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SUPPLEMENTARY INFORMATION

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Supplementary information:

Our 1080 km x 360 km model setup includes a 7 km thick oceanic crust adjacent to a 60 km thick, 250 km wide orogenic crust. Both the oceanic and the continental crust overly a homogeneous mantle reservoir. A compressible layer of air-like, low viscosity and low density overlies the model. This layer allows the model to move up and down to maintain isostatic equilibrium. A weak zone is introduced between the continent and the ocean to simulate the weak accretionary wedge and its underlying serpentinized oceanic crust and mantle. The thermal state of the model setup is such that the base of the orogenic crust is at 950^{o}C whereas the 1330^{o}C isotherm is at ca 45 km depth underneath the ocean, gradually deepening to 122 km underneath the continent. We introduce a thermal and compositional anomaly underneath the Cordillera simulating the hydrated, weaker and buoyant mantle wedge lying above the Benioff plane. Thermal and mechanical parameter values for all materials are listed in Table 1.

Away from the subduction zone, a constant velocity condition in the direction of the fixed continental plate is attached to a small section of the computational grid, transmitting motion to the oceanic lithosphere. Therefore, flow and dynamics at the junction between ocean and continent are self-consistent. No particle is added or removed from the model. In order to test whether or not volume forces alone can trigger extension, boudinage and the opening of marginal sea, a key feature of our model is that it does not include a subducting slab. A free slip boundary condition is attached to the upper and lower surface of the model as well as along its walls. There is no heat flow across the vertical sidewalls.

Each material has a visco-plastic rheology with a temperature and stress dependent viscosity for stresses below the yield stress, and a depth dependent plastic branch above the yield stress. We use power law relationships between strain rate and stress to describe

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dislocation creep. The viscosity varies with the temperature and stress according to: $\eta = \frac{1}{2} A^{-1/n} \cdot Exp(\frac{E}{n \cdot R \cdot T}) \cdot \dot{\varepsilon}^{(1-n)/n}$. For the upper continental crust, the lower continental crust and the mantle we use *A*, *n* and *E* from quartz-rich rocks (Brace and Kohlstedt, 1980), dry diabase (Mackwell et al., 1998) and olivine (Brace and Kohlstedt, 1980) respectively. Above the yield stress (τ), depth dependent plastic flow is described by a Mohr-Coulomb failure criteria: $\tau = C_0 + \mu_{eff} \cdot (\rho \cdot g \cdot z)$, where *Co* is the cohesion, μ_{eff} is the effective coefficient of friction, ρ is the density, *z* is the depth and *g* the gravitational acceleration. For all material, a strain-weakening factor lowers the yield stress linearly down to a maximum of 20% of its initial value at an accumulated plastic strain of 120% (cf. Winjs et al., 2005 for details). Finally, for semi-brittle effect with impose a maximum yield stress of 250 MPa for the upper crust, 350 MPa for the lower crust, 380 MPa for the mantle, and 100 MPa for the weak region in between the continental and the ocean.

Each material has a solidus and liquidus over which its viscosity linearly decrease over 2 orders of magnitude for the mantle (Hirth and Kohlstedt, 2003) and 3 orders of magnitude for the crust. The bulk of the decrease occurs in between 20 and 30% melt for the crust, and in between 3 and 8% melt for the mantle. In addition, the density of partially molten regions decreases with the melt fraction. As the melt fraction increases from the solidus to the liquidus, the density decreases by 13% for both the crust and the mantle.

We assign to the hydrated mantle wedge a reference density, at room temperature and pressure, of 3350 kg.m^{-3}, 20 kg.m^{-3} lower that the non-hydrated mantle. Its vapor-saturated solidus at atmospheric pressure is 1278ºC compared to 1350ºC for the surrounding dry mantle; this assumes a water content of the vapor-saturated melt of 2.5% (Grove et al., 2006). We assign to the hydrated basaltic melt a density of 2915 kg.m^{-3}; a very conservative value for hydrated basaltic magmas, which can be as low as 2750 kg.m^{-3} for a H_2O content of

2.5% (e.g. Gaetani and Grove, 2003). It is beyond the capability of the code to account for the partial melting evolution in the mantle wedge, which can be quite complex (Grove et al., 2006). We therefore choose a simple melting model allowing for a maximum percentage of melt of 8% in the mantle wedge above the vapor-saturated solidus. The solidus and liquidus of the dryer surrounding mantle is taken from McKenzie and Bickle (1988).

To simulate eclogitization, an increase in density (maximum of 16%) is imposed on the oceanic crust as the confining pressure rises beyond 1200 MPa.

We use Ellipsis, a Lagrangian integration point finite-element code to solve the fully coupled 2-D governing equations of momentum, mass and energy in incompressible flow. These equations are solved along with constitutive relations for visco-plastic body. The code is documented in Moresi, et al., 2002 and O'Neill et al., 2006 and freely available at www.geodynamics.org.

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Table 1: Thermal and mechanical parameter values

	Density at room temperature ($kg.m^{-3}$)	Thermal expansion (K^{-1})	Radiogenic Heat Production (W.kg-1)	Cohesion (MPa)	Coeff. of Friction	A ($MPa^{-n}\ s^{-1}$)	E ($J.mol^{-1}$)	n
Upper crust	2720	0	$2.52\ 10^{-10}$	10	0.577	$5\ 10^{-6}$	190000	3
Lower crust	2820	0	$2.52\ 10^{-10}$	20	0.577	8	485000	4.7
Mantle wedge	3350	$3\ 10^{-5}$	0	10	0.577	$1.78\ 10^{12}$	520000	3
Mantle	3370	$3\ 10^{-5}$	0	10	0.577	$7\ 10^{4}$	520000	3
Accretionary prism	2810	0	$2.52\ 10^{-10}$	0	0.05	$5\ 10^{-4}$	190000	3
Oceanic Crust	2900	0	0	10	0.268	1010	367000	2.4

For all materials: Thermal diffusivity = $0.9\ 10^{-6}\ m^2\ s^{-1}$; heat capacity = $1000\ J.kg^{-1}\ K^{-1}$