

Asthenosphere rheology and the continental-scale interseismic deformation driven by repeated megathrust earthquakes: Insights from the South American plate

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SUMMARY

This study presents a framework for quantifying large-scale interseismic deformation of a tectonic plate overlying a viscoelastic asthenosphere. We explain and compare four common methods for modelling upper-plate surface deformation induced by time-dependent slip on the subduction interface. One method is a widely used, purely elastic model. The other three are viscoelastic models, in which slip on the subduction interface is either steady over time, steady over time but with the impact of the last earthquake taken into account, or represented as periodic earthquake cycles.

For the three viscoelastic models, two asthenospheric rheologies are investigated. The first is a Maxwell rheology with a single viscosity, inferred from early postseismic deformation (3×10^{18} Pa.s). This model reasonably reproduces short-term deformation. The second is a Burgers rheology with a comparable short-term viscosity, but with a higher long-term effective viscosity (3×10^{19} Pa.s). This higher long-term viscosity is constrained by decadal-to-centennial observations. Therefore, this model alone accounts for deformation across all timescales and spatial scales. These observations include a trenchward deformation persisting 50 years after the 1960 M_w 9.5 Valdivia earthquake, and deformation from northern Chile, where deformation reflects coupling along segments that have not ruptured for decades.

All models are evaluated using a finite-element mesh spanning an 8000 km-wide region around the Chilean subduction margin. For the same coupling, near the coast, all models yield similar deformation, with differences smaller than 20%. However, discrepancies increase markedly around 200 km inland. In the continental interior, the elastic model consistently underestimates interseismic velocities. Only the Burgers rheology, combined with a realistic cyclic slip history, reproduces the observed deformation over the entire continent and across annual to centennial timescales. We constrain the acceptable parameter range of this simplified Burgers rheology. This requires a Kelvin-Voigt shear modulus 4 to 15 times smaller than the elastic shear modulus.

These results demonstrate that the effective viscosity of the asthenosphere increases markedly from annual to decadal or longer-term timescales. They also show that millennium-scale strain accumulation must be accounted for when interpreting interseismic deformation. They call into question estimates of coupling depth and sliver motion based on purely elastic models.

Key words: subduction zone; seismic cycle; viscoelastic models; geodesy; finite-element models

1 INTRODUCTION

The accumulation of stress, and consequently seismic hazard, in subduction zones is directly linked to surface deformation. This deformation reflects the elastic energy stored within the plates. This energy is released during large earthquakes. Slip rates over most subduction interfaces are highly irregular. Brief periods of significantly higher velocity occur during earthquakes or slow-slip events. These periods are followed by much longer intervals of slower movement. A steady slip rate equal to the convergence rate would induce no surface intraplate deformation. Instead, it is these fluctuations in relative motion that induce deformation in both the overriding and subducting plates (Savage, 1983).

Surface deformation of the overriding and subducting plates during the seismic cycle is commonly described in terms of three phases: **coseismic**, **postseismic**, and **interseismic**.

During the **coseismic** phase, deformation occurs over seconds to minutes. This deformation is well approximated by elastic models. The short duration justifies neglecting viscous effects.

In contrast, the **postseismic** phase involves deformation that persists for years to decades following an earthquake. This long-lasting, large-scale deformation extends far from the trench. It is widely interpreted as the result of viscoelastic deformation in the asthenosphere (e.g., Nur and Mavko, 1974; Thatcher and Rundle, 1984; Wang et al., 2012; Klein et al., 2016). Part of the early postseismic deformation may also be attributed to afterslip along the subduction interface (e.g., Marone et al., 1991; Perfettini et al., 2010; Rolandone et al., 2018; Tissandier et al., 2023). Observations show that, after major earthquakes, the overriding plate continues to stretch at large distances from the trench. At the same time, compression progressively develops near the trench, due to the reloading of the subduction interface, and propagates landward (Wang et al., 2012; Trubienko et al., 2013). This behaviour has been particularly well documented following the 1960 M_w 9.5 Valdivia earthquake in southern Chile (Klotz et al., 2001; Khazaradze et al., 2002; Wang et al., 2007; Moreno et al., 2009; Bürgmann et al., 2013). The postseismic phase refers to the period when velocities remain trenchward. This period depends upon the distance to the trench (Wang et al., 2012).

Finally, the **interseismic** phase is typically characterised by a general compression of the lithosphere. Continentward velocities decrease away from the trench, but vary little with time. This phase is most commonly modeled using a purely elastic Earth (e.g., Savage, 1983; Chlieh et al., 2008, 2011; Métois et al., 2012, 2013, 2014; Nocquet et al., 2014; Villegas-Lanza et al., 2016; Métois et al., 2016; Klein et al., 2018; Michel et al., 2019; Herman and Govers, 2020; Jolivet et al., 2020; Maubant et al., 2022; Yáñez-Cuadra et al., 2022; Jarrin et al., 2023; Maubant et al., 2023; Loversy et al., 2024; Oryan et al., 2026).

The general evolution of deformation in the overriding plate throughout the seismic cycle is observed by geodesy, e.g., GNSS and InSAR (e.g., Wang et al., 2012; Bürgmann et al., 2013). However, the methods used to model this deformation remain debated. In particular, treating the asthenosphere as viscoelastic during the postseismic phase, in the years following an earthquake, but purely elastic during the interseismic phase is difficult to justify conceptually. Elastic models are appealing. They are straightforward to implement and involve relatively few parameters. In contrast, viscoelastic models are more complex. They require assumptions about asthenosphere rheology and about the history of slip along the plate interface. Both of these assumptions remain debated. Key aspects of this debate include: **(i)** whether the slip history of the subduction interface should be explicitly accounted for, and, if so, how it should be represented, **(ii)** which rheological formulation best captures asthenosphere behaviour, and **(iii)** whether structural complexities, such as crustal blocks or slivers, should be explicitly included. These three points are discussed in detail below.

(i) Modelling the slip on the subduction interface. Four main classes of models have been proposed to model upper-plate deformation induced by intermittent slip on the subduction interface.

- The *elastic* model is by far the most widely used. In this model, the mantle beneath the lithosphere is assumed to behave as a purely elastic medium during the interseismic phase. This elastic framework has been widely used to generate coupling models over numerous subduction zones (Oryan et al., 2026). This coupling pattern is related to the present distribution of relative velocities along the plate interface (Savage, 1983).

- The *steady long-term* model, hereafter named the *steady* model, is employed to describe deformation long after the last large earthquake. In this model, the asthenosphere is assumed to be viscoelastic.

Interseismic deformation is estimated by assuming that the present-day coupling pattern has persisted unchanged over an infinitely long time (e.g., Li et al., 2015, 2018; Itoh et al., 2019). In practice, the constant coupling must be imposed on the interface over a time such that the surface velocities become stable. This time is proportional to the effective viscosity appropriate for the corresponding time-scale.

- A variant of the previous model includes the contribution of postseismic deformation from the most recent earthquake. In this variant, the predicted interseismic velocity is the sum of the steady long-term velocity and the decaying postseismic velocity from the last earthquake. This variant is hereafter shortened as *steady + last EQ* (Wang et al., 2001; Sun et al., 2018; Wang et al., 2012; Itoh et al., 2021). Both *steady* models imply the long-term increase of elastic strain in the lithosphere.

- The *cyclic* models, instead, assume repeated periodic earthquake cycles. These cycles are separated by phases during which the interface is fully or partially locked (e.g., Savage, 1983; Hyodo and Hirahara, 2003; Trubienko et al., 2013; Pollitz and Evans, 2017; Li and Chen, 2023; Pollitz, 2025). In these periodic cyclic models, the long-term average slip rate at every point on the interface equals the plate-convergence rate. This avoids unbounded lithospheric compression. It also provides a more realistic representation of fault motion. However, except for a few subduction zones such as Nankai, the perfect periodicity of earthquakes is not observed (Philibosian and Meltzner, 2020; Brodsky and Farge, 2026). The conditions under which the *cyclic* model yields reliable velocity estimates for irregular seismic cycles remain largely unexplored. Moreover, the *steady + last EQ* and *cyclic* models differ only in how they account for the penultimate and earlier earthquakes. Accordingly, they should yield similar deformation patterns only if the cumulative velocity contributions from those earlier events are negligible.

(ii) Rheology of the asthenosphere. Up to now, postseismic deformation, reflected in changes in surface velocity following an earthquake, has provided valuable constraints on mantle viscosity. These constraints cover mostly short timescales, of only a few years. The present study extends the investigation to longer timescales. It explores the relationship between long-term surface deformation and slip history, with a particular focus on the role of the rheology of the asthenosphere over decades to centuries. In particular, it remains unclear whether the viscosity inferred from postseismic observations over a few years (Pollitz, 2003; Pollitz et al., 2006; Hu and Wang, 2012; Klein et al., 2016) represents a true long-term material property. It may instead represent an apparent, transient viscosity, relevant only over timescales of a few years. This question is fundamentally linked to the timescales that different rheological models are designed to capture. A Maxwell rheology is characterised by a single viscosity. A Burgers rheology, in contrast, incorporates two characteristic timescales. This allows both a transient response and longer-term viscous behaviour to be represented. More complex temporal behaviour can be modeled using an Extended Burgers rheology or more generally, models where the shear modulus is a frequency-dependent complex number. These models allow transient deformation over multiple timescales. It thus offers greater flexibility in capturing deformation processes spanning a wide range of timescales (e.g., Benjamin et al., 2006; Jackson and Faul, 2010; Ivins et al., 2020, 2021; Lau et al., 2021). A Burgers rheology has frequently been used to describe differences between the effective viscosities for timescales of a few months versus a few years, during postseismic deformation (Pollitz, 2003; Wang et al., 2012; Klein et al., 2016; Sun et al., 2018). Here, because seismic cycles evolve over long timescales, we investigate the possibility of a transition in effective viscosities, from annual to decadal or centennial timescales. The choice of the most appropriate viscoelastic rheology, to describe these decadal to centennial timescales, lies at the core of an active scientific debate. This debate has broad implications, not only for seismic cycle modelling but also for climate-related processes. In particular, understanding whether short-term (a few years) viscosity estimates reflect long-term mantle behaviour is critical. This understanding is needed to accurately quantify cryospheric mass loss and the resulting non-steric sea-level rise, both in the present and since the Little Ice Age (e.g., Ivins et al., 2023; Adhikari et al., 2021; Paxman et al., 2023; Rousselet et al., 2025).

(iii) Plastic deformation of the lithosphere. Finally, overriding plates are often thought to experience not only elastic but also plastic deformation. This plastic deformation occurs within active mountain chains, rifts, transform faults, and slivers (Lavé and Avouac, 2001; Bevis et al., 2001; Brooks et al., 2003; Moreno et al., 2011; Métois et al., 2012, 2013, 2014; Nocquet et al., 2014; Métois et al., 2016; Villegas-Lanza et al., 2016; Klein et al., 2018; Saillard et al., 2017; Jolivet et al., 2020; Loverly et al., 2024; Oryan et al., 2024). GNSS data capture the total deformation of the plate. This deformation encompasses both plastic and elastic components. It is essential to analyse these two types of deformation in a complementary manner. Neglecting, or inaccurately modelling, the elastic deformation of the plates induced by irregular slip on the subduction interface can lead to misinterpretations of their long-term plastic deformation, and vice versa. For instance, several quantities rely on a comprehensive understanding of present-day deformation linked to the history of slip on the Chilean, Sumatra, and

Japanese subduction interfaces. These quantities include the long-term convergence rate on the eastern flank of the Andes (Brooks et al., 2003), the velocity of the Sunda block with respect to Eurasia (Bock et al., 2003), and the compression rate within the Japan Sea (Heki et al., 1999), all deduced from velocities measured by GNSS or InSAR. Purely elastic models of interseismic deformation resort to slivers or third microplates. This is because the observed significant far-field continentward velocity cannot be reproduced, in this simple elastic setup, without invoking an unrealistically deep locking depth Savage (1983). Would slivers or microplates still be needed after the introduction of viscoelasticity in models of interseismic velocities? Most studies of interseismic deformation involving a viscoelastic asthenosphere predict a velocity gradient much smaller than that predicted by elastic models (Wang et al., 2001; Trubienko et al., 2013; Li et al., 2015). They are thus compatible with a lower maximum coupling depth, without requiring additional plates or blocks.

This study aims to develop a consistent framework for modelling surface deformation across the seismic cycle. It also aims to assess the predictive performance of the four principal classes of slip models and of the two rheologies described above through a series of forward finite-element simulations. To determine the most appropriate rheology and model of slip along the interface, we compare the surface deformation predicted by the various models with GPS data. These data cover the southern part of the South American continent, facing the Chilean subduction zone. Most studies of interseismic deformation are based on near-field data (e.g., Chlieh et al., 2008, 2011; Moreno et al., 2011; Métois et al., 2012; Wang et al., 2012; Métois et al., 2013, 2014; Nocquet et al., 2014; Villegas-Lanza et al., 2016; Métois et al., 2016; Klein et al., 2018; Itoh et al., 2021). Our study, in contrast, mainly relies on far-field data. It is in continuity with the models proposed by Trubienko et al. (2013) in 2D. These models also investigate the impact of the Maxwell versus Burgers rheology on continental-scale deformation, with potential variations of effective viscosities from annual to decadal or centennial time-scales. They also mainly consider constraints provided by far-field data.

2 SETTING OF THE NUMERICAL COMPUTATIONS

2.1 Mesh and elastic parameters

All numerical computations presented in this paper are performed using the finite-element code Zset-Zebulon (<http://zset-software.com/>, last access: 25 August 2026). They are performed on the mesh *Chile_Mesh_v1.0*, described in Boulze et al. (2025). This mesh of the Chilean subduction zone spans 8500 km in longitude and 7300 km in latitude. It extends from the Earth’s surface to the Core-Mantle Boundary, at 2900 km depth. Its large extent allows for the study of continental-scale deformation associated with the seismic cycle in South America. All details about the mesh construction and earthquake implementation can be found in Boulze et al. (2025).

In our model, elastic parameters include a continental crust. These parameters increase with depth in the mantle, following PREM (Dziewonski and Anderson, 1981). This stratification has a strong and counterintuitive effect on purely elastic models. When elastic parameters increase with depth, large-wavelength surface deformation, which penetrate deeper, are attenuated. As a result, the predicted interseismic velocity signal decays much more rapidly away from the trench (Trubienko et al., 2013). As a consequence, neglecting this stratification actually improves elastic model predictions. The resulting velocity patterns more closely resemble those from viscoelastic models that do account for elastic stratification.

2.2 Viscoelastic rheologies used for numerical simulations

After megathrust earthquakes, GNSS observations indicate that viscous creep in the asthenosphere appears Newtonian during the postseismic phase. Indeed, the apparent viscosity does not vary with earthquake magnitude, and thus does not vary with the level of applied stress (Trubienko et al., 2014; Boulze et al., 2022). This apparent Newtonian behaviour may result from large background stresses. However, the strongest stresses of the seismic cycle occur immediately after rupture. This implies that the linear relationship between strain-rate and stress, documented during the postseismic phase, should hold throughout the entire cycle. Therefore, seismic-cycle deformation should be represented using linear viscoelastic rheologies. Examples include the Maxwell, Burgers, and Extended Burgers rheologies, or more generally, rheologies involving viscoelastic shear moduli expressed as complex numbers that are functions of frequency.

Postseismic deformation observed in the years following large earthquakes points to an apparent asthenosphere viscosity of about $3\text{--}4 \times 10^{18}$ Pa.s. This value assumes deformation occurs in a thin

asthenosphere (~ 200 km thick), overlain by a ~ 80 km lithosphere (Klein et al., 2016). Larger viscosities are inferred when thicker asthenospheres or thinner lithospheres are assumed (Trubienko et al., 2014).

It is now well established that the effective viscosity of the mantle depends strongly on the time-scale considered. Values inferred from short time-scales (e.g., seismic attenuation) are much lower than those inferred from geodetic or geologic deformation over years to millennia (e.g., Benjamin et al., 2006; Lau et al., 2021). A key question is therefore whether viscosities estimated from postseismic deformation over only a few years are representative of the rheological behaviour throughout the entire seismic cycle (100 yrs or more).

To address this, we use two viscoelastic rheologies for the asthenosphere: Maxwell and Burgers (Fig. 1). These rheologies are represented using simple mechanical elements, springs for elasticity and dashpots for viscous creep. A Maxwell rheology consists of a spring and a dashpot in series. A Burgers rheology adds a Kelvin-Voigt element in series to the Maxwell element. A Kelvin-Voigt element consists of a spring and a dashpot in parallel. For the Maxwell rheology, we use a viscosity of $\eta_M = 3 \times 10^{18}$ Pa.s. For the Burgers rheology, we use $\eta_M = 3 \times 10^{19}$ Pa.s and a transient viscosity of $\eta_K = 2.7 \times 10^{18}$ Pa.s. Note that the elastic modulus of the Kelvin-Voigt element we used, μ_K , is equal to $0.2 \times \mu_M$. Here, μ_M is the elastic modulus of the Maxwell rheology. The unusual ratio $\mu_M/\mu_K = 5$, often equal to 1 in other studies, is discussed in detail in Section 7.6. Further details on these rheologies and their responses to imposed stress or strain can be found in Karato (2008).

For simplicity, we did not include in our simulations geophysical zones with locally varying viscosity, such as a low-viscosity channel or wedge mimicking a serpentinized or hydrated upper mantle. We also did not include the short-term Kelvin-Voigt transient (with viscosity lower than 10^{18} Pa.s) commonly invoked to explain rapid postseismic deformation during the first few months after an earthquake (Pollitz, 2003; Pollitz et al., 2006; Hu et al., 2016; Klein et al., 2016). Viscous flow in the deep upper mantle (270-670 km) and lower mantle (670-2900 km) is represented by a Maxwell rheology (viscosities of 3×10^{20} Pa.s and 3×10^{21} Pa.s, respectively). Note that assuming free-slip or free boundary conditions at the base of the asthenosphere, i.e., inviscid deep upper and/or lower mantles, would lead to very different results (Boulze et al., 2025).

2.3 Modelling surface deformation of the overriding plate

Surface deformation of the lithosphere is modeled following the backslip formulation of Savage (1983). The asthenosphere behaves linearly over the relevant time scales and stress levels (Boulze et al., 2022). Because of this linearity, the resulting surface deformation can be expressed as the superposition of the responses of the uniform long-term slip and the complementary slip on the plate interface (Fig. 2). The uniform slip rate, associated with steady plate convergence, does not induce internal deformation within either plate. Thus, the deformation within the overriding and subducting plates is only due to the complementary slip. In the following sections, we discuss different ways of prescribing this complementary slip over time. We also relate it to present-day surface deformation. Note that the term 'slip' will designate this 'complementary slip' in the following.

In all numerical simulations, the complementary slip either corresponds to an earthquake or is prescribed to be proportional to the interplate coupling distribution. When coupling is equal to 0, the total slip velocity is equal to the convergence velocity, and the complementary slip velocity is equal to zero. When coupling is equal to 1, the interface is locked, and the complementary slip velocity is equal to minus the convergence velocity. Our coupling distribution combines three published models. We use Metois et al. (2016) between 18°S and 22.8°S , and between 28.5°S and 37.5°S . We use Klein et al. (2018) between 22.8°S and 28.5°S . South of 37.5°S , we use Moreno et al. (2011). The along-strike details of the coupling pattern yield little impact on far-field deformation. Far-field deformation is only sensitive to the long wavelengths of the coupling pattern, scaling with the distance from the trench. All coupling distributions used in this study are limited to depths shallower than 69 km. This avoids slip at the contact between the slab and the asthenosphere since the thickness of the continental lithosphere is set to 70 km. Along the entire subduction interface, the slip direction is assumed to be parallel to the plate-convergence direction, with an azimuth of 78°N and a convergence rate of 68 mm/yr (Vigny et al., 2009). For illustration, a fully coupled segment (coupling = 1), with this convergence rate, produces a coseismic slip of 10.2 m for a recurrence interval of 150 yrs.

Table 1. Definitions of the mathematical notations used in the paper.

Notation	Definition
(x, y)	coordinates of a point on the subduction interface
(λ, φ)	coordinates of a point on the Earth's surface
t	time
t_0	time when the surface velocity is calculated (present time)
$\vec{v}_s(\lambda, \varphi, t)$	velocity of a point on the Earth's surface with coordinates λ, φ
$v_{int}(x, y, t)$	slip velocity on the interface (corrected from long-term velocity)
$d_{int}(x, y, t)$	slip displacement on the interface (corrected from long-term velocity)
$\vec{v}_{s1}(\lambda, \varphi, x, y, t_0 - t)$	surface velocity in (λ, φ) induced by a unitary slip in (x, y) which occurred $(t_0 - t)$ years ago
$\vec{d}_{s1}(\lambda, \varphi, x, y, t_0 - t)$	cumulated surface displacement in (λ, φ) induced by a unitary slip in (x, y) which occurred $(t_0 - t)$ years ago
$\vec{d}_s(\lambda, \varphi, \tau)$	surface displacement during the time-lapse τ
$D_{int}(x, y)$	distribution of slip on the interface
$\vec{V}_{Eq-s}(\lambda, \varphi, t)$	surface velocity induced by the distribution of interface slip D_{int} which occurred t years ago
$\vec{D}_{Eq-s}(\lambda, \varphi, t)$	cumulated surface displacement induced by the distribution of slip D_{int} which occurred t years ago

3 MODELLING THE SURFACE DEFORMATION THROUGH THE SEISMIC CYCLE: THE CONVOLUTION FORMULATION

In this section, we describe the general formalism that relates the deformation at the surface of the overriding plate to the history of slip on the subduction interface. We assume a viscoelastic asthenosphere. To help the reader, Table 1 summarises the notation used. The mechanical properties of the asthenosphere, appropriate for describing deformation during the seismic cycle, are linear (Boulze et al., 2022). Thus, the superposition principle applies. The velocity at time t_0 , at any point on the surface (λ, φ) , can be computed as an integral over time and over the subduction interface. This integral sums the velocity induced by past and present slip on each elementary patch of the subduction interface. This relationship is expressed as

$$\vec{v}_s(\lambda, \varphi, t_0) = \iint_{interface} \int_{-\infty}^{t_0} v_{int}(x, y, t) \times \vec{v}_{s1}(\lambda, \varphi, x, y, t_0 - t) dt dx dy \quad (1)$$

Here, $\vec{v}_{s1}(\lambda, \varphi, x, y, (t_0 - t))$ is the time derivative, in $\tau = t_0 - t$, of the surface displacement $\vec{d}_{s1}(\lambda, \varphi, x, y, t)$. This displacement is induced by a unit slip located at (x, y) over a unit surface of the interface (Fig. 3). $v_{int}(x, y, t)$ denotes the complementary slip rate, i.e., the convergence-parallel velocity, at time t for the point on the subduction interface with coordinates (x, y) . This slip rate is corrected for the far-field convergence velocity. $d_{int} = v_{int}dt$ is then the slip over a short time span dt . By construction, v_{int} should have a zero long-term average at all points on the interface.

The kernel $\vec{v}_{s1}(x, y, \lambda, \varphi, \tau)$, with $\tau = t_0 - t$, represents the sensitivity of the surface deformation located at (λ, φ) to the unit slip. This unit slip occurred over a unit surface located at (x, y) on the interface, at a time lag τ earlier. This kernel has units of $s^{-1}m^{-2}$. At $\tau = 0$, $\vec{v}_{s1}$ behaves like a delta function in time, multiplied by the coseismic surface displacement induced by a unit slip over a unit surface on the interface. Its time integral, $\vec{d}_{s1}(x, y, \lambda, \varphi, 0)$, corresponds to a jump equal to the coseismic displacement in (λ, φ) , induced by this unit slip on the interface.

Note that, for simplicity, in Eq. 1, v_{int} is written as a scalar. This is appropriate if the slip is always in the same direction on the interface, for example, the direction of plate convergence. If this is not the

case, the same equations can be written for a slip rate in the perpendicular direction. The total surface displacement is then simply the sum of these two components.

The displacement of a point at the surface of the plate, between time t_1 and time t_2 , is equal to

$$\vec{d}_s(\lambda, \varphi, t_2 - t_1) = \int_{t_1}^{t_2} \vec{v}_s(\lambda, \varphi, t) dt \quad (2)$$

Fig. 3 shows the response of the surface of the overriding plate to a unit slip along the subduction fault plane (Fig. S1). Three features are key to the model predictions presented in the subsequent sections.

- The velocity $\vec{v}_{s1}(\lambda, \varphi, x, y, t)$, i.e., the time derivative of $\vec{d}_{s1}(\lambda, \varphi, x, y, t)$, becomes negligible after a duration. This duration depends on both the distance to the trench and the employed rheology. For the Burgers rheology (dashed lines in Fig. 3), this time is long, several thousand years, for points in the far field. This means that present-day deformation will be sensitive to very ancient slip history or earthquakes on the interface.

- We introduce the ratio R , which characterises the ratio of 'long-term' to 'elastic' displacements: $R = \frac{\|\vec{d}_{s1}(\lambda, \varphi, x, y, +\infty)\|}{\|\vec{d}_{s1}(\lambda, \varphi, x, y, 0)\|}$. This ratio increases with the distance from the coast. Near the coast, at less than 160 km from the trench, the ratio is close to 1. Thus, the elastic and viscoelastic rheologies predict a rather similar $\vec{d}_{s1}$ (Fig. 3B). For the same v_{int} , they therefore predict a similar velocity as a function of time. The elastic approximation can then be considered a reasonable first approximation for the viscoelastic solution. $\vec{d}_{s1}(\lambda, \varphi, x, y, +\infty)$ is independent of the rheology in the asthenosphere. This is because it represents the solution for a totally relaxed, i.e., inviscid, asthenosphere.

- A few decades, 50 years, after the slip on the interface, the far-field velocities predicted by the Maxwell rheology are much larger than those of the Burgers rheology. Thus, observations made in the far field, some decades after a large earthquake, allow discrimination between the two tested rheological models (Fig. 3D, red dashed line in the inset).

4 SIMPLIFICATIONS OF THE SLIP HISTORY ON THE INTERFACE USED TO MODEL THE INTERSEISMIC VELOCITY

Eq. 1 uniquely relates the present-day surface deformation velocity of the overriding plate to the slip history on the subduction interface. This holds regardless of the rheology of the mantle, whether elastic or viscoelastic. If the slip history on the interface was known over the past centuries and millennia, there would be no difficulty in calculating the induced surface velocity and displacements for a given Earth mechanical model, using Eqs. 1 and 2. The main problem is that, usually, only partial information about what has happened over the last century, or centuries, is available. The commonly employed models differ in whether and how they account for the slip history on the subduction interface: some attempt to reconstruct it from the available historical and geological constraints, whereas others neglect it altogether. We identified four different slip implementations on the interface in the scientific literature for estimating the interseismic velocity (Fig. 4).

4.1 Elastic model

The *elastic* model assumes that, at any instant t_0 , the surface deformation is generated only by the Earth's instantaneous elastic response to slip on the subduction interface at that same time. In other words, the model ignores any deformation contribution from viscoelastic deformation in the asthenosphere induced by past slip. This corresponds either to an asthenosphere so viscous that it remains effectively exempt from viscous deformation over the timescale of the seismic cycle, or to a scenario in which the interface was fully uncoupled before t_0 . In this second scenario, $v_{int} = 0$ for all times except t_0 (remember that v_{int} is the complementary velocity). In this case, Eq. 1 reduces to:

$$\vec{v}_s(\lambda, \varphi, t_0) = \iint_{interface} v_{int}(x, y, t_0) \times \vec{d}_{s1}(\lambda, \varphi, x, y, 0) dx dy \quad (3)$$

Here, $\vec{d}_{s1}(\lambda, \varphi, x, y, 0)$ was introduced in Section 3, before Eq. 2. The time evolution of this model is shown in Fig. 4A (blue line).

4.2 Steady model

The *steady* model assumes that the present-day slip on the interface has been constant for a very long time, without any earthquakes occurring. Thus, the interseismic surface velocity field is modelled by applying a constant backslip on the fault interface, for an infinitely long time. This backslip equals the product of the long-term convergence slip rate and the present-day coupling (purple line in Fig. 4B). This model is mainly used to describe the velocity pattern in the interseismic phase. As v_{int} is constant, Eq. 1 becomes:

$$\begin{aligned}\vec{v}_s(\lambda, \varphi, t_0) &= \iint_{interface} v_{int}(x, y) \int_{-\infty}^{t_0} \vec{v}_{s1}(\lambda, \varphi, x, y, t_0 - t) dt dx dy \\ &= \iint_{interface} v_{int}(x, y) \times \vec{d}_{s1}(\lambda, \varphi, x, y, \infty) dx dy\end{aligned}\quad (4)$$

Eq. 3 (*elastic* model) and Eq. 4 (*steady* model) are similar. The only difference is that $\vec{d}_{s1}(\lambda, \varphi, x, y, 0)$ has been replaced by $\vec{d}_{s1}(\lambda, \varphi, x, y, \infty)$, i.e., multiplied by R (Fig. 3B). Similarly, one can compute, at time t_0 , the surface velocity induced by a constant coupling that exists uniquely between times t_2 and t_1 , where $t_2 < t_1 < t_0$:

$$\begin{aligned}\vec{v}_s(\lambda, \varphi, t_0) &= \iint_{interface} v_{int}(x, y) \int_{t_2}^{t_1} \vec{v}_{s1}(\lambda, \varphi, x, y, t_0 - t) dt dx dy \\ &= \iint_{interface} v_{int}(x, y) \times (\vec{d}_{s1}(\lambda, \varphi, x, y, t_0 - t_2) - \vec{d}_{s1}(\lambda, \varphi, x, y, t_0 - t_1)) dx dy\end{aligned}\quad (5)$$

Eq. 5 is used in the supplementary material, where cases of constant coupling over a limited time interval are considered.

Three important properties of the *steady* model are worth noting.

(i) The total surface displacement, accumulated over an infinite time, $\vec{d}_{s1}(\lambda, \varphi, x, y, \infty)$, is completely independent of the chosen asthenosphere rheology. This state represents full relaxation of all stresses in the asthenosphere. An alternative way to reach the same conclusion is to use the scaling relationships between time and viscosity (see appendix S6-1 of Boulze et al. (2022)). Note that the value of $\vec{d}_{s1}$ at infinite time is equivalent to its value at finite time, when a null viscosity is assigned to the asthenosphere.

(ii) The surface velocity predicted by the *steady* model depends on the present-day velocities at the interface, $v_{int}(x, y)$, and on the elastic parameters of the continental lithosphere. It does not depend on the rheology of the mantle below the plate. Failure to recover this through numerical integration indicates that the integration has not been pursued far enough in time. Note that, in 2D problems (e.g., Trubienko et al., 2013), this leads to a constant stretching of the plate between the end of the coupling zone and the continentward limit of the mesh. The solution then depends strongly on the lateral size of the mesh.

(iii) In 3D simulations, the solution also depends on the mesh size (Boulze et al., 2025). If only one segment of a subduction interface is considered, geometrical dispersion overcomes this problem. This holds as long as the lateral extent of the mesh is significantly larger than the along-strike width of the segment. However, when coupling is applied over a long portion of a subduction zone, for example over the entire Chilean interface, the mesh must extend very far for the result to be independent of the meshed area size.

Both the *elastic* model and the *steady* model predict surface deformation that can be simply related to the present-day slip at the interface. The difference lies only in the nature of the material below the plate. The elastic model assumes that the mantle beneath the plates is elastic, with parameters compatible with seismological observations. The *steady* model, in contrast, is equivalent to a model with an inviscid fluid below the lithosphere, in which the slip at the interface may have been applied only since a short time. In other words, the viscosity of the asthenosphere plays no role in the *steady* model. Close to the coast, as R approaches 1, the *elastic* and *steady* models predict rather similar

deformation. Note that the *steady* model does not respect the backslip principle, which states that the slip at the interface, used to compute surface deformation, should have a zero long-term average.

4.3 Steady model corrected from the postseismic of the latest earthquake (*steady + last EQ*)

The *steady + last EQ* model is a variant of the *steady* model. It is also used to model postseismic surface deformation and velocity during the seismic cycle, in the continental plate reference frame (Sun et al., 2018). The contribution of the last known earthquake, as well as that of the associated afterslip on the fault plane, is then added to the *steady* model computed from Eq. 4 (orange line in Fig. 4C). As discussed later, this approach is correct as long as the steady velocity computed from Eq. 4 leads to a valid approximation. This requires that the influence of all earthquakes preceding the last one be negligible.

4.4 Cyclic model

The *cyclic* model (red line in Fig. 4D) belongs to a large class of models. These models assume that coupling, earthquake slip, and afterslip are all three homothetic. This implies that slip on the interface, at any time, can be written as $D_{int}(x, y) \times f(t)$. Here, $D_{int}(x, y)$ is a slip distribution on the interface, and $f(t)$ is the variation of slip with time on the interface. Note that D_{int} differs from d_{int} , the slip at a single point of the interface. The slip rate on the interface is then $v_{int}(x, y, t) = D_{int}(x, y) \times f'(t)$. $\vec{D}_{Eq-s}(\lambda, \varphi, t)$ denotes the surface response, expressed in (λ, φ, t) , of an earthquake characterized by a displacement distribution $D_{int}(x, y)$ on the interface. The velocity at time t_0 , generated by slip on the interface between times t_1 and t_0 , is then directly obtained from Eq. 1.

$$\vec{v}_s(\lambda, \varphi, t_0) = \int_{t_1}^{t_0} f'(t) \iint_{interface} D_{int}(x, y) \times \vec{v}_{s1}(\lambda, \varphi, x, y, t_0 - t) dx dy dt = \int_{t_1}^{t_0} f'(t) \times \vec{V}_{Eq-s}(\lambda, \varphi, t_0 - t) dt \quad (6)$$

$\vec{V}_{Eq-s}$ is the time derivative of $\vec{D}_{Eq-s}(\lambda, \varphi, t)$. To satisfy the condition that the long-term average interface slip rate, corrected for the convergence slip rate, tends towards zero, the average of $f'(t)$ over a sufficiently long time must be close to zero. The function $f(t)$ can be chosen so that it encompasses all that is known about past earthquakes, from historical or archaeological sources or geological records. In the absence of detailed knowledge, earthquakes can be assumed to be regular and similar. In this case, the recurrence time T equals the interval between the last two events. If $D_{int}(x, y)$ is the coseismic slip of one earthquake on the interface, then the backslip velocity on the interface is $-D_{int}(x, y)/T$, i.e., $f'(t) = 1/T$.

Unlike the *elastic* and *steady* models, which predict a constant surface deformation between earthquakes, the *cyclic* model yields a deformation that evolves continuously throughout the seismic cycle. This deformation is the sum of two contributions. The first is the deformation induced by the constant long-term interseismic slip, $(-D_{int}(x, y)/T)$. The second is the postseismic deformation induced by all earthquakes that occurred $t_0, t_0 + T, t_0 + 2T \dots t_0 + NT$ years ago. It is then equal to:

$$\vec{v}_s(\lambda, \varphi, t_0) = -\left[\frac{\vec{D}_{Eq-s}(\lambda, \varphi, t_0 + N_A.T)}{T}\right] + \sum_{k=0}^{N_B} \vec{V}_{Eq-s}(\lambda, \varphi, t_0 + k.T) \quad (7)$$

Here, N_A and N_B are two possible values for N . N_A appears in the first term of Eq. 7. It represents the contribution of steady coupling over $N_A.T$ periods. N_B appears in the second term. It represents the contribution of N_B past earthquakes to the present-day velocity. As these two terms are distinct, there is no obvious a priori reason for N_A and N_B to be equal. However, they must in fact be equal to satisfy a key condition. If all deformation is to be recovered after one complete seismic cycle, the total displacement of any point on the upper plate must be zero. We show, through numerical tests, that this condition is satisfied only when $N_A = N_B$ (Fig. S2). In practice, $N_A = N_B = 10$ provides a good compromise between computational cost and convergence of the solution (Fig. S3). As N_A and N_B tend toward infinity, the first term converges to a constant backslip, directly obtained from

Eq. 5. Meanwhile, $\vec{V}_{Eq-s}(\lambda, \varphi, t_0 + k.T)$ denotes the postseismic velocity induced by an earthquake that occurred $(t_0 + k.T)$ years ago.

The velocity at a given surface point, due either to a past earthquake or to a steady coupling during any time interval in the past, can easily be computed from $\vec{D}_{Eq-s}(\lambda, \varphi, t)$. The computation of $\vec{D}_{Eq-s}(\lambda, \varphi, t)$ is the only step that requires substantial computational time. For example, replacing the *steady* model with one that includes earthquakes compatible with an average slip on the interface equal to the convergence slip rate can be achieved with almost no extra computational cost. Changing the slip history as a function of time can also be performed without extra computational time. It is, for example, easy to implement a succession of earthquakes with variable magnitudes (black line in Fig. 4).

Note also that the velocity induced at time t_0 on the surface, by a constant coupling characterised by a velocity on the interface equal to $D_{int}(x, y)/T$ between times t_1 and t_2 , is expressed quite simply as:

$$\vec{v}_s(\lambda, \varphi, t_0) = -\left[\frac{(\vec{D}_{Eq-s}(\lambda, \varphi, t_0 - t_2) - \vec{D}_{Eq-s}(\lambda, \varphi, t_0 - t_1))}{T}\right] \quad (8)$$

Here, $\vec{D}_{Eq-s}(\lambda, \varphi, t_0 - t_2)$ and $\vec{D}_{Eq-s}(\lambda, \varphi, t_0 - t_1)$ are simply the displacements generated at times t_2 and t_1 by an earthquake that occurred at time t_0 , with a slip distribution $D_{int}(x, y)$ on the interface (like in Fig. 3 but for a distribution of displacement corresponding to the earthquake).

5 COMPARISON OF THE RESPONSES INDUCED BY VARIOUS SLIP HISTORIES ON A SINGLE SUBDUCTION SEGMENT

Using forward simulations, we compare interseismic velocity predictions from the four slip-history models described in the previous section. Simulations are performed for both the Maxwell and Burgers rheologies of the asthenosphere. They use the coupling distribution along the Maule segment of the Chilean subduction zone, between 33.5°S and 37.5°S (see Section 2). For the *elastic* and *steady* models, the surface velocities are directly deduced from the elastic structures of the lithosphere and the underlying mantle. They are also deduced from the interface velocity $v_{int}(x, y)$, i.e., through the coupling distribution. For the *cyclic* model, the earthquake slip distribution $D_{int}(x, y)$ is equal to $D_{int}(x, y) = -v_{int}(x, y) \times T$, where T is the recurrence time. Here, we choose $T = 150$ yrs. All velocities are expressed in a South American reference frame. Note that we did not include any after-slip in this simulation. The velocities for the first years after the earthquake should therefore not be directly compared with the data.

5.1 Comparison of interseismic velocities predicted by the different slip models

The *elastic* and *steady* models predict constant velocities. In contrast, the *cyclic* and *steady + last EQ* models predict time-variable velocities. For this reason, we choose to first compute the velocities for all models at the end of the seismic cycle, i.e., at $t_0 = 150$ yrs, just before the next earthquake (Figs. 5,6).

In the **near-field** (within approximately 200 km of the trench), all models predict velocities with very close amplitudes (Figs. 5,6). This consistency confirms that, in this narrow region, *elastic* models provide a reasonable approximation of interseismic deformation, when compared to *steady* or *cyclic* models.

In the **mid-field** (roughly 250-500 km from the trench), all three viscoelastic models predict a smaller velocity gradient than the *elastic* model (orange, purple, and red lines vs. blue line in Fig. 6). Both the *steady* and *steady + last EQ* models (purple and orange vectors in Fig. 5 and solid lines in Fig. 6), and the *cyclic* model (red vector/line), predict more gradual reductions in velocity. In the case of the Burgers rheology, the predictions from the three viscoelastic models differ significantly in the mid-field (Fig. 6B, orange and purple vs. red and dashed black lines). As mentioned in the introduction, these mid-field interseismic velocities lie at the centre of ongoing debates. These debates concern the depth of coupling on the interface, and the existence of a potential third plate and its velocity relative to the inland block.

In the **far-field** (beyond ~500 km from the trench), velocities predicted by the *elastic* model

become negligible. They are typically less than 1 mm/yr beyond 1000 km from the trench. Considering first the Maxwell rheology, all three viscoelastic models still predict measurable deformation in the far field. The *steady* model produces large far-field velocities (Fig. 5A, purple vector; Fig. 6A, purple line). The *cyclic* model, whether using regularised (red vectors/lines) or realistic recurrences (black dashed line), also produces similarly large far-field velocities. Going further, the *cyclic* model is the only one sensitive to the rheology. Both the steady models indeed predict the same far-field deformation whether in Maxwell or in Burgers (Fig. 6A&B, purple and yellow lines). On the contrary, the *cyclic* model within a Burgers rheology predicts a significant decrease with distance to the trench (Fig. 6B, red and dashed black lines).

With smaller earthquakes occurring over shorter recurrences, all other parameters being equal, the two rheologies yield quite different far-field velocities (Fig. S4). Interestingly, models using the Burgers rheology are less sensitive to these variations of earthquake size in the seismic cycle than models using the Maxwell rheology (orange and red lines in Fig. S4B vs. Fig. S4A). In short, the *steady* and *cyclic* models predict very different velocities when the effective viscosity of the asthenosphere is high enough. In this case, the displacement $\vec{d}_s$ has not yet reached its asymptotic value at the end of the cycle (see Fig. 3C,D).

The responses of the *cyclic* models are very similar, whether they use a realistic or a regularised earthquake recurrence time interval, and regardless of the rheology (Fig. 6, red line vs. dashed black line). This seems to indicate that the details of the slip history before the last two earthquakes have a negligible influence on surface velocity. This holds at least as long as the complementary slip has an average close to zero over the last millennium.

A thorough analysis of the impact of non-periodic slip history on surface velocities, i.e., irregular cycles alternating small and large earthquakes over irregular time windows, can be found in Section S4 (Figs. S6, S7, S8, S9, S10). The main conclusions derived from these tests are summarised in Section 7.2.

5.2 Time variability of surface velocities during the seismic cycle

The general evolution of surface velocities during the seismic cycle is well documented in previous studies (e.g., Wang et al., 2012; Trubienko et al., 2013; Sun et al., 2018). This evolution goes from predominantly trenchward motion, shortly after an earthquake, to entirely continentward motion, toward the end of the cycle. Rather than revisiting these established characteristics, we focus here on comparing how this evolution differs for two viscoelastic rheologies of the asthenosphere, Maxwell and Burgers. We also compare the two slip models that allow surface velocities to vary throughout the seismic cycle: the *steady + last EQ* model and the *cyclic* model. These temporal variations arise from the cumulative effects of viscoelastic deformation. For the *steady + last EQ* model, this deformation follows the last earthquake. For the *cyclic* model, it follows all past seismic events. To illustrate this time dependence, we compute surface velocities for both Maxwell and Burgers rheologies, at several stages of the seismic cycle: +5, +10, +30, and +75 years after the most recent earthquake (Figs. 7 and 8). For +150 yrs, see Fig. 5.

In the early postseismic phase, 5 yrs after the earthquake, the near-, mid-, and far-field velocity fields are predominantly oriented toward the trench (Figs. 7 and 8). Deformation of the overriding plate is then largely controlled by viscoelastic deformation in the asthenosphere, following the most recent earthquake. Continentward-oriented velocities are restricted to regions close to the coast (Fig. 7). This suggests that the contribution of the steady backslip term, interpreted as fault re-locking, locally exceeds that of viscoelastic deformation from past earthquakes.

At later stages of the seismic cycle, near- and mid-field velocities become uniformly oriented toward the continent, regardless of the distance from the trench (Fig. 7).

In the far field, in contrast, the timing of this reversal differs markedly between models (Fig. 8). Both the *cyclic* and *steady + last EQ* models, with the Maxwell rheology, reverse between 30 and 75 years. With the Burgers rheology, the *steady + last EQ* model reverses earlier, between 5 and 10 years. For the *cyclic* model, it predicts the smallest far-field velocities at 75 years, and an intermediate reversal time between 10 and 30 years.

5.3 Impact of the seismic cycle duration

Figs. 7, 8 illustrate various aspects of the predicted velocities for a period of the seismic cycle equal to $T = 150$ years. The recurrence time between earthquakes (here T) has a direct impact on surface

deformation. This impact arises through larger or smaller magnitudes implied by a longer or shorter cycle. We quantify this impact by comparing displacements, as a function of time, at three distances from the trench: 200, 400 and 1400 km. We use two largely different recurrence times, $T = 150$ years ($M_w 9.0$) and $T = 900$ years ($M_w 9.5$). Fig. 9 compares the displacement for three rheologies: elastic, Maxwell and Burgers rheologies. Recall that velocity is the time derivative of displacement. The minimum of each curve in Fig. 9 is thus reached at the reversal time, i.e., the moment at which velocities switch from trenchward to continentward.

In the near field, at 200 km from the trench, velocities are oriented continentward after less than 10% of the cycle duration. This holds for both cycle durations and for all rheologies (Fig. 9A). In the mid field, at 400 km from the trench, differences start to appear between models. However, the reversal time remains close to 20% of the cycle duration for all models and for both cycle durations (Fig. 9B). In the far field, at 1400 km from the trench, the behaviour differs markedly, both between the two cycle durations and between the two rheologies (Fig. 9C). For $T = 150$ years, the Maxwell rheology reverses at around 40 to 50% of the cycle, while the Burgers rheology reverses earlier, at around 20%. For $T = 900$ years, this ordering is reversed. The Maxwell rheology now reverses earlier, at around 20% of the cycle, while the Burgers rheology reverses later, at around 40 to 50%.

The relative difference between the predictions of the Maxwell and Burgers rheologies increases with distance from the trench. Examination of far-field velocities is therefore particularly important for discussing whether the Maxwell or the Burgers rheology is the more likely one. The time for switching from trenchward to continentward velocities depends on the distance to the trench. It also depends on the length of the cycle T , and on the magnitude of the last earthquake. For example, the far-field velocities in southern South America may have a component oriented towards the source of the Valdivia earthquake for many decades after 1960.

6 INTERACTION AMONG SEGMENTS OF THE SUBDUCTION ZONE: A FORWARD MODELLING OF MULTI-SEGMENT SEISMIC CYCLE ON THE SUBDUCTION ZONE

In reality, the deformation associated with the seismic cycle along a subduction zone, such as the Chilean margin, is more complex to model. Subduction zones are segmented into several portions called segments. These portions are characterised by earthquakes of different sizes and recurrence times. Their seismic cycles are not synchronised (Ruiz and Madariaga, 2018). In the far field, the velocities induced by slip on desynchronized segments may have identical or opposite signs. The resulting amplitude may increase or decrease. In this section, we show that properly accounting for all segments and their respective earthquake histories is necessary to correctly evaluate the deformation at the continental scale, associated with the seismic cycle of the various segments that make up the subduction zone.

We identify 10 segments along the Chilean subduction zone (Table 2). We assign each segment a recurrence time T , an earthquake magnitude M_w , and a coseismic slip distribution proportional to local coupling. When a recent large earthquake is known, the magnitude of the most recent event is rescaled accordingly. Details on the choice of T and M_w for each segment are given in Section S5. All simulations target surface deformation of the South American plate in 2010, before the Maule, Iquique, and Illapel earthquakes. They are run for both Maxwell and Burgers rheologies.

6.1 Differences between the predictions of the various models at continental scale

The predicted velocities, at the scale of the South American plate, are shown in Figs. 10 and 11. At this stage, the goal is not to fit the observations in detail.

First, the near and mid fields show some differences. They appear specifically where large earthquakes have occurred recently, or even less recently if they were very large. At the coast, differences between models are smaller than 10%, except facing the Valdivia coseismic rupture.

In 2010, the entire northern part of the subduction zone, north of 30° , had not experienced a notable earthquake for at least 100 years (Ruiz and Madariaga, 2018). With a Maxwell rheology (Fig. 10), the *steady + last EQ* and *cyclic* models behave similarly in this region. Remarkably, the impact of the Valdivia earthquake is clearly visible, with a quasi-rotation of velocities across the entire continent. In contrast, the influence of Valdivia seems to be suppressed with the Burgers rheology (Fig. 11). The

Table 2. Segmentation of the Chilean subduction zone and associated seismic cycle parameters used in this study. **Name:** name of the segment, **Extent:** latitude extent of the segment, $M_w(\mathbf{c})$: magnitude of the cyclic earthquake derived from the coupling model, cycle duration T and convergence velocity, $M_w(\mathbf{m})$: Magnitude of the last earthquake that broke the segment. When it is followed by (??), it means the absence of reliable records, and so the magnitude $M_w(\mathbf{c})$ is used in a first-order approximation. **T:** return time of earthquakes in years. **Last known event:** data and location of the last major earthquake known to have broken the segment.

Name	Latitude extent	$M_w(\mathbf{c})$	$M_w(\mathbf{m})$	T	Last known event
Patagonia	[-47.0, -37.5]	9.56	9.5	1000	1960 Valdivia
Maule	[-37.5, -33.5]	9.00	8.8	150	1835 Darwin earthquake
South Metropolitan	[-33.5, -32.0]	8.54	8.0	80	1985 Valparaiso earthquake
North Metropolitan	[-32.0, -30.0]	8.43	8.3	80	1943 Illapel earthquake
Atacama	[-30.0, -27.5]	8.55	8.5	100	1922
Chañaral	[-27.5, -25.2]	8.70	8.0	150	1918
Paranal	[-25.2, -22.8]	8.48	8.0	80	1995 Antofagasta
Loa	[-22.8, -20.5]	8.74	8.74(??)	150	1877
Camarones	[-20.5, -19.5]	8.56	8.56(??)	150	(??) set to 2010-150
Arica	[-19.5, -18.0]	8.60	8.60(??)	150	1868

cyclic model now predicts far-field velocities that are much lower than the *steady + last EQ* model, which in contrast, predicts far-field deformation very close to that of the *steady* model.

In the southernmost part of the continent, even at large distances from the trench, deformation is dominated by the large viscoelastic response to the 1960 Valdivia earthquake. This contribution is significantly larger for the Maxwell than for the Burgers rheology (red dashed lines in Fig. 3D). This is reflected in the *steady + last EQ* model predictions: They are oriented trenchward for the Maxwell rheology, and continentward for the Burgers rheology (orange vectors in Figs. 10 and 11).

6.2 Comparison with GNSS data in the southern part of the continent

In this subsection, we quantitatively compare model predictions with geodetic observations. We focus on the southern part of the continent, inland from four segments between 30°S and 47°S: the North Metropolitan, South Metropolitan, Maule, and Patagonia segments (Table 2).

We use GNSS time-series from the SOAM_GNSS_solENS_V6.0 solution (Klein et al., 2022). The raw GNSS data come from four databases: CSN (Baez et al., 2018), RAMSAC (Piñón et al., 2018), RBMC and IGS. We derive a GNSS velocity field over the 2000–2010 period, using continuous GPS stations (Table S2). This velocity field is expressed in a stable South American reference frame ($\lambda = -131.0918^\circ\text{E}$, $\varphi = -19.3817^\circ\text{N}$, $\Omega = 0.1205^\circ/\text{Ma}$), calculated for this study (Table S1). We then compare these velocities to predictions from all four slip models, for both Maxwell and Burgers rheologies (Fig. 12).

In the **near field**, the different models yield similar velocities. There, the *elastic* model provides a good fit to the observed velocities, with a mean residual of 3–4 mm/yr. This fit is equivalent to that of the *cyclic* model with the Burgers rheology. Differences appear mainly where large earthquakes have occurred recently, in particular using Maxwell rheology. For example, for the *cyclic* model, the residual increases to 6 mm/yr due to a large misfit at CONS, CONZ, ANTC and MAUL stations, impacted in the south by the Valdivia postseismic deformation. We therefore focus on the far field, where model predictions diverge most clearly.

In the **far field**, the observations show a clear pattern. In Argentina, the direction of the deformation rotates from eastward in the north (e.g., UCOR) to westward in the south (e.g., RWSN). Reproducing this rotation is the key test for any model.

The *elastic* model, which does not depend on asthenosphere rheology, yields a low residuals of 1.4 mm/yr. However, despite this low residual amplitude, the *elastic* model does not have the correct

direction. It predicts uniformly continentward velocities everywhere, at all epochs. The predicted vectors never reverse, regardless of location or the epoch considered. The *elastic* model, therefore, fails to reproduce the observed directional rotation.

The *steady* model fails to reproduce the observed rotation. It predicts eastward velocities throughout, with a residual of 11.2 mm/yr. This residual does not depend on the rheology used, since the *steady* model predictions are independent of asthenosphere rheology (Section 4.2).

The *steady + last EQ* model, with the Maxwell rheology, overpredicts westward motion in the south, leading to a large residual of 8.4 mm/yr (Fig. 12A). With the Burgers rheology, the residual remains large, at around 7.2 mm/yr. Accounting for only the most recent earthquake is therefore insufficient to explain the observed deformation, regardless of the rheology.

The *cyclic* model, with the Maxwell rheology, similarly overpredicts westward motion in the south, with a residual of 9.2 mm/yr (Fig. 12A). With the Burgers rheology, in contrast, the *cyclic* model closely reproduces the observed velocity field, including the north-to-south rotation, with a residual of 0.9 mm/yr (Fig. 12B).

In conclusion, the comparison between forward models and GPS observations demonstrates that only the *cyclic* model, using Burgers rheology, is able to reproduce both the amplitude and orientation of the interseismic deformation field that prevailed before 2010 at the continental scale. This model successfully captures the combined effects of interseismic deformation and the long-lasting postseismic deformation associated with the Valdivia earthquake. In contrast, all other slip models fail to reproduce the observed velocity field. Finally, the Maxwell rheology, with a viscosity on the order of 3×10^{18} Pa.s, fails to reproduce the observed deformation, regardless of the slip model used. This results in significant misfits in both velocity amplitude and orientation. This indicates that such low viscosities are not appropriate for describing deformation over decadal timescales following megathrust earthquakes.

6.3 Varying the moment of the Valdivia Earthquake and the maximum coupling depth

In the context of the *cyclic* model associated with the Burgers rheology that best fits the observations (Section 6.2), two parameters may influence the predicted velocity field. These parameters are the magnitude of the Valdivia earthquake and the coupling depth of the nearby segments. These parameters would also affect the *cyclic* models using the Maxwell rheology. They could improve the fit to the data. An inversion of all the parameters introduced in the model is beyond the scope of the present paper. However, we check whether our conclusion still holds for different values of the moment of Valdivia or with different coupling distributions.

Valdivia moment. All models using the Maxwell rheology yield large far-field deformation. We therefore examine whether artificially decreasing the Valdivia earthquake magnitude would compensate and allow the model to better fit the data (Fig. S12). With a M_w 9.4, we still obtain predicted deformation ten to twenty times larger than the observed values. Even reducing the magnitude to 9.3 still yields much too large velocities. So it is clear that a Maxwell rheology with a few 10^{18} Pa.s is not appropriate.

Variation of coupling depth. Here, we show how the coupling depth impacts interseismic velocity predictions within the framework of a *cyclic* model with a Burgers rheology (Fig. 13). As previously noted, all models perform almost equally in the near field. The exception is close to segments affected by a recent large earthquake (e.g., stations CONS, CONZ, ANTC, MAUL in Fig. 12A) due to the Valdivia earthquake. This means that the elastic coupling pattern is reasonably adequate, at least in its shallower part.

A change in coupling depth is not very well resolved by common elastic models. Such a change would mostly impact the mid- and far-field velocities, while having a limited impact on the near field. Therefore, a change in coupling depth may help some models perform better in the mid and far-field, while maintaining their fit in the near field.

Our test confirms this expectation. The impact of artificially reducing the coupling depth is indeed minimal in the near field, but significant elsewhere (Fig. 13). The mean residuals between observed and predicted interseismic velocities, in the mid and far fields, decrease from 0.9 mm/yr to 0.7 mm/yr when limiting the coupling to 45 km depth, instead of the actual ~ 69 km.

It is apparent that changing only one parameter characterising the coupling distribution, such as the depth of coupling, cannot yield a very precise fit. Fits are usually obtained by fine-tuning the value

of the coupling coefficient through inversion over every patch of the interface. Such an inversion is among our future projects but is beyond the scope of the present paper. Although not perfect, the simple test provided in Fig. 13 indicates that a close fit between predicted and observed velocities can be achieved with a *cyclic* model within a Burgers rheology. It will hopefully be possible to improve this fit further by tuning the coupling on each patch of the interface.

7 DISCUSSION

7.1 Asthenospheric rheology: *steady* vs. *cyclic* models on the interface

This paper lays the groundwork for identifying the differences between the predictions of the various models of interseismic plate deformation. These models are the *elastic*, *steady*, *steady + last EQ*, and *cyclic* models.

The effect of the slip model on the predicted surface velocities depends on the assumed rheology of the asthenosphere. In the case of the Maxwell rheology, the slip history older than about 150 years plays a minor role in present-day surface deformation. The Maxwell line reaches its asymptote around 100 years in the near field (Fig. 3C, black solid line). It reaches its asymptote around 200 years in the far field (Fig. 3D, black solid line). Thus, for times older than 150 yrs, two scenarios yield rather similar results. The first scenario is a permanent compressive loading of the interface by backslip, as in the *steady* model. The second scenario is a zero long-term average loading, where more or less periodic earthquakes release this compression, as in the *cyclic* model (see Fig. 5A, purple and red vectors).

In the case of the Burgers rheology, the *steady + last EQ* model predicts velocities much greater than those predicted by the *cyclic* model. The predictions of the Maxwell and Burgers rheologies differ in two ways. Between 30 and 100 years after an earthquake, the far-field postseismic deformation rate predicted by the Maxwell rheology is considerably larger than that predicted by the Burgers rheology (Fig. 3D). Also, during the last half of the interseismic phase, for a *cyclic* model, in the far field, the continentward velocity predicted by the Maxwell rheology is several times larger than that predicted by the Burgers rheology (Figs. 5, 6, 10, 11).

The postseismic deformation induced by the large Valdivia earthquake is clearly visible in the pre-2010 deformation rate observed by geodesy. However, this signal is considerably smaller than the one predicted by the Maxwell rheology. Additionally, further north, the observed eastward velocities are very small. Both features suggest that only the Burgers rheology can lead to predictions compatible with the observations. A smaller magnitude of the Valdivia earthquake (Fig. S12) and a reduced coupling depth (Fig. 13) would not alter this conclusion.

Lateral viscosity variations cannot explain the high effective viscosity inferred from the late post-seismic of Valdivia. A cooler or drier asthenosphere, situated below the stations registering the post-seismic deformation after Valdivia, could indeed explain a higher viscosity. This could explain the contrast between the viscosities inferred from the Valdivia and Maule earthquakes. However, this argument does not hold. In eastern Argentina, between 34°S and 40°S, velocities have been measured at the same stations both a few years after the 2010 Maule earthquake and fifty years after the 1960 Valdivia earthquake.

In short, the pronounced differences between the predictions of the two rheological models suggest that, even with additional refinements, the apparent viscosity of the asthenosphere must be significantly greater on decadal time scales than on annual ones.

7.2 Realistic versus regular cycles

With our Burgers rheology, the earthquake history over the last 1500 years influences the present-day deformation rate of the plate (Fig. 3). At first sight, this necessity to include the long-term history seems to preclude any prediction of present-day deformation, unless the earthquake and coupling history over the last 1500 years is well known. In the tests of Section 6, it is by far the *cyclic* model that provides the best fit to the data. However, one may argue that the *cyclic* model is a purely theoretical concept. It is the purpose of Section S4 to illustrate why this *cyclic* model is appropriate in most situations.

In most subduction zones, although earthquakes are not regular, the stress accumulated on the interface is released on a rather short time-scale, of a few hundred years.

Slow compressive accumulation and sudden coseismic extensive release of past accumulated

stresses have compensating effects on present-day surface deformation. This is illustrated in Fig. S5B. There, the surface velocity generated by an earthquake of magnitude 9, which occurred 150 years ago, combined with a compensating coupling from 150 to 300 years ago, is on the order of 0.2 mm/yr. For most subduction interfaces, which cannot sustain stresses corresponding to strain accumulated over long periods, the detailed slip history older than 50 years does not matter very much (Section S4, Figs. S6, S7, and S8). Thus, the present-day deformation is mainly a function of two things. The first is the present-day coupling, over a time-scale corresponding to the recurrence time of earthquakes on the studied segment. The second is the postseismic deformation generated by the last earthquake. For times older than that, imposing regular cycles, or imposing no displacement on the interface, affects the surface velocity negligibly. However, one must be careful not to impose, as proposed in the *steady* models, the coupling for a long period (1000 years) without releasing the accumulated strains by earthquakes.

In practice, we suggest using first whatever is known about the past slip history, in particular about the last major earthquakes. In the absence of data on ancient slip history, a *cyclic* model can be built from three ingredients. The first is the interface coupling, deduced from present-day near-field velocities. The second is cyclic earthquakes, with a period inferred from the last two major events. The third is a seismic moment such that the interface displacement from the earthquakes compensates for the displacement accumulated by backslip. In our tests (Section 6), we readjust the moment of the last earthquake to the known value. Such a model, even if it does not closely fit the actual slip on the interface as a function of time, is expected to correctly predict present-day surface velocities (see Fig. S6). This remains true as long as the long-term (1000 years) strain accumulation, on the considered interface, does not significantly exceed the strain accumulated during a cycle (see Fig. S9).

Indeed, some subduction segments may be prone to megacycles. In these megacycles, stresses accumulated over more than a millennium are suddenly released, across several segments that are apparently independent. Slow accumulation rates may favour these rather infrequent earthquakes. In such cases, a far-field velocity closer to that predicted by the *steady* model, compared to the *cyclic* model, is expected at the end of the mega-earthquake cycle (Fig. S5C; Fig. S9, history 7). This would indicate long-term compressive stress accumulation in the overriding plate, raising an alert concerning the possible future occurrence of a very large earthquake. Note that, in the case of the Valdivia segment, the velocities induced by the Valdivia earthquake are so large that an error in the slip on the interface several centuries ago, or in the recurrence period of mega-earthquakes, would negligibly affect the predicted velocities.

7.3 Elastic versus viscoelastic models and coupling depth on the interface

Although it is now generally accepted that the asthenosphere behaves as a viscoelastic material during the postseismic phase, *elastic* models are still widely used to model present-day coupling on subduction segments. It is well known that the predicted depths of coupling are much larger with elastic models than with viscoelastic models (Savage, 1983; Li et al., 2015). However, viscoelastic rheology is still rarely used to model interseismic deformation. Perhaps this is because, when using either steady or Maxwell rheologies, predicted velocities are unrealistically large at distances greater than ~500 km from the trench. The Burgers *cyclic* model proposed here predicts velocities compatible with observations. These velocities are larger than those predicted by *elastic* models, but smaller than those predicted by 'classical' viscoelastic models. Near-field deformation is comparatively consistent among all models (see Figs. 6 or 9A). However, the mid-field velocity (~400 km from the trench) predicted by *elastic* models is already significantly smaller than that predicted by our *cyclic* Burgers viscoelastic model, during most of the cycle (Fig. 9).

Several studies have inverted GNSS surface velocities in Chile to obtain the coupling distribution, using data from the coast up to 400 km from the trench (Metois et al., 2016), or even up to 700 km (Brooks et al., 2003; Bevis et al., 2001). At these distances from the trench, the forward models *elastic*, *steady*, and *cyclic* predictions yield different surface velocities for the same coupling distribution. The inferred maximum coupling depth is therefore larger for the *elastic* model, smaller for the *steady* model or the Maxwell viscoelastic model, and intermediate for our favourite *cyclic* Burgers rheology.

Here, to simplify the number of free parameters and cases, we did not introduce viscous channels along the interface, nor low-viscosity zones at the base of the lithosphere below the volcanic arc. However, for a given coupling, these channels and low-viscosity zones are expected to slightly increase the magnitude of interseismic velocities in the mid-field, but not in the far-field (Trubienko et al., 2014). Hence, taking into account the channels inferred from postseismic models (Klein et al., 2016) would further reduce the coupling depth inverted from the data, compared to the predictions of *elastic*

models. It would also somewhat affect the coupling depth inferred from the *cyclic* model with Burgers rheology.

7.4 Slivers and blocks

To fit the observed interseismic velocity gradient in the mid-field, simple two-plate elastic models require coupling at unrealistically deep depths Savage (1983). When the coupling depth is constrained to more realistic values, the predicted gradient is significantly lower than observed. To address this mismatch, more complex models were suggested, incorporating a sliver or a third microplate between the subduction zone and the interior of the main plate (Bevis et al., 2001; Brooks et al., 2003; Moreno et al., 2011; Métois et al., 2012, 2013, 2014; Nocquet et al., 2014; Métois et al., 2016; Villegas-Lanza et al., 2016; Klein et al., 2018; Jolivet et al., 2020; Jarrin et al., 2023; Loverly et al., 2024). The role of the sliver is to explain the significant inland velocities through its own block motion.

The present study may greatly help in defining these blocks and their velocities more clearly. As an illustration, the mid-field interseismic velocity, between 200 and 500 km from the trench, predicted by our *cyclic* model, is significantly different from that predicted by both the *elastic* and the *steady* models (Fig. 6B). As a result, the velocity of the sliver relative to the bulk of the continental plate, and in some cases even its existence, must be reconsidered.

We do not deny the necessity of plastic intraplate deformation in the hinterland of some subduction zones. However, there is no reason for the mid-field velocities predicted by an *elastic* model to be upheld. Moreover, if these blocks or microplates prove necessary, we suggest that they should be considered neither as rigid nor as elastic entities, overriding an elastic mantle. Instead, their internal deformation should depend on the slip history of their borders, and on the viscoelastic properties of the asthenosphere beneath them, just as found for the slip history along subduction plate boundaries.

7.5 Definition of stable regions for plate motion modelling

Our results further indicate that the South American plate does not behave as a rigid block. This holds even at distances of several thousand kilometers from the subduction zone. It also holds even though the last major earthquake occurred more than 50 years ago. Consequently, stations used to define the so-called low-deformation zones may be significantly affected. This is particularly relevant when determining plate-motion models, such as those used in the ITRF framework. For example, the stations VBCA and MPLA (Fig. 12), which have been included in the ITRF plate model since ITRF2020 (Altamimi et al., 2023), are not located sufficiently far from the plate boundary. Similarly, the station LPGS, used in ITRF2000, ITRF2005, and ITRF2008, is also affected (Fig. 12). The region in which these stations are situated cannot be considered stable and representative of the plate interior.

Recent ITRF realizations do account for postseismic deformation following well-documented, recent earthquakes. However, our results show that the interseismic signal is also affected by much older, very large events. In this region, this is the case for the 1960 Valdivia earthquake. Accounting for postseismic deformation from moderate Mw8+ earthquakes over the last few decades is therefore not sufficient. The much longer-lasting viscoelastic relaxation following the largest Mw9+ earthquakes, even more than half a century later, must also be accounted for.

7.6 The rheology of the asthenosphere

Although tested only in South America, we believe that the proposed rheology applies to the asthenosphere below most areas. Exceptions are cratons and regions affected by hot spots. The effective viscosity on the scale of a few years is very similar for the three major earthquakes of the last thirty years, namely Sumatra, Maule, and Tohoku (Trubienko et al., 2014). It is also very similar for the last three large Chilean earthquakes, namely Maule, Iquique, and Illapel (Boulze et al., 2022).

We used a single-Burgers rheology in this study. Although this single-Burgers rheology provides a good fit to GNSS data, we believe that an Extended Burgers rheology (Ivins et al., 2021) is more appropriate to describe the rheology of the asthenosphere over a wide range of timescales. For example, our Burgers rheology does not account for seismic attenuation. Additionally, a Kelvin-Voigt element with a large μ_K and a low viscosity (10^{18} Pa.s or smaller) has been proposed to explain the fast deformation observed in the first months after an earthquake. Such an element should be added to the Burgers rheology described in this study (Fig. 1) to explain the early postseismic deformation. In most studies of postseismic deformation using a Burgers rheology, the Kelvin-Voigt element is introduced to fit this initial short phase of fast transient creep (Pollitz, 2003; Klein et al., 2016). We

are rather convinced of the existence of this rapid initial creep, although it is sometimes difficult to separate its impact on surface velocities from that of afterslip on the interface. However, the velocities on decadal timescales studied here would not be affected by modifications of the rheology affecting short timescale deformation.

Note that the type of long-term transient creep introduced here requires studies of postseismic effects spanning more than 15 years. Most studies of postseismic deformation, so far, cover only a few years. This is why most studies consider only the short-term transient creep. Their so-called long-term viscosity is then simply equivalent to the Kelvin-Voigt viscosity defined in this study. As the data used here relate mainly to deformation over long timescales, we did not, for simplicity, introduce elements in our rheology describing deformation on timescales shorter than a year.

All tests reported here use only one specific set of parameter values for the proposed Burgers rheology (Fig. 1). However, our numerical experiments, together with previous studies, allow us to estimate plausible ranges for each of these parameters.

Kelvin-Voigt viscosity: The Kelvin-Voigt viscosity η_K is set here to 3×10^{18} Pa.s. This value is deduced from the one used in (Klein et al., 2016) for the case without afterslip. A value of 4.5×10^{18} Pa.s is preferred when slip on the interface is taken into account. Note that even larger values could be found if thinner lithospheres or thicker asthenospheres are hypothesised (Trubienko et al., 2014). For the same μ_K , a larger Kelvin viscosity would simply increase the duration of the transient creep phase without changing its amplitude. For example, the transition from the 'fast' to the 'slow' postseismic regime would occur around 23 yrs instead of 15 yrs, for a viscosity of 4.5×10^{18} Pa.s instead of 3×10^{18} Pa.s.

Maxwell viscosity: The Maxwell viscosity η_M is set here to 3×10^{19} Pa.s, nearly the maximum possible value for a 'steady-state' convective asthenosphere. Indeed, convective models predict that, with larger viscosities, plates in old oceans and non-cratonic continental regions would thicken to depths exceeding 100 km (Dumoulin et al., 1999). On the other hand, with a long-term viscosity smaller than 3×10^{19} Pa.s, the postseismic velocity from the Valdivia earthquake and the interseismic velocity in the far field would both increase, which would spoil the fit to the data.

The pioneering studies of long-term postseismic deformation induced by earthquakes date from the sixties. These studies, based on near-field and mid-field data, used an elastic model to describe the interseismic part of the deformation related to coupling on the interface. They used a Maxwell viscosity in the asthenosphere to model the postseismic component of the Alaska and Valdivia earthquakes. Although their modelling of the interseismic component might be improved, these studies agree with the 'high' viscosity values necessary to describe postseismic deformation several decades after an earthquake. In Alaska, the deformation following the 2002 Denali earthquake provides an estimate of the effective viscosity in the same region as the 1964 subduction earthquake, on a time-scale of a few years. This effective viscosity, for a time-scale of a few years, is found to be of the order of 3×10^{18} Pa.s (Freed et al., 2006). This value is about one order of magnitude below the effective viscosity for decadal time-scales, in very good agreement with the present study.

Ratio of elastic moduli: The large value of μ_M/μ_K proposed here (i.e., 5) is rather unusual (μ_M : Maxwell shear modulus, μ_K : Kelvin-Voigt shear modulus). In many postseismic studies, this ratio is instead set to 1 (Pollitz, 2003; Pollitz et al., 2006; Hu et al., 2016). The value of 5 used here lies within the range of experimental values found for olivine or dunite assemblages. Chopra (1997) proposes a μ_M/μ_K between 3 and 20. Ohuchi et al. (2024) do not provide μ_M/μ_K directly, but they interpolate their experimental values to the conditions of postseismic deformation. They plot estimates of effective viscosity as a function of time after an earthquake. When their effective viscosity for annual time-scales equals 3×10^{18} Pa.s, they propose a transition to a long-term viscosity (with values from 3×10^{19} to 10^{20} Pa.s) over a time-scale between 20 and 1000 years. This transition implies a μ_M/μ_K larger than 6. Thus, the large values of μ_M/μ_K proposed in our study are compatible with at least some transient-creep mechanisms observed experimentally. However, it might be inappropriate to apply values measured in the laboratory using small samples to mantle conditions without corrections. Using creep laws measured in the laboratory, one may be able to rescale experimental values to account for differences in temperature, stress, and strain-rate between laboratory experiments and what occurs in the asthenosphere during the seismic cycle. However, there could be other sources of transient behaviour in the asthenosphere, linked to heterogeneities at scales from a metre to 100 km,

that are ignored in laboratory experiments. In addition to the various mechanisms creating transients in small laboratory samples (e.g., Chopra, 1997; Wallis et al., 2021; Breithaupt et al., 2023; Ohuchi et al., 2024), some transient creep necessarily results from rheological heterogeneities linked to temperature, variations in water content, and depletion (Ivins and Sammis, 1996). This is why 'natural experiments', i.e., the measurement of large-scale deformation of the Earth after a stress perturbation, are particularly useful. Our study can help to put bounds on the value of μ_M/μ_K appropriate for large-scale deformation. If μ_M/μ_K were larger than 15, the postseismic deformation of the Valdivia earthquake would not yet have reached the slow regime dominated by the long-term viscosity in 2010. If μ_M/μ_K were smaller than 5 for a viscosity η_K of 3×10^{18} Pa.s, or smaller than 3.4 for a viscosity of 4.5×10^{18} Pa.s, the stations in the far field of Maule would currently be in the 'slow regime'. This does not appear to be the case yet. We propose a value of μ_M/μ_K between 3.4 and 15.

8 CONCLUSION

This work clarifies the differences between four main assumptions commonly used to model intermittent slip on the subduction interface: *elastic*, *steady*, *steady + last EQ*, and *cyclic*. We show that near-field deformation is comparatively consistent across all four approaches. In contrast, significant differences arise in the mid- and far-field, where predicted velocities are highly model-dependent. In particular, the *steady* and *steady + last EQ* models systematically overestimate continentward velocities compared to both the *elastic* and *cyclic* models. The *elastic* model, on the other hand, tends to underestimate these velocities toward the end of the seismic cycle, although it may not predict the trenchward velocities predicted by the viscoelastic models when a major earthquake has occurred within the past few decades. Overall, a *cyclic* model provides the most consistent representation of interseismic deformation at the scale of the entire continent, and it captures the temporal evolution of the deformation throughout the seismic cycle.

Our results, supported in particular by observations of postseismic deformation following the Valdivia earthquake, favour a Burgers rheology with a long-term viscosity on the order of a few 10^{19} Pa.s. In this framework, present-day surface deformation remains sensitive to slip events that occurred far in the past, although not to their details. Consequently, steady models are not appropriate for modelling interseismic deformation, as they neglect the cumulative effects of previous earthquakes. More generally, our findings highlight the need for a strong contrast between effective viscosities operating over short (years) and long (decades to centuries) timescales. This supports the use of an Extended Burgers rheology, including, in addition to the classical short-term (months) Kelvin-Voigt element, one or more additional Kelvin-Voigt elements with large μ_K . Further work is needed to refine these models, including improved constraints on deformation during the ~ 15 years following the Maule earthquake, better characterisation of the Valdivia earthquake source, and inversions of interplate coupling consistent with the proposed rheological framework.

Accurately capturing time-dependent mantle rheology is essential not only for understanding stress evolution and seismic hazards in subduction zones, but also for quantifying the solid Earth response to surface loading, for example, in the context of climate change. In particular, variations in hydrological and cryospheric loads over recent decades to a century (Adhikari et al., 2021; Ivins et al., 2023; Rousselet et al., 2025), as well as glacial isostatic adjustment following deglaciation (Caron et al., 2017), are highly sensitive to the presence of an early, rapid phase of viscous deformation. Thus, the results of this paper have important implications for modelling the seismic cycle, understanding the stress evolution and assessing seismic hazard in earthquake-prone regions, and examining intraplate deformation, as well as for the rheology of the asthenosphere.

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DATA AVAILABILITY

GNSS time-series are publicly available in (Klein et al., 2022) and (Blewitt et al., 2018). The time

series for station MPLA come from the SPOTGINS solution, distributed via the Geodesy Plotter (<https://www.poleterresolide.fr/geodesy-plotter/>).

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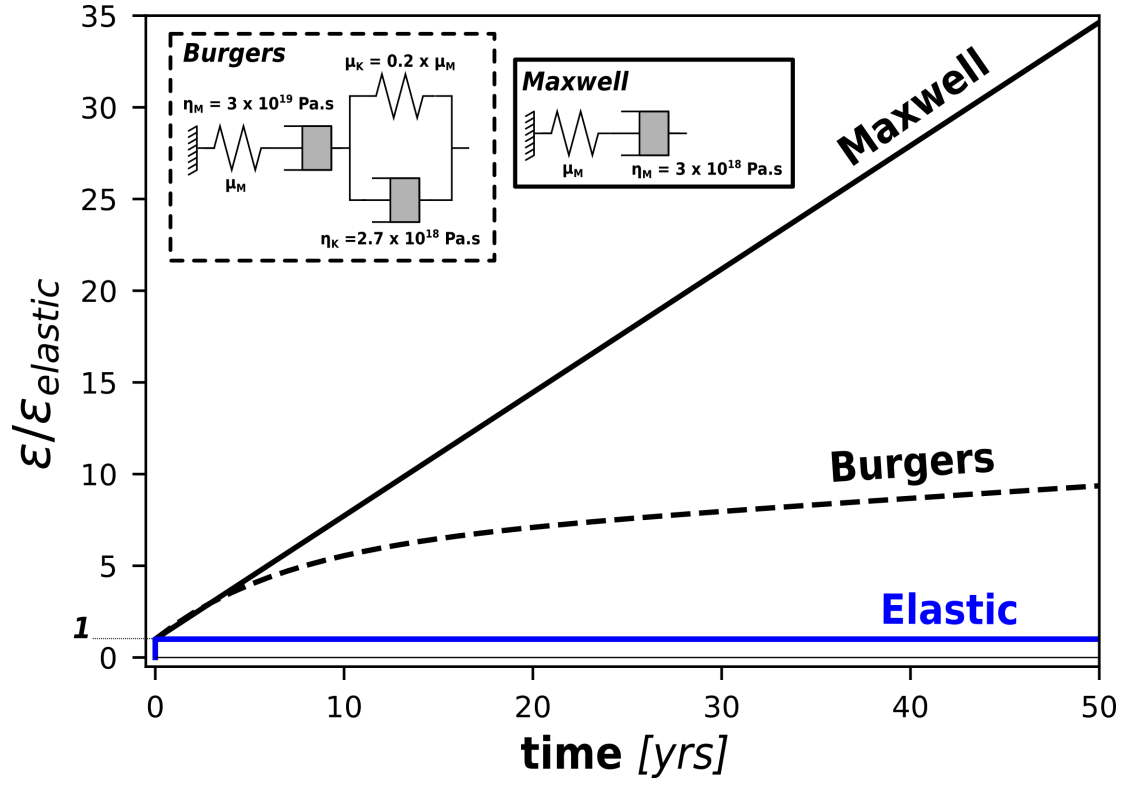


Figure 1. Evolution of the strain normalised by the elastic strain through time for elastic (blue line), Maxwell (black solid line), and Burgers (dashed black line) rheologies, in response to an instantaneous stress. For this calculation, $\mu_M = 64 \text{ GPa}$.

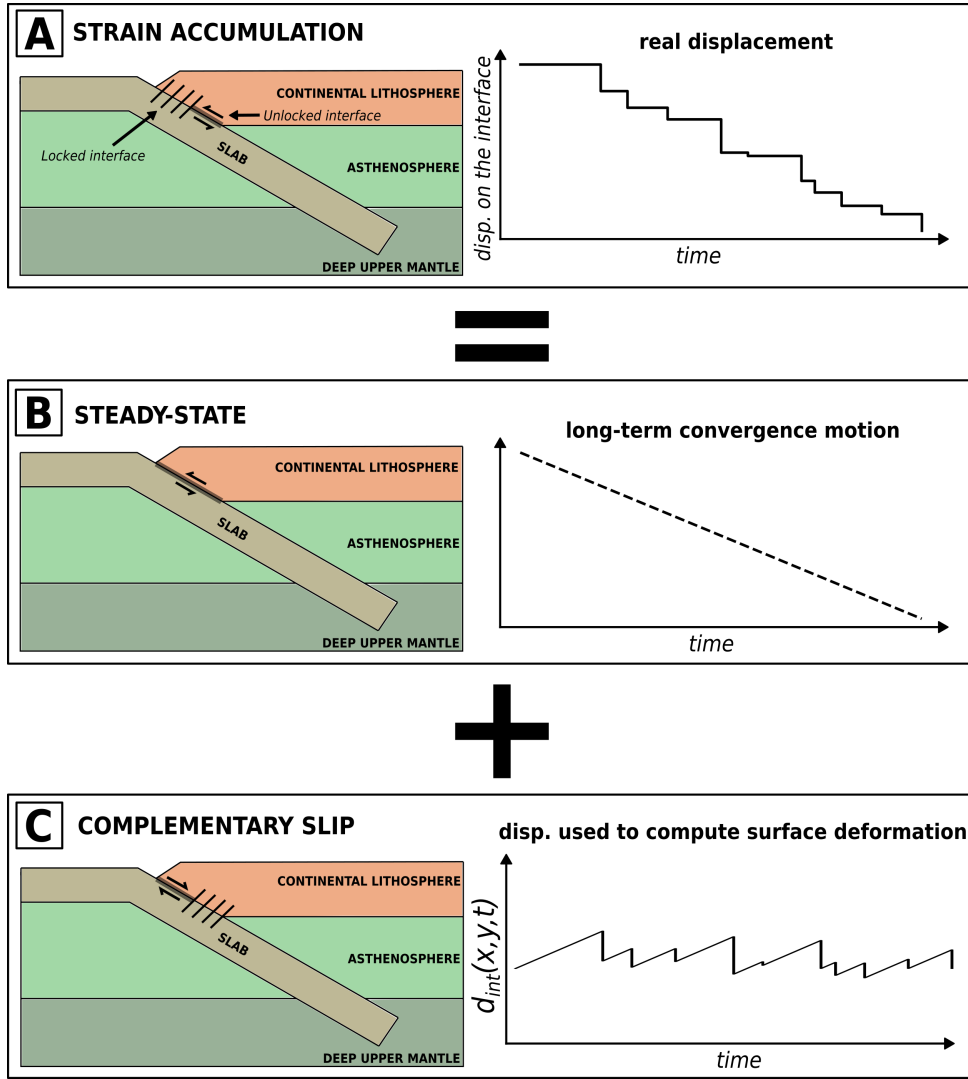


Figure 2. The backslip principle of Savage (1983), used to model interseismic deformation. Actual slip on the subduction interface (**A**) is decomposed into two components: (**B**) a uniform slip at the long-term plate convergence velocity, which produces no strain in the overriding plate; and (**C**) a complementary slip that captures the strong temporal variations associated with the seismic cycle (typically $\sim 10^2$ yrs). Over long timescales, this complementary slip has zero net displacement but drives deformation of the overriding and/or subducting plate.

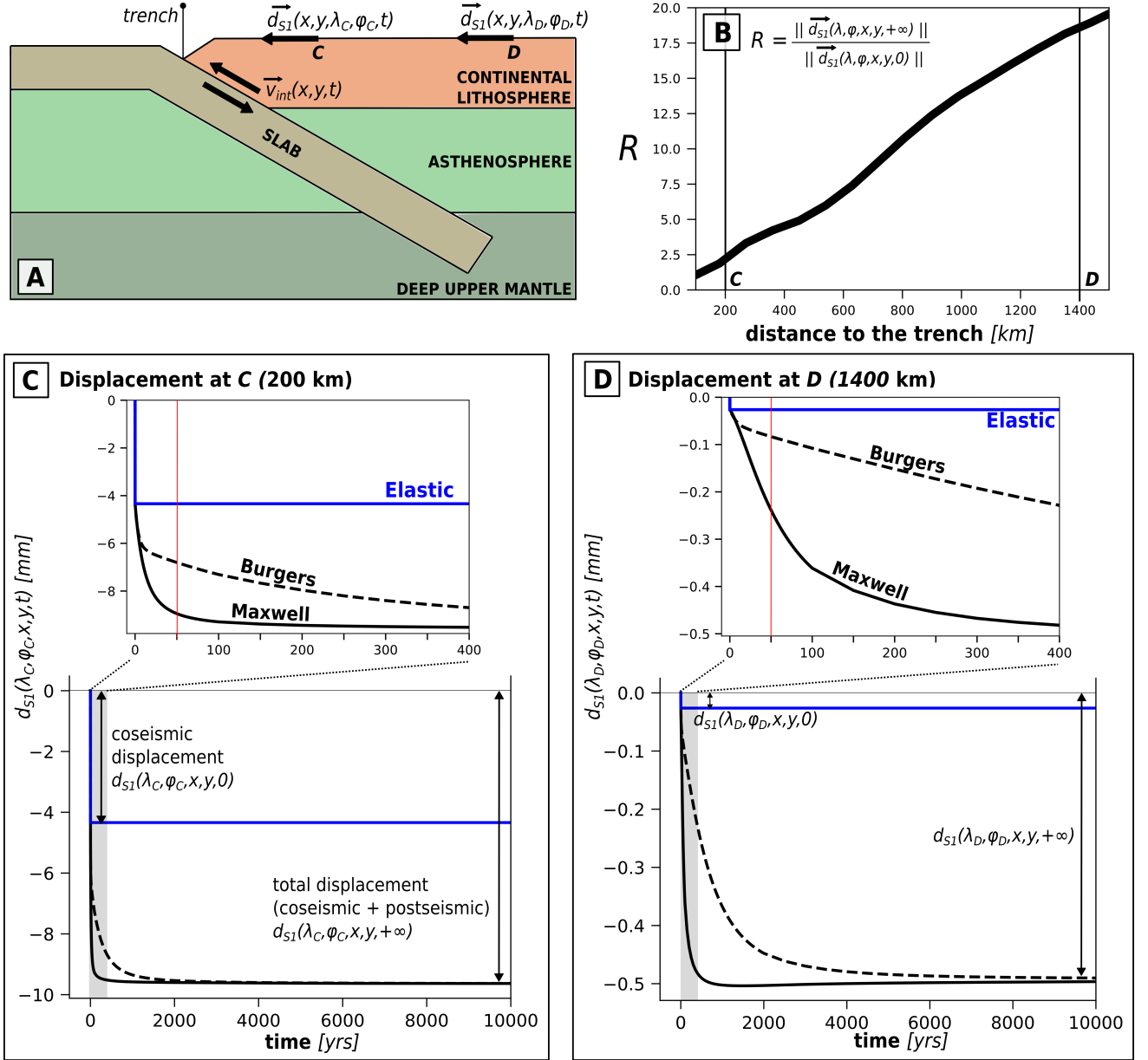


Figure 3. (A) Model setup. $v_{int}(x, y, t)$ is the velocity of a point on the subduction interface, defined as the time derivative of $d_{int}(x, y, t)$. $\vec{d}_{s1}(\lambda_C, \varphi_C, x, y, t)$ and $\vec{d}_{s1}(\lambda_D, \varphi_D, x, y, t)$ are the surface displacements at points C (λ_C, φ_C) and D (λ_D, φ_D), located 200 km and 1400 km from the trench, respectively, induced by an elementary slip $d_{int}(x, y, t)$ that occurred t years ago. (B) Ratio R of the long-term viscoelastic response (cumulative displacement at 10 000 yrs) to the elastic response (coseismic displacement), as a function of distance from the trench. This ratio is identical for Maxwell and Burgers rheologies. (C, D) Surface displacement over 10 000 yrs induced by a 1 m slip (parallel to convergence) on a node at 21 km depth, for three rheologies: elastic (blue line), Maxwell (solid black line), and Burgers (dashed black line). Displacement is shown at 200 km (C) and 1400 km (D) from the trench. The bottom plot shows a zoom on the first 400 yrs (gray zone in the main plot). The red dashed line marks $t = 50$ yrs.

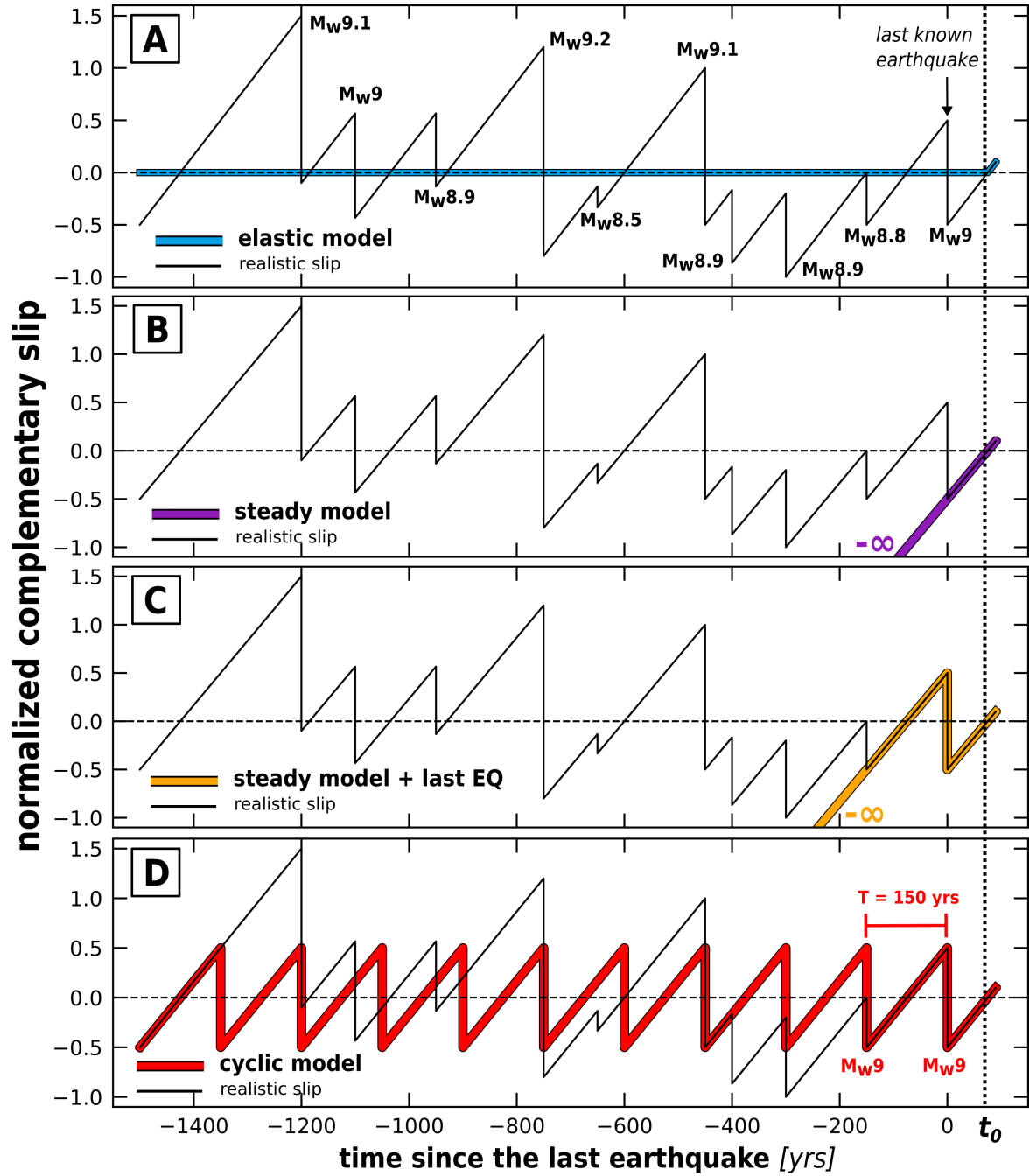


Figure 4. Slip histories on the plate interface (corrected for long-term convergence) assumed for the complementary slip by the four studied models: (A) *elastic*, (B) *steady*, (C) *steady + last EQ*, and (D) *cyclic*. The cyclic model features recurring $M_w 9$ earthquakes with a recurrence interval of $T = 150$ yrs, approximating the more realistic irregular slip history (black solid lines) with variable magnitudes and recurrence intervals. All models share the same present-day slip rate at t_0 .

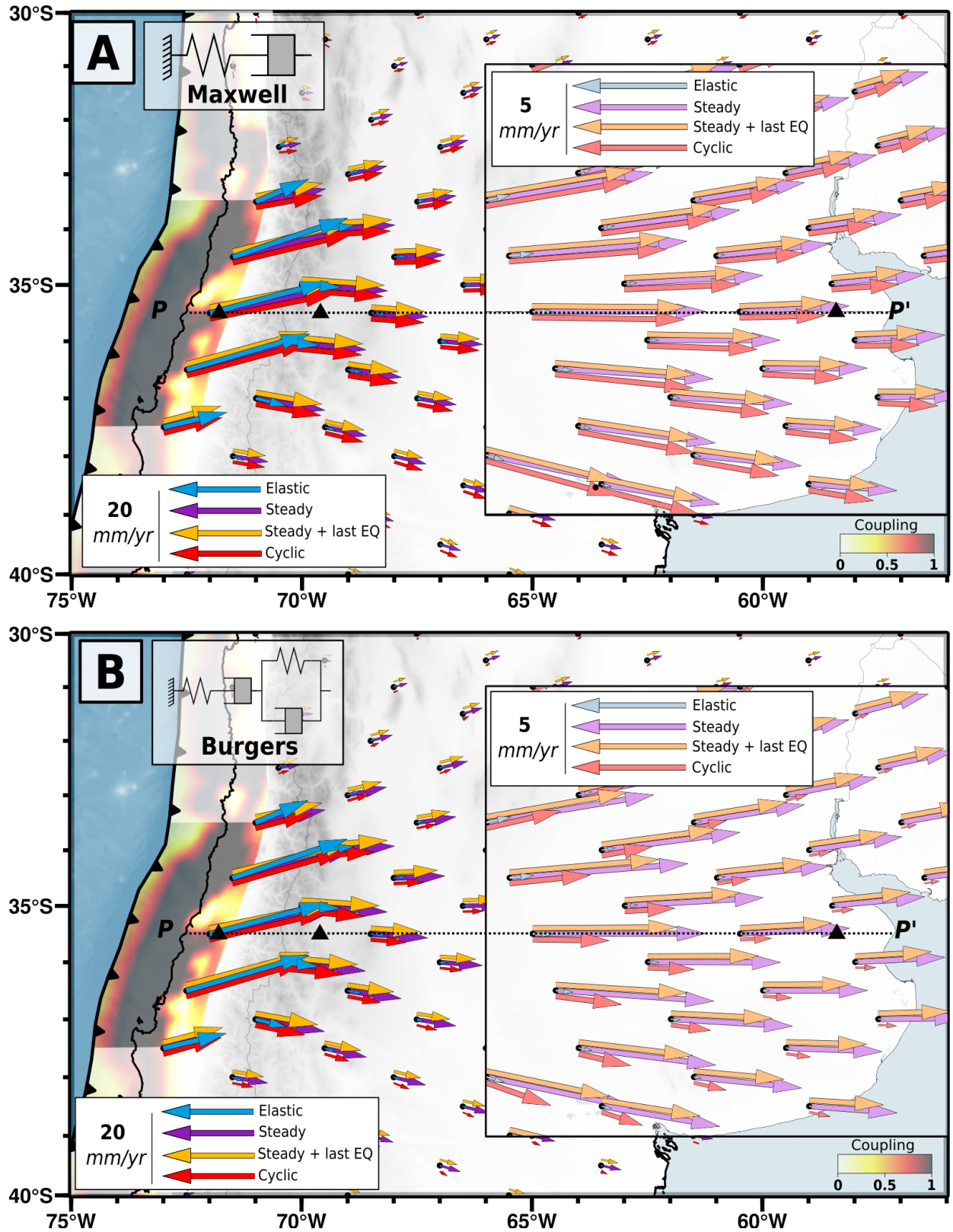


Figure 5. Predicted interseismic surface velocities for the Maule segment, expressed in a South American reference frame. All models use identical present-day interplate coupling. Three slip models are shown: *elastic* (blue), *steady* (purple), *steady + last EQ* (orange) and *cyclic* (red). (A) Maxwell rheology. (B) Burgers rheology. Velocities correspond to the end of the 150 yrs interseismic cycle. Note the change of vector scale inside the black square, lighter colors indicate the smaller scale.

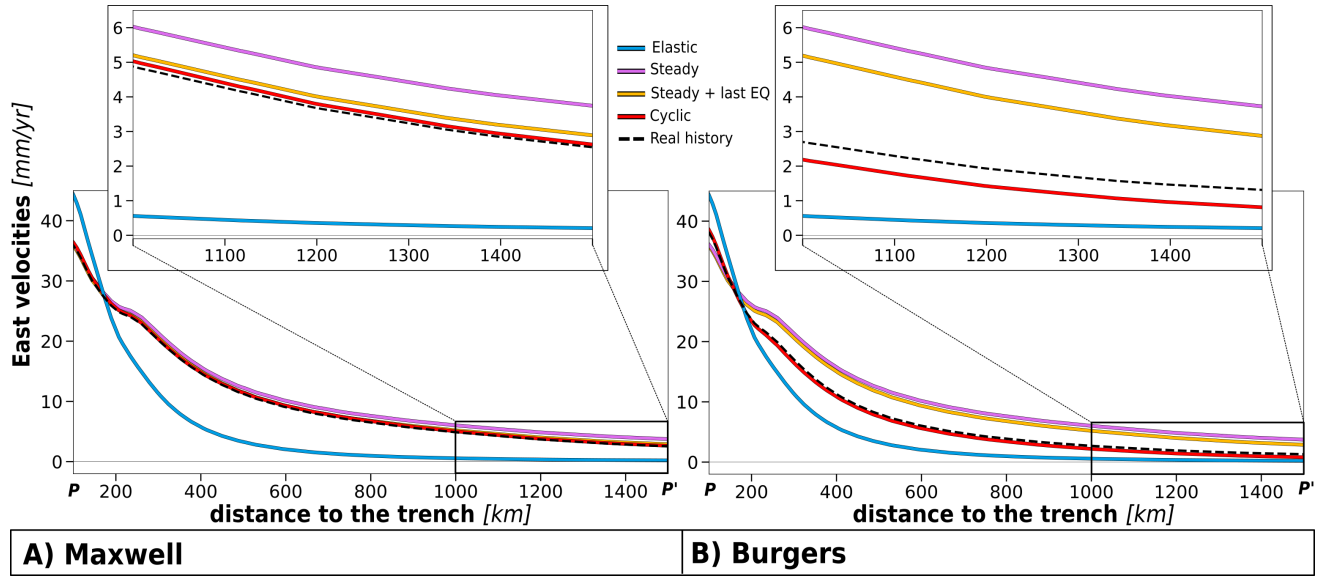


Figure 6. East-component interseismic velocity profiles along P-P' (see Fig. 5 for location). Results are shown for five slip models: *elastic* (blue), *steady* (purple), *steady + last EQ* (orange), *cyclic* (red), and *realistic* irregular recurrence (dashed, from Fig. 4). (A) Maxwell rheology. (B) Burgers rheology. The right panels show a zoom from 1000 to 1500 km from the trench.

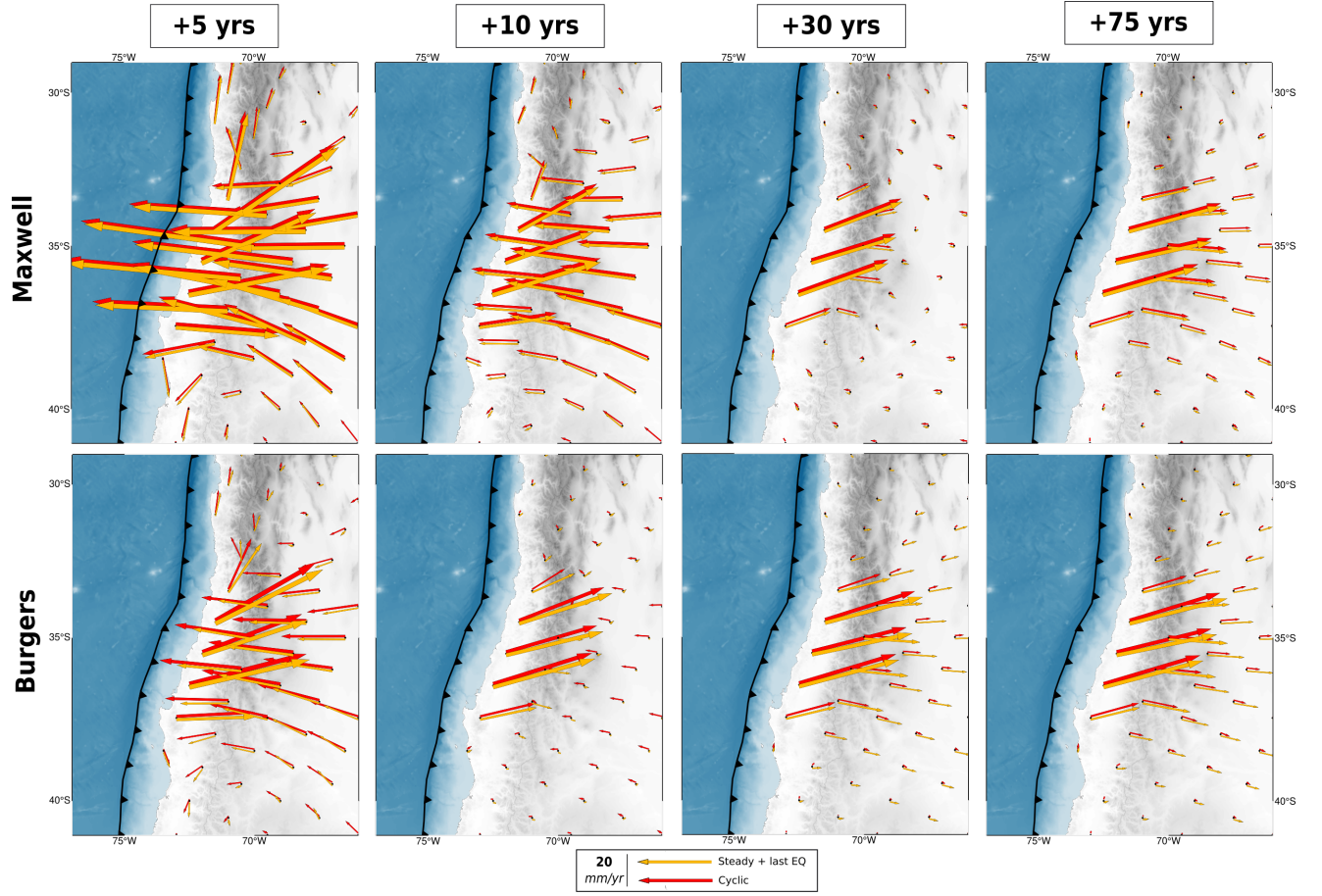


Figure 7. Evolution of near- to mid-field surface velocity following an earthquake, for the *cyclic* ($M_w 9$, $T = 150$ yrs) (red vectors) and *steady + last EQ* (orange vectors) models. **Top panels:** Maxwell rheology. **Bottom panels:** Burgers rheology. Coupling and slip histories are shown in Figs. 4 and 5. Velocities are expressed in a South American reference frame. Time elapsed since the last earthquake is shown at the top of each column.

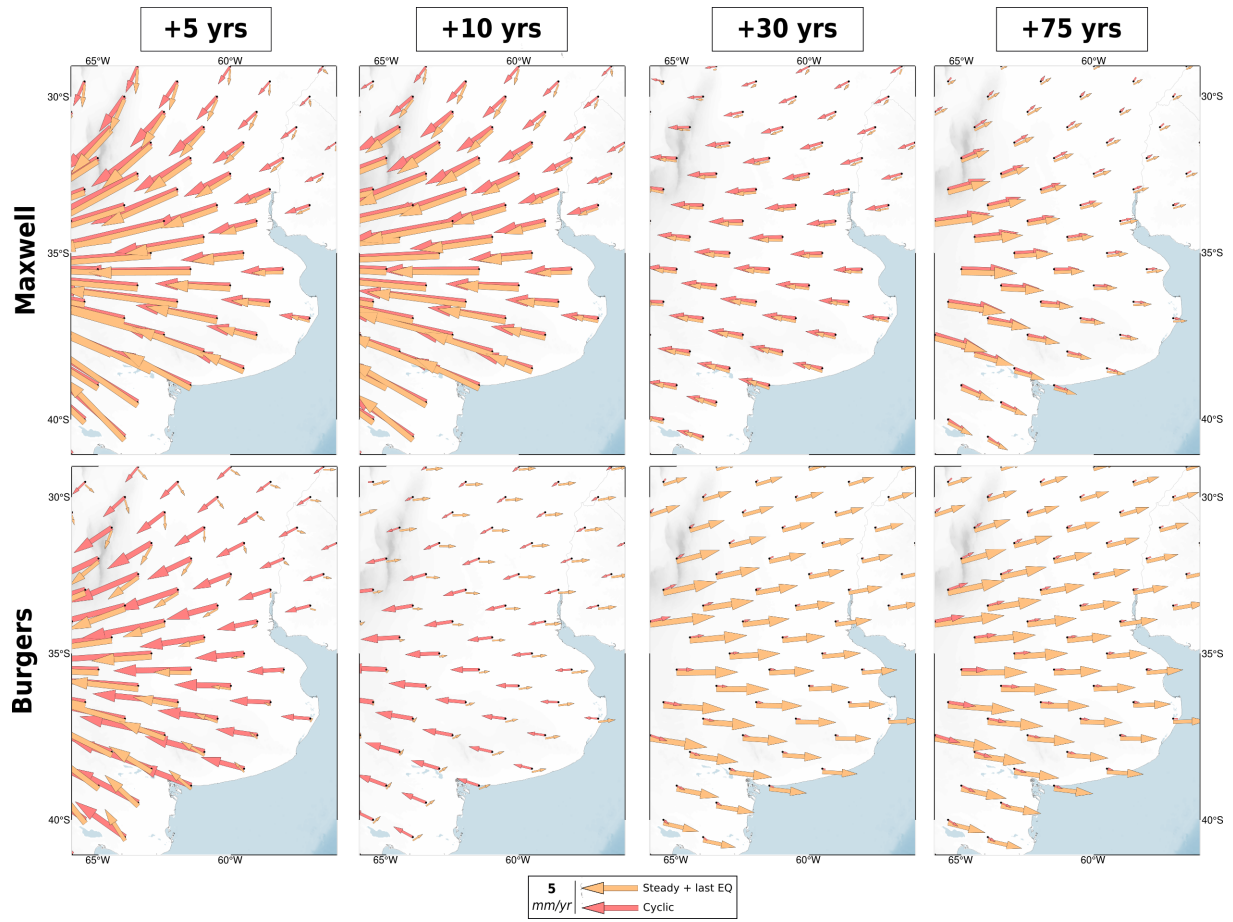


Figure 8. Same as Fig. 7 but for far-field surface velocities.

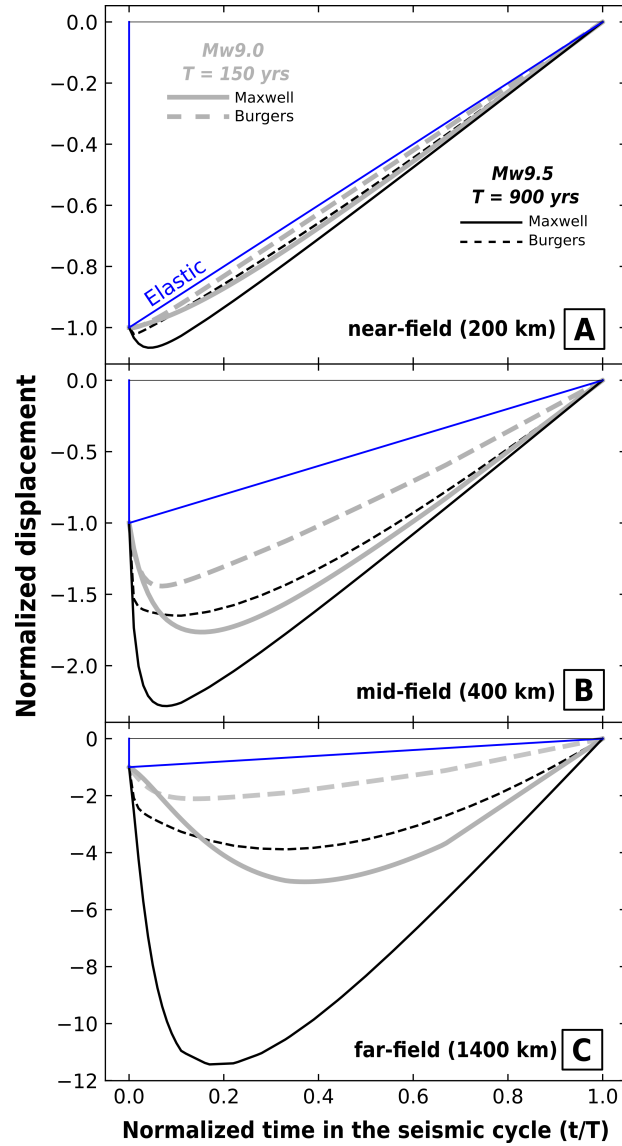


Figure 9. East-West displacement (normalised by the coseismic offset) through the seismic cycle, as a function of time normalised by the duration T of the cycle, for two *cyclic* models with the same coupling (Fig. 5) but different cycle periods: $T = 150$ yrs with $M_w 9$, and $T = 900$ yrs with $M_w 9.5$ (six times longer period and moment). Normalised displacements are shown at three locations: 200 km (A), 400 km (B), and 1400 km (C) from the trench (depicted by the triangles on Fig. 5A-B). Three rheologies are used: elastic (blue lines), Maxwell (solid lines), and Burgers (dashed lines).

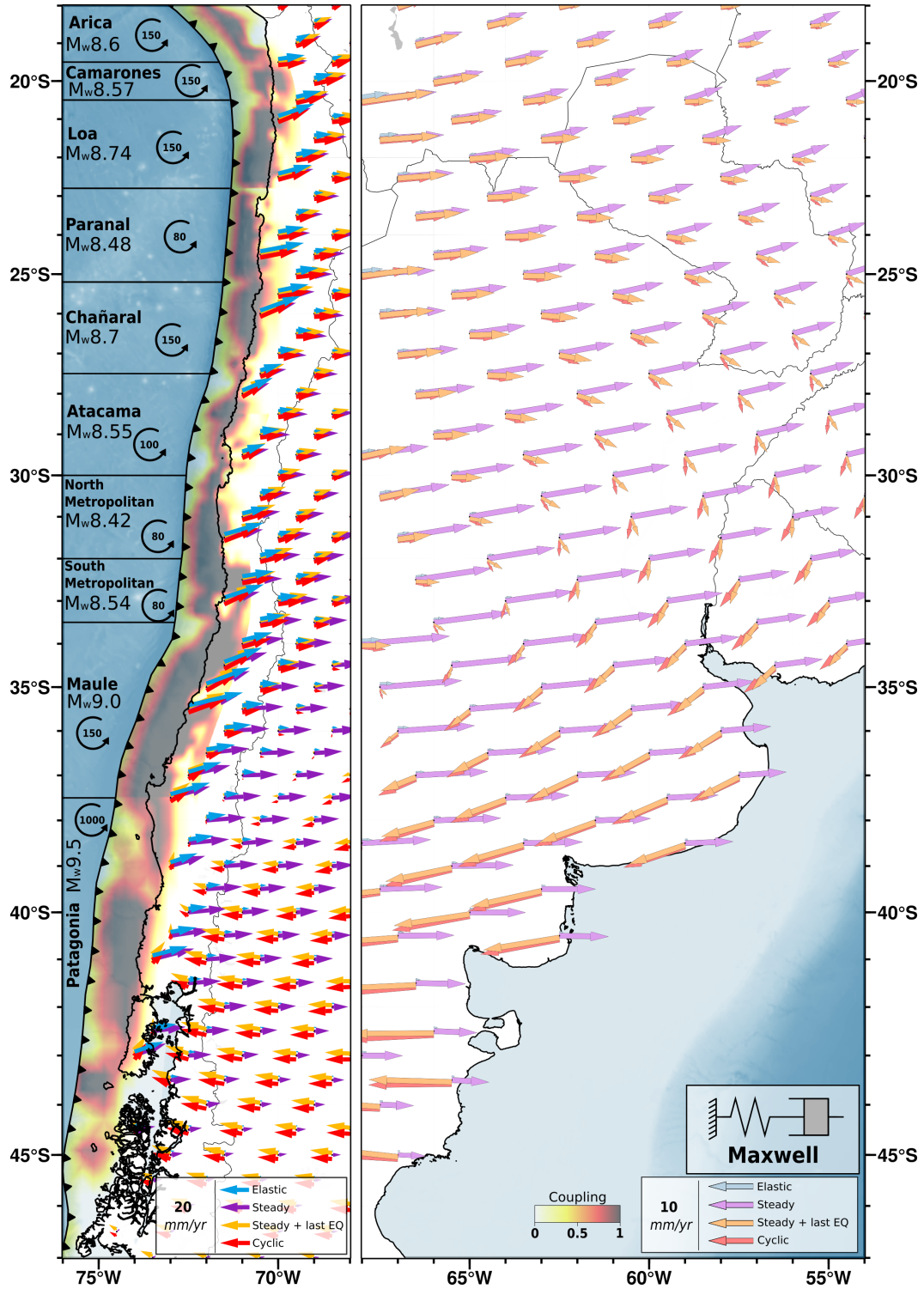


Figure 10. Pre-2010 continental-scale surface velocities predicted by the *elastic* (blue), *steady* (purple), *steady + last EQ* (orange), and *cyclic* (red) models, using the Maxwell rheology. Velocities are expressed in a South American reference frame. Note the change of vector scale between the two panels.

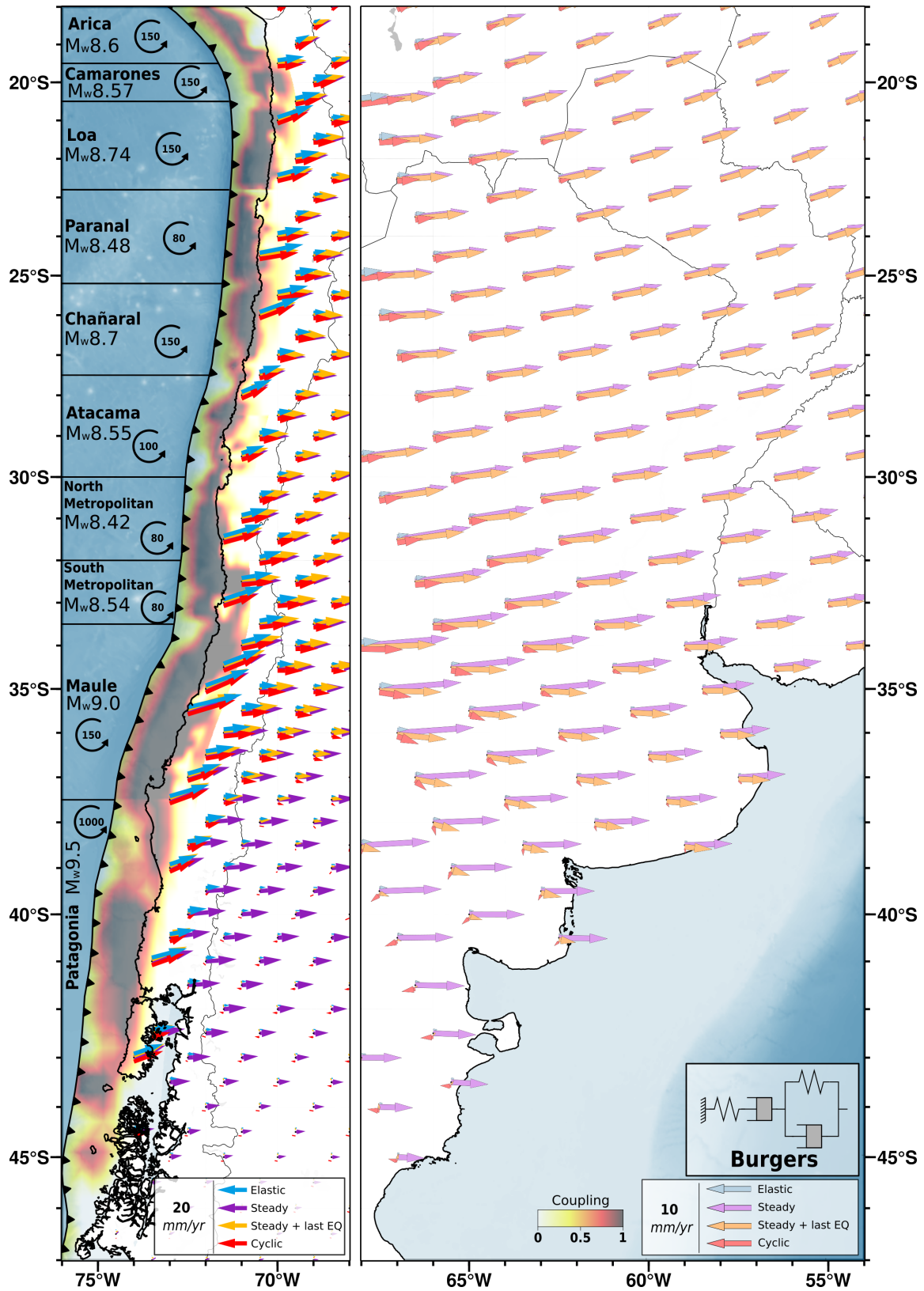


Figure 11. Same as Fig. 10, but using the Burgers rheology.

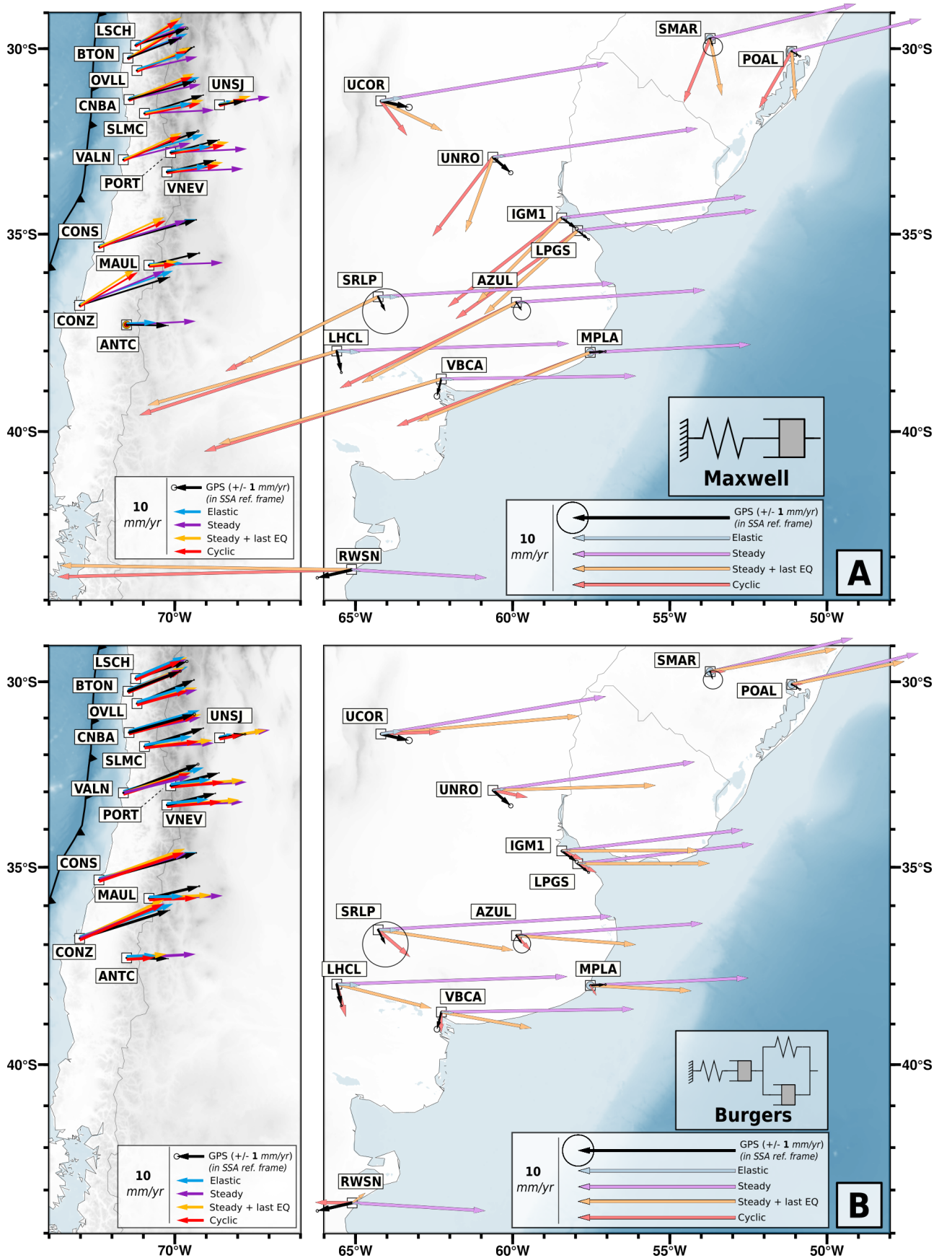


Figure 12. (A) Pre-2010 GPS velocities (black) compared to predictions from the *elastic* (blue), *steady* (purple), *steady + last EQ* (orange), and *cyclic* (red) models, using the Maxwell rheology. (B) Using the Burgers rheology. Velocities are expressed in a South American reference frame. Continuous GPS station locations are shown as white squares. Filled circles indicate stations where the corresponding model velocity amplitude is below 2 mm/yr (left panels) and 0.5 mm/yr (right panels); concentric circles denote multiple models below the threshold at the same site, ordered by decreasing amplitude from the largest (outer) to the smallest (inner) radius. Note the change of vector scale between panels.

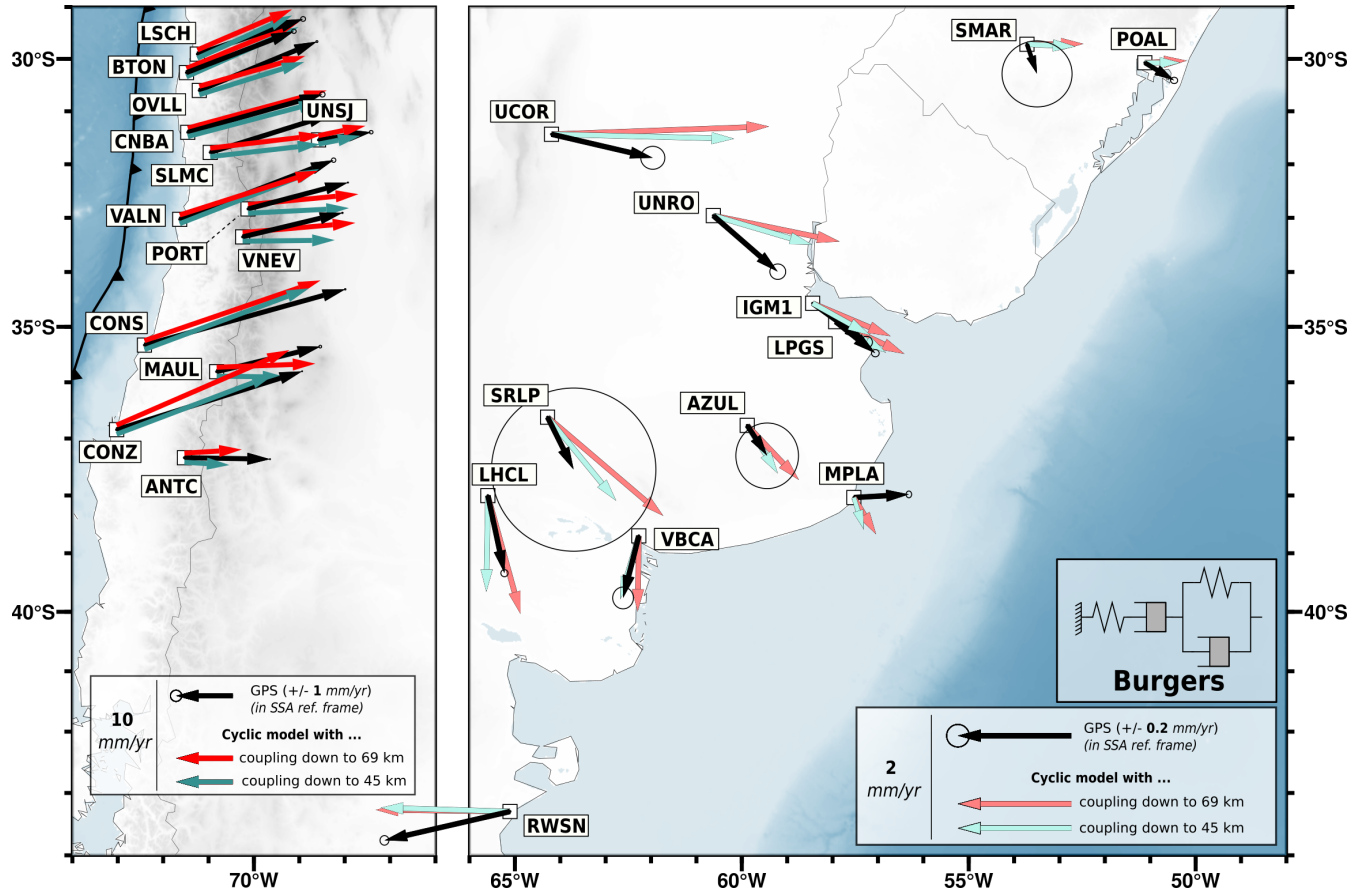


Figure 13. Effect of maximum coupling depth on predicted surface velocities. Pre-2010 GPS velocities (black) are compared to cyclic model predictions with coupling extending to 45 km (blue) and 69 km (red) depth, using the Burgers rheology. Velocities are expressed in a South American reference frame. Continuous GPS station locations are shown as white squares. Note the change of vector scale between panels.

Asthenosphere rheology and the continental-scale interseismic deformation driven by repeated megathrust earthquakes: Insights from the South American plate

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Supporting information

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S1 Single node simulation

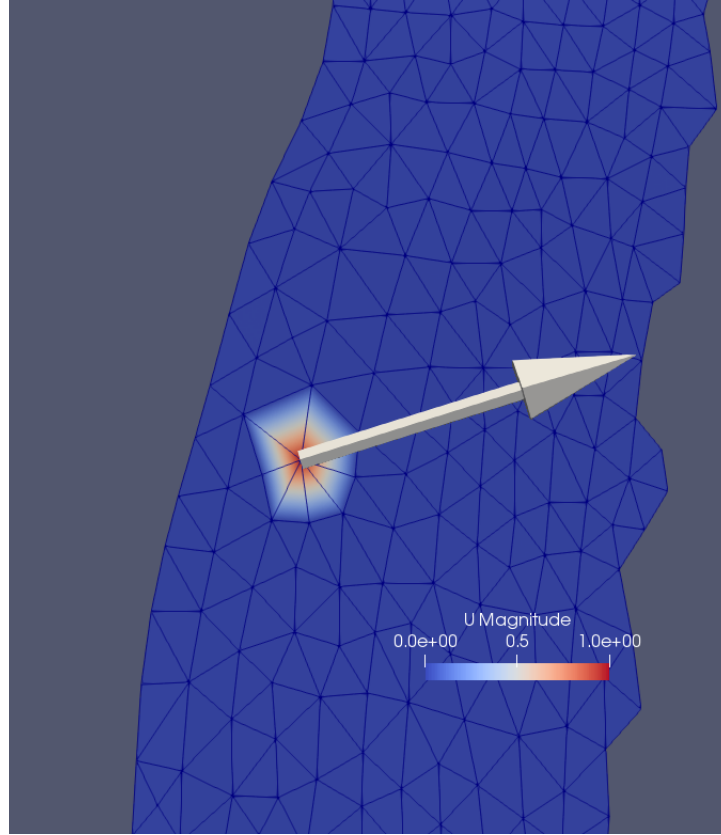


Figure S1: Node located on the fault plane of *Chile_Mesh_v1.0*, sliding 1 m in the convergence direction.

S2 Contribution of postseismic relaxation from ancient earthquakes and tips for a fast calculation for a cyclic model

$$\vec{v}_s(\lambda, \varphi, t_0) = -\left[\frac{\vec{D}_{Eq-s}(\lambda, \varphi, t_0 + N_A \cdot T)}{T}\right] + \sum_{k=0}^{N_B} \vec{V}_{Eq-s}(\lambda, \varphi, t_0 + k \cdot T) \quad (\text{Eq. S1})$$

The surface velocity given by the *cyclic* model is calculated as the difference between the *steady* model over N_A seismic cycles $\left[\frac{\vec{D}_{Eq-s}(\lambda, \varphi, t_0 + N_A \cdot T)}{T}\right]$ and the viscoelastic relaxation of N_B past earthquakes $\sum_{k=0}^{N_B} \vec{V}_{Eq-s}(\lambda, \varphi, t_0 + k \cdot T)$ (Fig. S2 and Eq. S1). Two numerical tests are used to determine the impact of N_A and N_B :

- Test 1: Should N_A and N_B be equal to each other?
- Test 2: What should be the value of $N_A = N_B$?

For both tests, we calculate the displacement of a point on the surface of the upper plate, located in Argentina, in the far field (~ 1400 km from the subduction trench).

Test 1 (Fig. S2). If no earthquake is taken into account ($N_B = 0$), the displacement of the point is, logically, exactly equal to that of the *steady* model. As N_B increases and approaches $N_A = 10$, the point displacement decreases over a given time. Only the model with $N_B = N_A = 10$ returns the point to its initial position (zero displacement) at the end of the seismic cycle, $t = 150$ yrs. All models with $N_A \neq N_B$ therefore fail to reach a steady state over time, resulting in a non-zero long-term displacement. It is therefore imperative that $N_A = N_B$.

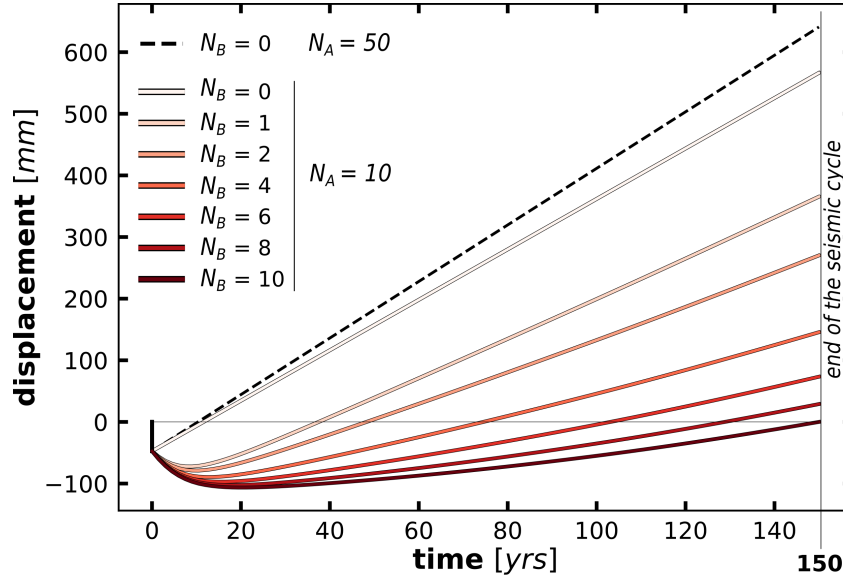


Figure S2: Illustration of **Test 1**. Contribution of N_B past earthquakes to the East displacement of a point at 1400 km from the trench, at 35.5°S , facing the Maule segment (black triangle in Fig. 5 of the main text). The *steady* model is calculated over $N_A = 10$ seismic cycles, using the Burgers rheology. Equilibrium (return to 0 at the end of the cycle) is only approximately reached when the backslip is calculated over the same time as past earthquakes ($N_A = N_B$). The dashed black line shows the displacement calculated over 50 cycles for the *steady* model.

Test 2 (Fig. S3). The backslip term is known to be linear, and it is also known that, at convergence, the displacement returns to zero at the end of the cycle. The asymptotic displacement during a cycle can then be obtained by adding the contribution of each earthquake, corrected for the linear trend between the initial and final values, as described in (Trubienko *et al.*, 2013). Convergence toward the asymptotic value is reached after only a few cycles, with good convergence already achieved after 5 cycles. In the case considered here, with the Burgers rheology, pursuing the computation of $\vec{D}_{Eq-s}(\lambda, \varphi, t)$ over 800 yrs is sufficient, unless a precision on the velocity below 0.1 mm/yr is required.

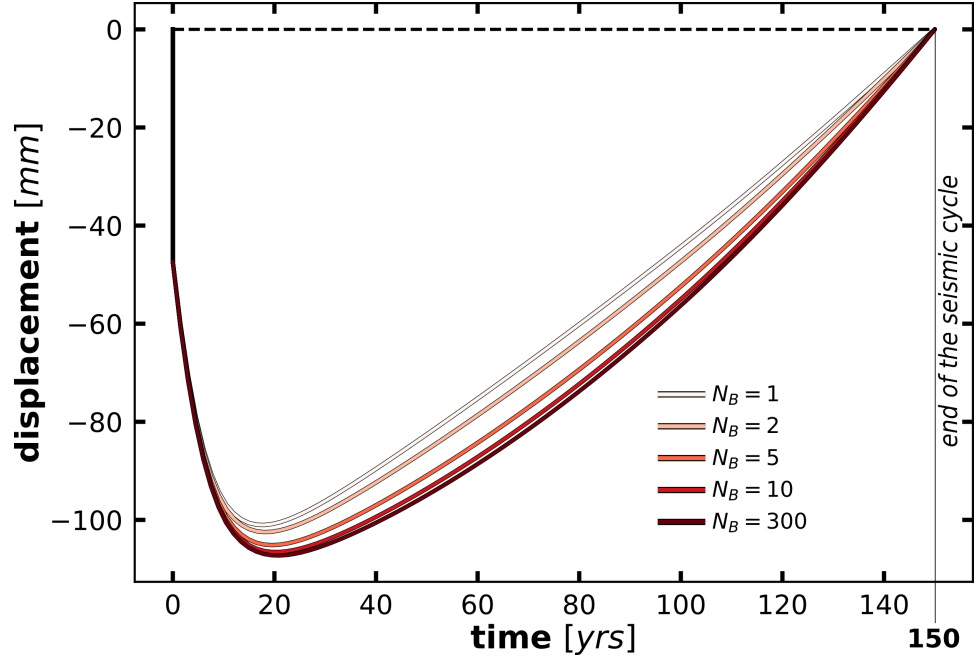


Figure S3: Illustration of **Test 2**. East time-series of displacement through the seismic cycle for a point at the surface of the overriding plate (black triangle in Fig. 5 of the main text), given $N_A = N_B$ past earthquakes taken into account in the computation of the *cyclic* model. N_A is the number of seismic cycles used to compute the backslip viscoelastic term of the *cyclic* model, and is fixed to $N_A = 10$. The return time of the earthquake is here set to $T = 150$ yrs.

S3 Impact of the period duration of the seismic cycle on the interseismic deformation

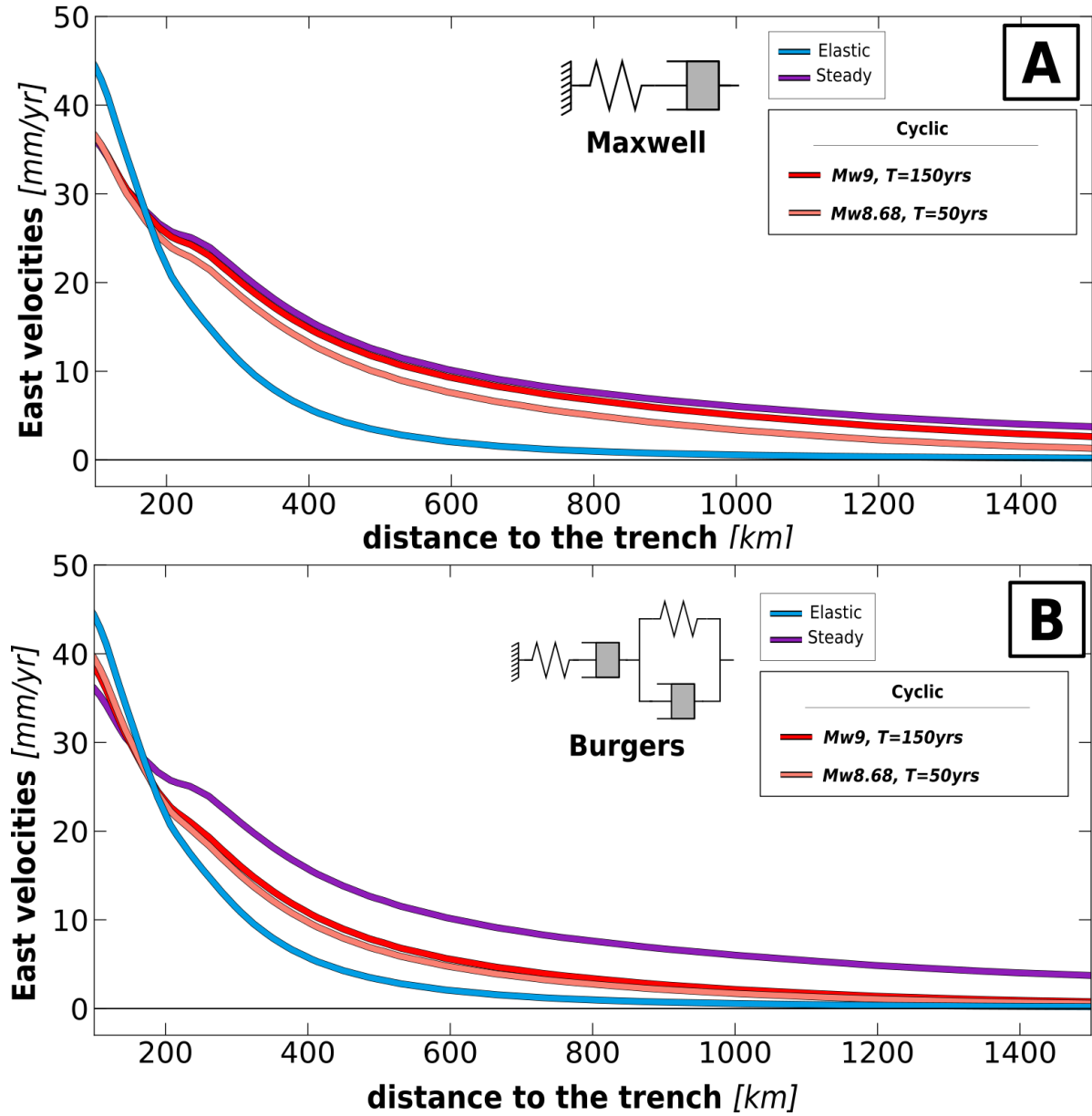


Figure S4: Comparison of two cycle parameters for the Maxwell (A) and Burgers (B) rheologies: for the red curve, a $M_w 9$ earthquake occurs every $T = 150$ yrs, for the light red curve, a $M_w 8.68$ earthquake occurs every $T = 50$ yrs.

S4 Non cyclic slip histories

While the seismicity of a few subduction zones is characterised by periodic earthquakes of similar magnitude, the time lapse between major earthquakes is usually rather variable, and there is a large dispersion in earthquake magnitude. Also, assuming the same coupling over several centuries may not be appropriate.

S4.1 About the impact on surface velocity of 'compensated' interface slips and some simple examples of slip histories: a graphical illustration

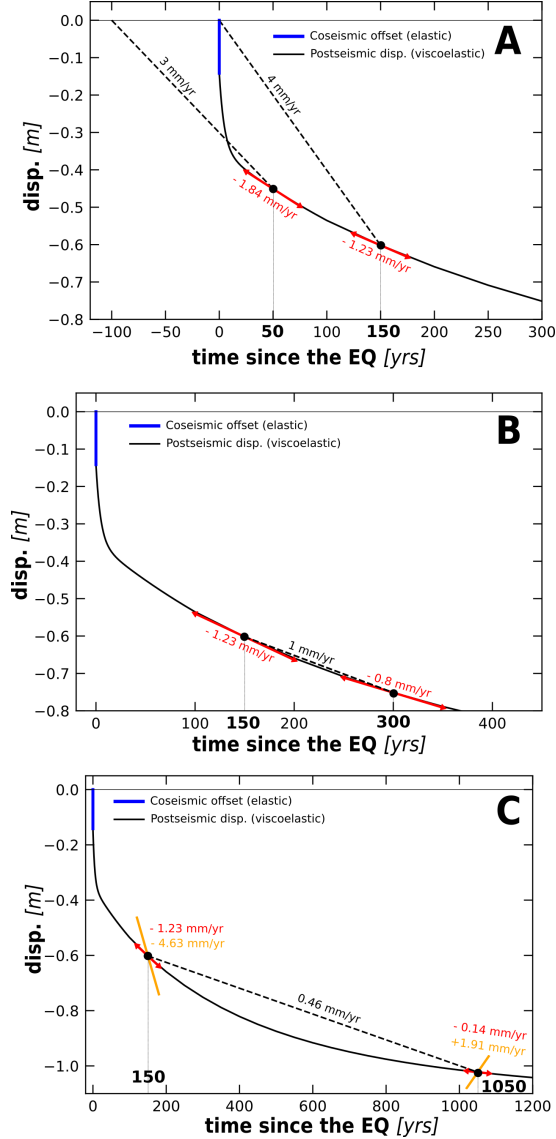


Figure S5: Full black line: Displacement through time for a point located at 800 km from the trench, facing the Maule segment, $\vec{D}_{Eq-s}(\lambda, \varphi, t)$, induced by an earthquake with a moment corresponding to the coupling distribution on the Maule segment, sustained for 150 years (see Fig. 5 of the main paper). The blue line indicates the coseismic displacement. See the paragraph below the figure for the other elements. The backslip coupling distribution is the same for all figures, and corresponds to the coupling distribution on the Maule segment plotted in Fig. 5 (main text). The slopes of the black dashed lines indicate the velocities induced by this backslip and are derived from Eq. 8. The slopes of the red lines represent velocities due to the postseismic effects of the earthquake, i.e. the derivative of the displacement with respect to time. Computed values of the velocities are given next to the curves. The yellow lines and velocities in case C indicate the velocities 150 yrs or 1050 yrs after the earthquake.

Through simple examples, we wish to illustrate the surface velocities induced independently by simple elements of the slip history.

A. Coupling between 0 and 50 years, and between 0 and 150 years. An earthquake with a slip corresponding to the backslip velocity times 150 years occurred either 50 or 150 years ago. The surface velocity due to coupling is much larger than the velocity due to the earthquake. The switch from trenchward to continentward occurs when the two slopes become equal, here in the region of transition between the low- and high-viscosity regimes.

B. Coupling between 150 and 300 years ago, with an earthquake either 150 or 300 years ago. The resulting velocity is very small and trenchward when the earthquake occurred 150 years ago, and continentward when it occurred 300 years ago. This example illustrates that, when the slip of the earthquake and that of the coupling compensate over a 'short' time, they induce small velocity perturbations.

C. The coupling is similar to the previous cases, but occurs between 150 and 1050 years ago, with the earthquake occurring either 150 or 1050 years ago. This earthquake has a moment six times larger than in case B, compensating for the coupling over 900 years. We chose to keep the same black line in Fig. S5-C as in cases A and B (i.e. the displacement generated by a magnitude 9 earthquake). According to Equation 8, the velocity generated by the backslip is equal to $(D(t_2) - D(t_1))/150$.

In Fig. S6, the slope of the black dashed line is equal to $(D(t_2) - D(t_1))/900$. So, according to Eq. 8, since the characteristic time T of the coupling is 150 yrs, it must be multiplied by 6 to give the velocity component generated by the backslip over 900 years. Similarly, the black curve corresponds to the displacement generated by a magnitude 9 earthquake, i.e., corresponding to the accumulation of backslip over 150 years. The velocity corresponding to this earthquake must be multiplied by 6, since the earthquake is 6 times larger in this case. The difference of slopes must therefore also be multiplied by 6 (yellow lines). In this case, the velocities generated by a slip compensated over 900 years are quite large.

Through these simple examples, we hope to have illustrated why a slip with zero average over past centuries may not affect present-day velocities, unless it occurred during past decades, or involves an accumulation of backslip over a long time, ≈ 1000 years.

S4.2 Some examples of irregular slip histories

Below, we present different slip histories on the subduction interface that are strongly perturbed compared to the *cyclic* model described in Section 4.4 of the main text. We show how these perturbations affect the predicted velocities. All cases have the same present-day coupling on the Maule segment (similar to the coupling used for Fig. 5 of the main text), and the Burgers rheology is used. For many of these histories (1, 2, 3, 4, 5, 7, 8), the velocities are computed 150 years after the last earthquake, which has a standard $M_w 9$ magnitude. The horizontal velocity as a function of distance to the trench is compared, for each new slip history, with the four 'standard' slip histories described in Section 4 (main text), i.e. *elastic*, *steady*, *steady + last EQ*, and *cyclic*. As described below, the numerically obtained velocity pattern for each of these histories can be understood by looking at Fig. 3, as in the previous subsection S5.1.

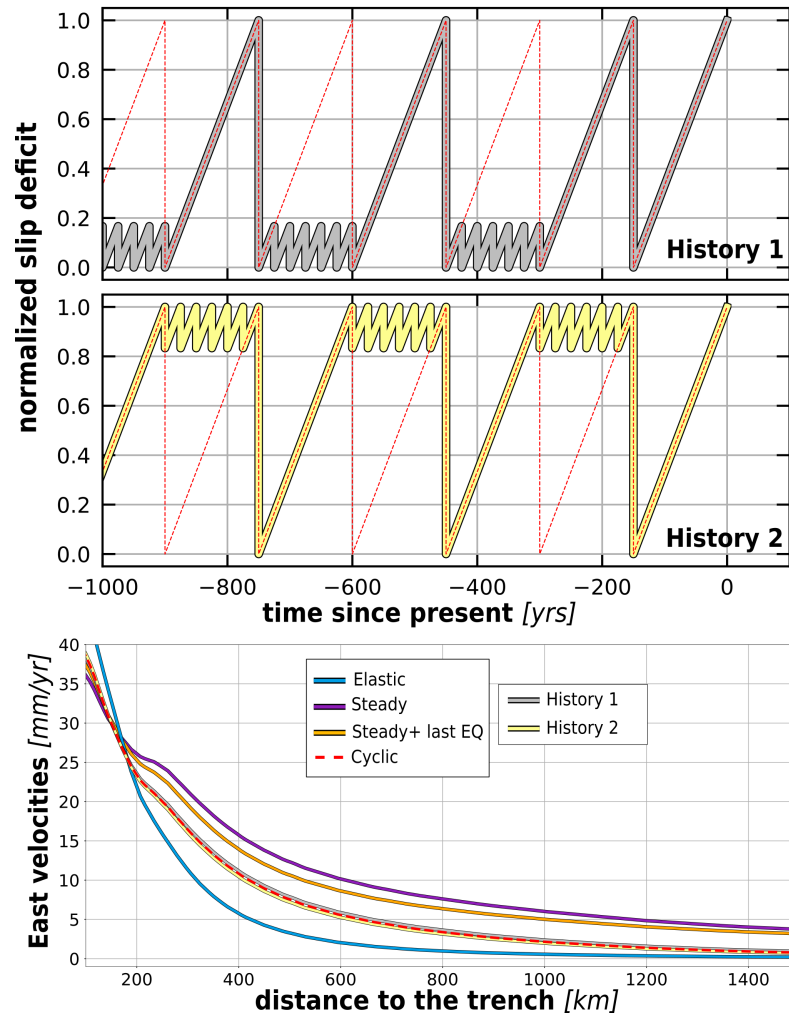


Figure S6: History 1 and 2.

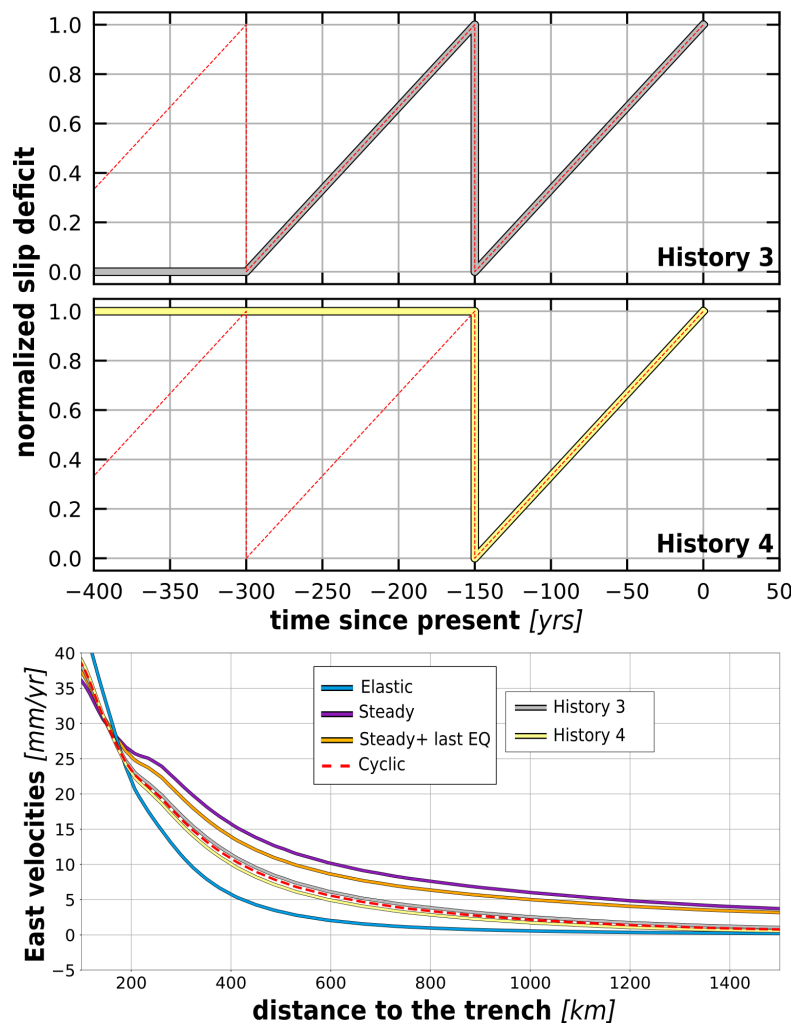


Figure S7: History 3 and 4

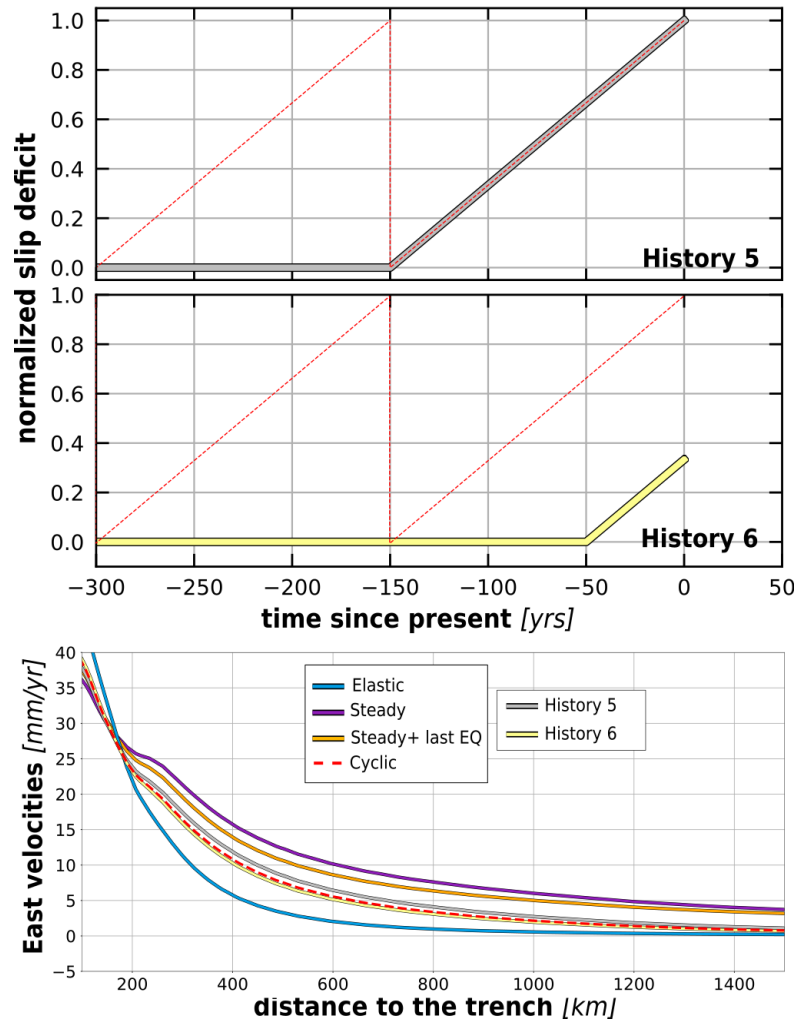


Figure S8: History 5 and 6

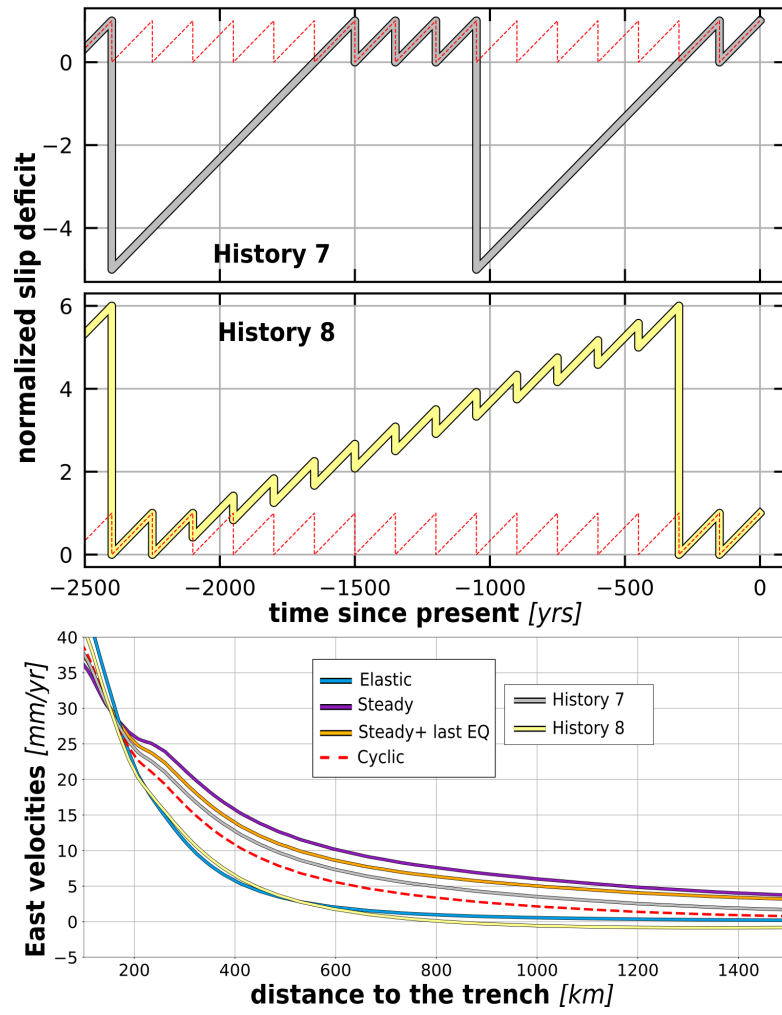


Figure S9: History 7 and 8

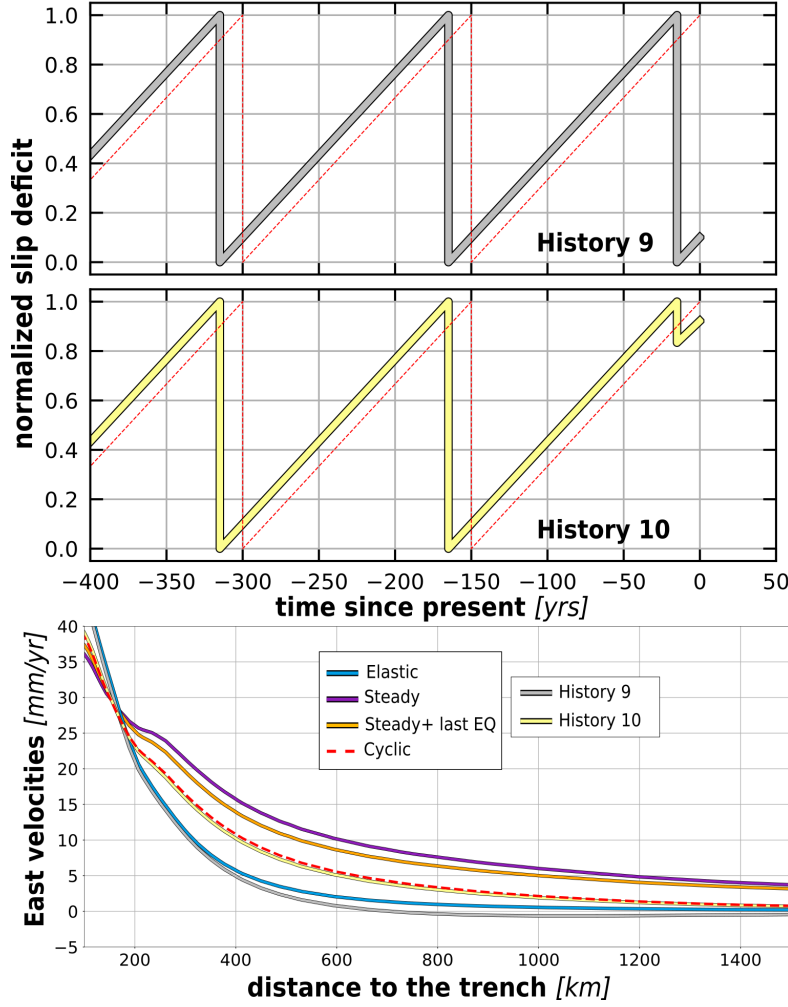


Figure S10: History 9 and 10

- **Histories 1 and 2** (Fig. S6): Between 300 and 450 years before present, or between 150 and 300 years before present, the coupling remains the same, but the $M_w 9$ earthquake has been replaced by 6 $M_w 8.5$ earthquakes. The succession of earthquakes, compensated by the coupling over a timescale of a few decades, induces a negligible signal. It is therefore as if the earthquake and the compensating backslip had been suppressed from the cyclic signal during those 150 years. As predicted from Fig. S5B, such a compensated slip on the interface affects the surface velocities very little. The velocity profile is thus very close to that predicted by the *cyclic* model: above it for history 1, and below it for history 2, because the earthquake is closer to the present for history 1, and further from it for history 2.
- **Histories 3 and 4** (Fig. S7). These histories are similar to histories 1 and 2, but both backslip and earthquakes have been suppressed before 300 years and 150 years, respectively. Again, the velocity profiles are very similar to those generated by the *cyclic* history.
- **Histories 5 and 6** (Fig. S8). History 5 is like history 4, but the earthquake at 150 years has been suppressed. As explained for Fig. S5A, the velocity induced by the earthquake is rather small compared to the velocity induced by the backslip between 0 and 150 years. In history 5, the coupling is only 50 years long, which somewhat reduces the velocity. The velocity predictions for histories 5 and 6 remain rather close to the *cyclic* case.
- **Histories 7 and 8** (Fig. S9). These histories involve megacycles, with mega-earthquakes occurring with a repeat time of the order of a millennium. In history 7, standard backslip took place over 10,000 years, except for a standard $M_w 9$ earthquake 150 years ago. The predicted velocity profile approaches that of the *steady* backslip. In history 8, a large earthquake occurred 300 years ago. The large postseismic velocity induced by that earthquake is not yet compensated by the backslip, so velocities remain trenchward in the far field.
- **Histories 9 and 10** (Fig. S10). These histories differ from the previous ones, as the surface velocity is modelled 15 years after the last earthquake, i.e. during the postseismic phase dominated by Kelvin–Voigt viscosity.

History 9 is the reference case: it is ‘normal’, and corresponds to the beginning of the cycle in the *cyclic* model. In this case, oceanward velocities are predicted at distances greater than 600 km from the trench.

In history 10, the earthquake magnitude is five times smaller. As a result, it perturbs the velocity field much less than the $M_w 9$ event of the regular cycle. This shows that oceanward velocities are not necessarily observed a few years after a moderate-magnitude earthquake (e.g. in 2010 after the moderate, $M_w 8$, 1985 Valparaíso and 1995 Atacama earthquakes).

The time required for velocities at a given distance to switch from oceanward to trenchward depends strongly on the timing of the last earthquake, i.e. on the time needed for the postseismic velocity induced by that earthquake to become smaller than the interseismic velocity.

Conclusion: If the subduction interface is unable to sustain large stresses, i.e. stresses corresponding to strains induced by the present-day coupling for periods exceeding a few hundred years, the slip history before about 100 years does not strongly affect present-day surface velocities. This holds as long as the interface cannot bear the accumulation of backslip over a long period. In practice, for computing present-day velocities induced by the slip history on the interface, we recommend including knowledge of the present-day coupling and of at least the last earthquake. At earlier periods, cycles such as those proposed here lead to surface velocity predictions that are not very different from those of a null ‘complementary slip’. Long-term accumulation of backslip without earthquakes releasing the stresses (*steady* models), on the other hand, seems inappropriate.

S5 A brief description of the various segments of the Chilean subduction zone used for the synthetic test describing the interaction between segments

S5.1 Patagonia segment

It would appear that an earthquake ruptured part of the interface in 1837 and 1737, but of much lower magnitude ($M_w 7.5 - 8$, (Lomnitz, 2004; Cisternas *et al.*, 2005)) than the 1960 earthquake ($M_w 9.5$). Paleoseismology indicates that an earthquake approaching the magnitude of that in 1960 occurred in 1575, implying a recurrence interval of 385 years. Using the coupling models provided by space geodesy for this segment (Moreno *et al.*, 2011), at a convergence velocity of 68 mm/yr, this would generate an earthquake of magnitude $M_w 9.2$ (Moreno *et al.*, 2011), implying several hypotheses. For example, energy not released by previous earthquakes could have been superimposed on that which has accumulated since 1575. Another hypothesis could be that the 1575 earthquake is still not the counterpart of the 1960 earthquake, and that the latter is even older. For numerical modelling, the earthquake is taken to be proportional to the coupling of (Moreno *et al.*, 2011). With such a coupling distribution, a recurrence time of $T = 1000$ yrs is required to produce an earthquake of $M_w 9.5$ at a convergence velocity of 68 mm/yr.

S5.2 Maule segment

It last ruptured in 2010 with the Maule $M_w 8.8$ megathrust earthquake. According to historical seismicity, the previous earthquake occurred in 1835 (e.g Darwin, 1851; Lomnitz, 2004; Ruiz et Madariaga, 2018); although it appears to have ruptured only the southern part of the Maule segment, which would explain its lower estimated magnitude ($M_w \sim 8.5$) than that of 2010 ($M_w \sim 8.8$). The observed return time would then be 175 yrs. The northern part of the segment could have ruptured during the 1730 earthquake (e.g. Lomnitz, 2004; Ruiz et Madariaga, 2018). For the numerical simulation, an earthquake proportional to the coupling of (Metois *et al.*, 2016) is considered, with a return time of 150 yrs. This value differs from the 175 years observed (although it was actually observed only once) between 2010 and 1835. However, at a convergence velocity of 68 mm/yr, and assuming an earthquake proportional to the coupling, a return time of only 90 yrs would be required to produce an earthquake of magnitude $M_w 8.8$ equivalent to that of the 2010 earthquake. On the other hand, a return time of 175 yrs would produce an earthquake of $M_w 9.05$, too high for the magnitude of the earthquakes that appear to be breaking the MAUL segment. A return time of 150 yrs is therefore chosen, giving an earthquake of $M_w 9$.

S5.3 South Metropolitan segment

The segment last ruptured in 1985 with the Valparaiso earthquake $M_w 8.0$ (Barrientos, 1988). This segment would experience events not exceeding magnitudes $M_w 8.0$ to $M_w 8.5$ with return times of the order of 80 years: in 1906 ($M_w 8.2$), 1822 ($M_w 8 - 8.5$) (Lomnitz, 2004). The 1985 earthquake would, therefore, be the smallest earthquake to have ruptured the interface of the South Metropolitan segment (Eisenberg *et al.*, 1986). The 1730 earthquake would also appear to have ruptured this segment, as well as the Metropolitan North segment, and would be associated with a much higher magnitude, $M_w \sim 9$, suggesting a seismic super-cycle involving several segments of the subduction. For numerical modelling of the seismic cycle on this segment, a return time of 80 yrs is therefore chosen. By choosing a seismic cycle with an earthquake proportional to the coupling of (Metois *et al.*, 2016) and a convergence velocity of 68 mm/yr, this results in an earthquake of $M_w 8.5$ every 80 years. As the 1985 Valparaiso earthquake was of significantly lower magnitude ($M_w 8$, i.e. a factor of 5.6 on seismic moments), the viscoelastic relaxation of the last 1985 earthquake is scaled to a $M_w 8$.

S5.4 North Metropolitan segment

The segment last ruptured in 2015 during the Illapel earthquake ($M_w 8.3$). The previous earthquake with a similar magnitude occurred in 1943. The segment also experienced an earthquake in 1880, with a magnitude likely much lower ($M_w 7.5$) than in 1943 and 2015 (Lomnitz, 2004). Based on historical seismicity, this segment also ruptured with the METS segment and the northern part of the MAUL segment during the great earthquake of 1730 ($M_w \sim 9$). If the METN segment did rupture in 1730, this would imply a period of quiescence of 213 years between 1730 and 1943, during which no earthquakes of magnitude $M_w \geq 8$ occurred (according to current knowledge). However, taking a convergence rate of 68 mm/year (Vigny *et al.*, 2009) and an earthquake proportional to the coupling (Metois *et al.*, 2016), the 1880 earthquake should have had a magnitude of $M_w 8.5$ (accumulating deformation between 1730 and 1880, over 150 yrs), i.e., one order of magnitude greater than the one that actually occurred ($M_w 7.5$) (Lomnitz, 2004). In contrast, the time elapsed between 1880 and 1943 (63 yrs) is consistent with an earthquake of magnitude $M_w 8.3$. Several hypotheses can therefore be considered to explain the absence of an $M_w 8.3$ earthquake for nearly 213 years between 1730 and 1943: the absence of

any recorded earthquakes between 1730 and 1880, or the underestimation of the magnitude of the 1880 earthquake. The return period is set at $T = 70$ years. The earthquake, therefore, has a magnitude of $M_w 8.3$, which is the same order of magnitude as those in 1943 and 2015.

S5.5 Atacama segment

The Atacama Segment is bordered to the south by the Metropolitan-North Segment (METN) and to the north by the Chañaral Segment (CHAN). To the south, the La Serena weak coupling zone (30°S) delimits the segment and seems to have stopped the two magnitude $M_w 8.5$ earthquakes that ruptured the segment in 1922 and 1819 (Willis, 1929; Abe, 1979; Beck *et al.*, 1998; Lomnitz, 2004; Ruiz et Madariaga, 2018). To the north, the segment is bounded by the weakly coupled Baranquilla zone (28°S). It would appear that the 1819 and 1922 earthquakes may have passed through it (Kelleher, 1972), although this is still debated. As with all segments of the subduction, it is once again difficult to establish a precise return time, as no earthquake of similar intensity predating that of 1819 has been clearly identified from period documents. For the numerical simulation of the seismic cycle, we take $T = 100$ yrs as the return time value, i.e. the time separating the events of 1819 and 1922. Such a return time gives a characteristic earthquake of magnitude $M_w 8.5$, if the co-seismic slip is proportional to the coupling of (Metois *et al.*, 2016) and (Klein *et al.*, 2018) and if the convergence velocity is 68 mm/yr. We note that the magnitude estimated for this segment is exactly that estimated for the 1819 and 1922 events, suggesting a good understanding of coupling in this region of Chile due to extensive GPS instrumentation work (Metois *et al.*, 2016; Klein *et al.*, 2018).

S5.6 Chañaral segment

The Chañaral segment is bounded to the south by the weakly coupled Baranquilla zone (28°S) and to the north by the Taltal anomaly (see Chapter ??). There are numerous earthquakes of magnitude $M_w \sim 7.7$: Copiapó (1918, $M_w 7.8$), Taltal (1966, $M_w 7.7$), Copiapó (1983, $M_w 7.7$) (Ruiz et Madariaga, 2018). Surprisingly, no $\geq M_w 8$ mega-earthquake has been recorded in light of historical knowledge of this segment (an almost uninhabited region until the 19th century, when mining began). This segment constitutes an anomaly on the scale of Chilean subduction. For the numerical simulations, we choose a return time of $T = 150$ yrs and an associated magnitude of $M_w 8.6$. The earthquake is then proportional to the (Klein *et al.*, 2018) coupling.

S5.7 Paranal segment

This segment last ruptured in 1995 during the Antofagasta earthquake of magnitude $M_w 8.0$ (e.g Ruegg *et al.*, 1996; Delouis *et al.*, 1997; Chlieh *et al.*, 2004). No earthquake similar to that of 1995 is known to have occurred in the past. It is therefore difficult to establish any recurrence time and magnitude for this segment. For the numerical simulations, we assume a recurrence time of $T = 80$ yrs. This gives an earthquake of magnitude $M_w 8.4$ with a convergence velocity of 68 mm/yr.

S5.8 Loa segment

It has long been suggested that this segment broke with the southern part of the Camarones segment in a magnitude $\sim M_w 9$ mega-quake in 1877. A recent study (Vigny et Klein, 2022) has shown that over time, the initial rupture of ~ 250 km (Kausel, 1986) has only increased in the literature to ~ 510 km (Jara *et al.*, 2018). A 510 km rupture would imply that the earthquake crossed the weakly coupled zone of Iquique. The initial rupture of ~ 250 km corresponds very well to the coupling zone of the Loa segment and would be assimilated to an earthquake of lower magnitude, of the order of ~ 8.5 . For the numerical simulation, the earthquake associated with this segment is considered to be proportional to the coupling of (Metois *et al.*, 2016). As the return time is difficult to determine (no previous earthquake similar to that of 1877 has been identified in the literature), the value of 150 yrs is chosen to provide an earthquake of $M_w 8.7$ (at the convergence velocity of 68 mm/yr).

S5.9 Pisagua segment

This segment ruptured in 2014 with the Iquique earthquake $M_w 8.1$ (e.g. Ruiz *et al.*, 2014; Lay *et al.*, 2014; Duputel *et al.*, 2015). For numerical simulations, we assume a return time of $T = 150$ yrs, i.e., a magnitude of $M_w 8.5$.

S5.10 Arica segment

This segment is thought to have ruptured in 1868 during the great Arica earthquake ($M_w \sim 8.5$). Determining the coupling in this region is difficult, as the subduction trench is significantly farther from the coast (200 km). This reduces the resolution of coupling models, as currently only land-based GPS stations are used. For numerical simulations, we assume a return time of $T = 150$ yrs, or a magnitude of $M_w 8.5$.

S6 GPS data

S6.1 Euler pole determination

Table 1: Velocities expressed in ITRF2020 (Altamimi *et al.*, 2023) used to determine the stable South American frame. **ens** for time-series from the (Klein *et al.*, 2022) solution. **unr** for time-series from (Blewitt *et al.*, 2018) solution.

site	lon (°E)	lat (°N)	v_E (mm/yr)	v_N (mm/yr)	σ_E (mm/yr)	σ_N (mm/yr)	solution
BABR	-44.9949	-12.1500	-3.7288	12.2159	0.0918	0.0769	ens
BAVC	-40.8027	-14.8883	-4.2646	12.2733	0.0203	0.0184	unr
BELE	-48.4625	-1.4088	-4.5490	12.8690	0.0614	0.0514	ens
BOAV	-60.7011	2.8452	-4.4233	11.8105	0.0575	0.0470	ens
BRAZ	-47.8779	-15.9475	-3.9813	12.4100	0.0255	0.0205	ens
BRFT	-38.4255	-3.8774	-4.7526	12.6894	0.0492	0.0414	ens
CEFE	-40.3195	-20.3108	-4.1894	12.7079	0.0159	0.0153	unr
CHPI	-44.9852	-22.6871	-4.0808	12.2164	0.0271	0.0226	ens
CUIB	-56.0699	-15.5553	-3.1389	11.8404	0.0229	0.0178	ens
FORT	-38.4256	-3.8774	-4.6918	12.3010	0.2271	0.2048	ens
GOJA	-51.7261	-17.8833	-3.5438	11.9262	0.1022	0.0859	ens
GVA1	-41.9576	-18.8556	-4.2809	12.5999	0.0180	0.0165	unr
KOUR	-52.8060	5.2522	-4.9416	12.1101	0.0377	0.0289	ens
MABA	-49.1223	-5.3624	-3.8503	12.7174	0.0643	0.0528	ens
MAPA	-51.0973	0.0467	-4.0000	12.4747	0.0713	0.0581	ens
MGMC	-43.8583	-16.7164	-3.7440	13.5729	0.0212	0.0192	unr
MGUB	-48.2560	-18.9192	-3.6750	12.4358	0.0204	0.0188	unr
MTCO	-55.4563	-10.8039	-3.6015	12.7997	0.0239	0.0206	unr
NAUS	-60.0550	-3.0229	-3.9936	12.3675	0.0470	0.0382	ens
PAIT	-56.0364	-4.2877	-4.0478	12.1818	0.1041	0.0802	ens
PBCG	-35.9071	-7.2137	-4.7788	11.9885	0.0690	0.0591	ens
PISR	-42.7028	-9.0307	-4.7150	12.6860	0.0220	0.0199	unr
SAGA	-67.0578	-0.1439	-4.4457	11.2877	0.0713	0.0601	ens
SALU	-44.2125	-2.5935	-4.2704	12.7674	0.0906	0.0759	ens
SAVO	-38.4323	-12.9392	-4.7012	12.5017	0.0514	0.0420	ens
SSA1	-38.5165	-12.9752	-4.3749	12.2984	0.0584	0.0485	ens
TOPL	-48.3307	-10.1711	-4.4593	12.4107	0.0564	0.0451	ens

S6.2 GPS velocities over the 2000-2010 period

Table 2: GPS velocities over the period 2000-2010 expressed in ITRF2020. **ens** for time-series from the (Klein *et al.*, 2022) solution. **spotgins** for time-series from SPOTGINS solution.

site	lon (°E)	lat (°N)	v_E (mm/yr)	v_N (mm/yr)	σ_E (mm/yr)	σ_N (mm/yr)	solution
ANTC	-71.53205	-37.33870	15.44827	10.64696	0.06488	0.06511	ens
AZUL	-59.88128	-36.76702	-0.77326	11.43344	0.55531	0.58149	ens
BTON	-71.48720	-30.26322	18.33112	18.18745	0.40394	0.37795	ens
CNBA	-71.45796	-31.39822	23.26883	17.58398	0.47079	0.45793	ens
CONS	-72.41212	-35.33096	35.56610	20.63631	0.11962	0.12281	ens
CONZ	-73.02548	-36.84376	33.15704	21.02301	0.06428	0.06412	ens
IGM1	-58.43932	-34.57224	-0.57304	11.38294	0.10436	0.10383	ens
LHCL	-65.59525	-38.00266	0.01618	10.13875	0.06403	0.06332	ens
LPGS	-57.93230	-34.90675	-0.85234	11.54574	0.06834	0.06438	ens
LSCH	-71.24604	-29.90819	17.94630	17.12355	0.44112	0.41692	ens
MAUL	-70.82145	-35.80960	18.32623	15.39614	0.24222	0.24314	ens
MPLA	-57.53114	-38.03560	-0.32894	12.18215	0.04927	0.05990	spotgins
OVLL	-71.20390	-30.60375	20.13683	19.55687	0.17779	0.16567	ens
POAL	-51.11976	-30.07404	-2.22880	12.13930	0.06204	0.05609	ens
PORT	-70.13002	-32.83523	17.19994	15.70519	0.08493	0.07659	ens
RWSN	-65.10725	-43.29888	-1.92156	11.04005	0.07885	0.08429	ens
SLMC	-70.96280	-31.77702	20.90153	17.64323	0.13016	0.12177	ens
SMAR	-53.71659	-29.71892	-2.31528	11.81531	0.62097	0.58584	ens
SRLP	-64.28040	-36.62150	-0.13910	10.69754	1.45677	1.44628	ens
UCOR	-64.19351	-31.43496	0.57809	11.21498	0.21370	0.20343	ens
UNRO	-60.62843	-32.95935	-0.29518	10.92776	0.14038	0.13691	ens
UNSJ	-68.57706	-31.54126	8.54088	12.57242	0.32764	0.30801	ens
VALN	-71.63503	-33.02793	26.90146	21.31781	0.39142	0.38768	ens
VBCA	-62.26923	-38.70077	-0.89314	10.69549	0.18354	0.19346	ens
VNEV	-70.24906	-33.35357	17.26715	15.28115	0.11188	0.10727	ens

S7 Impact of coupling depth on surface velocity

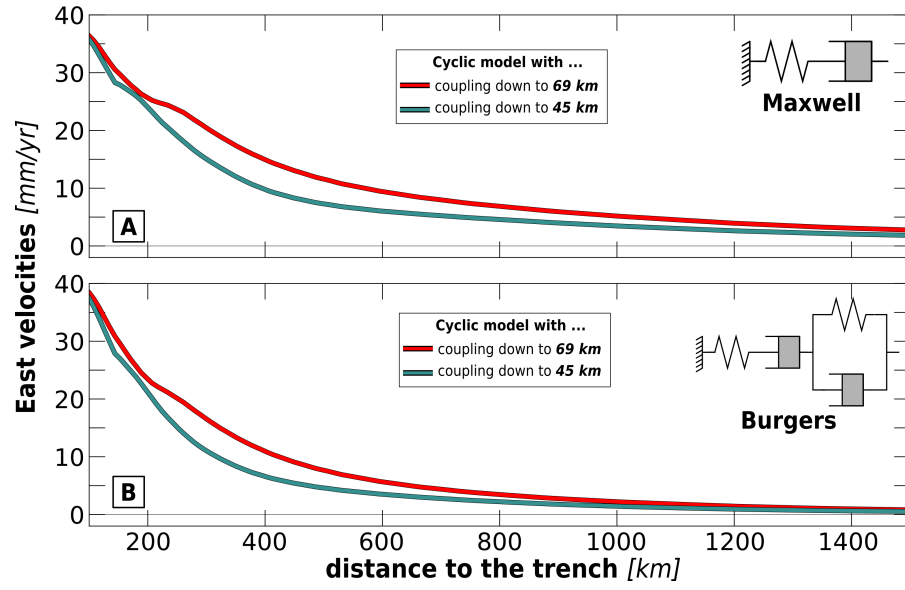


Figure S11: Impact of the assumed down-dip limit of interplate coupling on the predicted velocities for the Maule segment, shown along the P-P' profile (see Fig. 5 of the main text). The Maxwell (A) and Burgers (B) rheologies are used.

S8 Impact of the magnitude of the Valdivia earthquake on surface velocity

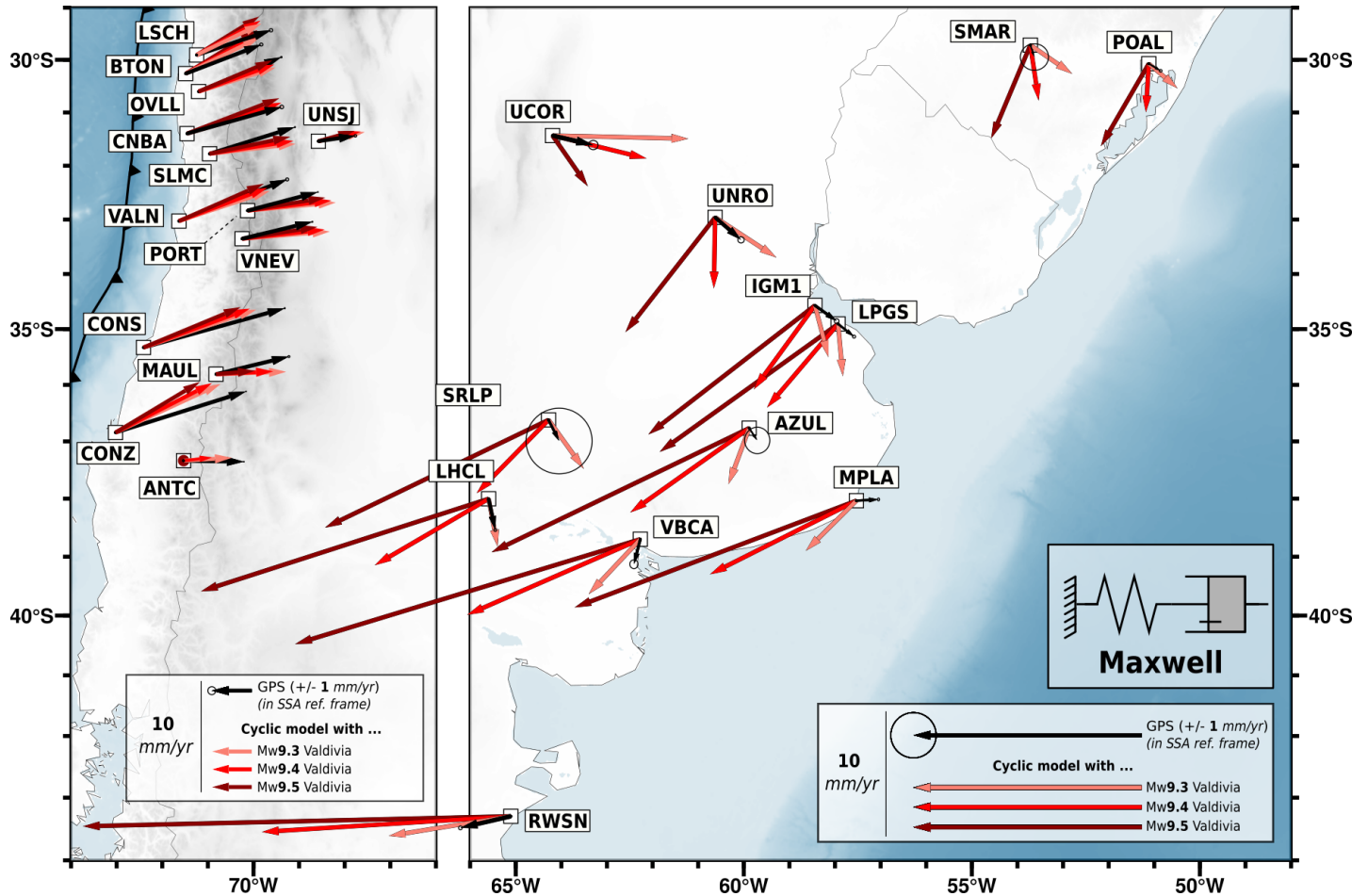


Figure S12: Impact of the magnitude of the 1960 Valdivia earthquake on the 2010 velocities induced by the *cyclic* model with a Maxwell rheology. Filled circles indicate stations where the corresponding model velocity amplitude is below 2 mm/yr (left panel).

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