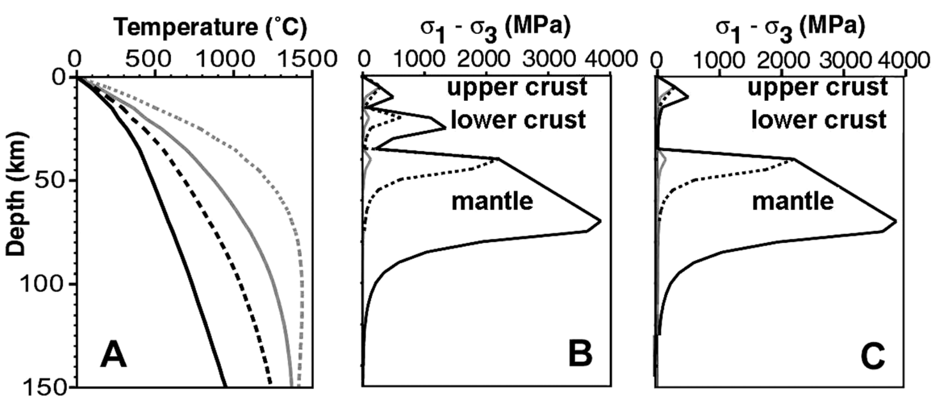

Figure 2. A: Initial geotherms and associated rheological strength profiles for lithosphere with 35-km-thick crust deforming at strain rate of 10^{-15} s^{-1}. B: Weak lower crust. C: Strong lower crust. Black line—cold lithosphere (thermotectonic age = 450 m.y., T_{Moho} = 400 °C); black dashed lines—intermediate lithosphere (150 m.y., 550 °C); gray line—hot lithosphere (75 m.y., 700 °C); gray dashed line—very hot lithosphere (25 m.y., 1000 °C).

The first universal variable parameter of all experiments was geotherm or thermotectonic age (Turcotte and Schubert, 1982), identified here with Moho temperature, T_{Moho} (Fig. 2A). The geotherm or age defines major mechanical properties of the system, e.g., the rheological strength profile (Figs. 2B, 2C). Thus, by varying the geotherm we account for the entire possible range of lithospheres, from very old, cold, and strong plates to very young, hot, and weak ones. The second variable parameter was the composition of the lower crust, which, together with the geotherm, controls the degree of crust-mantle coupling. We considered both weak (quartz dominated) and strong (diabase) lower-crustal rheology. In most experiments we assumed a fixed convergence rate of 2 × 3 cm/yr and no metamorphic changes at depth. Nevertheless, we also tested smaller convergence rates (e.g., two times smaller, four times smaller) and the influence of crustal eclogitization (at pressure, $P > 1.5$ GPa and $T > 550$ °C; see Table 1).

EXPERIMENTS WITH A WEAK LOWER CRUST

Influence of Thermotectonic Age or Geotherm

In this experiment set we assumed weak (quartz-dominated) lower-crustal rheology, high initial convergence rate (2 × 3 cm/yr), and no phase changes. Four types of collision

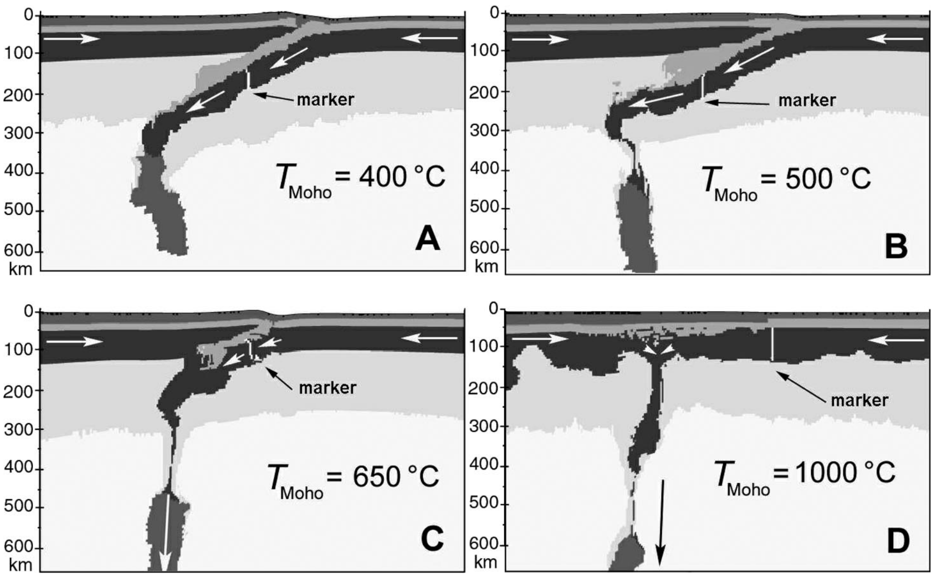

Figure 3. Experiments with weak lower crust at 5.5 m.y., no high-pressure–high-temperature (*P, T*) or ultrahigh-*P*–ultrahigh-*T* (see text) metamorphism, high convergence rate (2 × 3 cm/yr). A–D: Moho temperatures are, respectively, 400 °C, 500 °C, 650 °C, and 1000 °C. Only central parts of models are shown. Black color represents sediments. Lengths of arrows are proportional to material velocity.

scenarios were revealed as a function of thermotectonic age or geotherm.

Cold geotherm (T_{Moho} < 450 °C). An initially cold geotherm (Fig. 3A) allows the collision to evolve into stable, oceanic-type subduction. Almost all shortening is accommodated by subduction of both the continental lower crust and mantle. Because of low T_{Moho} the lower crust is resistant to decoupling and keeps ''glued'' to the lithospheric mantle. It thus can be dragged as deep as 250 km in spite of its positive buoyancy. However, the mechanical resistance of the upper crust to subduction remains lower than the buoyancy-induced stresses: it separates early from the lower crust and remains at the surface.

Intermediate geotherm (T_{Moho} = 450–550 °C). For intermediate geotherms (Fig 3B), the shortening is still largely accommodated by subduction, but the positively buoyant lower crust separates from the negatively buoyant lithospheric mantle and stagnates at some intermediate level (between 100 and 200 km), sometimes forming a double crustal zone (a possible analogy is the Northern Apennines; Ponziani et al., 1995). The geometry of the downgoing lithospheric mantle is affected by the ascent of the buoyant lower crust: the slab adopts a very low angle of subduction. As a consequence, the oceanic slab detaches and sinks into the deep mantle. Small-amplitude (1000 m) long-wavelength (350–400 km) lithospheric folding also accommodates some part of the shortening, specifically in the upper plate.

Hot geotherm (T_{Moho} = 550–750 °C). At a T_{Moho} of 650 °C (Fig. 3C), pure-shear thickening and moderate-amplitude (1500 m) lithospheric folding (wavelength 200–250 km) start to accommodate a significant part of shortening. This behavior is a result of the thermally induced weakening of the lithosphere that makes volumetric thickening mechanically easy. The base of the overriding lithospheric plate also becomes weak and can be dragged downward with the sinking lower plate.

Very hot geotherm (T_{Moho} > 750 °C). For very hot, weak lithosphere (Fig. 3D), stable subduction and lithospheric folding are not possible: all the convergence is accommodated by pure-shear thickening and RT instabilities. Because of high temperatures, the effective viscosity at the base of the lithosphere is reduced, whereas its density is still higher than that of the asthenosphere; these two factors promote rapid (in <1 m.y.) development of RT instabilities. The slab thins in a ''chewing gum'' way, and a cold spot forms (a possible natural example is the Vrancia body in the Carpathians; e.g., Wenzel, 2002). The rate of subduction in this case is not controlled by convergence rate, but by the internal growth rate of the RT instability. We call this style of deformation ''unstable subduction.''

Influence of Convergence Rate

At smaller convergence rates (<5 cm/yr), normal subduction is still possible in cold cases (T_{Moho} < 450 °C), whereas for intermediate-temperature and hot lithospheres, internal shortening dominates as a consequence of more effective conductive heating and, consequently, weakening of the lithosphere at decreasing convergence rates.

Influence of Metamorphic Changes

In these experiments, the major difference from the previous scenarios appeared to be a steeper subduction angle of the continental lithospheric slab. For example, the case shown in Figure 3B became similar to that in Figure 3A; in the case shown in Figure 3C, the slab became nearly vertical, and so on. These experiments suggest that phase changes do not significantly improve chances for normal subduction: when T_{Moho} exceeds 550 °C, subduction is not a dominant mechanism, whatever the degree of metamorphism. Yet this result is assumed to be valid for the weak eclogite rheology (same as quartz) used in the experiments. This assumption may be questioned because the rheology of eclogites is not well constrained. Additional experiments hint that the assumption that eclogite rheology is as strong as diabase would be equivalent to a shift in the T_{Moho} by ~ −200 °C, which greatly improves the chances for stable subduction.

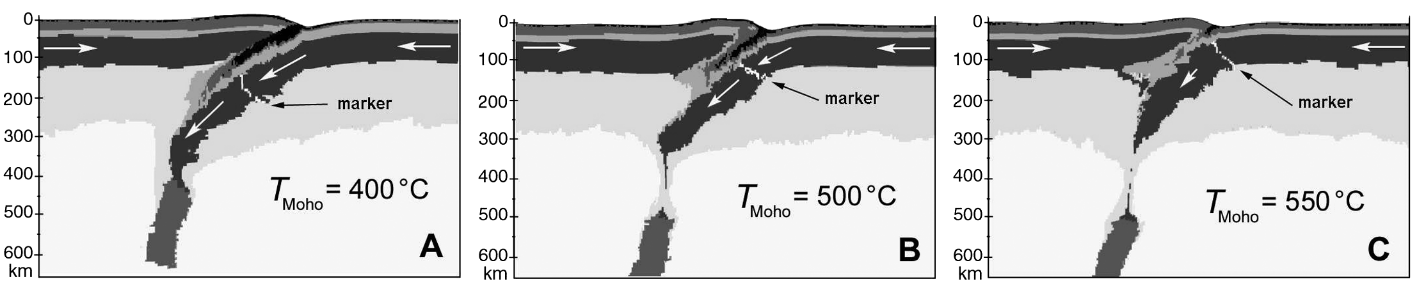

Figure 4. Experiments with strong lower crust at 5.5 m.y. Parameters and details are as in Figure 3.

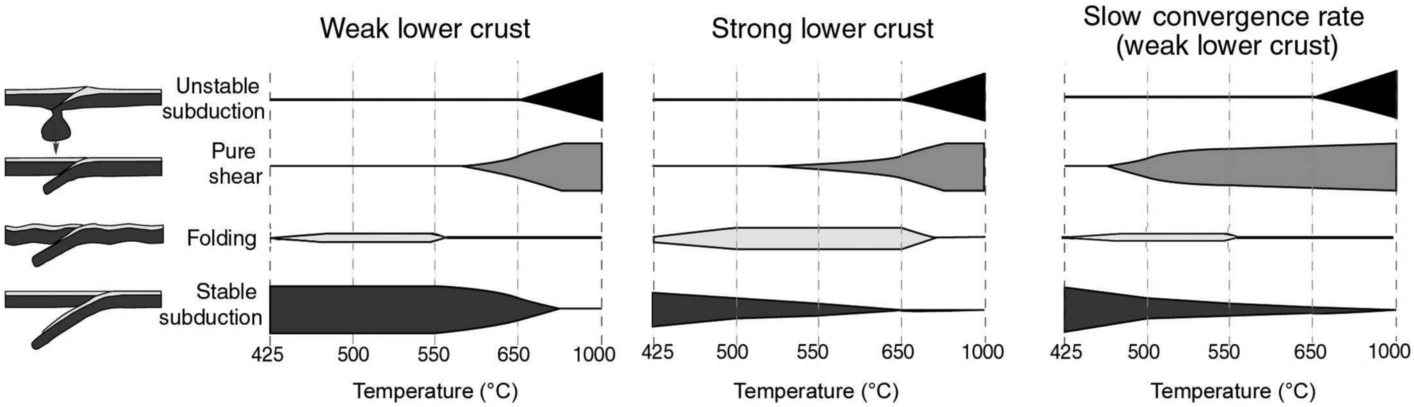

Figure 5. Summary of observed continental convergence scenarios. Diagrams show contribution of each particular deformation mechanism.

EXPERIMENTS WITH STRONG LOWER CRUST

The preceding experiments were repeated by using strong diabase rheology for the lower crust (Fig. 4). The resulting end-member scenarios (stable subduction vs. unstable subduction) are roughly the same as in the previous experiments. However, there are noticeable differences in the intermediate cases.

Cold lithosphere (Fig. 4A). For experiments with very cold lithospheres (T_{Moho} < 450 °C), the convergence yields stable subduction. However, the results of this experiment differ in many ways from its homologue with weak lower crust (Fig. 3A): in this case, subduction involves the entire continental crust, including the upper crust and its sedimentary rocks. The lithosphere also has a much higher tendency for folding.

Intermediate-temperature lithospheres (Figs. 4B and 4C). For higher Moho temperatures (T_{Moho} = 450–750 °C), stable subduction is progressively replaced by pure-shear thickening and by large-scale lithospheric folding. Folding is favored by the stronger rheology of the lower crust, which ensures its mechanical coupling with the lithospheric mantle. Note that for the same temperature range, but for a weak lower crust, subduction was a dominant mechanism of deformation (Figs. 3B and 3C).

Very hot lithosphere. The results of very warm experiments (T_{Moho} > 750 °C) are similar to those with the weak lower crust: unstable continental subduction and pure-shear thickening.

CONCLUSIONS

Compared to previous studies, the account for free upper boundary and erosion strongly modifies predictions for continental collision.

The results (summarized in Fig. 5) suggest a wide variety of scenarios for development of young collision zones, as a function of the initial convergence rate and the thermo-rheological state of the lithosphere: (1) Continental subduction is possible only for strong lithospheres characterized by T_{Moho} < 550 °C and at fast initial convergence rates (>5 cm/yr). In the case of weaker lithospheres (or slower convergence rates), alternative deformation mechanisms begin to prevail, i.e., (2) lithospheric folding (500 °C < T_{Moho} < 650 °C); (3) pure-shear thickening (550 °C < T_{Moho} < 650 °C), and (4) RT instabilities (T_{Moho} > 650 °C).

Strong diabase lower crust may allow for entrainment of light upper crust and sediments to a great depth (200 km), which is the domain of high-pressure–high-temperature metamorphism. The metamorphism influences deep slab geometry, but does not promote stable subduction, at least in weak eclogite rheologies.

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