Exhumation and subduction erosion in orogenic wedges: Insights from numerical models

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[1] At oceanic margins, syn-convergent exhumation, subduction erosion, and inter-plate coupling are intimately related, but ample questions remain concerning their interaction and individual mechanisms. To analyze these interactions for a thick-skinned, visco-elastic wedge, we focus on properly modeling stresses, energies, and topographies at the inter-plate and wedge bounding interfaces using a Coulomb frictional contact algorithm. In this innovative plane-strain, free surface, Lagrangian finite element model, fault dynamics is modulated by retreating subduction. Subduction is dynamically driven by slab-pull due to a slab sinking in a semi-analytic, computationally favorable approximation of three-dimensional induced mantle flow. Nodal trajectories show that continuous underthrusting of a slab induces a steady state corner flow through forced underplating and subsequent trenchward extrusion due to gravitational spreading. This flow pattern confirms early-proposed models of syn-orogenic deep-seated rock exhumation propelled by coexisting extension and continuous shortening at depth. A distinct reduction in upward flowing material and accompanying decrease of exhumation velocities, to millimeters per year as observed in nature, is induced by a diversion of orogenic wedge material toward the mantle once a subduction channel is formed. The key parameter affecting model evolution and spontaneous formation of a subduction channel is basal friction, which modulates the amount of erosion. However, formation of a subduction channel entrance needs to be ensured through the deformability of the overriding plate, which is influenced by applied pressure at the overriding plate tip and material properties. The down dragging of the overriding plate is sufficient above a threshold inter-plate shear stress of about 2–7 MPa.

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1. Introduction

[2] An understanding of the dynamic interactions of an orogenic accretionary wedge evolving at the will of two converging plates is vitally important for comprehending the mechanisms governing syn-convergent exhumation, accretion, and subduction erosion at oceanic convergent margins. These mechanisms are only partially understood [e.g., Agard et al., 2009; von Huene and Scholl, 1991], since limited intact or no evidence is returned to the surface.

[3] Exhumation is the relative motion of once deep-seated metamorphic rocks toward the topographic surface. This requires both the upward movement of these deep-seated rocks and the removal of the overburden [Platt, 1993]. Three distinct levels of metamorphic rock circulation have been identified by, e.g., Burov et al. [2001]; shallow wedge circulation (down to 40 km), upper crustal chamber circulation (down to 70 km), and lower crustal to mantle chamber circulation (down to 150 km). We focus on the shallowest circulation for the syn-convergent exhumation of oceanic blueschists within an orogenic wedge (for a review see Agard et al. [2009]). An orogenic wedge, as modeled here, dominantly consists of sediments accreted against the back-thrust, which belong both to the accretionary prism and part of the fore arc region.

[4] Insights into the deformation and growth of orogenic wedges were derived from analogue models [e.g., Cowan and Silling, 1978; Gutscher et al., 1998; Kukowski et al., 2002; Gladny et al., 2005; Bonnet et al., 2008; Luján et al., 2010; Konstantinovskaya and Malavieille, 2011], to amongst others facilitate the development of critical wedge theory for smaller-sized, plastic accretionary wedges [Chapple, 1978; Davis et al., 1983; Dahlen et al., 1984]. Numerical contributions can be divided into early thermo-mechanical models with a kinematically constrained geometry [e.g., Cloos, 1982; Allemand and Ladeaux, 1997; Giunchi and Ricard, 1999; Gerya et al., 2002], and those based on a basal drag or S-point approach [e.g., Willett et al., 1993; Beaumont et al., 1999]. These analogue and numerical models prescribe either their kinematics or geometry or both and thereby artificially modify the evolution of the wedge. To capture the authentic response of a wedge and downward extending subduction channel a fully dynamic approach, as implemented in e.g., Burov et al. [2001], Toussaint et al. [2004], Yamato et al. [2007], and Li et al. [2010] is more appropriate. These self-organizing, rheologically advanced models represent a big step forward in subduction zone modeling, including metamorphic changes, surface processes, and hydration. The introduction of these natural and rheological complexities, however, makes the model inherently complicated, and difficult to comprehend. The model of Yamato et al. [2007] is of specific interest, since it focuses on rheological parameters affecting shallow metasedimentary circulation and resultant exhumation at oceanic accretionary margins. Their fluid-dynamic model is, however, not continuously driven by constant, maximum slab-pull, but still requires far-field kinematic boundary conditions. Moreover, it predefines the presence of a wide subduction channel.

[5] Exhumation, and upward movement of deeply buried meta-sedimentary rocks, is intrinsically related to the deeper downward movement of these rocks. To understand how much and which material is transported to the surface, we need to understand what is instead transferred down into the mantle and eroded from the wedge through a subduction channel (a concept first contemplated by England and Holland [1979] and Shreve and Cloos [1986]). Subduction or tectonic erosion as used here refers to the process by which underthrusting of oceanic wedge material (here considered as part of the overriding block) is removed from the underplating wedge by entrainment into a subduction channel, toward the mantle [e.g., Cloos and Shreve 1988a]. The existence of subduction erosion is first convincingly observed in drillings described in the Initial Reports of the Deep Sea Drilling Project (Scientific Party, Legs 56 and 57, Washington, D.C., U.S. Government Printing Office, 1980), which are summarized in, e.g., von Huene et al. [1982]. Currently, subduction erosion is generally accepted and believed to be an important process in convergent margins [e.g., Scholl et al., 1977; Cloos and Shreve, 1988a; von Huene and Scholl, 1991; Clift and Vannucchi, 2004]. However, the structure and dynamic behavior of the subduction channel and its impact on wedge deformation and coupling remains poorly constrained.

[6] Orogenic wedge models can roughly be divided into three categories. Those who neglect sediment output toward the mantle [e.g., Cowan and Silling, 1978; Davis et al., 1983; Rossetti et al., 2000] and those that predefine a subduction channel, either using a fixed width or geometry [e.g., Kukowski et al., 1994; Mancktelow, 1995; Gutscher et al., 1998; Lohrmann et al., 2006], and those allowing for an evolving geometry [e.g., Shreve and Cloos, 1986; Cloos and Shreve, 1988a; Beaumont et al.,
In this work we present a fully dynamic, solid-mechanical, free surface approach that starts without subduction of sediment or weak material into a channel, and therefore allows for the spontaneous formation and evolution of a subduction channel. Compared to most exhumation models our model is mainly driven by slab pull from the start, allowing for a dynamically self-consistent evolution of the geometry, motions, and coupling of the inter-plate and wedge bounding faults.

The purpose of this paper is to determine the response of an orogenic wedge to large-scale, slab-pull-driven subduction dynamics in terms of wedge evolution, internal kinematics, and resulting mechanisms for syn-orogenic oceanic blueschist exhumation. These results and inter-plate coupling are affected by the dominant basal deformation mechanism, either being subduction erosion or underplating, for which we determine the controlling frictional and internal strength parameters. Finally, we also investigate the dominant energy sink of an orogenic wedge and its kinematic and geometric impact on the original subducting and overriding plate subduction model.

2. Numerical Modeling Approach

In this work an orogenic wedge is added to the two-dimensional, implicit, free surface Lagrangian subduction model developed in van Dinther et al. [2010]. In a generic intra-oceanic setting, the orogenic wedge (Figure 1a) interacts dynamically with a downgoing slab, overriding plate, and laterally unbounded mantle (Figure 1b), following the solution of the mechanical conservation equations, i.e.,...
conservation of mass and momentum, using the finite element package ABAQUS Standard [Hibbitt, Karlsson, and Sorensen, Inc., 2007]. The approach is an extension of the self-consistent single slab approaches of Funiciello et al. [2003], Morra and Regenauer-Lieb [2006], and Capitanio et al. [2007].

[10] The initial setup consists of a layered, linear viscoelastic slab that lies steadily at the 660-km discontinuity as analog of a mature subduction process. The slab is retreating due to slab pull resulting from a lithosphere-upper mantle density contrast of 80 kg·m⁻³. The overriding plate is a thin, viscoelastic oceanic lithosphere that is driven by a bottom traction of \(5.8 \times 10^{12} \text{N} \cdot \text{m}^{-1}\), representing the combined effects of mantle wedge corner flow and ridge push. Both lithospheres interact with a semi-analytic, laterally unbounded mantle in which the third-dimensional toroidal- and poloidal flow components are roughly captured by analytically calculated horizontal and vertical drag forces. Additionally, the mantle is represented by an upward Archimedes body force (displacement w.r.t. an equipotential surface) and a pressure foundation (isostasy and induced flow). The strength of the fault separating the two lithospheres is defined by a Coulomb frictional law (equation (1)), where an average shear stress up to 10 MPa is applied.

\[
\tau_{\text{crit}} = \mu \sigma_n
\]

where \(\tau_{\text{crit}}\) is yield shear stress, \(\mu\) is frictional sliding coefficient, and \(\sigma_n\) is normal pressure at the contact. This quasi-static solid-mechanical framework has a freely moving slab, trench, and inter-plate fault and is mainly driven by slab buoyancy. The employed parameters are provided in Table 1.

[11] The orogenic wedge is composed of 5183 6-node quadratic triangular elements (Figure 1a). Quadratic elements guarantee a robust solution also with large deformation, as they are not subject to locking, a fundamental problem with triangular elements employed in Funiciello et al. [2003] and Morra and Regenauer-Lieb [2006]. We model the syn-orogenic evolution of a mature wedge with an initial base of 78 km and height of 17 km (Figure 2a). This initial configuration is designed to morphologically match the self-consistent subduction geometry and is thereafter allowed to evolve.

### Table 1. Model Parameters

<table>
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<th>Parameter</th>
<th>Symbol</th>
<th>Value</th>
<th>Unit</th>
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<td>km</td>
</tr>
<tr>
<td>Dimensions overriding plate</td>
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<td>Density subducting plate</td>
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<td>Density upper mantle</td>
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*In numerical simulations applied densities are relative to 2600 kg·m⁻³.

*Averaged over depth.

*Averaged over time and depth. See Figure A1.
freely in response to the self-consistent, retreating motions of the stiffer subducting and overriding plates.

[12] We find that contact shear stresses vary from about 3 to 13 MPa for the basal interface, and from 0 to 22 MPa for the back-thrust interface. These results are extrapolated from friction coefficients (0.01 to 0.05 and 0.00 to 0.05, respectively), which correspond to modeled values of 0.1–0.5, and 0.0–0.5, respectively (see Table 2). The renormalization factor of ten results from modeled densities that are ten times lower than nature (see Appendix A and van Dinther et al. [2010]). The two wedge interfaces are formed by surface-to-surface contact formulations, which allow for arbitrary motions and separation. Numerical stability is ensured by implicit time stepping and sophisticated contact algorithms partially explained in Appendix B.

[13] The orogenic wedge rocks are composed of a melange of shales, with blocky structures containing oceanic crustal and mantle rocks, which are collectively modeled with a constant Newtonian viscoelastic bulk rheology. This rheology is justified by our interest in the large-scale average behavior of a single, mechanically continuous unit such as a thick-skinned wedge [Platt, 1987]. Neglecting small-scale deformation is a strong simplification that limits the validity of our internal

Figure 2. Integral orogenic wedge development from (a) an initial configuration depicts three phases; (b) phase 1: initial uplift with dominant underplating, (c) phase 2: coeval compression and extension with underplating and lateral displacement, and (d) phase 3: steady state circular flow with constant taper. Note that the steady state phase depicted here is taken from a different model (hence $t = 3.83$ My), compared to two reference frames; on the left compared to the geoid, and on the right compared to a moving reference (the Reference Bottom Wedge as defined in Figure 1). Colors show Mises stresses within the wedge (lithospheres are shaded) on top of that red signs indicate the stress state: + is compression and – is extension. Pink arrows indicate the main displacement direction. The reference model underwent 38% of slab-parallel shortening in one million years.

<table>
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<tr>
<th>FUL</th>
<th>FLC</th>
<th>FUC</th>
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<td>10.1</td>
<td>6.10</td>
<td>AC</td>
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*First three columns give renormalized friction coefficients. The next three columns provide measured shear stresses $\tau$ in MPa at the interplate (UL), basal (LC) and back-thrust (UC) interfaces, averaged both over contact length and time (0.0–0.48 My). The last column provides the end-member behavior mode; AC = accretion, ER = erosion. Additionally, SS denotes when a steady state configuration is reached.

*Steady state reference model.
*Accretion reference model.
*Erosion reference model.
*Wedge viscosity $\approx 5 \times 10^{19}$ Pa·s to $5 \times 10^{20}$ Pa·s.
velocity gradients, but still allows for a reliable analysis of velocity pattern within the wedge [Allemand and Lardeaux, 1997], also because the overall dynamics of viscous versus plastic rheologies is similar [Ranalli, 1995]. Furthermore, the simplified internal wedge rheology allows us to focus on the complex non-linear interactions between lithosphere and crust, while small-scale plastic localizations are mimicked by a constant, low viscosity of $5 \cdot 10^{19}$ Pa-s.

Besides this strong, but justified limitation, other limitations can be justified as follows. Strength contrasts between the moving plates and wedge, although too sharp and localized, are in size comparable to nature. This means that the lithospheres merely deform significantly slower than the wedge, and that the contact with the overriding plate along the back-thrust resembles a backstop, a common assumption when modeling orogenic wedges [e.g., Davis et al., 1983; Byrne et al., 1993; Buiter et al., 2006]. Another difference compared to the natural system is the lack of incoming sediments, which renders this model applicable to steady state mature wedges that are in a critical state in which frontal sediment accretion is balanced by internal deformation [cf. Platt, 1986]. Finally, the lack of deformation in the third lateral dimension also limits our model to routinely investigated and commonly occurring convergent margins with minor lateral changes in geometry and kinematics. Although we are fully aware that nature is more complex, our simplified model allows us to focus on the impact of fault interaction and self-consistent, retreating subduction on large-scale deformation patterns.

### 3. Results and Analysis

To understand orogenetic wedge dynamics in response to larger-scale, slab-pull-driven subduction dynamics we analyze the role of the friction coefficient on three interfaces (inter-plate, basal wedge, and back-thrust), and wedge viscosity (Table 2). Two end-member wedge behavior modes are observed in our simulations. In the accretion mode, all material is underplated below the wedge, i.e., no material leaves the supra-subduction wedge. In the erosion mode, a subduction channel is spontaneously formed and part of the wedge material is eroded toward the mantle. In the following two sections these two modes are described in more detail, rendering their geometric and kinematics features and identifying their dynamic causes.

#### 3.1. Wedge Accretion Mode

#### 3.1.1. Wedge Development

In the accretion mode the geometrical and stress evolution of an initially triangular wedge (Figure 2a) consists of three evolutionary stages (Figures 2b–2d). The characteristic development is illustrated here for a reference model with an inter-plate shear stress of 1.8 MPa, basal shear stress of 7.5 MPa, and a back-thrust shear stress of 7.7 MPa (for conversion to applied numerical friction coefficients see Appendix A).

During the initial phase, lithospheric convergence due to slab-pull causes viscous compression. Horizontal compression leads to underthrusting of wedge material, which is forced upward against a steep overriding plate and underplates the wedge (arrow 3, 4 and 7 in Figure 3a). Underplating thickens the wedge from underneath, increasing the wedge slope, and forming an outer-arc antiformal high (Figure 2b). This early evolutionary stage is in agreement with the orogenic wedge evolution described in Willett et al. [1993]. The outer-arc high or axial region is located nearly above the trench, and divides the wedge in a pro- and retro-wedge, respectively on the trench and overriding sides (following definitions of Willett [1999]).

In the second phase, the maximum elevation of the wedge remains constant, as arising longitudinal extensional stresses within the shallow axial region compensate the continuing horizontal contraction and underplating at the wedge base (Figure 2c). These extensional stresses result from a lateral flow due to a gravitational instability of an overthickened wedge slope (Figure 3a). Combined with continued underplating, and forced vertical uplift against a backstop, a biflected viscous flow arises. This flow pattern shows a reversal of velocity and transport direction with depth (Figure 3a).

In the final, third phase a steady state wedge configuration with a straight slope is reached, once the antiformal fore-bulge has spread laterally and shortening of the wedge has ceased (Figure 2d). Ongoing subduction, however, continues to exert shear stresses along the base and thereby drive a forced flow and minor, very shallow extension (Figure 3b). The critical wedge shape is a result of a critical balance between the applied basal stresses and resulting gravitational, and internal stresses [Davis et al., 1983; Dahlen et al., 1984]. The critical configuration of the model shown has a frontal
slope, $\alpha$, of 6.5°, and an average angle between subducting plate and the horizontal, $\beta$, of 13.5° (for definition see Figure 1a). These slopes coincide with observations from non-accretionary wedges, such as the Nankai wedge [Lallemand et al., 1994]. Note that, in our models, a steady state configuration is only attained, when subduction erosion is absent and a low enough basal friction is used to prevent too extreme numerical mesh distortion.

### 3.1.2. Internal Deformation

Detailed, quantitative analysis of particle trajectories relative the originally lowest point of a wedge (point A) on top of a retreating slab provides us with information on local velocities, turnover time, and exhumation location (Figures 3a and 3b). Local, upward velocities during transient stage 2 are about 1.7 cm·yr$^{-1}$, which is 35% of a subduction velocity of 4.8 cm·yr$^{-1}$ (Figure 3a). Almost six times lower velocities are observed for the steady state configuration model, which has a three times lower basal friction and a two-and-a-half times higher back-thrust friction. This reveals both absolute and relative to the surface vertical velocities of almost 0.26 cm·yr$^{-1}$ (6% of subduction velocity) and trenchward, surface velocities of 0.44 cm·yr$^{-1}$ (Figure 3b). Estimates for the turnover time of burial-exhumation cycles for a steady state wedge are in the order of 15–25 My.

In both the steady- and non-steady state cases, velocities progressively increase away from the center of circular flow (green dot, Figures 3a and 3b), unless when slowed down by friction in proximity of the overriding plate or located in two near surface corners that do not participate in the main flow (near point B and C in Figure 3a). This increase can be explained both geometrically, and by the two driving forces being located along the
interface (ongoing subduction), and free surface (extension). During transient periods, velocities decrease as material approaches the backstop and is slowly forced upward.

[22] The flow pattern shows that meta-sedimentary rocks that were originally located near the subduction interface, are forced upward in the proximity of the backstop. These rocks reach the surface at a location near the rear of the wedge, just on the retro-side of the axial region, as their flow paths are diverted from the backstop by gravitationally-driven spreading toward the trench.

3.2. Subduction Erosion Mode

[23] The evolution of the orogenic wedge is strongly affected by the spontaneous formation of a subduction channel. Initially, the evolution of an orogenic wedge follows the first two phases described for the accretion mode (Figures 2b and 2c). Once conditions for the formation of a subduction channel are met, however, orogenic wedge material can be transported toward the mantle and eroded from the wedge, instead of being underplated and accreted. In the following section, characteristics of the evolving wedge are described based on a reference model that has a different inter-plate shear stress of 7.5 MPa (Figure 4).

[24] Subduction erosion occurs through a downward thinning subduction channel with an average width of 2–3 km (Figure 4a), and eroded material reaches maximum depths of 92 km. The downward deflection and entrainment of material causes upward velocities to decrease significantly to $0.25 \text{ cm} \cdot \text{yr}^{-1} (\sim 85\% \text{ w.r.t reference model}).$ The deepest 3 km of the wedge, which are dragged toward the mantle, depict high channel velocities of $3.9 \text{ cm} \cdot \text{yr}^{-1}$ in the vicinity of the overriding plate that increase to subduction velocities near the slab. Consequently, pro-wedge surface velocities increase by 15% along the axial region to

Figure 4. Characteristics of the subduction erosion mode for a model with higher inter-plate friction ($\tau_{UL} = 7.5 \text{ MPa}$). (a) Mises stress distribution within the wedge, and (b) internal wedge kinematics from nodal point trajectories with respect to a moving reference of RBW over a period of 0.95 My. For further explanation see the captions of Figures 2 and 3.
compensate the resulting gap. Moreover, this affects the direction of flow, as throughout the entire wedge trajectories are diverted downward by about $0^\circ$–$120^\circ$. This directional, erosion promoting switch decreases the amount of rocks that underplate and are forced upward toward the surface. That significantly decreases the amount of exhumation, although an increased subsidence still allows rocks to exhume close to the backstop (compare arrow 8 and D in Figure 4b). The onset of erosion also causes subsidence of the top of the wedge (black arrow in Figure 7a).

3.3. Dynamics of Wedge Behavior

3.3.1. Role of Frictional and Material Parameters

[25] This section analyzes the parameters governing the transition from the accretion to the erosion end-member mode. The spontaneous formation of a subduction channel, and the amount of eroded material, is mainly controlled by frictional properties of the three fault interfaces (Figure 5a). Figure 5a shows the maximum vertical extent of wedge material beneath the tip of the overriding plate, as an indicator for amount of eroded material in each model. Our results demonstrate the existence of a minimum threshold of inter-plate friction between about 0.01 and 0.05. Above these frictions, rocks are able to entrain, as a channel entrance is formed by deforming and reversing the tip of the overriding plate. Increased overriding plate deformation and channel entrance formation at higher inter-plate coupling was already illustrated in models excluding a wedge, published by van Dinther et al. [2010] (Figure 5b). Beyond this threshold, basal friction is the key friction determining how much and how fast material is eroded (Figure 5a). A stronger coupled subduction interface, increasingly drags more material down. Friction at the back-thrust has an insignificant, though slightly positive, effect on the amount of eroded material (Figure 5a, compare red and green dots). The minor effect results from increased coupling, which causes an increased down drag along the top of the subduction channel.

[26] A summary of parameters promoting either underplating or subduction erosion is provided in Figure 6. Besides interface frictions, a higher viscosity, i.e., stronger orogenic wedge promotes the formation of a channel entrance (Appendix C and Figure C1), while this is inhibited by a stronger overriding plate (yellow versus purple line in two-plate model of Figure 5b). The relative strength of the orogenic wedge over the overriding plate increases the ability of the overriding plate to deform, thereby enhances the spontaneous
formation of a subduction channel entrance, and promotes the erosion of wedge material. The observation of enhanced erosion for a high strength wedge and low strength overriding plate is in agreement with numerical model results of Beaumont et al. [1999] and Yamato et al. [2007], respectively. Finally, higher tractions and trench-ward overriding plate velocities induce an upward pushed overriding plate that rolls over slightly [van Dinther et al., 2010, Figure 7b], and forms a channel entrance.

3.3.2. Wedge Height and Potential Energy

A description of the geometry of the orogenic wedge is provided in two different ways. First, we observe that the maximum absolute height reached by the wedge remains constant through time for all not distinctly eroding models, once gravitational spreading starts to counteract convergence-driven basal underplating (Figure 7a).

Secondly, we measure this steady state peak wedge height with respect to the trench $h$, as defined in Figure 1a. A positive, exponentially decaying relationship between wedge height $h$ and basal friction $FLC$, which grows until a basal friction of about 0.03, is observed (Figure 7b). As basal strength and coupling increase, more slab energy is transferred to the wedge to carry more material toward the backstop, where it is underplated and forced upward. This increases the height of the wedge, and stores the energy as potential energy. These results are in agreement with numerous numerical [e.g., Willett, 1992; Burbidge and Braun, 2002] and analogue [e.g., Gutscher et al. 1998] results. The limited vertical distribution of all models at each basal friction in Figure 7b indicates that inter-plate and back-thrust friction do not play an important role for wedge height and pro-wedge surface slope.

The potential energy $E_p$ stored in the height of wedge’s peak above the trench $h$ (right axis in Figure 7) is calculated as

$$E_p = C_{geom} P_{cugh} h$$

Figure 6. Cartoon summarizing the parameters (in red) promoting (a) accretion and underplating, and (b) entrainment and subduction erosion. Characteristics flow paths and circulation center for each mode are provided in green, while stress state is shown in purple signs. Entity colors; blue is subducting plate, yellow is overriding plate, and green is orogenic wedge.

Figure 7. (a) Depth of wedge top below geoid through time for all friction experiments, demonstrating that a steady state maximum is reached after 0.51 My (indicated by a vertical line), (b) wedge height above the trench $h$ versus basal friction $FLC$ at 0.51 My for all models that do not show distinct subduction erosion. These models are fitted with an exponential function for which an equation is given in terms of corresponding potential energy (shown in the right vertical axis).
where a triangular wedge shape is assumed \( C_{geom} = 0.5 \). The exponentially decaying relationship between basal strength and stored potential energy indicates that energy added to the system through an increase in basal shear stress, will be mainly dissipated within the wedge through internal viscous, brittle, and plastic deformation. Only a minor part is used for vertical wedge growth and potential energy storage.

### 3.3.3. Feedback With Inter-plate Coupling

[30] A quantitative comparison of inter-plate stresses for models with the same basal and back-thrust friction indicates the presence of a feedback mechanism of wedge behavior and entrainment onto inter-plate coupling. This is clear from two sets of models for which subduction erosion is induced by stepping from inter-plate friction coefficients of 0.01 to 0.05. Their corresponding inter-plate shear stress are increased with 16 and 23% less than expected from the applied linear relation between shear and normal stress (in Table 2 compare row 7 and 13, and 6 and 12, respectively). This means that the presence of a subduction channel with eroded sediments relatively reduces normal interface stresses and mechanical plate coupling, as it inhibits the direct transmission of slab pull from the slab to the overriding plate.

### 3.4. Impact on Large-Scale Subduction Geometry and Velocity

[31] The presence and frictional resistance of a small orogenic wedge does not result in apparent changes in large-scale subduction geometry (Figure 8a) and trench and subduction kinematics (Figure 8b). This is evident from comparing the current model that includes a wedge to the subducting and overriding plate model published in van Dinther et al. [2010]. Subduction is expected to speed up by a distinct decrease of stresses between the subducting and overriding plate (−150% on average), caused by the presence of an additional pressure resisting the approaching overriding plate. On the other hand, subduction is expected to slow down by an increased interface energy dissipation due to a larger interface length (+120% on average). The combination of these two opposite effects leads to a scenario in which mature subduction dynamics controls the crustal evolution at the convergent margin, with minimum feedback. This confirms
the validity of assumptions made in classical, more simplified subduction models as collectively described in, e.g., Billen [2008].

4. Discussion

This paper analyses the deformation of a free-surface orogenic wedge within a set of dynamic, frictional, slab-pull driven lithospheric boundaries. These sharp and strong, but freely moving and deforming boundaries partially postulate the means for flow inside the softer wedge. Furthermore, the numerical solution of the linear visco-elastic rheology adopted for the thick-skinned wedge cannot solve for internal faults and shear zones, resulting in a smoothened overall deformation pattern. Despite these limitations, and no incoming sediments, our model accurately quantifies stresses, energies, and topographies and thereby provides a basis for investigating long-term crustal deformation and exhumation from several levels of circulation. Hereafter, we will focus on the shallowest wedge circulation, and discuss (i) mechanisms for syn-convergent exhumation of oceanic blueschists, (ii) the parameters affecting the amount of subduction erosion and the spontaneous formation of a subduction channel, and (iii) two aspects relating to the dynamics of wedge behavior and growth in terms of inter-plate coupling.

4.1. Syn-convergent Exhumation at Oceanic Margins

The observed deformation pattern (Figure 3) suggests an effective mechanism to return deeply buried oceanic blueschists toward the surface and exhumate them in a syn-convergent scenario. The upward transport of deeply buried rocks is realized by underplating due to a forced corner flow along mechanically stronger boundaries. The removal of surface material that allows for actual exhumation is provided by coeval extension due to a gravitationally unstable slope [England, 1983]. The compelling presence of extensional stresses driving exhumation has also been observed in both numerical [e.g., Allemand and Lardeaux, 1997; Willett, 1999] and analogue models [e.g., Haq, 2008; Glodny et al., 2005], as well as in nature, both in continental orogens [e.g., Dewey, 1988; Platt, 1987; Jolivet et al., 1998] and in fore-arc wedges [e.g., Adam and Reuther, 2000].

Extension together with the observed wedge reversal in velocity direction with depth could in nature, and if a plastic or brittle rheology was included, lead to failure and formation of extensional shear zones. These would remove the overburden and result in normal-sense metamorphic gaps between the more efficiently exhumed, deep-seated rocks within the footwall, and slightly metamorphosed rocks within the hanging wall. These pressure gaps along extensional contacts are characteristic of exhumation complexes in the hinterland of orogenic regions [Platt, 1986]. This interpretation is compatible with an analogue model interpretation of Luján et al. [2010], which shows a reversal of non-coaxial transport direction with depth.

The observed composite exhumation mechanism is consistent with general ideas of the forced upward movement of material in the ‘corner flow’ concept, developed by Cowan and Silling [1978] and Cloos [1982], and extended for overcritical wedges by Platt [1986]. The resulting shallow-level circulation of meta-sediments in an orogenic accretionary wedge is comparable to numerical [Allemand and Lardeaux, 1997; Feehan and Brandon, 1999; Gerya et al., 2002; Yamato et al., 2007; Li et al., 2010] and analogue [Glodny et al., 2005] model observations.

The flow pattern and syn-orogenic mode of deep-seated rock exhumation are similar to what is predicted by Platt [1986] and Platt [1987] (compare Figure 3 to Figure 7 in Platt [1987]), providing a firm confirmation that syn-orogenic extension can be an effective exhumation mechanism in orogenic wedges. The continuous corner flow pattern is a mechanical consequence of contrasting boundaries that drive continuous underplating at depth, and extension through gravitational spreading at the top. The relative importance and continuity of the driving mechanisms, however, differs slightly from that suggested by Platt [1987]. Continuous subduction over long timescales keeps driving underplating and forced corner flow suggesting a virtually continuous process in which steady state growth and exhumation may coexist in space and time, if erosion is present. Natural occurrences of oceanic sedimentary exhumation also suggest a long-lasting process with estimates of about 50 My for the Chile and Franciscan (Western USA) complexes, and at least 25–30 My for the Western Alpine Schistes Lustrés complex [Agard et al., 2009].

The presence of exhumation during steady state growth does not, however, exclude the possibility that the majority of relative upward transport might take place during transient stages in which dynamic boundary conditions, e.g., subduction velocity and
fault coupling, change. These transient changes may follow and modulate the observed evolutionary phases. In particular, if the stored potential energy is increased by an increase in convergence velocity, major uplift (phase 1, Figure 2b) is expected, followed by more intensive extension and lateral spreading (phase 2, Figure 2c). These changes are suggested to be important in natural orogens, and natural data support a more discontinuous, episodic exhumation history of oceanic blueschist exhumation [Agard et al., 2009].

The mixed flow pattern results in exhumation near the rear of the wedge, in vicinity of the axial region of the orogenic wedge. Paths are diverted from a trajectory directly adjacent to the backstop, as predicted by pure corner flow models, by near-surface trenchward spreading. This exhumation location is in accordance with Platt’s [1986] predictions, numerical models of Yamato et al. [2007], and fits fairly well with natural occurrences as high-pressure low-temperature rocks are commonly observed near the rear of the wedge (e.g., Western Alps [Rubie, 1984], Franciscan and Sambagawa terrains [Platt, 1993]).

The observed steady state and erosion mode exhumation rates are within the range observed in nature for oceanic blueschists (about 0.1–0.5 cm yr\(^{-1}\)) according to Agard et al. [2009]). Too high velocities (around 1.7 cm yr\(^{-1}\)) for the current model parameters are only observed during the transient phases of the model. It is important to note, however, that there is a trade-off with a neglected non-linear rheology that would utilize part of the energy for locally increased stresses and deformation, instead of velocity. A steady state turnover circulation of 10–15 My is consistent with natural constraints from the Schistes Lustrés [Agard et al., 2002], Franciscan [Ring and Brandon, 1999], and Chilean complexes [Glodny et al., 2001]. Furthermore, model exhumation rates are within the same range as results of Giunchi and Ricard [1999] (0.3 cm yr\(^{-1}\)), Yamato et al. [2007] (0.1–1 cm yr\(^{-1}\) and a turnover time of 10–15 My), and smaller than often observed near subduction velocities in most pure corner flow models. These exhumation rates might be enhanced by not included processes such as erosion [Beaumont et al., 2001] and thermal buoyancy.

Summarizing, our exhumation pattern and velocities are in agreement with natural examples of fossil accretionary wedge complexes, where meta-sediments were scraped off the subducting oceanic crust. These paleoaccretionary wedges can amongst others be found in the Franciscan complex [Hamilton, 1969], the Schistes Lustrés complex [Agard et al., 2001], and the Western Series in the Valdivia area, Chili [Glodny et al., 2005]. These accretionary complexes formed while collision was still absent.

4.2. Subduction Erosion and Channel Presence

The spontaneous switch from a backstop-dominated to a channel-dominated flow due to subduction channel formation leads both to a reduction in upward velocities, and to a down-bend in path trajectories (Figure 3a versus Figure 4b). These two effects distinctly reduce the efficiency of syn-orogenic exhumation. This reduction in terms of velocity is required to bring transient exhumation rate estimates for our set of subduction velocity and frictions within the range observed in nature (about 0.1–0.5 cm yr\(^{-1}\)) according to Agard et al. [2009]). To deduce the existence and extent of a subduction channel from these results requires knowledge of friction values in nature in order to distinguish it from steady state exhumation rates. This information is not unequivocally available in the required resolution. The reduction of an upward material flow is expected and in agreement with results for activation of the conduit-mode (C) in Beaumont et al. [1999]. Ultimately, subduction erosion also leads to subsidence, which is in agreement with results for episodic basal erosion in Lohrmann et al. [2006].

The switch from dominant underplating to the occurrence of subduction erosion is mainly controlled by inter-plate friction, while the amount of erosion is mainly determined by basal friction (Figure 5). A higher wedge-slab coupling increases channel velocities and amount of eroded material, as suggested by, e.g., Cloos and Shreve [1988a] and Lallemand [1998]. Inter-plate coupling mainly acts as a minimum threshold which should be passed to ensure overriding plate deformation and downward bending to allow for the spontaneous formation of a subduction channel entrance. Two critical factors affecting the deformability of the overriding plate for channel formation are the material strength of the two competing entities and the pressure applied at the tip of the overriding plate. One can question the applicability of channel entrance formation in nature, since it is possibly related to our initial geometry and usage of explicit boundaries. Though we conjecture that in more transient, natural conditions the deformability of
the overriding plate can affect the width of the channel, increasing width and amount of eroded material with higher coupling. An extreme and temporal example of how overriding plate coupling and shape can affect backstop deformation is the entrance of a seamount. This increases interface roughness and thereby coupling, and induces slight overriding plate deformation, allowing for more material to be eroded.

4.3. Dynamics of Wedge Behavior

First, a lower than expected inter-plate normal stress increase was observed for models in which a subduction channel was present (Figure A1 and section 3.3.3). Inter-plate normal stress is thought to be the dominant stress transfer component in some regions for inter-plate stress transmission and upper plate strain regime [Heuret et al., 2011], while shear stress has also been analyzed in relation to plate interface nature by Kostoglodov [1988] and de Franco et al. [2008]. A relatively lowered inter-plate stress suggests a lower inter-plate coupling when sediments are present in the channel, as is supported by natural observations from Heuret et al. [2011]. These show the absence of compressive, highly coupled back-arc regimes in subduction zones characterized by trench sediment thicknesses larger than 2 km (assuming upper plate strain and sediment thickness at the trench as reasonable proxies for plate coupling [Lallemand et al., 2008] and subduction channel thickness, respectively). This leads us to speculate about the possible existence of a feedback mechanism between the presence and dimensions of the subduction channel, and normal inter-plate stresses. Despite the inherent difficulty to identify which is the initial triggering contribution (see Heuret et al. [2011] for a discussion), it is tempting to support the idea of a self-reinforcing mechanism in which an increase in inter-plate coupling increases the amount of subduction erosion, which in turn reduces coupling within the channel, as previously suggested by Cloos and Shreve [1988b]. This mechanism promotes less subduction erosion and channel fill, increasing inter-plate coupling once more. As direct consequence of this continuous process channel thicknesses in nature might be more inclined to be limited to a finite width, as in nature we observe maximum widths of 5 to 10 km (Shreve and Cloos [1986] and Guillot et al. [2009], respectively). We emphasize, however, that the amount of erosion is also significantly influenced by other important factors, like sediment supply and subduction velocity [e.g., Cloos and Shreve 1988a].

Secondly, we observed a positive, exponentially decaying relationship between the stored (potential) energy and the intensity of basal strength (Figure 7), as extrapolated from wedge height above the trench. Because of the linear rheology chosen for the orogenic wedge, the measured potential energy is proportional to the seismic potential available at the interface between the downgoing plate and wedge. Our model therefore offers a simple relationship between wedge height and the expected seismic energy release at a convergent margin. Such a relationship can be verified through a systematic statistical analysis of wedge uplift and subsidence, as for example done by Meltzner et al. [2006] for the 2004 M9.2 Sumatra earthquake.

5. Conclusions

In a slab-pull driven, free surface, solid mechanical subduction model with explicit frictional fault coupling, we investigate the mechanisms propelling syn-convergent exhumation at oceanic accretionary margins, basal erosion, and subduction channel formation.

After a period of dominant, rapid underplating and central uplift, a shallow wedge circulation arises, which is driven by continuous lithospheric convergence. Ongoing subduction underplates wedge material, forcing it up along the overriding plate toward the surface, where gravitational spreading causes subsidence or flattening and bends trajectories down and back toward the trench. This steady state corner flow accretionary process with coeval extension and compression is an effective mechanism for syn-orogenic oceanic blueschist exhumation, demonstrating the relevancy of the model proposed by Platt [1987]. The resulting observables show syn-orogenic exhumation occurs near the rear of the wedge, and in the presence of normal-sense pressures gaps along extensional shear zones, as observed for several paleoaccretionary wedges.

The efficiency of this exhumation mechanism is distinctly reduced to within a natural range of exhumation rates of a few millimeters per year due to the spontaneous formation of a subduction...
channel. This decreases upward velocities, and causes subsidence and an additional down-bend of particle trajectories of $0^{\circ}-120^{\circ}$ as material is transported to mantle depths instead of pushed upward.

[49] The initiation of subduction erosion through the formation of an entrance to a subduction channel is determined both by the possibility to deform and drag down the overriding plate and by the pressure applied at the tip of the overriding plate. Channel entrance formation is possible once a threshold of inter-plate shear stress of about $2-7$ MPa is passed, which is increased if the ratio of strength of the wedge over that of the overriding plate is decreased. The most important frictional parameter for the amount of erosion is basal friction, which drags material down at higher velocities if increased.

Appendix A: Renormalization of Friction Coefficients

[50] The numerical stability in our full-scale subduction simulations was hampered by realistic pressure. To consistently decrease pressures all densities have been reduced by $2600$ kg·m$^{-3}$, leaving differential densities as estimated from nature. Because the Coulomb law states that the stress on the faults is proportional to pressure and to fault friction for the same velocity, frictional coefficients have been reduced in our model to compensate for the reduced pressure and same differential stresses. Renormalized constants are given by the ratio of intended, natural wedge density over applied density, which is about 0.1 (300/2900). The renormalization procedure of friction coefficient is also explained in van Dinther et al. [2010, section 4.2.1]. All values mentioned throughout the paper renormalize the friction coefficient to natural values, and keep the observed stresses unmodified. Although one could alternatively multiply measured wedge shear stress in Figure A1 by almost 10 to acquire an estimate of natural shear stresses for applied numerical friction coefficients.

Appendix B: Contact Formulation

[51] The numerical stability of the contact formulations for wedge interaction is ensured through the usage of several ABAQUS functions. Penetration of nodes is prevented by a master surface definition for the slab and overriding plate, and transmission of an exponentially growing over-closure pressure from 0 GPa at 250 m to 10 GPa at 0 m. Contact smoothness is further improved by viscous contact damping using a damping proportionality coefficient of $7.8 \cdot 10^{16}$, which is decreasingly effective over a distance of 200 m. This option damps abrupt horizontal and tangential relative motions during approach and separation. Additional damping of all motions within the...
wedge is attained through the use of small-value, non-dissipative dash-pots. Finally, friction between external surfaces of the wedge itself is neglected.

Appendix C: Wedge Viscosity

[52] The effect of an increased wedge strength is analyzed by increasing viscosity by one order of magnitude (Figure C1). These results show that a stronger wedge deforms the tip of the overriding plate more easily, and thereby enhances channel entrance formation and the amount of subduction erosion.

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