Geodynamic models of Cordilleran orogens:
Gravitational instability of magmatic arc roots

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ABSTRACT

Cordilleran orogens, such as the central Andes, form above subduction zones, and their evolution depends on both continental shortening and oceanic plate subduction processes, including arc magmatism and granitoid batholith formation. Arc and batholith magma compositions are consistent with partial melting of continental lithosphere and magmatic differentiation, whereby felsic melts rise upward through the crust, leaving a high-density pyroxenite root in the deep lithosphere. We study gravitational removal of this root using two-dimensional thermal-mechanical numerical models of subduction below a continent. The volcanic arc position is determined dynamically based on thermal structure, and formation of a batholith-root complex is simulated by changing the density of the arc lithosphere over time. For the model lithosphere structure, magmatic roots with even a small density increase are readily removed for a wide range of root strengths and subduction rates. The dynamics of removal depend on the relative rates of downward gravitational growth and lateral shearing by subduction-induced mantle flow. Gravitational growth dominates for high root densification rates, high root viscosities, and low subduction rates, resulting in drip-like removal as a single downwelling over 1–2.5 m.y. At lower growth rates, the root is removed over >3 m.y. through shear entrainment as it is carried sideways by mantle flow and then subducted. In all models, >80% of the root is removed, making this an effective way to thin orogenic mantle lithosphere. This can help resolve the mass problem in the central Andes, where observations indicate a thin mantle lithosphere, despite significant crustal shortening and thickening.
INTRODUCTION

The central Andes represents the type example of a Cordilleran orogen, in which crustal shortening has produced a continental mountain belt above an active subduction zone. Geological studies indicate that the central Andes was primarily built through 200 to >500 km of shortening of the western South American plate during Cenozoic subduction of the oceanic Nazca plate (e.g., McQuarrie et al., 2005; Oncken et al., 2006; Barnes and Ehlers, 2009). This resulted in the formation of the Altiplano-Puna Plateau, an internally drained plateau with an average elevation of ~4 km. Seismic receiver functions show present-day crustal thicknesses of 50–80 km below much of the plateau (Beck et al., 1996; Yuan et al., 2002; Beck and Zandt, 2002; McGlashan et al., 2008; Bianchi et al., 2013). At present, subduction continues at a plate convergence rate of 7–8 cm/yr, and shortening is localized on the eastern side of the plateau (e.g., Brooks et al., 2003; Oncken et al., 2006).

During orogenesis, shortening of the upper crust should be accompanied by thickening of the deeper lithosphere. However, several observations indicate that the mantle lithosphere is not anomalously thick beneath most of the central Andes. Seismic tomography studies show that many parts of the orogen have anomalously low velocities in the shallow mantle (<100 km depth; e.g., Myers et al., 1998; Beck and Zandt, 2002; Schurr et al., 2006; Bianchi et al., 2013). The regions of lowest velocities underlie regions of recent volcanism (e.g., northern Puna and eastern Altiplano). In addition, receiver functions show that the lithosphere-asthenosphere boundary is at 100–150 km depth below the central Altiplano, indicating a 30–50-km-thick mantle lithosphere (Heit et al., 2008). Other indicators of a thin, hot mantle lithosphere include elevated 3He in groundwater of the mantle lithosphere (Heit et al., 2008). Other indicators of a thin, hot mantle lithosphere include elevated 3He in groundwater of the Altiplano (Hoke et al., 1994), the occurrence of magmatism across the width of the plateau (e.g., Trumbull et al., 2006; Kay and Coira, 2009), and high crustal temperatures from surface heat flow (Springer and Forster, 1998; Springer, 1999) and seismic velocities (e.g., ANCORP Working Group, 2003).

One explanation for the lack of a thick lithosphere, despite significant upper-crustal shortening, is that the mantle lithosphere was anomalously thin prior to orogenesis. Geological evidence indicates that this region was close to sea level during the early Cenozoic, at the start of orogenesis (Sempere et al., 1997). If the lithosphere was hot and thin at that time, a surface elevation greater than 1 km would be expected, owing to thermal isostasy (e.g., Hyndman and Currie, 2011). The alternate possibility is that the mantle lithosphere has undergone thinning during orogen development. Observations of pulses of basaltic and andesitic magmatism (e.g., Kay et al., 1994; Kay and Coira, 2009; Duca et al., 2013) and periods of abrupt surface uplift from paleo-elevation data (e.g., Garzione et al., 2006) have been interpreted to reflect episodic removal of mantle lithosphere, and possibly lower crust, during orogenic shortening. In addition, seismic tomography images show small-scale (50–100-km-wide) high-velocity bodies at ~100 km depth below the Puna region (Schurr et al., 2006; Bianchi et al., 2013), which have been interpreted as fragments of detached continental lithosphere.

In this contribution, we investigate the dynamics of lithosphere removal in Cordilleran orogens. Various mechanisms for removing lithosphere have been proposed for the central Andes (Fig. 1). Continuous removal may occur through ablation (Fig. 1A), as continual mantle lithosphere in the mantle wedge corner is entrained by the oceanic plate through viscous drag at a rate that balances orogen shortening (Tao and O’Connell, 1992; Pope and Willett, 1998). However, ablative removal is a continuous process that does not result in an overall thin lithosphere. This is at odds with observational evidence for episodic removal events in the central Andes. These observations appear to require gravitational foundering of the lithosphere. Removal can be driven by the negative buoyancy of the mantle lithosphere, as it is cooler and therefore denser than the underlying material. Eclogitization and densification of the lower crust during orogenic shortening, and especially during magmatism, may provide an additional driving force (e.g., Kay and Kay, 1993; Duca and Saleeb, 1998; Jull and Kelemen, 2001; Sobolev and Babeyko, 2005; Krystopowicz and Currie, 2013; Wang et al., this volume). Two distinct modes of gravitational foundering are: (1) Rayleigh–Taylor-type (RT) instability ("drip"; Fig. 1B), possibly induced by lithospheric shortening combined with magmatic extraction at deep levels under the arcs (e.g., Houseman et al., 1981; Houseman and Molnar, 1997; Molnar et al., 1998) and (2) delamination (Fig. 1C), in which mantle lithosphere peels along a weak crustal detachment layer (e.g., Bird, 1979). The two modes can be differentiated through patterns of surface deformation, uplift, and magmatism (Göğüş and Pysklywec, 2008), and both have been applied to explain observations of magmatism and surface uplift for the Altiplano-Puna Plateau (e.g., Kay and Kay, 1993; Garzione et al., 2006; Molnar and Garzione, 2007; Kay and Coira, 2009; Duca et al., 2013).

In Cordilleran orogens, subduction-related magmatism at the volcanic arc or in the back-arc region may also induce lithosphere removal (Fig. 1D). Arc volcanism is produced through melting of the mantle wedge above a subducting plate at a depth of 100–150 km. A key problem is that arc magmas extruded at the surface have an andesitic and dacitic, not basaltic, composition (e.g., Castro et al., 2010) have been proposed to explain this paradox, but the isotopic record of all major Cordilleran arcs rules out such an origin (e.g., Duca and Barton, 2007). Mantle-derived magmas undergo differentiation during ascent (e.g., Hildreth and Mooar, 1988; Duca and Barton, 2007), especially in the lowermost parts of the crust, in what is referred to as the MASH zone (mixing-assimilation-storage and homogenization zone; Hildreth and Mooar, 1988). Upwelling
basaltic magmas stagnate at the base of the crust, where melt fractionation and assimilation of upper-plate lithosphere result in felsic magmas that are the source of granitoid batholiths and andesitic-dacitic volcanism. This leaves a high-density garnet-bearing or garnet-free pyroxenite (the garnet-bearing type being an eclogite rock sensu lato) residue in the deep lithosphere below the arc, which is then prone to gravitational removal (e.g., Kay and Kay, 1993; Lee et al., 2006).

Isotopic studies show that Cordilleran batholith formation is not a steady-state process but instead occurs during 5–15 m.y. high-flux events during which the magmatic flux is 3–4 times higher than in the intervening periods (e.g., Ducea, 2001; DeCelles et al., 2009, and references therein). Isotopic data also indicate continental lithosphere within the magma source during a high-flux event (e.g., Ducea, 2001; Ducea and Barton, 2007). These data have been explained using a model in which upper-plate shortening introduces fertile continental mantle lithosphere into the volcanic arc region, fueling a magmatic flare-up (e.g., Ducea, 2001; DeCelles et al., 2009). This lithosphere is partially melted and differentiated into a felsic melt, which rises up to form the batholith and a residual “root,” which is primarily pyroxenitic/eclogitic (arclogitic) with lesser amounts of granulite (feldspathic) material. Subsequent foundering of the pyroxenitic part of the root is thus an important process for removing continental lithosphere.

At both island arcs and Cordilleran arcs, removal of dense roots has been inferred from geological field evidence, xenolith data, and seismic tomography studies (e.g., Ducea, 2002; Saleeby et al., 2003; Zandt et al., 2004; Behn and Kelemen, 2006). However, there have only been limited geodynamic studies of the removal process. Jull and Kelemen (2001) modeled removal for island arcs as a Rayleigh-Taylor drip and concluded that this can occur over time scales of less than 10 m.y. if temperatures are relatively high (>500 °C at the Moho for shortening lithosphere). Their models did not include sublithospheric mantle flow, which, as we show herein, has an important effect on the dynamics of gravitational instabilities. Behn et al. (2007) considered the foundering arc roots as falling spheres. Their models demonstrated that large, dense spheres strongly perturb slab-induced corner flow, leading to complex three-dimensional (3-D) flow that may explain observations of trench-parallel seismic anisotropy in the arc region.

In this study, we use numerical models to study the formation and removal of a dense mafic-ultramafic root at a Cordilleran volcanic arc. Using a simplified parameterization as a proxy for the development of the root, we examine the dynamics of its removal within an active subduction environment. We then assess the implications for Cordilleran orogen evolution.

### NUMERICAL MODELS OF SUBDUCTION

#### Model Geometry and Governing Equations

The thermal-mechanical numerical models are regional-scale two-dimensional vertical cross sections through a subduction
The model domain has a width of 2000 km and extends from Earth’s surface to a depth of 660 km. The initial geometry of the numerical models is shown in Figure 2A. The oceanic lithosphere is 90 km thick, with a 9 km crust. The continental plate is also 90 km thick, and crustal thickness varies from 48 km to 42 km. In this study, we examine the volcanic arc dynamics for a case without simultaneous shortening and formation of the orogen. Therefore, we use a prethickened crust (48 km) in the vicinity of the subduction zone, which results in pressures of ~1.4 GPa at the base of the crust, well within the stability field for the development of a garnet pyroxenite residue (Ducea, 2002; Saleeby et al., 2003).

The finite-element method is used to calculate the coupled thermal-mechanical evolution of the lithosphere–upper-mantle...
system, under the assumptions of plane strain, incompressibility, and zero Reynolds number. The governing equations are: (1) conservation of volume when incompressible, (2) force balance, and (3) energy balance:

$$\frac{\partial \nu}{\partial x} = 0, \quad (1)$$

$$\frac{\partial \sigma}{\partial x} + \rho g = 0, \quad (2)$$

$$\rho C_v \left( \frac{\partial T}{\partial t} + \nu_i \frac{\partial T}{\partial x} \right) = k \frac{\partial^2 T}{\partial x^2} + A_x + \sigma' \dot{\varepsilon}_0 + v_x \dot{\varepsilon}_0, \quad (3)$$

where $\nu_i$ are spatial coordinates ($ij = 1,2$), $\nu_i$ are components of velocity, $\rho$ is density, $g$ is (vertical) gravitational acceleration, $C_v$ is specific heat, $T$ is absolute temperature, $t$ is time, $k$ is thermal conductivity, $A_x$ is volumetric radioactive heat production, and $\sigma_0$ is the volumetric thermal expansion coefficient. Repeated indices imply summation. In Equation 3, the last two terms on the right-hand side correspond to shear heating, assuming that all dissipated mechanical energy associated with deformation is converted to heat (term 3), and the temperature correction for adiabatic heating for vertical velocity $v_z$ (term 4).

The associated stress tensor is:

$$\sigma_{ij} = -P \delta_{ij} + \sigma'_{ij} = -P \delta_{ij} + 2\eta_{\text{eff}} \dot{\varepsilon}_{ij}, \quad (4)$$

where $P$ is pressure (mean stress), $\sigma'_{ij}$ is the deviatoric stress tensor, $\eta_{\text{eff}}$ is effective viscosity, $\delta_{ij}$ is the Kronecker delta (1 for $i = j$ and 0 otherwise), and the strain rate tensor is:

$$\dot{\varepsilon}_{ij} = \frac{1}{2} \left( \frac{\partial \nu_i}{\partial x_j} + \frac{\partial \nu_j}{\partial x_i} \right). \quad (5)$$

These equations are solved using arbitrary Lagrangian-Eulerian (ALE) finite-element techniques (Fullsack, 1995), subject to the boundary conditions described in the following discussion and internal buoyancy forces. Mechanical and thermal calculations are carried out on an Eulerian mesh that stretches vertically to conform to the upper model surface. The Eulerian mesh has 200 elements in the horizontal direction (10 km width) and 108 elements vertically (3 km height in the upper 180 km, and 10 km height below). Material properties are tracked on a Lagrangian mesh and additional Lagrangian tracer particles that are advected with the model velocity field. The Lagrangian particles are used to update the Eulerian model material distribution at each time step. In the calculations, the thermal and mechanical fields are coupled through the temperature dependence of material densities and viscous rheologies, the shear and adiabatic heating terms in Equation 3, and redistribution of radioactive heat–producing materials by material flow.

The numerical modeling code (SOPALE) has been fully benchmarked for studies of gravitational instabilities (e.g., Pysklywec et al., 2002). Our own tests show that the finite-element mesh used here can resolve the growth rate of Rayleigh-Taylor instabilities to within 6% of the analytic values (e.g., Houseman and Molnar, 1997). Additional tests with a viscous Stokes cylinder show that there is less than 7% difference in the sinking velocity computed using the preferred mesh and one with 2 km square elements.

### Material Properties

Table 1 lists the thermal-mechanical properties of each model material in the reference model shown below. Materials have a viscous-plastic rheology and a temperature-dependent density. Frictional-plastic deformation follows a Drucker-Prager yield criterion:

$$J' = P \sin \Phi + c_0 \cos \Phi \phi_{\text{eff}}, \quad (6)$$

where $J'$ is the square root of the second invariant of the deviatoric stress tensor ($J' = \frac{1}{2} \sigma'_{ij} \sigma'_{ij}$), $c_0$ is the cohesion, and $\phi_{\text{eff}}$ is the effective internal angle of friction, which includes the effects of pore-fluid pressure (e.g., Huismans and Beaumont, 2003; Beaumont et al., 2006). Plastic deformation is modeled by defining an effective viscosity that places the state of stress on yield (Fullsack, 1995; Willett, 1999). All materials undergo frictional-plastic strain softening through a decrease in $\phi_{\text{eff}}$ from 15° to 2° over accumulated strain ($\Gamma_z$) of 0.5–1.5, as an approximation of material softening or an increase in pore-fluid pressure during deformation (Huismans and Beaumont, 2003).

At stresses below frictional-plastic yield, deformation follows a viscous power-law rheology, with effective viscosity ($\eta_{\text{eff}}^\nu$) given by:

$$\eta_{\text{eff}}^\nu = f(B^*) (\dot{\gamma}^1)^{(1+\eta)/\eta} \exp \left( \frac{Q+PV^*}{nRT_k} \right), \quad (7)$$

where $f$ is a scaling factor (see following), $\dot{\gamma}$ is the square root of the second invariant of the strain rate tensor ($\dot{\gamma}^2 = \frac{1}{2} \dot{\varepsilon}_{ij} \varepsilon_{ij}$), $R$ is the gas constant (8.3145 J mol$^{-1}$ K$^{-1}$), and $B^*$, $n$, $Q$, and $V^*$ are the pre-exponential viscosity parameter, stress exponent, activation energy, and activation volume from laboratory data. The parameter $B^*$ includes a conversion from the uniaxial laboratory experiments to the tensor invariant state of stress used in the models (Table 1).

The materials in our models are based on several well-constrained laboratory-derived viscous rheologies, and we use the scaling factor $f$ (Eq. 7) to linearly scale the effective viscosity of the model materials relative to these base rheologies to approximate changes in strength owing to minor changes in composition or degree of hydration (Beaumont et al., 2006). For example, experimental data for olivine show a nearly linear decrease in effective viscosity with increasing water content (Hirth and Kohlstedt, 2003, and references therein), such that dry olivine is 5–10 times stronger than water-saturated
olivine at a depth of 50–100 km (Karato and Wu, 1993; Hirth and Kohlstedt, 2003). The chosen rheologies and scaling factors follow those used in earlier studies (e.g., Beaumont et al., 2006; Krystopowicz and Currie, 2013). The oceanic crust uses the parameters of dry Maryland diabase (Mackwell et al., 1998), with \( f = 0.1 \), assuming that the crust is hydrated, and therefore weaker than the base rheology. For simplicity, the entire continental crust has a wet quartzite viscous rheology (Gleason and Tullis, 1995), with \( f = 5 \), to approximate a stronger composition than pure wet quartzite (Beaumont et al., 2006). Mantle materials follow a wet olivine rheology (Karato and Wu, 1993), with \( f = 1 \) in the sublithospheric mantle and \( f = 5 \) and \( f = 10 \) in the continental and oceanic mantle lithosphere, respectively. The higher values of \( f \) reflect dehydration and melt depletion of the mantle lithosphere during formation. Figure 2B shows the variation in effective viscosity with temperature for the chosen rheologies.

Modeling Approach and Boundary Conditions

Models are run in three phases. In phase 1, the initial two-dimensional (2-D) thermal structure of the models is computed using the material thermal properties (Table 1), temperatures of 0 °C and 1560 °C for the top and bottom boundaries of the model domain, and a no-heat-flux (insulating) condition for the side boundaries. This yields temperatures of 830–950 °C for the continental Moho (Fig. 2A) and a continental surface heat flux of 65.5–71.5 mW/m², with the higher values associated with the region of thicker crust. The models then undergo isostatic adjustment, to allow the oceanic and continental plates to come into equilibrium. As a result, the oceanic plate sinks, such that its surface is at a depth of ~5.2 km.

In phases 2 and 3, plate convergence and subduction occur. The thermal and mechanical boundary conditions are given in Figure 2A. The thermal boundary conditions consist of prescribed temperatures along the top boundary (0 °C) and bottom boundary (1560 °C), and a no-heat-flux (insulating) condition for the side boundaries of the continental lithosphere and sublithospheric mantle. The side boundary of the oceanic lithosphere has prescribed temperatures that are given by a geotherm for a 50-m.y.-old oceanic plate (Stein and Stein, 1992). The mechanical boundary conditions include a stress-free top boundary and...
a free-slip bottom boundary. Plate convergence at 7 cm/yr is imposed through assigned velocities for the oceanic (5 cm/yr) and continental (2 cm/yr) plates at the side boundaries of the model. To maintain mass balance in the model domain, a small uniform outflux ($V_{\text{out}}$) through the side boundaries of the sublithospheric mantle occurs, equally distributed on each boundary. Models are solved in the continental reference frame by adding 2 cm/yr to all side boundaries. No surface processes (e.g., erosion) are included in the models.

In phase 2, subduction is initiated. Subduction is aided by a narrow, inclined zone of weak material between the oceanic and continental plates (Fig. 2A). This material is subducted with the oceanic plate and does not affect later model evolution. In addition, a high viscous strength is assigned to the continental crust ($f = 50$; Eq. 7) and continental mantle lithosphere ($f = 10$). This phase is run for a total of 10 m.y. (total convergence of 560 km). All the model experiments shown herein start at this point (phase 3), and times are reported as the time since the start of phase 3. During this phase, the viscous strength of the continental crust and mantle lithosphere is set to the reference values ($f = 5$; Table 1), and magmatic processes at the volcanic arc are imposed.

**SUBDUCTION ZONE BASE MODEL**

We first present a model in which magmatic processes are not included. Figure 3A shows the model at the end of phase 2, after subduction has been established. The oceanic plate descends into the mantle along a well-defined shear zone, with little deformation of the overlying continent. At this time, the tip of the subducted plate is at the bottom of the model domain and is deflected horizontally along this impermeable boundary. Over the next 3–4 m.y., the oceanic plate undergoes retreat as a consequence of the reduction in the upper-plate strength at the start of phase 3. The trench shifts seaward by ~90 km (Fig. 3B), and there is minor distributed extension in the upper plate. In the mantle wedge corner (above a slab depth of 60–150 km), there is a slight decrease in the dip angle of the slab from ~40° to ~35°. After this, the subduction zone stabilizes, and there is steady subduction with little change in geometry.

Figure 3. Evolution of a model with no volcanic arc. The entire model domain is shown on the left, and a close-up of the mantle wedge corner is shown on the right. Model parameters are those used in the reference model (Table 1). Material shading is same as in Figure 2. Times are relative to the start of phase 3 of the models (10 m.y. after subduction initiation).
Figure 3C shows the model at 15 m.y., which is the end of the model run. At this point, 1750 km of plate convergence has occurred, and the subducted slab has reached the side boundary of the model domain. Interactions with this boundary produce buckling of the slab in the deep part of the model domain, but there is little effect on the mantle wedge corner.

During model evolution, subduction and the associated mantle wedge flow cause shearing of the continental mantle lithosphere, resulting in minor ablation and thinning in the mantle wedge corner. Overall, the mantle lithosphere is relatively stable, with a thickness of ~15 km in this area.

**IMPLEMENTATION OF MAGMATIC PROCESSES**

Arc magmatism and the formation of a dense pyroxenite root are the result of a series of complex processes within the subduction zone. As the oceanic plate subducts, metamorphic phase changes in the subducting oceanic crust and mantle lithosphere lead to the release of aqueous fluids into the overlying mantle wedge, reducing its solidus temperature and promoting partial melting (e.g., Schmidt and Poli, 2003; Tatsumi, 2005). The basaltic melts then intrude the continental lithosphere, where fractional crystallization of the melt and localized partial melting of the host rocks result in a silicic melt, which rises to form crustal batholiths and surface magmatism, and a pyroxenite residue root, formed as either a cumulate or restite. Phase equilibria calculations of Jull and Kelemen (2001) predict that root densities are 50–250 kg/m³ greater than that of mantle, depending on composition. Direct samples of pyroxenite xenoliths from the Sierra Nevada arc have densities of 3500–3600 kg/m³ (300 kg/m³ more dense than mantle), owing to their high garnet content (Ducea, 2002).

We do not attempt to model the details of arc magmatism but instead focus on the development of a high-density eclogitic pyroxenite root during a high-flux event at a Cordilleran arc (Fig. 1D). High-flux events are inferred to reflect times of enhanced magmatism related to the emplacement of melt-fertile continental lithosphere below the arc (Ducea, 2001; DeCelles et al., 2009), and therefore root formation through differentiation of continental lithosphere should be greatest during these events. We use a simplified model to simulate the development of the root. This is implemented at each time step in model calculations using a two-step approach: (1) determination of the location of the active volcanic arc (this is the region where localized basaltic melts are assumed to intrude the deep continental lithosphere), and (2) densification of the continental lithosphere in the arc region.

In the models, the arc is assumed to overlie the area where both: (1) the upper 30 km of the subducting plate is at a temperature less than 800 °C and (2) the mantle wedge has a temperature greater than 1200 °C (Fig. 4). The first condition accounts for the maximum stability temperature of hydrous phases in the oceanic mantle (e.g., chlorite) and allows for kinetic delay in dehydration reactions in oceanic crust and subducted sediments (e.g., Schmidt and Poli, 2003; Hacker et al., 2003). The latter condition is based on the solidus temperature for partially hydrated mantle (e.g., Schmidt and Poli, 2003) and constraints on arc magma source temperatures from geochemical analyses (e.g., Kelemen et al., 2003). By tying the definition of the volcanic arc to the thermal structure of the subduction zone, the location of the arc can migrate during model evolution. As shown in Figure 4 and subsequent figures, the arc region is located above a slab depth of ~80 to ~150 km, as observed in nature (e.g., Tatsumi, 2005; Syracuse and Abers, 2006).

The second step is to change the density of the lithosphere in the volcanic arc region to simulate magmatic differentiation, i.e., the formation of a high-density eclogite root and the emplacement of low-density felsic magmas within the arc crust. It is assumed that the eclogitic pyroxenite root resides in the continental mantle lithosphere below the arc (~15 km thick in our models) and that this entire region progressively becomes denser owing to differentiation through partial melting. The simplified model assumes that infiltration of small-volume basaltic mantle-derived melts produces sporadic, localized zones of partial melting in the continental lithosphere. The time scales of arc formation are tens of millions of years or more for large Cordilleran arcs. The Sierra Nevada arc, for example, consists of plutons with ages ranging from 230 Ma to ca. 80 Ma, and all of these plutons are confined to a surface area that is ~120 km wide. While the arc did form in a non-steady-state fashion, with short-lived 10–15 m.y. high-flux events separated by longer periods of lower magmatic flux (magnetic lulls), it is clear that the arc was built through incremental addition of mantle-derived melts and small batches of felsic differentiates in the upper crust (Coleman et al., 2004). The lifetime of a plutonic suite such as the classic Tuolumne Suite in the Yosemite region, central Sierra Nevada, is in the range of 10 m.y. (Coleman et al., 2004), and the composite pluton was assembled...
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via small batches of melt at any given time. That observation implies that the root itself, while hot, was for the most part solid at any given time, and contained only small areas of partial melt. Melt fractions in excess of 20%, and in some cases up to 50%, are documented in studies of xenoliths representing the pyroxenitic part of the root (Lee et al., 2006), but they represent local areas of high-percent melt at any given time. Similarly, one-dimensional (1-D) thermal models show that basaltic intrusion may produce only limited melting of the arc complex (Annen and Sparks, 2002). Thus, at the million-year time scale or more, the root can be considered solid during the evolution of the arc.

In nature, the rate of root densification depends on the rates of basaltic magma intrusion, partial melting of the surrounding rock, melt differentiation, and extraction of the felsic melt component. Constraints on this come from the apparent intrusive flux of plutonic rocks and arc magmas. Typical island arcs have a flux rate of 20–40 km³/m.y. per kilometer along strike (Reymer and Schubert, 1984); this range is commonly referred to as 1 Armstrong Unit (AU) (DeCelles et al., 2009). At Cordilleran arcs, the rate during a high-flux event is 3–4 times greater than this (Ducea, 2001). Assuming that the high-flux event corresponds to the sum of the background mantle-derived magmatic flux (1 AU) and melt productivity associated with the incorporation of fertile continental lithosphere into the magma (2–3 AU), the volume of felsic intrusives derived from continental lithosphere is only limited melting of the arc complex (Annen and Sparks, 2001). Thus, at the million-year time scale or more, the root can be considered solid during the evolution of the arc.

The reference model uses the material properties in Table 1, and the density of the arc root region increases at 20 kg/m³/m.y. The assumption is that the process of root formation only involves short-lived small-volume partial melts (e.g., Annen and Sparks, 2002; Coleman et al., 2004), and therefore the average root region can be considered as a solid. Figure 5 shows the evolution of this model. With rollback and slight shallowing of the slab during the first ~4 m.y. of phase 3, the arc location migrates seaward, and its width increases to 80–90 km. After this, it remains relatively stationary. As the model progresses, the base of the root is sheared by subduction-induced mantle flow, which causes a slight perturbation and initiates gravitational instability of the dense arc root. In the early stages of instability, the downward velocity is fairly slow, and therefore the perturbation is entrained by mantle flow and carried toward the wedge corner. The growth rate is initially exponential, but once the strain rate associated with instability exceeds the strain rate of shearing, the growth rate becomes superexponential, owing to the power-law rheology of this material (Molnar et al., 1998; Currie et al., 2008). As a result, the downward velocity of the root increases, and the root rapidly descends through the mantle wedge. It detaches from the upper plate and is carried downward with the subducting plate.

Overall, the removal process is rapid and efficient. The root is removed within 8 m.y. of the start of magmatic processes, and the dripping event (i.e., when the downward velocity exceeds the rate of lateral entrainment) occurs within 3 m.y. of the onset of instability. The instability involves nearly the entire root region. After removal, only a thin layer (<5 km) remains below the arc crust; some of this material is back-arc mantle lithosphere that was carried into the arc region by mantle flow.

RESULTS: MODELS WITH A HIGH-DENSITY ARC ROOT

Reference Model

The rheologies of the root and batholith are the same as those of the original material (mantle lithosphere and crust, respectively). The assumption is that the process of root formation only involves short-lived small-volume partial melts (e.g., Annen and Sparks, 2002; Coleman et al., 2004), and therefore the average root region can be considered as a solid. Figure 5 shows the evolution of this model. With rollback and slight shallowing of the slab during the first ~4 m.y. of phase 3, the arc location migrates seaward, and its width increases to 80–90 km. After this, it remains relatively stationary. As the model progresses, the base of the root is sheared by subduction-induced mantle flow, which causes a slight perturbation and initiates gravitational instability of the dense arc root. In the early stages of instability, the downward velocity is fairly slow, and therefore the perturbation is entrained by mantle flow and carried toward the wedge corner. The growth rate is initially exponential, but once the strain rate associated with instability exceeds the strain rate of shearing, the growth rate becomes superexponential, owing to the power-law rheology of this material (Molnar et al., 1998; Currie et al., 2008). As a result, the downward velocity of the root increases, and the root rapidly descends through the mantle wedge. It detaches from the upper plate and is carried downward with the subducting plate.

Overall, the removal process is rapid and efficient. The root is removed within 8 m.y. of the start of magmatic processes, and the dripping event (i.e., when the downward velocity exceeds the rate of lateral entrainment) occurs within 3 m.y. of the onset of instability. The instability involves nearly the entire root region. After removal, only a thin layer (<5 km) remains below the arc crust; some of this material is back-arc mantle lithosphere that was carried into the arc region by mantle flow.

The rate of root formation. For a 150–300 kg/m³ density contrast (Jull and Kelemen, 2001; Ducea, 2002) and a root area of 1200 km², the volume of felsic intrusives derived from continental lithosphere is only limited melting of the arc complex (Annen and Sparks, 2001). Thus, at the million-year time scale or more, the root can be considered solid during the evolution of the arc.

In nature, the rate of root densification depends on the rates of basaltic magma intrusion, partial melting of the surrounding rock, melt differentiation, and extraction of the felsic melt component. Constraints on this come from the apparent intrusive flux of plutonic rocks and arc magmas. Typical island arcs have a flux rate of 20–40 km³/m.y. per kilometer along strike (Reymer and Schubert, 1984); this range is commonly referred to as 1 Armstrong Unit (AU) (DeCelles et al., 2009). At Cordilleran arcs, the rate during a high-flux event is 3–4 times greater than this (Ducea, 2001). Assuming that the high-flux event corresponds to the sum of the background mantle-derived magmatic flux (1 AU) and melt productivity associated with the incorporation of fertile continental lithosphere into the magma (2–3 AU), the volume of felsic intrusives derived from continental lithosphere is 40–120 km³/m.y./km. Given a melt to residue ratio of 1:1–1:3 for a Cordilleran batholith (Ducea, 2001; DeCelles et al., 2009), it is estimated that the root will form at a rate of 40–360 km³/m.y. (per km along strike), as felsic components are extracted.

This is modeled by increasing the average (bulk) density of the continental mantle lithosphere within the arc root region (width of ~80 km, thickness of ~15 km), assuming that the entire region undergoes homogeneous melt extraction and densification. The root zone initially has the density of continental mantle lithosphere, 3250 kg/m³, and the rate of densification is:

$$\frac{\delta \rho}{\delta t} = \frac{(\rho_p - \rho_m) R_i}{A_i},$$

where \(\rho_p - \rho_m\) is the density contrast between pyroxenite and mantle, \(A_i\) is the cross-sectional area of the root, and \(R_i\) is the rate of root formation. For a 150–300 kg/m³ density contrast (Jull and Kelemen, 2001; Ducea, 2002) and a root area of 1200 km², the rate of densification is 5–90 kg/m³/m.y. In the models, the densification rate is a free parameter. The reference model shown next uses a value of 20 kg/m³/m.y., and later we present models with densification rates of 10–80 kg/m³/m.y. During each time step, the density of the root region is increased by the prescribed amount. At the same time, the average density of the volcanic arc crust (“batholith”) is decreased at a rate that maintains mass balance in the models. This approximates the emplacement of felsic magmas in the shallow crust.

Magmatic processes are implemented at the start of phase 3 of the models, i.e., after a well-developed subduction zone is established. The start of this phase is taken to be the time at which basaltic melts start to intrude the continental mantle for this cross section through a subduction zone. This may represent either the along-strike migration of arc volcanism to this location or the reestablishment of magmatism in a region that underwent prior root removal, followed by rapid upper-plate shortening and the emplacement of fertile lithosphere below the arc. Note that the models do not include magmatic addition to the arc lithosphere. Rather, we assume that the main contribution to the root density comes from partial melting and differentiation of in situ continental lithosphere, as required by isotopic data for Cordilleran arcs (Ducea, 2001; DeCelles et al., 2009). Presumably, the heat source for partial melting comes from magmas derived from the mantle wedge, which are not explicitly included in our models.
Variations in Root Densification Rate

The root densification rate is a free parameter in the models, and it is estimated that it may vary from 5 to 90 kg/m³/m.y. in nature (see previous). Figure 6 shows models with densification rates of 10 and 40 kg/m³/m.y., keeping all other parameters the same as in the reference model. In both cases, the base of the root is sheared by mantle flow, causing an initial perturbation. The growth rate of the perturbation is proportional to the density contrast between the root and underlying material (Houseman and Molnar, 1997). With a densification rate of 10 kg/m³/m.y. (Fig. 6A), the growth rate is sufficiently slow that it does not exceed the rate of lateral entrainment. In this case, the root is swept into the mantle wedge corner and is removed through an ablation-like process as it is mechanically subducted with the subducting plate. This occurs over ~4 m.y., and approximately half of the thickness of the root is removed. The shallower part of the root remains intact and is not entrained in the early stages of the model, likely because it has a high viscosity due to lower temperatures. After the initial removal event, the remaining root is gradually thinned through shearing and ablation.

In contrast, a higher densification rate of 40 kg/m³/m.y. leads to a greater instability growth rate, enabling the initial perturbation to rapidly amplify (Fig. 6B). The root falls nearly vertically through the wedge and has a drip-like appearance, with a width of ~20 km. In comparison to the reference model, the removal event occurs over a shorter time and involves a greater area of root. The residual material in the tail of the instability is carried into the mantle wedge corner and is subducted, leaving very little root in the arc region. As the dense root is removed, the crust within the arc is slightly thickened (5 m.y. panel) and then relaxes after removal.

Lower-Viscosity Root

In these models, the root has a rheology that is 5 times stronger than wet olivine, which approximates the rheology of relatively dry olivine (Karato and Wu, 1993). However, this material may be weaker, due to factors such as higher temperatures related to magma emplacement in the arc root, the presence of melt, or a change in composition as the root region becomes garnet pyroxenite. Two laboratory studies have examined the rheology of dry eclogite (Jin et al., 2001; Zhang and Green, 2007). Their flow laws predict effective viscosities that are 2.5–5 times lower than that used in the reference model (Fig. 2B).

Figure 7 shows two models in which the root region is taken to have a wet olivine rheology with $f = 2$, which has a similar viscosity to that of the Jin et al. (2001) eclogite. The unaltered continental mantle lithosphere has the same rheology as the reference model (wet olivine with $f = 5$). With the weaker rheology, the root is easily sheared and perturbed by mantle wedge flow. With a densification rate of 20 kg/m³/m.y. (Fig. 7A), the perturbation grows at a relatively slow rate, and the root is carried into the wedge corner by mantle flow as it detaches. This removes
approximately two-thirds of the root area, and the remaining root is removed through later shearing. By increasing the densification rate to 40 kg/m³/m.y. (Fig. 7B), the growth rate of the instability is increased, enabling a drip-like instability to form. This occurs earlier than in the comparable model with a stronger root (Fig. 6B).

Models with a weaker rheology (wet olivine with $f = 1$) experience even greater shearing of the root by mantle flow, with shearing velocities of ~4.2 cm/yr, i.e., ~0.4 cm/yr more than for $f = 2$. For a densification rate of 20 kg/m³/m.y., root removal occurs primarily through sideways entrainment (similar to Fig. 7A), whereas higher densification rates result in removal with a greater vertical velocity (similar to Fig. 7B).

**Higher-Viscosity Root**

We now test models in which the root viscosity is twice that of the reference model (wet olivine with $f = 10$). Rheological studies of eclogite show that the strength of eclogite may be controlled by the presence of omphacite, which is much weaker than garnet (Jin et al., 2001; Zhang and Green, 2007). As there is no omphacite in samples from the Sierra Nevada root (e.g., Ducea, 2002), pyroxenite roots may be stronger than predicted by the eclogite flow laws. Alternatively, a stronger rheology may correspond to a root that is dehydrated or where root temperatures are 50–100 °C cooler than in our models (Karato and
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With the greater strength, the lithosphere is less easily perturbed by mantle flow, and the growth rate of the ensuing instability is decreased (Fig. 8). For densification rates of 20 and 40 kg/m³/m.y., the root removal time is increased relative to models with a weaker rheology. In both cases, the perturbation is initially swept sideways by mantle flow but then falls along a near-vertical trajectory once its growth rate is high enough. This removes almost all root material. In addition, the stronger root rheology leads to greater viscous coupling between the root and overlying arc crust, leading to slight crustal thickening above the detaching root.

Variations in Subduction Rate

The previous models demonstrate that mantle wedge corner flow plays an important role in the gravitational removal of the arc root region. We now investigate variations in the subduction rate, as this is what drives mantle wedge flow. For these models, the continental plate velocity is fixed at 2 cm/yr, as in the reference model. The oceanic plate velocities of 3 cm/yr and 7 cm/yr are examined, giving subduction rates of 5 cm/yr and 9 cm/yr, respectively; the reference model has a subduction rate of 7 cm/yr. In these models, the root parameters are those...
in the reference model (wet olivine with $f = 5$; densification rate of 20 kg/m$^3$/m.y.).

With a lower subduction rate (Fig. 9A), the initial lithosphere perturbation and development of the instability are slightly delayed. However, once instability occurs, the root falls nearly vertically through the wedge and has a more drip-like appearance than in the reference model (Fig. 5). In this case, the downward velocity exceeds the mantle wedge flow velocity. In contrast, root instability develops at an earlier time with a higher subduction rate (Fig. 9B). This is a consequence of both greater shearing on the base of the root, as well as a minor decrease in the root viscosity, as the root has a non-Newtonian rheology, such that its viscosity decreases with increased strain rate (Currie et al., 2008). However, the initial growth rate of the instability is not large enough to exceed the rate of mantle flow, and the perturbation is carried into the wedge corner before detaching.

**DISCUSSION**

**Dynamics of Arc Root Removal**

The models presented here highlight the range of dynamics that may occur as a high-density root forms below the frontal arc of a Cordilleran subduction system. We have tested a range of root
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Figure 9. Model evolution for a subduction rate of: (A) 5 cm/yr and (B) 9 cm/yr. All other parameters are those of the reference model, with a root densification rate of 20 kg/m³/m.y. and root rheology of wet olivine (WO) ×5. Material shading is same as in Figure 2 and Figure 4.

The removal of the dense arc root occurs primarily as a gravitational instability, driven by the density contrast between the root and the mantle. The initiation of root removal is defined as the time at which the root region has a negative (downward) velocity at a depth of 60–63 km (i.e., the base of the root). As shown in Figure 10A, initiation of removal occurs earlier in models with a high densification rate. Removal is initiated within 2–3 m.y. after the start of densification for rates of 60–80 kg/m³/m.y. The rheology of the root also controls the onset of instability; stronger roots tend to have a later initiation time. The effect of rheology is most significant at low densification rates. In all models, initiation occurs when the density of the root is 50–250 kg/m³ greater than that of the mantle (corresponding to absolute densities of 3300–3500 kg/m³), within the range of densities for arc-related pyroxenites (Jull and Kelemen, 2001; Ducea, 2002). Once initiated, root removal occurs within 5 m.y., with more rapid removal for roots with a greater density contrast relative to the mantle and for roots with a weaker rheology (Fig. 10B).

The removal of the dense arc root occurs primarily as a gravitational instability, driven by the density contrast between
TABLE 2. SUMMARY OF MODEL RESULTS

<table>
<thead>
<tr>
<th>Root rheology</th>
<th>Rate of densification (kg/m³/m.y.)</th>
<th>Initiation of removal (m.y.)</th>
<th>Root density at initiation (kg/m³)</th>
<th>Duration of removal (m.y.)</th>
<th>Removal style*</th>
<th>Figure number</th>
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<td></td>
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<td>Shear</td>
<td>7A</td>
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<td>3401</td>
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<tr>
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</tr>
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<tr>
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<td>9A</td>
</tr>
<tr>
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</tr>
<tr>
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<td>4.4</td>
<td>3338</td>
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<td>Shear</td>
<td>9B</td>
</tr>
</tbody>
</table>

Note: WO × f = wet olivine rheology (Karato and Wu, 1993), scaled by factor f (Eq. 7).

* Determined by comparing the horizontal velocity (V_h) and vertical velocity (V_v) at the base of the arc root lithosphere during root removal. Shear entrainment has V_h > V_v throughout removal; drip-style removal has V_v > V_h in the latter stages of removal.

Figure 10. (A) Time of initiation of removal of the arc root for different densification rates (horizontal axis) and root strengths (symbols). Dashed gray lines are the density contrast of the root relative to a mantle density of 3250 kg/m³. (B) The duration of removal for each model, as a function of the average density contrast between the root and mantle during removal. Black symbols use a subduction rate of 7 cm/yr; gray symbols use the rate given on the figure. Solid symbols are models that were removed as a drip-like instability; open symbols are those that underwent shear entrainment. Model results are listed in Table 2.
the root and underlying mantle. However, subduction-induced mantle wedge flow plays a key role in both the onset and dynamics of root removal (Fig. 11A). Mantle flow shears the base of the root, which causes a perturbation that initiates instability (Currie et al., 2008). The perturbation then grows in amplitude, at a rate determined by the density contrast between the root and underlying mantle and the viscosity of the root (Houseman and Molnar, 1997). As it grows, the perturbation is entrained by mantle flow and carried toward the wedge corner. The effect of mantle flow is greatest for weak roots, as they are more easily perturbed and readily entrained by mantle flow.

Two styles of removal are observed: (1) drip-like removal, in which the root falls subvertically through the mantle wedge, and (2) shear entrainment, in which the root is swept into the wedge corner by mantle flow. The style of removal is determined by the relative rates of gravitational growth of the instability and shearing by mantle flow. Drip-like removal occurs if gravitational growth dominates. These two styles are similar those of Behn et al. (2007), who approximated a detached arc root as a falling sphere. In their analysis, spheres with a large negative buoyancy (due to high density and/or large diameter) had a vertical trajectory, whereas spheres with a lower negative buoyancy were entrained by mantle wedge flow.

Figure 11B summarizes the style of removal observed for different root densification rates and strengths (viscosities) in our models. Drip-like removal is favored for roots with a high densification rate and high viscosity. Removal in this manner occurs within 1–2.5 m.y. of the onset of instability (Fig. 10B), which is consistent with the time scale observed for gravitational removal of a 1–10-km-thick eclogite layer in the models of Jull and Kelemen (2001). In most of these models, the average density contrast between the root and mantle is >150 kg/m³ during removal. Shear removal takes up to 5 m.y., with longer times associated with stronger roots. These models have a lower density contrast relative to the mantle. It should be noted that our models are based on a wet olivine rheology. A comparison of olivine and eclogite rheologies indicates that eclogite may have a rheology similar to the weaker models in our study (Fig. 2B). In this case, we predict that drip-like removal will occur for root densification rates of 40 kg/m³/m.y. or more (Fig. 11B).

Scaling Analysis for Root Removal

These results can be quantified by comparing the time scales for gravitational growth and shear entrainment. The main factor driving removal is gravitational instability of the high-density root. At the same time, mantle flow shears the base of the arc lithosphere, carrying the gravitational instability toward the wedge corner. In order for removal to occur as a drip, gravitational growth of the instability must result in a downward velocity that exceeds the horizontal velocity of the underlying mantle before the perturbation reaches the mantle wedge corner.

Figure 11. (A) Schematic diagram showing the relationship between root dynamics and subduction-induced mantle wedge flow. Flow perturbs the base of the root, initiating instability. The trajectory of the root as it destabilizes is determined by the relative rates of downward gravity-driven instability growth \( V_v \) and horizontal entrainment by mantle flow \( V_h \). (B) Diagram showing the style of removal for variations in root densification rate and viscosity (as a linear scaling of the base wet olivine [WO] rheology) for a subduction rate of 7 cm/yr. Drip-like removal (filled circles) occurs where \( V_v \) is greater than \( V_h \) within the volcanic arc region. Shear entrainment (white squares) occurs for \( V_h > V_v \). Dashed lines show the boundary between style of removal for variations in subduction rate. The shaded gray region is the strength of dry eclogite from laboratory studies.
For a layer sheared by mantle flow, gravitational instability initially has an exponential growth rate (Molnar et al., 1998; Currie et al., 2008) given by:

\[ q = \frac{c \Delta \rho g h}{2 \eta_z}, \]  

where \( c \) is a nondimensional factor, \( \Delta \rho \) is the density contrast between the root layer and underlying material, \( g \) is gravitational acceleration, \( h \) is the thickness of the layer undergoing instability, and \( \eta_z \) is the layer viscosity (e.g., Houseran and Molnar, 1997). At a time \( t \) after initiation, the downward velocity of the instability is:

\[ V_s(t) = (Z_0q)e^{qt}, \]  

where \( Z_0 \) is the amplitude of the perturbation that initiates instability (Houseran and Molnar, 1997).

We define the gravitational threshold time \( t_s \) as the time at which \( V_s = 7 \) cm/yr, as this corresponds to the maximum horizontal velocity of mantle wedge flow for a 7 cm/yr subduction rate. The arc root is approximated as a layer with constant density contrast and viscosity, and we assume an initial perturbation of 10% of the layer thickness. As the arc root layer that undergoes instability is 5–15 km thick, variations in density and viscosity related to temperature are relatively small. In this case, \( c = 0.32 \) (Houseran and Molnar, 1997). We neglect the transition from exponential to superexponential growth that occurs due to the non-Newtonian rheology of the arc root (Molnar et al., 1998). We also assume that Equations 9 and 10 apply throughout the evolution of the instability, although numerical experiments by Houseran and Molnar (1997) show that the growth rate of a Newtonian material rapidly increases once the perturbation amplitude equals the original layer thickness. These two factors result in enhanced instability growth, and therefore our calculated \( t_s \) values are overestimates.

Figure 12 shows \( t_s \) as a function of density contrast for different combinations of layer thickness and viscosity. The root viscosity is the product \( f \times 10^{19} \) Pa s. This is consistent with the average viscosity in the lowermost root of the numerical models. In all cases, \( t_s \) decreases with increasing density contrast and layer thickness and decreasing root viscosity, as expected based on Equation 9.

The second critical time scale is that associated with lateral shearing of the root. As the perturbation grows gravitationally, it is swept sideways by mantle flow. We define the threshold time for shearing \( (t_s) \) as the time for a perturbation to be carried a horizontal distance of 80 km, the nominal width of the volcanic arc region. The rate of shearing depends on both the velocity of mantle wedge flow (set by the subduction rate) and the viscosity of the arc root, with a greater shearing for weaker (lower \( f \)) material (Currie et al., 2008). In our models, the observed shearing velocity is \(-4.2 \) cm/yr for a root with \( f = 1 \), resulting in \( t_s = 1.9 \) m.y. (horizontal gray line on Fig. 12). As \( f \) increases, the shearing velocity decreases and \( t_s \) increases; values of 2.1 m.y., 3.7 m.y., and 4.7 m.y. correspond to \( f = 2, 5, \) and 10, respectively.

To determine the style of root removal, \( t_s \) is compared to \( t_c \). For \( t_s < t_c \), the gravitational growth rate is sufficiently high that the root material is removed as a drip before it reaches the mantle wedge corner. Where \( t_s > t_c \), shear entrainment occurs. Our calculations (Fig. 12) predict that a 5-km-thick root with \( f = 1 \) rheology requires a density contrast of at least 90 kg/m³ in order to be removed gravitationally. For \( f = 10 \), drip-like removal requires greater layer thicknesses and densities.

This analysis can be used to understand the styles of removal observed in the models. In models with a high densification rate, drip-like removal was observed for all root viscosities that we tested (Fig. 11B). In these models, the rapid increase in root density produces a high growth rate and low \( t_s \), allowing gravitational removal of the root (Fig. 12). For a modest root densification rate (20–40 kg/m²/m.y.), the key factor controlling the style of removal is the strength of the root. For a weak root (low \( f \)), the greater shearing by mantle flow leads to low \( t_s \). The initial
perturbation of the root and onset of instability occur earlier, when the root has a relatively low density (Fig. 10A), and a thinner layer that participates in the instability (Currie et al., 2008). As a result, the drip has a low gravitational growth rate, despite its low strength, and removal occurs through shear entrainment ($t_s < t_r$). In contrast, a stronger root is less susceptible to perturbation by mantle flow, resulting in a later perturbation, and thus higher density and growth rate and a reduced rate of shearing, and thus drip-like removal ($t_s < t_r$).

**Effect of Subduction Rate**

As observed in the models, another important parameter controlling root dynamics is the rate of subduction-induced mantle flow. As shown on Figure 10B, a lower subduction rate results in drip-like removal over a larger range of root densities and viscosities. In this case, the initial perturbation of the root is delayed due to low flow velocities (cf. Figs. 5 and 9A), and therefore the initiation of removal occurs at a higher density, producing higher gravitational growth rates. More importantly, the reduced shearing leads to a large $t_s$, and the root has sufficient time to be removed as a drip. With a higher velocity, $t_s$ is reduced, and the root is rapidly carried into the wedge corner before it can grow gravitationally, except in cases where its density rapidly increases.

**Assumptions and Limitations of the Models**

The root zone in our models corresponds to the full thickness of the continental mantle lithosphere (~15 km thick), and its density is increased using a simplified parameterization that simulates the formation of a garnet pyroxenite residue. The assumption is that the entire region undergoes homogeneous densification at a constant rate, and that the root contains only small volumes of partial melt that are rapidly extracted. In nature, this is a complex process, and differentiation occurs heterogeneously in both space and time. Clearly, magmatism is focused under arcs in centers that are spaced typically at 50–80 km along the strike of the arc. This is seen both in active arcs where stratovolcanoes are aligned at this spacing, as well as in old arcs, where subsurface batholiths are exposed. These factors should result in small-scale variations in density and viscosity within the root region. These variations are below the resolution of our models, and it is unclear how they will affect the dynamics of the root.

Furthermore, our models do not include addition of mantle-derived magmas to the arc lithosphere, although there is an implicit assumption that magma intrusion provides the necessary heat for partial melting that leads to lithosphere differentiation. For Cordilleran arcs, up to 50% of the mass of the arc may be derived from mantle wedge magmatism (Ducea, 2002; DeCelles et al., 2009, and references therein). Using a basaltic magmatic flux of 20–40 km²/m.y. per kilometer along strike and a melt to residue ratio of 1:1–1:3 (DeCelles et al., 2009), the mantle-derived magmas would produce an eclogite pyroxenite root that thickens at 0.25–1.5 km/m.y. for an 80-km-wide arc. This should enhance the potential for gravitational removal of the entire root complex, but the removal dynamics will depend on the distribution of basaltic melts in the root zone (i.e., whether they underplate or intrude). Behn et al. (2007) argued that a 2–4 km layer of underplated material should be removed within 1–10 m.y., which is on the same time scale as the instabilities observed in this study. Models that include magmatic addition are the focus of ongoing work. Such models will also be important for understanding dynamics of island arcs, where there is less involvement of upper-plate materials in arc magmatism.

Two other factors that may affect the behavior observed in the models are the geometry of the subducting plate and the structure of the continental lithosphere. In our models, the subduction geometry and velocity are nearly constant during model evolution. Temporal variations in these parameters are expected to affect the dynamics of the root. In particular, an episode of low-angle subduction could enable a dense arc root to remain in place until the slab is removed. Flat-slab and ultrashallow subduction under the arc, corresponding to the Laramide orogeny, has been proposed to explain the ~70 m.y. delay between the end of magmatism and inferred root removal for the Sierra Nevada arc in California (Saleeby et al., 2003; Zandt et al., 2004).

The structure of the continental lithosphere will also affect the dynamics of root removal. In the volcanic arc region, our models have a hot, prethickened orogenic crust, and the root forms within the underlying ~15-km-thick continental mantle lithosphere. The overall thermal structure of the orogen lithosphere (>900 °C at the arc Moho) is consistent with petrological and seismic constraints from modern arcs (Kelemen et al., 2003), and the crustal rheology is based on a wet quartzite flow law, consistent with seismic studies indicating that the central Andes has a dominantly felsic crust (e.g., Beck and Zandt, 2002; ANCROP Working Group, 2003). If the arc crust were weaker than in our models, possibly due to local heating by intruding melts and the presence of the melts themselves, it may be easier for the arc root to detach, as the viscous coupling between the root and crust is decreased. A weaker crust may also be more easily entrained during root removal, leading to thickening of the deep crust (e.g., Pyskylywec and Beaumont, 2004; Wang et al., this volume). In contrast, the root removal process may be less efficient for regions where the continental crust is thinner or more mafic, or the lithosphere is cooler (and therefore thicker) than assumed here. A thinner, mafic crust and/or lower Moho temperature would result in an increase of both the root viscosity and the coupling between the crust and root, making it more difficult for the root to be removed. In addition, if the lithosphere were thicker, the root may occupy only the uppermost mantle lithosphere (top 10–20 km), and its removal may be further hindered by the underlying lithosphere. Additional models that explore a range of lithosphere structures are needed to determine how the removal style and amount of root removed are affected.

A major limitation of the models is that they are two dimensional. This may have several important consequences. For
example, in 2-D models, a gravitational instability grows by pulling in material from within the model plane. In contrast, a 3-D instability involves material from the volume surrounding the initial perturbation, allowing instabilities to grow faster (e.g., Kaus and Podladchikov, 2001; Hasenclever et al., 2011). Furthermore, in 2-D models, mantle wedge flow is restricted to the model plane, and therefore, it can easily shear the base of the root. In 3-D models, flow may be diverted around any perturbation in the overriding lithosphere (e.g., Hasenclever et al., 2011). With the enhanced growth rates and reduced shearing, 3-D arc roots may favor a drip-like style of removal for a wider parameter space than predicted by the 2-D models (Fig. 11B).

Implications for Cordilleran Orogens

The formation and removal of dense eclogitic residues at volcanic arcs have been argued to be a fundamental processes in the evolution of Cordilleran orogens (DeCelles et al., 2009, and references therein). The dense root appears to occupy much of the lower lithosphere under arcs, where it is inferred that fertile continental mantle lithosphere undergoes melting and differentiation during high-flux events that have a duration of 5–15 m.y. (Ducea, 2001; DeCelles et al., 2009). Our models demonstrate that even a small increase in density of continental mantle lithosphere can result in its destabilization and removal for a wide range of root strengths and subduction rates (Fig. 11B). In the models, removal occurs within 15 m.y. of the start of densification, consistent with the duration of high-flux events. Over 50% of arc root material can be removed, making this an effective way to dispose of continental mantle lithosphere from a Cordilleran orogen. This may be an important mechanism to resolve the mass problem in Cordilleran orogens, where the mantle lithosphere is much thinner than expected on the basis of crustal shortening rates (e.g., Wernicke et al., 1996).

The models provide insight into surface observables that may be associated with this process. As root material is removed relatively quickly, it does not thermally equilibrate with the surrounding mantle wedge, and, therefore, it may be detected in seismic tomography studies as a high-velocity region (e.g., Beck and Zandt, 2002; Schurr et al., 2006; Bianchi et al., 2013). In our models, this material has different geometries, from 20- to 30-km-wide drips to small-scale (<10 km) features that are stretched and distorted by mantle flow (e.g., Fig. 6). If removal is primarily through shear entrainment, the size of the features is likely below the resolution of most seismic tomography studies.

The removal of a dense arc root may also be reflected in the surface topography. Figure 13 shows the surface elevation at the volcanic arc for models with different root densification rates. These models have the reference rheology structure (Table 1) and a convergence rate of 7 cm/yr. In the early stages of the models, there is little variation in surface topography. As the volcanic arc root detaches, the surface undergoes uplift of 300–900 m, as an isostatic response to the removal of the dense root. Surface uplift is greater for models with a high root densification rate, as it is assumed that this is associated with a greater rate of crustal emplacement of low-density felsic magmas. The formation and drip-style removal of magmatic roots may be reflected by the anomalous surface deflections that are recorded in local “bobber basins” in the central Andes, such as the Arizaro Basin (DeCelles et al., this volume). One interesting observation is that uplift occurs during drip-style removal, but if the root is removed through shear entrainment, there is little surface response until after removal. In drip-style removal, the greater vertical velocity of the root may lead to stronger entrainment of the ductile lower crust, resulting in crustal thickening above the descending root, which may explain the earlier onset of uplift (e.g., Fig. 6B; Pysklywec and Beaumont, 2004; Wang et al., this volume).

Two additional surface observables are variations in surface magmatism and orogenic shortening. Magmatism is directly associated with the formation and removal of the eclogite pyroxenite root. Isotopic signatures and magma flux rates have been used to argue for the formation of arc roots during high-flux events (e.g., Ducea, 2001; DeCelles et al., 2009). Root removal is thus constrained to take place after major high-flux events. Recent numerical models suggest that the root material itself may undergo partial melting as it descends (Elkins-Tanton, 2007). Ducea et al. (2013) demonstrated that this may explain the upper-plate chemical signature that is observed in mafic magmas from central Andes. This requires that the edge of the descending root is conductively heated above its solidus within the time scale that it takes for the material to be removed. Elkins-Tanton (2007) showed that sufficient heating can occur within 1–2 m.y. for a descending lithosphere drip, similar to the time scale of removal in our models. However, the details of melting depend on the

Figure 13. Relative surface elevation of the volcanic arc region for models with the reference material parameters (Table 1), subduction rate of 7 cm/yr, and different densification rates for the arc root. The shaded regions show the time scale for removal (Table 2).
initial temperature structure and composition of the removed lithosphere. A quantitative assessment of the relationship between the model dynamics and the composition of arc root material is needed to address whether significant melting could occur during arc root removal.

DeCelles et al. (2009) argued that the overall shortening rate of the orogen may be modulated by the formation and removal of arc roots. In their conceptual model, shortening rates decrease as continental mantle lithosphere fills the mantle wedge corner. As melt-fertile lithosphere enters the volcanic arc region, it undergoes partial melting and differentiation during a high-flux event. Subsequent gravitational removal of the root clears the wedge corner, enabling renewed shortening of the orogen and the introduction of new fertile continental material to the arc region. In addition, surface uplift associated with root removal (Fig. 13) may cause an outward propagation in the locus of orogenic shortening. In this manner, Cordilleran orogens may develop through a cyclical pattern of eclogite root buildup and removal, which may be recorded in the magmatic and shortening history of the orogen. High-flux events are observed to occur every 25–50 m.y. for Cordilleran orogens, which is interpreted as the time needed for continental mantle lithosphere to be thrust below the arc during orogen shortening (DeCelles et al., 2009). Our models assume that continental mantle lithosphere is already in place at the start of a high-flux event, which corresponds to a time of 0 m.y. in our models. They do not include upper-plate shortening, and therefore they do not address the relationships among orogen shortening, arc root removal, and the temporal spacing of high-flux events. This will be explored in the next generation of models.

CONCLUSIONS

The numerical models in this study have addressed one aspect of the proposed Cordilleran orogenic cycle (DeCelles et al., 2009), namely, the dynamic removal of a high-density eclogitic root under the frontal arc. We used a simplified parameterization to simulate chemical differentiation of continental lowermost crust and mantle lithosphere in the arc region, resulting in the formation of a dense, gravitationally unstable, garnet pyroxenite root. For the continental lithosphere structure used in this study, foundering of this material occurs readily for a wide range of densification rates, root viscosities, and subduction rates. Removal occurs within 15 m.y. of the start of densification, which is consistent with the 5–15 m.y. duration of high-flux events at Cordilleran arcs (Ducea, 2001; DeCelles et al., 2009). The dynamics of removal are strongly affected by subduction-driven flow in the underlying mantle wedge. This flow creates the initial perturbation that induces instability, and it can entrain the dense root as it destabilizes. Two styles of removal are observed: (1) drip-like removal, in which the root is removed as a single downwelling over 1–2.5 m.y.—this is observed for high densification rates, strong roots, and low subduction rates; and (2) shear entrainment, where the lower part of the root is swept sideways by mantle flow and then subducted with the oceanic plate—this is a more gradual process that can take >3 m.y. In both cases, nearly the entire root region can be removed (>80% for the parameters tested here). In general, our models demonstrate that with minor densification, the root region of a volcanic arc is susceptible to gravitational foundering, making this an efficient way to remove mantle lithosphere from a Cordilleran orogen. By clearing the lithosphere from the mantle wedge corner region, this process may modulate the overall shortening rate of the orogen and may be observable in magmatic, paleoelevation, and seismic data (DeCelles et al., 2009; Ducea et al., 2013).

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