Recurrence Time and Down-Dip Size of Chilean Earthquakes Influenced by Geological Structure

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Recurrence Time and Down-Dip Size of Chilean Earthquakes Influenced by Geological Structure

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Abstract.

In 1960, the giant Valdivia earthquake (Mw 9.5), the largest earthquake ever recorded, struck the Chilean subduction zone, rupturing the entire depth of the seismogenic zone and extending for 1,000 km along strike. The first sign of new seismic energy release since 1960 occurred in 2017 with the Melinka earthquake (Mw 7.6), which affected only a portion of the deepest part of the seismogenic zone. Despite the recognition that rupture characteristics and rheology vary with depth, the mechanical controls behind such variations of earthquake size remain elusive. Here, we build quasi-dynamic simulations of the seismic cycle in Southern Chile including frictional and viscoelastic properties, drawing upon a compilation of geological and geophysical insights, to explain the recurrence times of recent, historic, and paleoseismic earthquakes and the distribution of fault slip and crustal deformation associated with the Melinka and Valdivia earthquakes. The frictional and rheological properties of the forearc, primarily controlled by the geological structure and distribution of fluids at the megathrust, govern earthquake size and recurrence patterns in Chile.

Article.

Subduction zones produce 90\% of the world’s natural seismicity and host Earth’s largest earthquakes accompanied by devastating tsunamis, outlying the key importance to understand subduction dynamics during the seismic cycle. This process is controlled by the structure of the accretionary prism and the fabric of the megathrust – the seismogenic interface that separates the down-going oceanic lithosphere from the upper plate. Earthquakes originate from a frictional instability at the plate interface, primarily affected by the lithology, temperature, and fluid content of the fault zone, leading to a stratification of source properties\textsuperscript{1–3}. Numerous studies have shown variability in rupture width and slip of earthquakes in different depth domains, including great ruptures (moment magnitude Mw > 8.5) with a trench-breaking feature\textsuperscript{4}, followed by blind and large ruptures (7 < Mw < 8.5) nucleated within the interseismically locked region as illustrated by paleoseismic and instrumental record at the Japan\textsuperscript{5}, Sunda\textsuperscript{6}, and Aleutian\textsuperscript{7} trenches. The succession of large and great ruptures at the subduction megathrust modulates the size and frequency of earthquakes, but the underlying mechanics is uncertain.
The Southern Chile Subduction Zone (SCSZ) produced the Mw9.5 Valdivia 1960 megaquake\textsuperscript{12}, the largest earthquake on record, which was followed 56 years later by the Mw7.6 Melinka earthquake\textsuperscript{13} (Fig. 1). The 2016 Melinka rupture concentrated in the down-dip limit of the Valdivia earthquake, below the continental shelf, sparing the central part of the seismogenic zone that is currently locked\textsuperscript{13}. The sequence of a full-depth rupture followed by seismic resurgence in the deep segment of the seismogenic zone, along with the extensive geodetic, geophysical, and paleo-seismological record associated with the 1960 mainshock, is fundamental to explore the factors that control the slip evolution and recurrence of subduction earthquakes.

Here, we implement a numerical model of the seismic cycle at the SCSZ with depth-variable rate- and state-dependent frictional properties on the megathrust and thermally activated rheological properties in the subducting oceanic asthenosphere and over-turning mantle wedge underneath the upper plate. We base the constitutive properties on our own thermal model of the SCSZ (see Methods: Section 2) and the lithology of the subducted channel and surrounding lithosphere. We construct a cross-section of the SCSZ with a structural layout compatible with geological and geophysical observations. We then explore how the
recurrence periods, slip, and deformation of large and great earthquakes are controlled by the frictional properties of the megathrust. Our optimal model, based on a sensitivity analysis of the main parameters controlling frictional resistance, such as rheology and pore pressure ratio ($\lambda$), allows us to match multiple measurements at the earthquake cycle in Chile, providing new insights into the mechanical control of subduction earthquakes.

**Modeling the subduction seismic cycle of the Southern Andes.**

We simulate the occurrence of great (Valdivia-like) and large (Melinka-like) ruptures throughout the seismic cycle of the SCSZ using quasi-dynamic numerical simulations that resolve the mechanical interactions between a frictional megathrust, elastic plates, and the temperature-controlled viscoelastic asthenosphere. The model resolves all phases of the seismic cycles consistently, including the nucleation and propagation of earthquakes, but also the afterslip and viscoelastic relaxation that follow the mainshocks. We build a model of subduction dynamics along a representative two-dimensional cross-section (Fig. 2a) assuming a 61.8 mm/year convergence rate between the Nazca and South American tectonic plates in the trench-perpendicular direction. The structural layout is constrained by the Slab2 model, seismicity, seismic wave tomography and reflections profiles, and gravity-constrained density models. The outer wedge is defined by the highly deformed frontal Accretionary Prism. The inner wedge extends from the outer arc high to the continental slope, featuring inactive accretion structures and a sedimentary apron below the seafloor. The paleo-prism extends down to the Lower Seismogenic Zone that intersects the Moho, incorporates underplating structures, and features seismic wave velocities and density anomalies suggesting the presence of a static backstop and a high $V_p/V_s$ ratio zone (Fig. 2a). To the east, the continental crust transitions from Permo-Triassic metamorphosed units of the Metamorphic Belt to the extensional Ancud Basin and granitoids of the North Patagonian batholith below the active volcanic arc. This last feature is deformed by the Liquiñe-Ofqui Fault System (LOFS) that bounds the eastern part of the Chiloe Sliver (Fig. 1a). Beneath the continental crust, the megathrust underlies the Metamorphic Belt and the Serpentinized Mantle Wedge, terminating at the 650°C isotherm at about 85 km depth. From there, the downgoing slab is permanently coupled to the mantle wedge that deforms by viscous flow.

We assume that the frictional behavior of the megathrust is controlled by a rate- and state-dependent friction law (see also Eq. 3 in Methods): 

$$\tau = \mu_0 \bar{\sigma} \left( \frac{V}{V_0} \right)^a \left( \frac{\theta V_d}{L} \right)^b \mu_0$$

where $\mu_0$ is the effective coefficient of friction, $\bar{\sigma}$ is the effective normal stress, $V$ and $V_0$ are the instantaneous and reference velocities at the fault interface, $\theta$ is a state variable that follows an evolution law allowing healing at stationary contacts, and $L$ is a characteristic slip distance. The direct and steady-state velocity dependence of friction is controlled by the parameters $a$ and $a-b$, respectively, with $a-b<0$ corresponding to velocity-weakening, potentially unstable friction, and $a-b>0$ giving rise to velocity-strengthening friction, which inhibits rupture nucleation and propagation. Guided by the expected mineralogical composition at the plate interface and considering the thermal structure, we use laboratory-
derived information extracted from fault gouges to constrain the frictional parameters assigned to the megathrust (Table 1). The depth-dependent thermal and compositional stratification leads to a structural model for the fault with five down-dip segments with distinct properties (Fig. 2b).

Seismic wave profiles and sedimentary fluxes indicate a trench fill varying from 1 to 3 km. The shallow sediments were deposited during the exhumation of the North Patagonian batholith and the denudation of the Andean Cordillera during the last Pliocene glaciation. As a result, we assume that friction in the outer wedge is controlled by a mixed phyllosilicates and quartz composition inherited from the erosion of the Andes. We use the average values from laboratory results on fault gouges composed of about 50% quartz and 50% clay minerals at low temperatures, yielding \( a-b=2 \times 10^{-3} \). As ~80% of those sediments enter the subduction channel, we assume that the seismogenic zone is controlled by a granitic composition rather than by the altered basalts from the subducting slab (Fig. 2a), leading to \( a-b=1 \times 10^{-2} \) in the Upper and Lower Seismogenic Zone (Fig. 2b). Finally, as the bottom depth of background seismicity and the transition from a locked to creeping fault occurs at 35 km depth, we assume that the megathrust shifts to velocity-strengthening properties beyond the continental Moho. As fluid-driven serpentinization takes place in the cold mantle wedge, developing a serpentinization front at the plate interface, we select the frictional parameters of serpentinite to characterize the two deepest megathrust segments. This accounts for the transition from lizardite to antigorite at ~350°C, leading to \( a-b=1.5 \times 10^{-2} \) underneath the Metamorphic Belt and \( a-b=5 \times 10^{-2} \) below the Serpentinized Mantle Wedge (Fig. 2b).

### Table 1. Frictional parameters of the preferred and initial models

<table>
<thead>
<tr>
<th>Segment</th>
<th>Mineralogy</th>
<th>References</th>
<th>a x10-2</th>
<th>b x10-2</th>
<th>L (cm)</th>
<th>a-b x10-2</th>
<th>σ (MPa)</th>
<th>σ (MPa) initial</th>
<th>μ</th>
<th>W (km)</th>
<th>λ</th>
<th>λ initial</th>
</tr>
</thead>
<tbody>
<tr>
<td>AP</td>
<td>phyllosilicates + quartz</td>
<td>34, 35</td>
<td>1</td>
<td>0.8</td>
<td>7.5</td>
<td>0.2</td>
<td>45</td>
<td>45</td>
<td>0.4</td>
<td>20</td>
<td>0.7*</td>
<td></td>
</tr>
<tr>
<td>LSZ</td>
<td>granite</td>
<td>37</td>
<td>0.3</td>
<td>1.3</td>
<td>1.5</td>
<td>-1</td>
<td>113</td>
<td>19</td>
<td>0.6</td>
<td>73</td>
<td>0.7</td>
<td>0.95</td>
</tr>
<tr>
<td>USZ</td>
<td>granite</td>
<td>37</td>
<td>0.3</td>
<td>1.3</td>
<td>1</td>
<td>-1</td>
<td>56</td>
<td>15</td>
<td>0.6</td>
<td>30</td>
<td>0.92</td>
<td>0.98</td>
</tr>
<tr>
<td>MB</td>
<td>lizardite</td>
<td>40-42</td>
<td>2</td>
<td>0.5</td>
<td>5</td>
<td>1.5</td>
<td>130</td>
<td>65</td>
<td>0.3</td>
<td>85</td>
<td>0.9</td>
<td>0.9</td>
</tr>
<tr>
<td>SMW</td>
<td>antigorite</td>
<td>40-42</td>
<td>1</td>
<td>0.5</td>
<td>0.5</td>
<td>0.5</td>
<td>220</td>
<td>110</td>
<td>0.4</td>
<td>102</td>
<td>0.9</td>
<td>0.9</td>
</tr>
</tbody>
</table>

The effective normal stress on the megathrust depends on the local value of the partial fluid pressure \( \lambda \) via \( \bar{\sigma} = \sigma_n (1 - \lambda) \), where \( \sigma_n \) is the local normal stress based on gravity (Eq. 14 in Methods: Section 3). Our initial model setup considers published estimates of \( \lambda \) at the accretionary prism and we assume an overpressure region at the bottom of the seismogenic zone as previous studies suggest (Table 1). To obtain the pressure-temperature path along the fault (Extended Data Fig. 1b), we build a finite-element thermal model based on near-lithostatic pore-fluid pressure at megathrust depth, where the temperature satisfies the steady-state heat advection-diffusion equation (Methods: Section 2). The resulting temperature distribution (Extended Data Fig. 1a, Fig. 2a) shows the same depth of the 650°C isotherm (~85 km) than previous models, but features a deeper 350°C isotherm.
Figure 2. Summary of data acquired and setup of the model. A) Cross-section of the SCSZ with the compilation of available data. Crustal faults are from reference\(^27\), interseismic seismicity is from reference\(^16\), outer and inner wedge structures are from reference\(^18\). Horizontal lines and names at the top of the figure define the down-dip megathrust segments defined in our work. Colored thin lines with numbers are isotherms from our thermal model (see Methods). The red curve is the interpolated temperature at the plate interface from our thermal model (Methods). The Melinka slip was obtained from reference\(^9\) while the Valdivia slip is the result of our new inversion (Methods).

For the viscoelastic domain, we assume the rheological parameters of olivine aggregates for transient and steady-state dislocation creep\(^{48,49}\) and divide the mantle wedge into a serpentinized wet region below the arc and a dry region farther east, establishing a cold-nose configuration\(^50\) (Extended Data Fig. 1a and Extended Data Table 1). Viscoelastic deformation on each volume element is captured by a Burgers assembly of springs and dashpots\(^{49}\) whereby the anelastic strain rates in the Maxwell and Kelvin elements follow a power-law rheology (Eq. 7 in Methods: Section 1). The rheology is activated with the background temperature distribution of the finite-element thermal model (Eq. 13 in Methods: Section 2).

Finally, we conduct simulations spanning 10,000 years of seismic activity following the distribution of parameters constrained by the geological structure, providing the surface displacement associated with various sequences of total and partial ruptures in the SCSZ (Fig. 3a) as well as the distribution of down-dip slip. The simulations employ adaptive time steps to capture the details of rupture initiation and propagation and the more quiescent periods in between. To mitigate bias from initial conditions\(^51\), we focus on the last 7,500 years of each simulation during which the patterns of the rupture sequences are stable over time.
Earthquake cycle observations

Decadal records of crustal deformation are fundamental to characterizing the earthquake cycle. During this time, vertical tectonic displacements modulate the local relative sea-level by several meters, while horizontal deformation evolves on a continental scale. The surface deformation associated with the Valdivia earthquake is constrained by an extensive array of biological and anthropogenic markers that capture coseismic and postseismic vertical displacements, involving slip on the megathrust, but also an 8-year-long viscoelastic flow in the surrounding lithosphere. Coseismic deformation of the Melinka earthquake was registered with GNSS data and intertidal biotic indicators that describe up to 0.3 m of uplift at the Lower Seismogenic Zone. The occurrence of paleoseismic trench-breaking ruptures followed by large earthquakes in the following years has also been recorded in sediments at lake and fjords of the SCSZ. The recurrence time for earthquakes with Mw>8 is close to 300 years, while events with 7≤Mw≤8 happen every 139 years.

To calibrate the model, we compare our model predictions with the estimations of recurrence times for large and great events at the SCSZ and with surface deformation of the Melinka and Valdivia earthquakes (Extended Data Fig. 2). We compute the root mean square error (RMSE) between the model results and the available dataset (Extended Data Fig. 3). Using the distribution of pore fluid pressure from the initial model (Table 1), we cannot simultaneously reproduce the recurrence time and the surface deformation after large and great ruptures. We evaluate different configurations of pore fluid pressure at each frictional segment (Extended Data Table 2) while keeping the rest of the frictional parameters constant. The pore fluid pressure distribution that minimizes the RMSE features λ ≥ 0.9 in the Lower Seismogenic Zone, 0.7 ≤ λ ≤ 0.75 in the Upper Seismogenic Zone, and λ = 0.9 in the Creeping Zone. We use this configuration hereafter as our preferred model (Table 1) to describe the rupture mechanics and consequent deformation at the SCSZ.

Mechanics and surface deformation of large and great earthquakes.

We analyze the seismic cycle at the SCSZ by viewing a representative example of fault dynamics of our preferred model between 6,000 to 7,000 years (Fig. 3a). Most of the seismicity nucleates at the transition from the keel of the continental crust to the cold-nose of the mantle wedge corner. Large earthquakes are associated to partial ruptures (Fig. 3a) that includes Melinka-like events (Fig. 3b), which nucleate at the base of the lower seismogenic zone, propagating towards the upper seismogenic zone but stopped on their way to the trench. Great ruptures are events that break the whole seismogenic zone and the outer wedge (Fig. 3c). Among these, we distinguish earthquakes that nucleate at the base of the lower seismogenic zone as Valdivia-like events (Fig. 3c) due to similarities on the style of rupture and trench-breaking propagation with the 1960 mega-quake. Despite the assumed velocity-strengthening friction in the outer wedge, all great ruptures break the trench, consistent with up to 4 m of seafloor uplift of a recent model derived from tsunami data after the Valdivia earthquake, and compatible with most of the SCSZ paleo events with Mw>8.5 generating tsunami deposits. For both kinds of earthquakes, the rupture front propagates through creep zones caused by stress concentration at the tip of minor and past events (Fig. 3b-c), indicating a mechanical link between large and great earthquakes.
This interaction allows the trench to host coseismic slip during great ruptures and creeping at the interseismic period (Fig. 3b-d).

The preferred model predicts recurrences for large (82 ± 43.4 yr) and great (271 ± 70 yr) ruptures matching the estimates from paleo-turbidites and tsunami deposits within uncertainties. For large earthquakes, the mean recurrence time of simulated events approaches the median recurrence of Mw>7 derived from the geological record and the 56 years lag between the 1960 and 2017 earthquakes. In the simulations, the recurrence times and the coseismic slip of each rupture depend mostly on $\sigma$, $a-b$, and the width of the seismogenic zone. There is a trade-off between fitting large recurrence times and data-constrained maximum slip, but our model with the optimized pore fluid pressure generates results within the 95% confidence interval of estimations derived from paleoseismicity (Extended Data Fig. 3, Fig. 3e-f).

We compare simulated surface deformation of great and large ruptures with the available uplift dataset 8 years after the Valdivia earthquake and the coseismic deformation observed in the GNSS data after the Melinka earthquake (Extended Data Fig. 2). For each great earthquake, we extract the simulated vertical displacement produced by coseismic rupture and 8 years of postseismic deformation to obtain the surface vertical profile. The median value follows the surface deformation patterns observed 8 years after the Valdivia earthquake, characterized from west to east by offshore uplift, coastal subsidence, and uplift of the volcanic arc (Fig. 4). The postseismic deformation results from a combination of afterslip in the Lower Seismogenic Zone and behind the Metamorphic Belt, and viscoelastic uplift of the Accretionary Prism and the volcanic arc (Fig. 4). The uplift beneath the volcanic arc is caused by the rheological contrast between the cold-nose and a high temperature mantle below the arc, as also proposed by previous authors. Postseismic uplift near the trench is caused by viscoelastic flow in the oceanic asthenosphere. The magnitude of uplift is not fully compensated by the interseismic subsidence in the years following the main event (Extended Data Fig. 5), explaining the similar magnitude between coseismic and postseismic uplift (Fig. 4a).
Figure 3. Fault dynamics and recurrence times at the SCSZ. a) Slip rate over time of the simulation along the megathrust. b) Snapshot of Melinka-like earthquakes at a). c) Snapshot of Valdivia-like earthquakes at a). d) is the cumulative slip over the same period as a), black lines represent the cumulative slip curve every 25 years at the interseismic period. e) and f) are violin plots of recurrence times of simulated events against recurrence times of paleo-seismicity. Middle lines and dashed gray lines represent the median values of the model and previous measurements respectively. Gray boxes are the standard deviation of the median value of recurrence times of paleo-seismicity. Red dashed line on e) is the time that passed between the Valdivia and the Melinka earthquake.

The slip distribution of the simulated great ruptures 8 years after each mainshock is consistent with a Bayesian estimate of coseismic slip for the Valdivia earthquake\cite{63} (Eq. 15 in Methods: Section 4) that is inverted form the same land-level changes data captured 8
years after the main event. Our kinematic inversion shows up to 7 m of coseismic slip near the trench (Fig. 4b), like previous results from estimations incorporating tsunami data, and has the same maximum slip (around 20 m) as past models that include changes in slab geometry. The minimum uncertainty is observed between 75 to 150 km from the trench, solving the slip distribution at the end of the rupture and thus coinciding with the end of the seismogenic zone at 35 km depth (Fig. 2b and 4b). A small proportion of simulated giant ruptures generate a slightly lower coseismic slip in the seismogenic zone compared to the geodetic model. However, all simulated events and their median profile (black bold line at Figure 4b) match the slip distribution within uncertainties. All simulated great earthquakes generate coseismic slip up to the trench, consistent with the proposed trench-breaking feature of the Valdivia event.

Figure 4. Upper crust deformation. a) Comparison between Valdivia-like ruptures with the land-level changes data 8 years after the Valdivia earthquake. b) Coseismic slip generated by Valdivia-like ruptures compared against our new coseismic slip model (red lines) with uncertainties (red bars). c) Comparison between Melinka-like ruptures and the East component of the GNSS data from the Melinka earthquake. d) Comparison between Melinka-like ruptures and the vertical component of the GNSS data from the Melinka earthquake and the intertidal biotic indicators from reference. e) Coseismic slip generated by simulated Melinka-like ruptures compared against the model of reference. Light blue, gray and light green curves are surface deformations and coseismic slip distributions for each simulated rupture in both cases.

Finally, we assess the crustal deformation associated with large ruptures comparing the model predictions with the continuous GNSS displacements and land-level changes data.
of the 2016 Melinka earthquake. The median curve of our simulated Melinka-like ruptures (Fig. 4c and d) is larger than the coseismic pattern observed by the continuous GNSS data acquired immediately after the main event\textsuperscript{9}. Nevertheless, simulated cases with the smaller surface deformation explain the horizontal and vertical components of the observed data, reproducing the same coseismic uplift above the Lower Seismogenic Zone inferred from intertidal biotic indicators\textsuperscript{54}. The median value of simulated coseismic slip from all the events has the same amplitude as published models\textsuperscript{9,13} but slightly shifted towards the down-dip limit of the seismogenic zone. Fig. 3a and 4c-e show that large ruptures are a family of moderate events that generate surface deformation from tens to hundreds of cm and up to 4 m of coseismic slip concentrated at the Lower Seismogenic Zone. Our two-dimensional models differ slightly from the horizontal component from the GNSS data of the Melinka earthquake (Fig. 4c) due to the difference in decay with source-receiver distance with a three-dimensional case, particularly for such small ruptures. Along-strike rupture propagation is an important aspect of megathrust dynamics\textsuperscript{64}. In nature, we expect other Melinka-like earthquakes to initiate at similar depths at different latitudes along the trench axis, but this is not captured within our two-dimensional approximation.

Subduction cycle at the Valdivia segment

By incorporating the down-dip variation in the rock composition of the subduction channel, the temperature distribution in the surrounding lithosphere, and the rheology of oceanic and continental mantle, our synoptic model (Fig. 5) can explain the succession of great and large earthquakes, their respective recurrence times, and the associated crustal deformation at the SCSZ. Our models suggest that the geological structure and the pore fluid pressure play an important role in the frictional behavior that controls the characteristics of large and great earthquake sequences (Extended Data Figs. 3 and 4), explaining the recurrence intervals, the surface deformation after each rupture, and the coseismic slip. If no increase in pore fluid pressure is considered, the time between each kind of earthquake will not match the observations, and the amplitude of uplift and subsidence will not reproduce the coseismic and postseismic data after the Valdivia earthquake. Our models do not explain the origin of fluid concentration at the Lower Seismogenic Zone, but our pressure-temperature path (Extended Data Fig. 1b) is consistent with \textit{in-situ} fluid release from prograde metamorphism of metapelites\textsuperscript{65,66} at the same depth at the observed high $V_p/V_s$ ratio zone\textsuperscript{26}, outlying the effect of dehydration processes on earthquake mechanics.

The mineralogy at the plate interface also plays an important role. If we neglect the input of sediments from the erosion of the continental crust, and thus letting altered basalt of the oceanic plate control friction at the megathrust, the model cannot explain the 8-year deformation curve after great earthquakes (Extended Data Fig. 4a). Finally, if we enlarge the width of the seismogenic zone, then the pattern of surface deformation and coseismic slip after simulated ruptures is shifted landwards compared to observations. At the SCSZ, the bottom of the seismogenic zone could be controlled by the presence of fluids and/or the continental Moho (Fig. 5), rather than by the temperature distribution at the megathrust (Extended Data Fig. 4b-c).

Numerical models constrained by geological and geophysical data, such as presented here, provide an ideal medium to integrate and test geodynamics processes throughout the
seismic cycle of megathrust earthquakes, with possible applications at other subduction zones worldwide. Given the spatial relationship between fluid concentration depth and forearc structure at the SCSZ (Fig. 5), we propose that the seismic cycle depends on the thermodynamic conditions at the plate interface associated with the resulting depth-dependent lithology, hydrothermal regime, and geological structure, leading to a fluid-rock interaction along the seismogenic megathrust that impacts the stability of frictional sliding.

**Figure 5.** Conceptual model of the SCSZ integrating insights from quasi-dynamic simulation and compiled data. λ values are the ones calibrated to our preferred model. Nucleation depth is referred to the depth at which large and great ruptures start. Blue area marks water concentration region at depth of in-situ fluid release. Black lines at the cross section are basal accretion structures interpreted from seismic wave profiles. Isotherms are obtained from our modeled thermal distribution. Light blue arrows represent the fluid-flow after mineral dehydration.

**References**


**Methods**

**1 Fault friction and Mantle viscosity**

The stages of the seismic cycle in faults can be described using constitutive relationships of rate – and state – friction that explain the evolution of sliding velocity over time. These laws had been used to model several rupture styles in nature and to characterize the seismic cycle at different tectonic settings. This leads to the possibility of understanding fast and slow ruptures at different fault configurations. Here, we used a constitutive framework obtained from the assumption of a microphysical model of the rate – and state – dependent friction under isothermal conditions. In this context, the sliding velocity depends on the density of the real area of contact as follows:

\[
V = V_0 \left( \frac{\tau}{\mu_0} \right)^{\frac{\mu_0}{\alpha}} \exp \left[ -\frac{\alpha}{R} \left( \frac{1}{T} - \frac{1}{T_0} \right) \right]
\]

where \( V \) is the sliding velocity, \( \tau \) is the norm of the shear stress resolved on the fault plane, \( T \) is the absolute temperature, \( \mu_0 \) is the static coefficient of friction at \( V_0 \) and \( T_0 \), \( \alpha \) is the real area of contact density (real area of contact divided by the nominal surface area), \( \chi \) is the
indentation hardness, \( a << 1 \) is the direct effect, \( Q \) is the activation energy and \( R \) is the gas constant. The real area of contact depends on the effective normal stress and changes as a function of the shape of the surrounding contact junctions\(^6\). This last feature is characterized by a state variable that represents the age of the grain which coincides with the age of contact defined by reference\(^7\). The relationship between the effective normal stress and the real area of contact is given by\(^8\):

\[
\mathcal{A} = \frac{\mu_0 \sigma_{\text{eff}}}{X} \left( \frac{\theta V_0}{L} \right)^b \mu_0^{-b} \tag{2}
\]

where \( \sigma_{\text{eff}} \) is the effective normal stress, \( \theta \) is the state variable, \( L \) is the characteristic weakening distance associated with the gouge thickness or fault roughness and \( b << 1 \) is the evolution effect. By combining Eqs. (1) and (2) the multiplicative form of rate – and state – friction at isothermal conditions is obtained\(^9, 68\):

\[
\tau = \mu_0 \bar{\sigma} \left( \frac{V}{V_0} \right)^a \mu_0^{-a} \left( \frac{\theta V_0}{L} \right)^b \mu_0^{-b} \tag{3}
\]

The state variable under isothermal conditions follows an evolution law which allows healing at the stationary contacts\(^5, 68\):

\[
\theta' = \exp \left[ -\frac{H}{R} \left( \frac{1}{T} - \frac{1}{T_0} \right) \right] - \frac{\theta V}{L} \tag{4}
\]

Fault slip dynamics is controlled by the spatial distribution of \( \mu_0, \bar{\sigma}, a, b \) and \( L \), which are governed by the thermal state at the fault interface and by the mineral assembly of the respective stable lithological facie under that state. We assume that those parameters do not change over time and are not affected by heat and fluid transport, which means that the value that controls the frictional behavior on the fault at steady state will be \( a - b \). Depending on its temperature distribution, effective normal stress and rock composition, any given fault can experience a velocity weakening \((a - b < 0)\) or a velocity strengthening \((a - b > 0)\) behavior\(^7\). Velocity weakening areas will experience unstable slip, while velocity strengthening regions should slip aseismically\(^7\). Nonetheless, previous works had shown that velocity strengthening regions can propagate ruptures\(^7\).

As can be anticipated from Eq. (3) and Eq. (4), there is a very wide spectrum of sliding mechanisms that will be generated at the fault over time. This introduces high uncertainties about the distribution of the parameters that govern rate – and state – friction relations. Nevertheless, from a dimensional analysis of the frictional equations, reference\(^29\) defined two dimensionless parameters that can be understood as two degrees of freedom for the frictional behavior of a given fault:

\[
R_u = \frac{(b-a)\sigma_{\text{eff}} W}{G} \frac{W}{L} \tag{5}
\]

\[
R_b = \frac{b-a}{b} \tag{6}
\]

Where \( R_u \) is known as the Dieterich-Ruina-Rice number and depends on the geometry of the fault given by the width \( W \) of the velocity weakening patch, on the frictional behavior given by \( a - b \), on the effective normal stress and on the shear modulus \( G \) at plane strain conditions. For a velocity weakening segment \( R_u \) takes positive values and represents the
relation between the segment dimension and the nucleation size. Greater values of $R_6$ will leads to the nucleation of smaller instabilities with respect to $W^{29}$. 

The other parameter is $R_b$, which controls the transient evolutionary effects and the strengthening behavior. Velocity strengthening patches will have $R_b < 0$, velocity weakening domains are characterized by $0 < R_b < 1$ and velocity neutral regimes are defined when $R_b = 0$. $R_b$ also controls the rupture propagation in the same way as the ratio $a/b$ $^{74}$. At velocity weakening domains and high sliding velocities, when $a/b$ is close to the unity, then the rupture will tend to extend towards the whole segment.

The combination of $R_6$ and $R_b$ gives a two-dimensional space where different rupture styles can arise. By constraining the spectrum of frictional parameters, the actual slip behavior of a given fault can be approximated. We used this approach to feed the initial state of all the parameters in our model. We collected a wide range of datasets to constrain the pressure-temperature conditions and the slip mechanism at the plate interface. Then, we assumed an initial composition for the oceanic plate and for the sediments at the trench. From that point, we used previous stability fields of subduction related metamorphic facies to define the metamorphic state along dip. Finally, we defined the frictional parameters using previous gouge data from different lithologies that represent the metamorphic facies at the plate interface.

The viscoelastic behavior was calculated using a power-creep law that assumes pure dislocation creep over minerals with a constant water concentration$^{48}$:

$$\dot{\varepsilon} = A \sigma^n C_{OH}^r d^{-m} \exp \left( \frac{-Q+p \Omega}{RT} \right)$$  \hspace{1cm} (7)$$

Where $\varepsilon$ is the strain rate, $A$ is a pre-exponential factor, $\sigma$ is the deviatoric stress, $C_{OH}$ is the water concentration, $d$ is grain size, $Q$ is the activation energy, $p$ is the lithostatic pressure, $\Omega$ is the activation volume, $R$ is the constant for ideal gasses, $T$ is the temperature and $n$, $r$ and $m$ are experimentally derived exponents. To yield an expression independent of the grain size, we assumed $m = 0$ and $n > 1^{48,75}$.

Assuming this configuration for the subduction fault, we made a meshed subduction system (Extended Data Fig. 1a) to run a numerical model of the subduction seismic cycle. We discretized the system into a triangular mesh of 857 cells with a size of 20 km (Extended Data Fig. 1a), while the points elements at the fault were separated by 200 m along the Slab2 geometry$^{15}$ allowing numerical convergence. Deformation at each triangle was computed considering a Burgers assembly where the total anelastic strain rate is given by $\dot{\varepsilon}_{T} = \dot{\varepsilon}_M + \dot{\varepsilon}_K$, in which $\dot{\varepsilon}_M$ and $\dot{\varepsilon}_K$ are the instantaneous strain rates in the Kelvin and Maxwell elements respectively$^{49}$. Power-creep law parameters for the oceanic and continental mantle (Extended Data Table 2) were assumed following previous experimental$^{48,49}$. We used the numerical code Unified Cycle of Earthquakes (UniCyclE) that employs the integral method$^{5,76,77}$, including surface and volume elements. The changes in fault traction will depend on the surrounding fault slip and distributed strain. We assumed a background strain rate in the horizontal direction of $\varepsilon = -1 \times 10^{-14}$ s$^{-1}$ and predefined a set of synthetic GNSS stations at the surface of our mesh to compare those results with the actual data.

### 2 Temperature model
The subduction thermal model was constructed considering an oceanic slab subducting with a kinematically prescribed velocity (66 mm/year) beneath a fixed and layered continental plate. We used the continental lithospheric structure of reference considering an upper and lower crust, an oceanic crust, and a continental and oceanic mantle. In this framework the downgoing slab drives flow within the overlying viscous mantle wedge.

Velocity and pressure at the mantle wedge are found by solving the equations of mass and momentum conservation. Following reference, we solve the equation of conservation of mass:

\[ \nabla \cdot \mathbf{v} = 0 \]  

And the conservation of momentum:

\[ \nabla \cdot \tau - \nabla P = 0 \]

Where \( \tau \) is velocity, \( \tau \) is the deviatoric stress tensor, and \( P \) the dynamic pressure. The deviatoric stress tensor is given by:

\[ \tau = 2\eta \varepsilon \]

In which \( \eta \) is the effective dynamic viscosity and the components of the strain rate tensor \( \varepsilon \) are:

\[ \varepsilon_{ij} = \frac{1}{2} \left( \frac{\partial v_i}{\partial x_j} + \frac{\partial v_j}{\partial x_i} \right) \]

We assume that viscosity is obtained by dislocation creep:

\[ \eta = A \exp \left( \frac{q}{nRT} \right) \frac{\varepsilon^{1-n}}{n} \]

where the rheological parameters are the same as in Equation (7). We consider a temperature- and stress-dependent wet olivine rheology for the mantle wedge.

The boundary conditions for mantle wedge are (1) no slip below the rigid overriding plate and (2) constant velocity with a value equal to plate convergence along the top of the slab and below the maximum decoupling depth.

The thermal field within the entire subduction zone was computed by solving the steady-state heat advection-diffusion equation, so temperature is obtained by solving:

\[ \rho c_p (v \cdot \nabla) T = \nabla \cdot (k \nabla T) + f \]

In which \( \rho \) is density, \( c_p \) is the specific heat, \( k \) is thermal conductivity, and \( f \) is a parameter that describe possible heat sources.

To solve the system, we assumed a constant temperature at the surface of the model (0°C) and at the bottom of the oceanic lithosphere (1,450°C). On the slab inflow boundary, we prescribed a geotherm obtained using a half-space cooling model considering an age of 14 Ma. At the right edge of the model, we assumed a conductive geotherm calculated using the thermal properties of Extended Data Table 3, which is then applied along the landward boundary of the overriding plate. Along the inflow part of the wedge, we prescribed a geotherm which was obtained using an adiabatic gradient of 0.4 °C/km and a mantle
potential temperature of 1,300°C. For the outflow boundary of the wedge, zero heat flux is assumed.

Finally, we solve equations (8), (9) and (11) using a stabilized finite element\textsuperscript{45} which allows us to use equal order of polynomial interpolation at each unknown point.

3 Pressure and temperature path

To compute the pressure-temperature path, we used the lithospheric structure of reference\textsuperscript{23} and we took the average density structure of reference\textsuperscript{83}. We compute the vertical stress with:

$$\sigma_v = g \rho h$$  \hspace{1cm} (14)

Where \( g \) is the gravity acceleration constant, \( \rho \) is density and \( h \) the thickness of the lithospheric layer (upper crust, lower crust, and lithospheric mantle). Then, we compute the normal component over each element of the fault surface to obtain the \( \sigma_n \). By combining the temperature distribution at the plate interface and the \( \sigma_v \) over it, we obtained the pressure-temperature path from Extended Data Figure 1b.

4 Coseismic slip model.

We inverted the coseismic slip for the dip direction using the uplift displacements on the surface from the work of reference\textsuperscript{12} in between 42°S and 45°S.

To find the solution, we applied the Bayesian inversion approach of reference\textsuperscript{63}, which allows us to obtain positive model parameters and associated uncertainties. This constraint allows us to obtain a more realistic slip solution as the slip direction is positive with respect to the dip direction, given the subduction stress regime. We impose a slip solution to be a multivariate folded normal distribution, and simultaneously we seek to find a solution for \( G_m = d \), where \( d \) are the deformation data, \( m \) is the backslip on each subfault in the plate contact and \( G \) is the matrix which contains the Green’s functions. The Bayesian formulation is given by:

$$p(s|d, H) = \frac{p(d|s)p(s|H)}{p(d|H)}$$  \hspace{1cm} (15)

with

$$p(s \vee H) = (2\pi)^{-\frac{N_m}{2}} |S|^\frac{-1}{2} \exp \left[ \frac{1}{2} (s^p - s)^T S^{-1} (s^p - s) \right]$$  \hspace{1cm} (16)

$$p(d \vee s) = (2\pi)^{-\frac{N_d}{2}} |D|^\frac{-1}{2} \exp \left[ \frac{-1}{2} (d - G|s|)^T D^{-1} (d - G|s|) \right]$$  \hspace{1cm} (17)

To enforce positivity constraints on the model parameters, we impose the following changes of variables:

$$m(s) = (|s_1|, |s_2|, ..., |s_m|)^T$$  \hspace{1cm} (18)
which guarantees us a positive solution and results in a Folded Normal Distribution in the posterior distribution.

We used the maximum evidence criteria to perform Bayesian model selection, namely, to search for the hyperparameters of a particular hypothesis $H$. $D$ and $S$ are the covariance matrices of the likelihood and prior, respectively. The mean of the prior is $s^p$. We then obtain the correlation between the slip parameters, based on the information of the same data.

References


Extended Data

### Extended Data Table 1. Power-creep law properties

<table>
<thead>
<tr>
<th>Volume</th>
<th>A (MPa−n s−1 ppm H/Si)−n</th>
<th>r</th>
<th>C_ox (ppm H/Si)−n</th>
<th>n</th>
<th>Ω (cm³/mol)</th>
<th>Q (Kj/mol)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Oceanic Mantle</td>
<td>10^{0.56}</td>
<td>1.2</td>
<td>1000</td>
<td>3</td>
<td>1.30E-05</td>
<td>5.10E+05</td>
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<tr>
<td>Continental Mantle (dry conditions)</td>
<td>10^{0.56}</td>
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<td>1000</td>
<td>3</td>
<td>1.10E-05</td>
<td>4.75E+05</td>
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<tr>
<td>Continental Mantle (wet conditions below the arc)</td>
<td>10^{0.56}</td>
<td>1.2</td>
<td>1200</td>
<td>3</td>
<td>1.10E-05</td>
<td>5.10E+05</td>
</tr>
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</table>

Extended Data Table 1. Power-creep law properties. Initial values were obtained from reference 48,49.

### Extended Data Table 2. Pore pressure ratio and effective normal stress variation

<table>
<thead>
<tr>
<th>USZ</th>
<th>LSZ</th>
<th>CZ</th>
</tr>
</thead>
<tbody>
<tr>
<td>λ</td>
<td>σ (Mpa)</td>
<td>λ</td>
</tr>
<tr>
<td>MB</td>
<td>AB</td>
<td>MB</td>
</tr>
<tr>
<td>-----</td>
<td>-----</td>
<td>-----</td>
</tr>
<tr>
<td>0.6</td>
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<td>0.65</td>
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<tr>
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<tr>
<td>0.95</td>
<td>19</td>
<td>0.95</td>
</tr>
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</table>

Extended Data Table 2. Pore pressure ratio and effective normal stress variation. USZ = Upper seismogenic zone, LSZ = Lower Seismogenic Zone, MB = Metamorphic Belt, AB = Ancud Basin, CZ = Creeping Zone.
Extended Data Table 3. Thermal properties of the temperature model.

<table>
<thead>
<tr>
<th>Layer</th>
<th>Thermal conductivity (W/mK)</th>
<th>Radiogenic heat production (µW/m³)</th>
<th>Density (kg/m³)</th>
<th>Specific heat (J/kgK)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upper continental Crust</td>
<td>2.9</td>
<td>1.3</td>
<td>2700</td>
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<tr>
<td>Lower Continental Crust</td>
<td>2.9</td>
<td>0.4</td>
<td>3300</td>
<td>1000</td>
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<tr>
<td>Oceanic Crust</td>
<td>3.1</td>
<td>0.02</td>
<td>3300</td>
<td>1000</td>
</tr>
<tr>
<td>Oceanic and Continental Mantle</td>
<td>3.1</td>
<td>0.02</td>
<td>3300</td>
<td>1000</td>
</tr>
</tbody>
</table>

Extended Data Figure 1. Temperature distribution, mesh, and pressure-temperature path. a) black arrow denotes the convective asthenosphere flux. Purple curves are contours of the Vp/Vs value. Triangles are the mesh used for the continental and oceanic mantle. b) Gray dashed lines are the limits of metamorphic facies for a metapelite rock. LB = Lawsonite Blueschist, epB = Epidote Blueschist, Amph = Amphibolite, G = Greenschist, PP = Prehnite-Pumpellyite, E = Eloglute. Intraplate and Interplate refer to the location of the interseismic seismicity.
Extended Data Figure 2. GNSS and intertidal biotic indicator data. a) Velocity vectors from reference for the coseismic slip of the Melinka Mw7.6 2016 earthquake. b) Uplift data from reference measured after 8 years of the Valdivia Mw9.5 1960 earthquake.
Extended Data Figure 3. Root mean squared error values for systematic variation on $\lambda$. In the left column are plotted RMSE values with respect to the surface deformation obtained from GNSS data of Melinka and Valdivia earthquakes. In the right column are plotted RMSE values with respect to the median value of great and large earthquakes obtained from the paleoseismic record. Creeping zone in c) refers to the Metamorphic Belt and Serpentinitized Mantle Wedge segments.

Extended Data Figure 4. Effects of petrology and depth of the end of the seismogenic zone. Isotherm in b) is obtained from our computed thermal distribution. Gray and green dashed curves mark the depth in which ends the seismogenic zone for each model. The red curve is the interpolation at 43°S of the 3D coseismic slip model from reference^9 ends the seismogenic zone for each model. The red curve is the interpolation at 43°S of the 3D coseismic slip model from reference^9. Frictional parameters for each non preferred model are described in Extended Data Table 1.
Extended Data Figure 5. Time series of the east and vertical components of the synthetic gps from our model. We select the same time interval from Figure 3. Red triangles mark the position of the volcanic arc.
Supplementary Files

This is a list of supplementary files associated with this preprint. Click to download.

- ExtendedData.docx