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Seismotectonic Setting of the Andes along the Nazca Ridge Subduction Transect: New Insights from Thermal and Finite Element Modelling

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Abstract: The structural evolution of Andean-type orogens is strongly influenced by the geometry of the subducting slab. This study focuses on the flat-slab subduction of the Nazca Ridge and its effects on the South American Plate. The process of flat slab subduction impacts the stress distribution within the overriding plate and increases plate coupling and seismic energy release. Using the finite element method (FEM), we analyse interseismic and coseismic deformation along a 1000 km transect parallel to the ridge. We examine stress distribution, uplift patterns, and the impact of megathrust activity on deformation. To better define the crust’s properties for the model, we developed a new thermal model of the Nazca Ridge subduction zone, reconstructing the thermal structure of the overriding plate. The results show concentrated stress at the upper part of the locked plate interface, extending into the Coastal and Western Cordilleras, with deeper stress zones correlating with seismicity. Uplift patterns align with long-term rates of 0.7–1 mm/yr. Cooling from flat-slab subduction strengthens the overriding plate, allowing far-field stress transmission and deformation. These findings provide insights into the tectonic processes driving stress accumulation, seismicity, and uplift along the Peruvian margin.

Keywords: subduction zones; slab geometry; thermal structure; finite-element method; uplift; stress and strain



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1. Introduction

The structural evolution of Andean-type orogens is strongly influenced by the geometry of the subducting slab. Several researchers [1–6] have widely investigated how anomalies in the thickness of the oceanic crust, such as ridges and plateaus, have significant implications for the geodynamics of subduction margins and the overriding plate [3,6]. In this study, we focus on the Nazca Ridge and the effects of its subduction on the overriding South American Plate.

The Nazca Ridge, shown in Figure 1, is a portion of oceanic crust with an estimated thickness of 18 ± 3 km [7] that is currently subducting beneath the South American Plate between 14° and 16° S [7]. The origin of this aseismic ridge is related to the spreading centre of the Pacific–Farallon/Nazca plates, and it was formed during the early Cretaceous [4,8,9].

The collision and the consequent subduction of the ridge with the South American Plate occurred around 11.2 Ma at a latitude of 11° S. Having an orientation of $N42^\circ$ E, the ridge tends to migrate southward, and it reached the current subduction zone around 4 to 5 Ma ago [10,11]. The ridge extends approximately 200 km in width and over 1100 km in length, with an additional estimated 900 km of the ridge already subducted beneath

the South American Plate [9]. The presence of this thickened crust has probably induced the geological evolution of the Peruvian margin and a flat-slab subduction phenomenon beneath the South American continental margin [12]. According to tomographic data, the flat slab extends for nearly 300 km at depths between 80 and 100 km [13–15]. The presence of a flat slab in this region leads to several significant consequences. The stress distribution within the overriding plate is altered, resulting in a reduction of extensional stresses and localised shortening in the forearc region. In these conditions, the coupling between the overriding and subducting plates is strengthened and induces a greater release of seismic energy [16]. Despite this, there are notably few intermediate-depth earthquakes within the lower plate and along its upper boundary [17]. Additionally, the flat slab causes the asthenospheric wedge to withdraw from the trench, producing a westward pinching out of the asthenosphere beneath the overriding plate, and significantly impacting the thermal structure of the upper plate. Indeed, the withdrawal produces the eastward migration of arc magmatism and, sometimes, the complete deletion of volcanic activity [16]. Nevertheless, the impact of the Nazca Ridge's flat subduction on the geological evolution of the Peruvian Cordillera, as well as the geodynamic behaviour of the Peruvian margin and the South American Plate, remains a subject of ongoing debate. In particular, the factors driving crustal shortening are still uncertain, with interplate coupling [16,18,19] and structural reactivation [20,21] both proposed as potential influences.

In this study, we investigated the stress distribution on the overriding plate and, in particular, during the interseismic and coseismic stages of the subduction zone and the megathrust. With the aim of simulating strain accumulation and interseismic stress, we discretized a 2D FEM model along a transect of 1000 km that runs almost parallel to the Nazca Ridge and the flat slab (Figure 1). The transect intersects the main features of the South Peruvian Cordillera until the undeformed area of the Amazonian Basin. The coseismic slip was constrained, taking into consideration the earthquakes with $M_w \geq 6$ that occurred during the last 50 years in a buffer of 100 km around the study area [22]. These events are related to the megathrust and the distribution describes the subduction interface. To perform the FEM modelling of the megathrust, we used a modern and asserted methodology already employed in several studies [23] and in different geological contexts, (i.e., Chile [24], China [25]; Val D'Agri (Italy) [26]; Zagros (Iran) [27]. The procedure consists of an analysis of interseismic deformation in zones of plate convergence, which results from the elastic strain in a brittle layer above a ductile half space [28]. Understanding this deformation is crucial for comprehending the tectonic processes driving seismic activity and surface uplift. This model assumes that the brittle layer encompasses the entire model thickness, exhibiting stick-slip behaviour. Furthermore, we modelled coseismic deformation linked with certain seismic occurrences using the same numerical technique. To constrain our model, we utilized available geological information derived from both published works [11–15] and our fieldwork in the study area. Moreover, investigating the stress distribution and geodynamics of the area requires understanding the thermal structure of the overriding plate [29–33]. Using this approach, data about rheology are crucial for performing an accurate analysis of the co-seismic field, identifying areas with high strain rates, and determining where stress tends to accumulate. Therefore, in this study, we propose an upgrade to the thermal model proposed by Ciattoni et al. [34] for the same area.

The new thermal model considers new parameters of density and a more accurate geometry of the overriding plate. This combined approach enabled us to study recent uplift through topographic analysis and lithologic boundary conditions. By using FEM methodology, we aim to explore the interseismic and coseismic deformation processes and compare them with the observed uplift patterns. This comprehensive analysis not only enhances our understanding of crustal deformation but also provides insights into the underlying tectonic mechanisms.

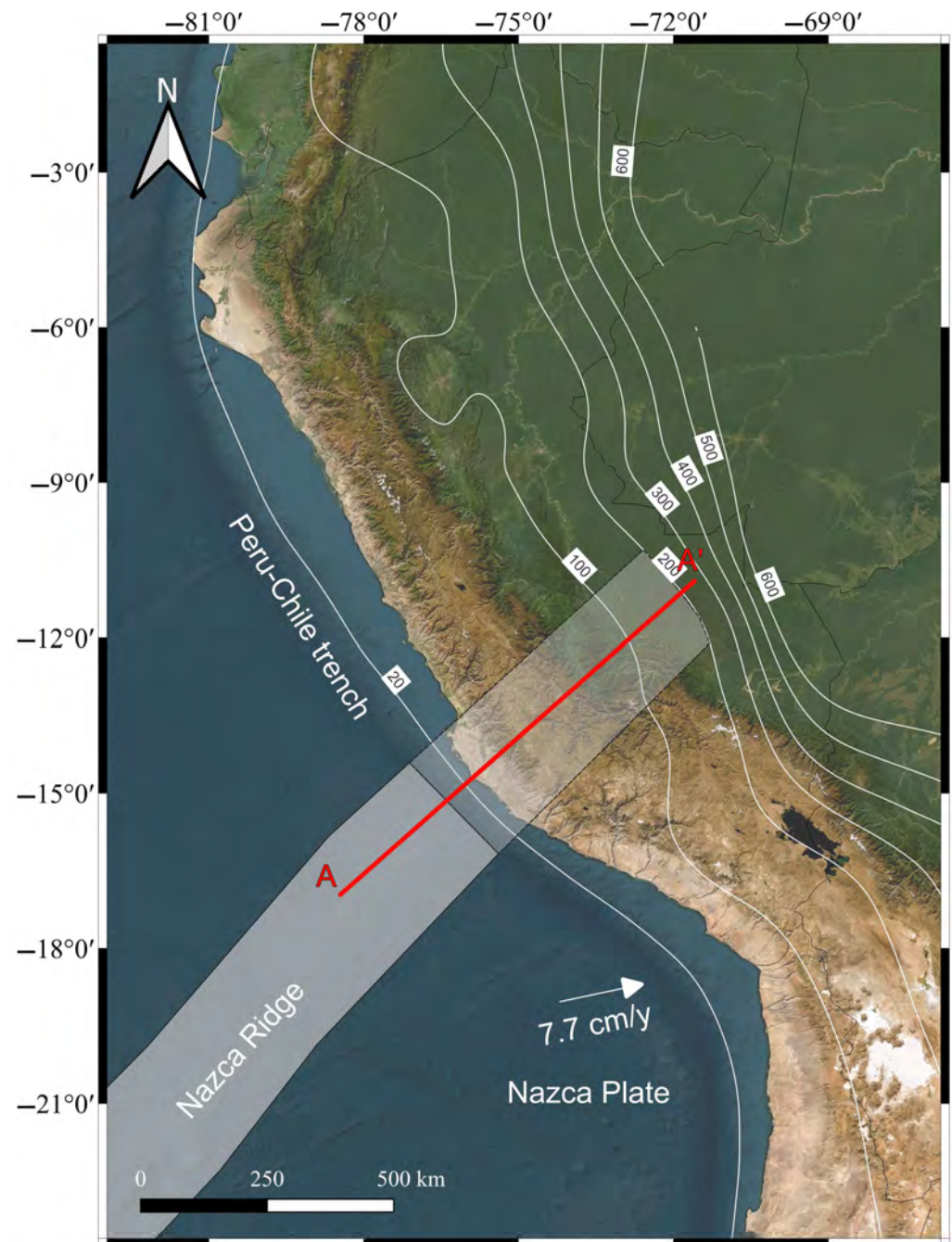


Figure 1. Map of the study area (Esri World Imagery basemap). The red line (A–A′) is the trace of the studied transect; the white lines are the 100 km contour line indicating the depth of the slab from Hayes et al. [15]. The grey area shows the Nazca Ridge, while the light-grey area represents the subducted portion of the Nazca Ridge.

2. Geological Setting

The Andean Cordillera is a continental mountain belt that ranges from 8° N to 52° S and is the result of the effect of the Nazca Plate subduction beneath the South American Plate [35]. The Peruvian Andes are a part of the Central Cordillera and, more specifically, belong to the southern portion of the Northern Central Andes and the northern branch of The Central Andes Orocline. The current configuration of this portion of the Cordillera is the expression of a compelling geological history [36,37]. During the Paleozoic, the first deformation step occurred during the Hercynian orogenic phase (Late Devonian and Early Mississippian), which strongly deformed the high-grade metamorphic basement

(Precambrian) and the overlying marine sediments (Middle Ordovician to the Upper Devonian) [36]. The second crucial deformation stage, called the Andean orogenic cycle, began during the Cenozoic, and it is still active. In this tectonic phase, we can identify three compressional periods: (i) Peruvian (Late Cretaceous); (ii) Incaic (Eocene); and (iii) Quechua (Miocene). The Quechua compressional phase is subdivided into three subphases: (i) Quechua 1 (Lower Miocene), (ii) Quechua 2 (Tortonian), and (iii) Quechua 3 (Messinian). All of these compressional periods are combined with plutonic activity, volcanism, and periods of sedimentation, and part of the deformation occurred along the inherited structures of the Hercynian phase [36]. The most recent history of this part of the Cordillera is strongly linked to the Nazca Ridge subduction, and to its southwest migration. Since the beginning of its subduction 11 Ma at 11° S, the presence of the Nazca Ridge considerably affected the development of the Peruvian forearc system and the deformation of the innermost part of the overriding plate along the whole Peruvian margin until it came to its present position (14° – 16° S) [9,38]. The most evident effect of the Nazca Ridge subduction is the evolution and uplift of the East Pisco forearc Basin [39,40] and the deformation of the Amazonian Basin with the formation of the Fitzcarrald Arch [9,19,41]. In this study, we focused our attention on the current subduction zone of the Nazca Ridge. The study area (Figure 2) goes from 78° W 16° S to 71° W 10° S and includes (i) the East Pisco Basin, which represents the subaerial portion of the Peruvian Forearc System; (ii) the Coastal Cordillera and the Marañón fold-and-thrust belt, which compose the Western Cordillera; (iii) the Eastern Cordillera; (iv) the Sub Andean Zone; and (v) the Fitzcarrald arch that is part of the Amazonian Basin.

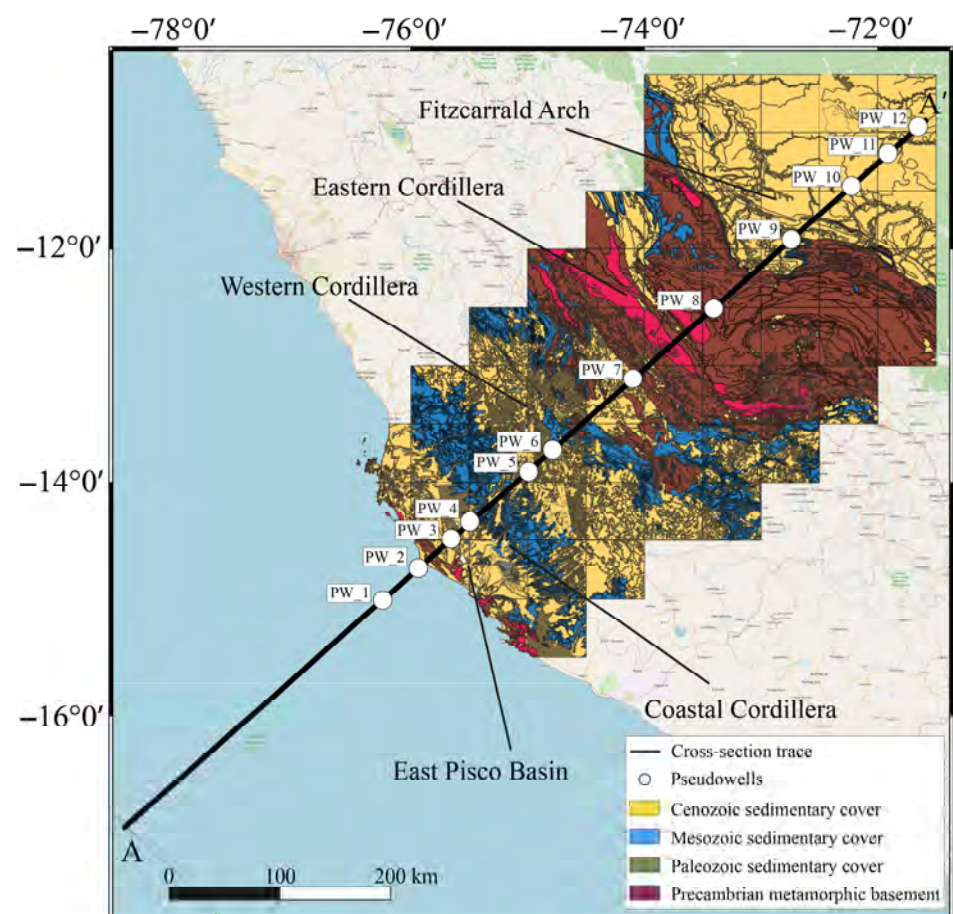


Figure 2. Geological map of the study area based on the 1:100,000 scale geological maps from INGEMMET database. Modified after Ciattoni et al. [34]. The black line (A–A') is the trace of the studied transect, and the white dots indicate the pseudowell location.

3. Methods

In this study, FEM analysis is used to examine both interseismic and coseismic deformation along the section. This procedure consists of the geometric modelling of the fault and the crustal geological section, discretizing the model by 2D finite-element method (FEM) by MSC-Marc software (2012) [42] and recreating the main geological features of the tectono-stratigraphy units of the study area. Since the strength and the rheology of the lithosphere are functions of the temperature and the heat flow, we also analysed the thermal structure of the overriding plate. In particular, the model proposed by Ciattoni et al. [34] has been improved and integrated with new available data. Density data for the crustal layers were obtained from the models provided by Ciattoni et al. [43].

3.1. Thermal Model

The method used to develop the model is the same as that applied by Ciattoni et al. [34], where they used the one-dimensional calculation approach to 11 pseudo-wells placed along the section. In this research, we added a further pseudo-well to improve the model's constraints in the Western Cordillera. The pseudowells reach the Moho interface, so their depths range from a minimum of 21 km (PW_2) to a maximum of 51 km (PW_5). Next, the heat flow obtained was interpolated through a sixth-order polynomial equation. Although the procedure and the zone remained unchanged from Ciattoni et al. [34], adjustments were made to the density and geometry of the layers based on new insights suggested by Ciattoni et al. [43]. These authors [43] produced a new integrated model of the crustal structure up to depth of 130 km along the transect across the Peruvian Forearc and Andean Cordillera. They integrated geological and geophysical data and performed forward modelling of the Bouguer anomaly to develop a 2D density model. They propose a more detailed geometry of the geological structures present in the area, including the slab and the Moho. Moreover, the density data are validated by a gravimetric model that takes into account the petrological properties of the materials and the P-T conditions. The adjustments provided in this study to the thermal model consist of changes to the basement thicknesses by 10% and values of the densities that vary by less than 5%. A thickness change can affect the heat flow amount, while a different density can affect the frictional heat calculation. To calculate: (i) heat flow at the surface, (ii) temperature gradients, and (iii) lines of equal temperature within the continental crust of South America, we utilized the method previously applied in various regions by Basilici et al. [44], Santini et al. [45], and Ciattoni et al. [34]. The thermal model described here accounts for four layers that represent the primary elements of the tectono-stratigraphy in the studied region: (i) sedimentary units from Mesozoic to Cenozoic, (ii) sedimentary units from Paleozoic, (iii) upper basement, and (iv) lower basement. The Paleozoic sequence is primarily made up of a mix of schists, sandstones, and limestones, while the Cenozoic–Mesozoic succession is mostly composed of micritic limestones and calcareous sandstones, with occasional quartz-rich sandstones, conglomerates, and volcanic materials like andesitic lavas and tuffs [46–49]. The upper basement consists of high-grade metamorphic rocks like granulites and amphibolites, while the lower basement is made up of mafic rocks. Figure 3 represents the entire plan.

Calculating surface heat flow (Q_s) for the model in Figure 3, we considered the mantle heat flow (Q_m), radiogenic heat in the sedimentary cover (Q_{SC}), and in the basement (Q_B), as well as frictional heat (Q_F) in the megathrust zone, resulting in the following equation:

$$Q_s = Q' + \sum_{i=1}^2 Q_{SC_i} + Q_B \quad (1)$$

In the formula, Q' represents the heat transfer at the bottom of the model caused by the heat transfer from the mantle and the frictional heat from the thrust if present. In the megathrust area, we took into account $Q' = (Q_0 + Q_F) S^{-1}$, where Q_0 represents the heat flow of the oceanic mantle and $S = 1 + b \left(z_f v \sin \delta / \kappa \right)^{1/2}$, with $b = 1$, is the decreasing

factor close to the megathrust fault [4]. For the continental zone, we took Q' to be equal to Q_m . Instead, $Q_{SC_i} = H_{SC_i}h_{SC_i}$ represents the contributions from individual sedimentary layers with a constant radiogenic source, as mentioned in references [50–53], while $Q_B = H_B D \left(1 - e^{-\frac{h_{UB}+h_{LB}}{D}}\right)$ denotes the contribution of the basement with an exponential trend due to a radiogenic source, as discussed in references [54,55].

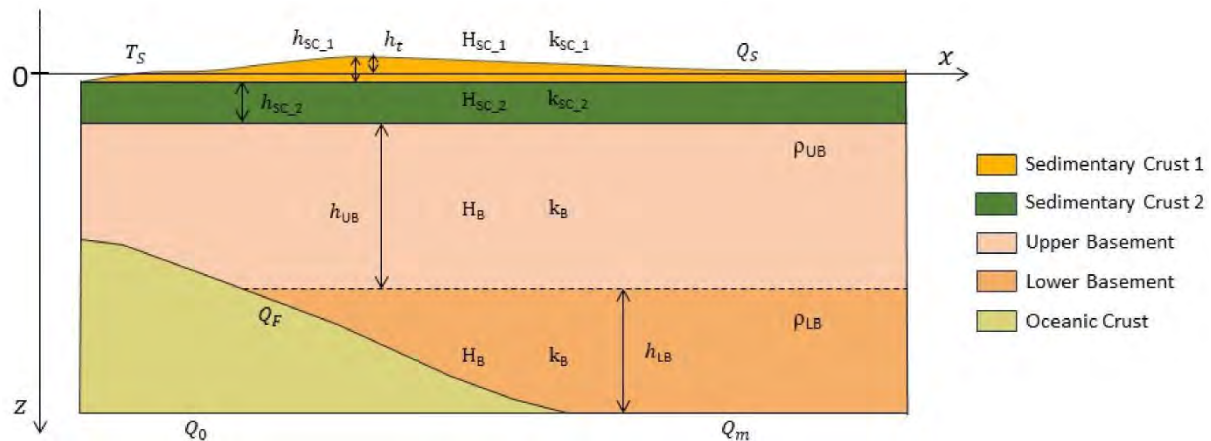


Figure 3. Sketch of the layers used in the analytical procedure (the parameters are listed in Table 1).

With a constant radiogenic source, the temperature in sedimentary cover layers is determined as follows:

$$T_{SC_i}(z) = T_{SC_i} + \frac{Q_s - \sum_{i=1}^2 Q_{SC_{i-1}}}{k_{SC_i}} \left(z - z_t - \sum_{i=1}^2 h_{SC_{i-1}} \right) - \frac{H_{SC_i}}{2k_{SC_i}} \left(z - z_t - \sum_{i=1}^2 h_{SC_{i-1}} \right)^2 \quad (2)$$

The temperature at the top of layer i is denoted as T_{SC_i} .

Taking into account a radiogenic source exhibiting an exponential pattern, we calculate the temperature found within the basement as follows:

$$T_B(z) = T_B + \frac{Q'}{k_B} \left(z - z_t - \sum_{i=1}^2 h_{SC_i} \right) + \frac{H_B D^2}{k_B} \left(1 - e^{-\frac{(z-z_t-\sum_{i=1}^2 h_{SC_i})}{D}} \right) \quad (3)$$

T_B represents the temperature at the top of the basement, while D represents the depth scale. The definitions of the terms in the equations can be found in Table 1.

Table 1. Definitions of terms considered in the thermal model [56–61].

Equation Parameters	Model Data
z_t (m)	Ground Elevation (Negative) or Sea Depth (Positive) related to m.s.l.
h_{SC_1} (m)	Sedimentary cover (Cenozoic–Mesozoic) thickness
h_{SC_2} (m)	Sedimentary cover (Paleozoic) thickness
h_{UB} (m)	Upper basement thickness
h_{LB} (m)	Lower basement thickness
	$H_{SC_1} = 0.83 \text{ Wm}^{-3}$ RHP * of sedimentary cover (Cenozoic–Mesozoic) [56]
	$H_{SC_2} = 1.2 \text{ } \mu\text{Wm}^{-3}$ RHP * of sedimentary cover (Paleozoic) [56]
	$H_B = 1.6 \text{ Wm}^{-3}$ RHP * of basement thickness [56]
	$k_{SC_1} = 2.1 \text{ Wm}^{-1}\text{K}^{-1}$ Thermal conductivity of sedimentary cover (Cenozoic–Mesozoic) [57]
	$k_{SC_2} = 2.4 \text{ Wm}^{-1}\text{K}^{-1}$ Thermal conductivity of sedimentary cover (Paleozoic) [57]

Table 1. Cont.

Equation Parameters	Model Data
	$k_B = 2.7 \text{ Wm}^{-1}\text{K}^{-1}$ Thermal conductivity of the basement [57]
	$\kappa = 10^{-6} \text{ m}^2\text{s}^{-1}$ Thermal diffusivity [58]
	$\mu = 0.7$ Coefficient of static friction [58]
	$\rho_{SC_1} = 2.4 \times 10^3 \text{ kgm}^{-3}$ Density of sedimentary cover (Cenozoic–Mesozoic) [58]
	$\rho_{SC_2} = 2.4 \times 10^3 \text{ kgm}^{-3}$ Density of sedimentary cover (Paleozoic) [59]
	$\rho_{UB} = 2.83 \times 10^3 \text{ kgm}^{-3}$ Density of the upper basement [59]
	$\rho_{LB} = 2.98 \times 10^3 \text{ kgm}^{-3}$ Density of the lower basement [59]
$x_f \text{ (km)}$	Horizontal coordinate for points on the megathrust fault
$z_f \text{ (km)}$	Vertical coordinate for points on the megathrust fault
$\delta = \arctg\left(\frac{z_f}{x_f}\right)$	Dip angle of the megathrust fault
	$\lambda = 0.96$ Pore fluid factor [60]
	$v = 7.7 \text{ cm yr}^{-1}$ Relative plate velocity [60]
$Q_F = \rho gh(1 - \lambda)v$	Frictional heat flow density
	$Q_o = 60 \text{ mWm}^{-2}$ Oceanic heat flow density from the mantle [61]
	$Q_m = 20 \text{ mWm}^{-2}$ Continental heat flow density from the mantle [61]
$Q_S \text{ (mWm}^{-2}\text{)}$	Surface heat flow density
	$T_S = 15 \text{ }^\circ\text{C}$ Surface temperature

* Radiogenic heat productivity.

3.2. Finite-Element Modelling

This study examines the coseismic and interseismic deformation along the segment using finite element analysis. Using MSC-Marc software [42], the crustal geological section has been discretized in order to create a 2D FEM model that replicates the major tectono-stratigraphic features of the study region. As represented in Figure 4, in the model, the crust was considered to be composed of four main layers: (i) sedimentary cover, which includes the entire sedimentary succession without any distinction in age or composition; (ii) basement, composed of rocks of granitoid composition (including gneiss and granulites); (iii) oceanic crust; and (iv) lithospheric mantle. To define the rheological behaviour of the basement, we computed its strain rate using the empirical flow law in the well-known power law from Kirby [62]:

$$\dot{\epsilon} = A\sigma^n e^{-Q/RT} \quad (4)$$

where σ is differential stress, A is a material property, Q is activation energy, R is the molar gas constant, and T is absolute temperature. The most appropriate values for the parameters included in the equation above were chosen by comparing both the observed Vp (recorded by seismic stations) and lithology type with published Vp values and available crustal rocks, as discussed by Ranalli and Murphy [63] and Ranalli [64]. When information about rock density is unavailable, it is often estimated from P-wave velocity (Vp) using an empirical relationship, such as Gardner's equation [65]. For the basement layer, we selected a value of $1.8 \cdot 10^{-9} \text{ MPa}^{-n} \text{ s}^{-1}$ for Parameter A, 123 kJ mol^{-1} for the activation energy (Q), and 3.2 for the power-law exponent. By defining both temperature and differential stress as a function of depth in the calculation, we can describe the variations of strain rate with depth.

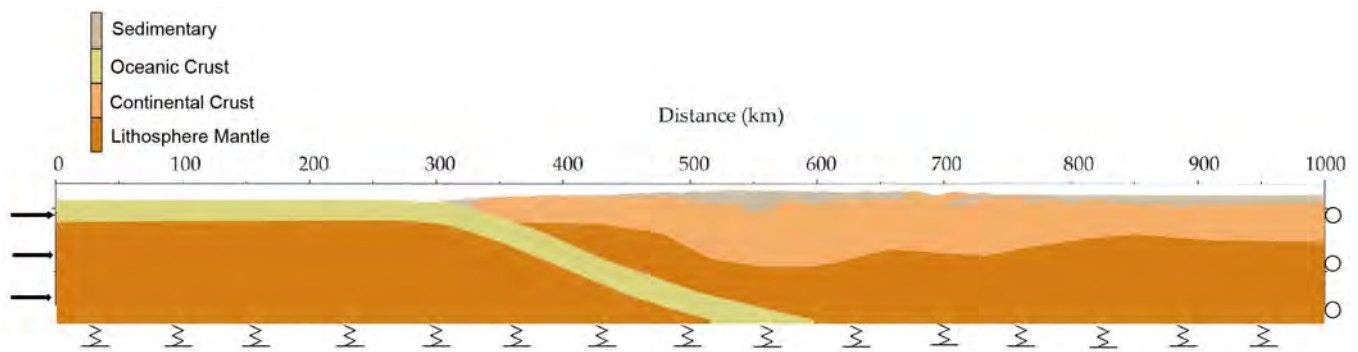


Figure 4. Sketch of the FEM model, showing the layers used in modelling.

The proportion between the viscous and elastic behaviour of a particular element is governed by the local crustal geotherm derived from heat flow measurements. Equation (4) can be expressed in terms of a simple viscous relation:

$$\dot{\epsilon} = \frac{1}{2\eta} \sigma \quad (5)$$

This is defined by an effective viscosity:

$$\eta = \frac{1}{2A} \sigma^{1-n} e^{Q/RT} \quad (6)$$

which varies with depth as a function of temperature T [66,67].

The friction coefficient was set as $\mu = 0.7$ according to Cermak et al. [58]. The model was divided into 65,670 quadrangular elementary cells and 66,740 nodes. Numerical modelling included two independent FEM simulations conducted using the same mesh: a first procedure was used to analyse interseismic stress and strain accumulations, while a second procedure was applied to investigate surface vertical motions associated with the megathrust coseismic stage (by modelling the characteristic earthquake). Modelling of the interseismic stage was divided into two different steps: the first step consisted of setting the boundary conditions of the model in order to observe the effects of gravity alone. In the model, the following boundary conditions were applied: (i) the surface was considered free to move in all directions; (ii) the SW and NE boundaries were considered free to move in the vertical direction and locked in the horizontal direction; and (iii) the base was treated as a Winkler's foundation [68]. In the simulation of the hydrostatic pressure of the Earth's mantle, the model was set in order to allow free horizontal movement, while the vertical motion was subject to the influence of an elastic spring with stiffness coefficient K equal to $K = A/L \cdot E$, where A is the base length, L is the thickness of the model, and E is the average of Young's modulus of the rocks included in the model.

Using this method, we were able to estimate the amount of equivalent Von Mises stress (σ_{VM}):

$$\sigma_{VM} = \sqrt{\frac{3}{2} \left(\sum_{ij} \sigma'_{ij} \sigma'_{ij} \right)}, \text{ with } \sigma'_{ij} = \sigma_{ij} - \frac{1}{3} \sum_k \delta_{ij} \sigma_{kk} \quad (7)$$

where σ_{ij} is the stress component and δ_{ij} is the Kronecker delta, and of equivalent strain (ϵ_{eq}):

$$\epsilon_{eq} = \sqrt{\frac{2}{3} \left(\sum_{ij} \epsilon'_{ij} \epsilon'_{ij} \right)}, \text{ with } \epsilon'_{ij} = \epsilon_{ij} - \frac{1}{3} \sum_k \delta_{ij} \epsilon_{kk} \quad (8)$$

where ϵ_{ij} is the strain component [69]. This procedure also yields values of surface motion in the vertical direction, accumulated over a particular time interval during the interseismic stage. Modelling of the coseismic stage shares the first step with the interseismic simulation procedure. However, in this case, the second step consisted of the sudden movement of a

part of the fault plane to simulate stick-slip behaviour. This behaviour was achieved by imposing a slip value consistent with the magnitude of the seismic event to be modelled (characteristic earthquake in our instance).

Studies on the historical and recent seismicity of the area show that the most significant events do not exceed M_w 8, in particular, an event of M_w 7.7 occurred on 12 November 1996, along our section at a depth of 33 km. In addition, other studies indicate a period of 100 years as seismic periodicity [70]. Based on these two indications, we created our FEM model (Figure 5) and calculated surface displacement, stress, and strain. For the coseismic phase, we considered a fault with a width of about 50 km, imposing a constant slip of 70 cm on it to simulate, as the FEM model was made, an event with a M_w 7.5. To define the fault width, we considered the portion of the slab where most events with magnitudes greater than 6 occur. The depth range for our target fault is between 30 km and 50 km. During the interseismic phase, we applied the plate convergence rate of 7.7 cm/yr [61], which, when projected along our section, provides a velocity of 6.5 cm/yr. This was applied to the southwestern edge of the FEM model for a duration of 100 years, locking the upper part of the slab down to the simulated fault's base while allowing the slab beneath to move freely, representing aseismic movement.

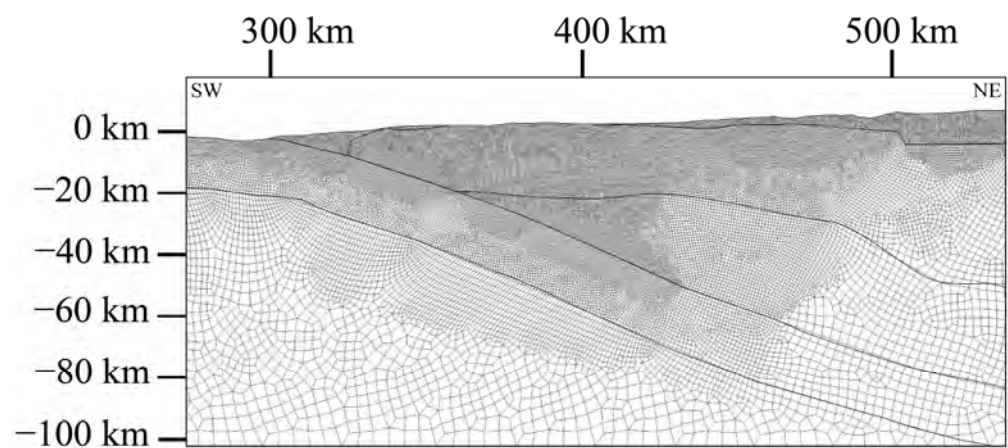


Figure 5. Detail of the FEM model, showing the constructed mesh pattern. The cells are of smaller size in areas that are more critical for modelling deformation and stress accumulation.

4. Results

4.1. Thermal Model

The computed geotherms (Figure 6) are characterised by lower temperature values for Wells PW1 to PW4, while higher temperatures are obtained for Well PW5 at the same depth. Notably, there is a similar trend in the geothermal profiles for Wells PW7 to PW11. The variation of temperature as a function of depth along the entire geological profile leads to a complex trend of the isotherms. In Figure 7, moving from SW to NE, we observe: (i) a deepening of the isotherms near the cold plate beneath the megathrust; (ii) an arching of the curves in the central part of the profile, where there is a thickening of the sedimentary layers and the basement; and finally (iii) a trend subparallel to the surface for the remaining part.

The trend of the heat flux at the surface was defined by calculating the Q_s of the 12 pseudo-wells through the analytical treatment described in Section 3.1, and the results were compared with the observed data available in the literature by projecting them orthogonally on the profile (Figure 7). In particular, the results show low values of Q_s along the southwestern shallow part of the megathrust ranging from about 24 mW/m² to 36 mW/m², and then it continues to increase to about 47 mW/m² at Pseudo-well 5. In the central part of the section, the heat flux decreases slightly and then remains constant between about 40 mW/m² and 44 mW/m². The low values of Q_s (30 ± 5 mW/m²) are due to the presence of oceanic plate subduction, between the trench and the volcanic arc [34].

The higher values of the heat flux ($45 - 50 \text{ mW/m}^2$) depend mostly on the thickening of the crustal structure since the contribution of mantle flux is constant.

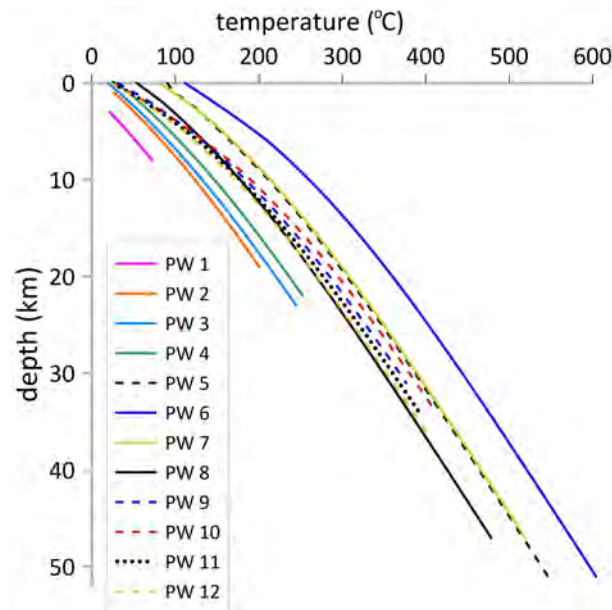


Figure 6. Computed geotherms for the 12 pseudo-wells.

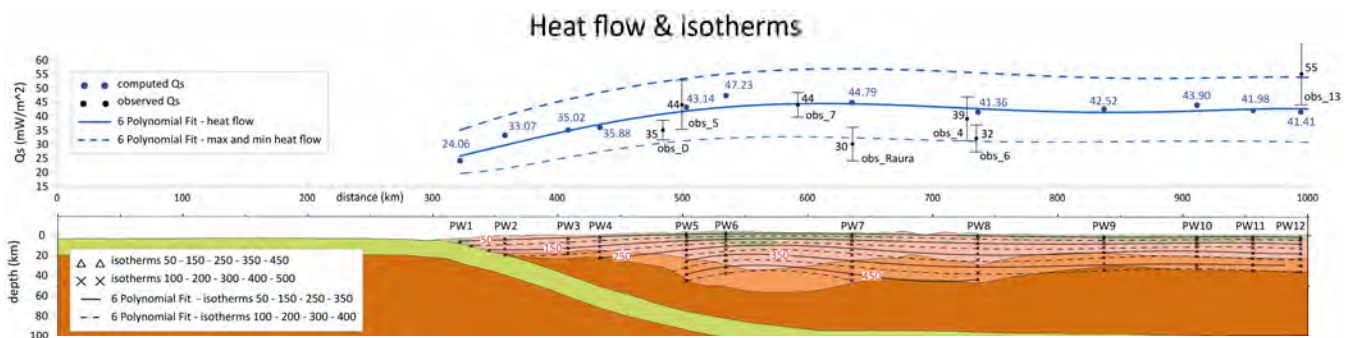


Figure 7. The values of heat flow Q_s (blue dots) are interpolated using a sixth-degree polynomial (solid blue line) to discern the trend across the crustal cross-section. A graphical comparison is made with the observed data (black dots) and their associated relative errors (black bars), taking into account the maximum and minimum variations in radiogenic heat production, heat flow, and slab dip during the heat-flow calculations (dashed blue lines). The isotherms are displayed in the section at intervals of 50°C (dashed and solid black lines). For comparative purposes, the lithospheric section, provided by Ciattoni et al. [43], is included in the background.

4.2. Finite-Element Model

Figure 8 shows the surface displacements for the coseismic and interseismic phase, as well as the cumulative coseismic–interseismic phase. The cumulative displacement has a maximum of about 0.3 m, over a period of 100 years (0.3 mm/yr), at the top of the fault similar to the coseismic displacement but decreases less rapidly than the coseismic vertical displacement due to contribution from the interseismic displacement, which ranges from 0.05 to 0.2 m. The interseismic results of the equivalent Von Mises stress (Figure 9) and the equivalent strain (Figure 10a) show an accumulation of stress and strain in the locked part of the slab, as well as in the deepest part. In order to compare stress and strain accumulation with the pattern of seismicity, the hypocenters of $M_w \geq 5$ seismic events that occurred within 100 km over the last 40 years are shown in the geological section (Figure 10b). Further higher values of stress and strain are observed in the accretionary prism and in some areas of the interface between the basement and the sedimentary cover.

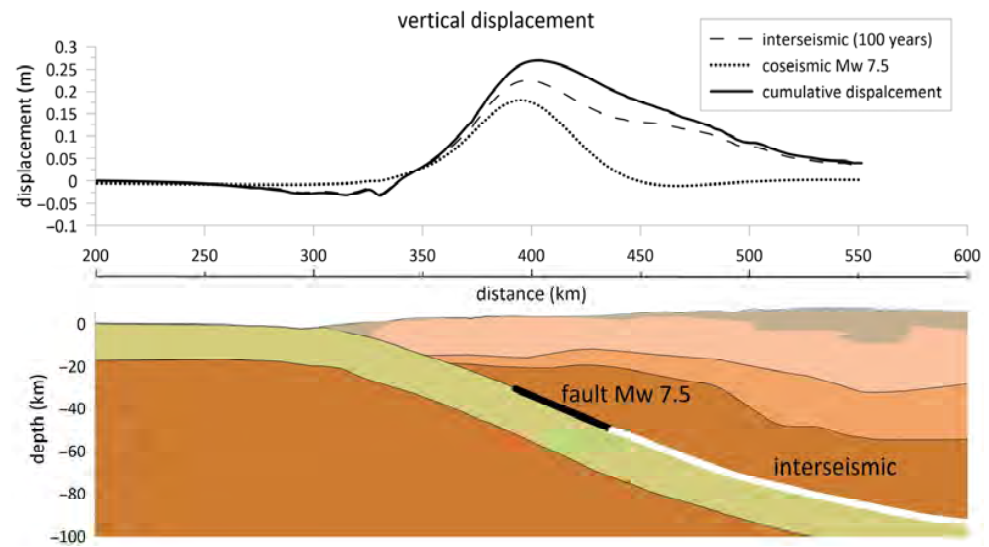


Figure 8. Modelled results for coseismic, interseismic, and cumulative vertical displacement.

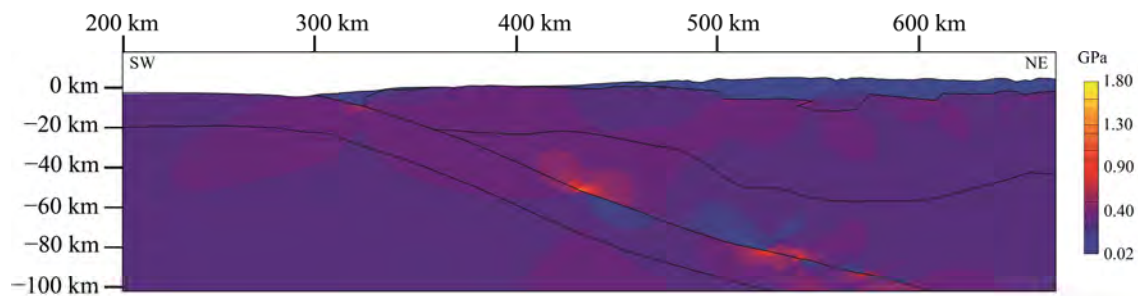


Figure 9. Von Mises stress results. White fault segment is unlocked (i.e., free to slip).

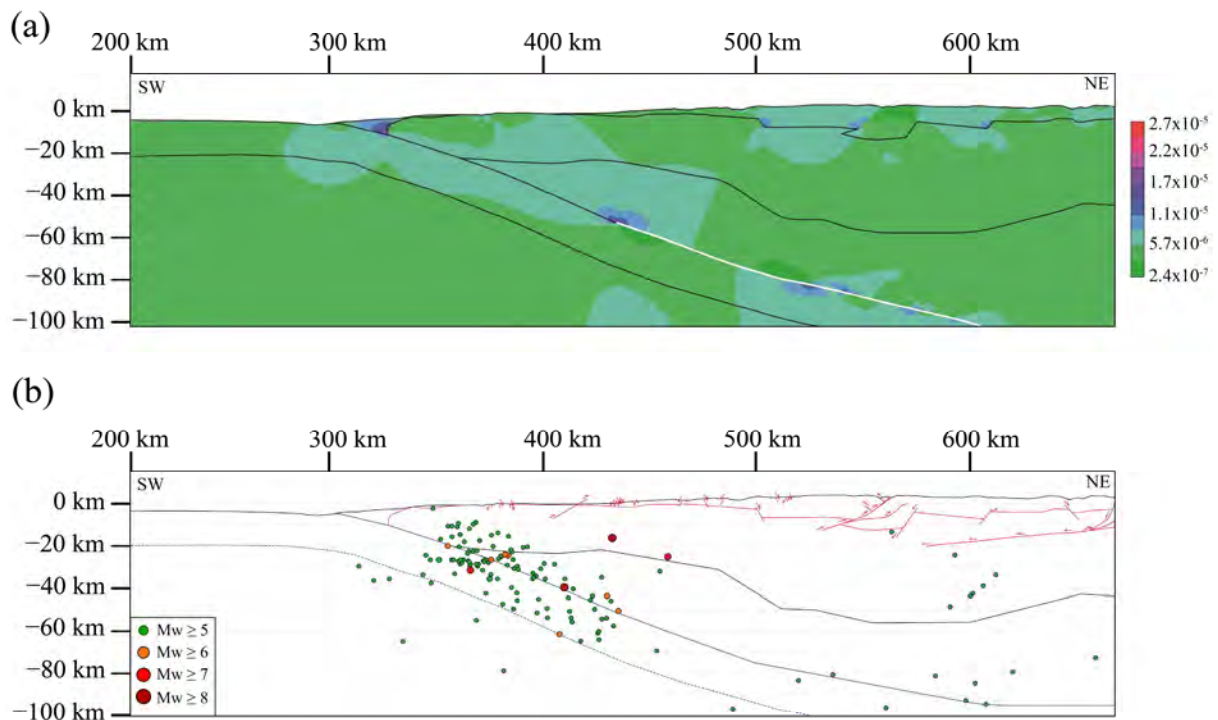


Figure 10. (a) Equivalent total strain results. White fault segment is unlocked (i.e., free to slip). (b) Earthquake hypocenters of events with $M_w \geq 5$ that occurred in the last 40 years, projected from a 100 km wide region along the study section. Data from the USGS catalogue [22] and ISC catalogue [71].

5. Discussion and Conclusions

The integration of recent data from Ciattoni et al. [43] has resulted in variations in both the heat flow and the geothermal trend, predominantly observed in the initial segment of the section. Notably, in contrast to the model proposed by Ciattoni et al. [34], heat flow decreases from PW2 to PW5, with the most significant variation reaching about 4 mW/m² at PW3. Additionally, the isotherms are positioned at greater depths due to the reduction in temperature at those depths. In general, the thermal structure of the South American continental crust is characterised by unusually low temperatures at the Moho depth. This is consistent with the abrupt cooling of the upper plate, which is believed to accompany flat-slab subduction [16,72]. Therefore, the recent (i.e., Pliocene) establishment of flat-slab subduction along the study transect bears major implications for upper-plate rheology, as cooling produces a strengthening of the overriding plate. Upper-plate strengthening may allow far-field stress transmission, which in turn leads to foreland deformation. Therefore, it may be envisaged that “young” flat-slab subduction in the study area controls deformation, uplift, and recent (<4 Ma) [9,19,41] exhumation as far as the Fitzcarrald Arc foreland domain to the NE. Our study shows for the first time the marked effect of the intensely debated, recent (<5 Ma) Nazca ridge subduction in perturbing the thermal structure of the overriding South American crust. The thermal model conducted in this study was also instrumental in subsequent finite element modelling, as temperature controls effective viscosity in the FEM and, hence, have a significant impact on modelled rheology.

Finite-element modelling conducted in this study was focussed on permanent deformation and uplift patterns, neglecting any coseismic elastic rebound effect, which is not modelled in our FEM. The accumulation of elastic strain is well known to occur where plate motion is hindered by frictional coupling associated with asperities along the plate interface, representing a fundamental aspect of the well-established earthquake cycle model for megathrusts located at subduction plate boundaries [73]. Rupturing of the asperities and recovery of the accumulated strain occurs by elastic rebound of both subducting and overriding plates during megathrust earthquakes, as testified by the fact that most directly observable coseismic strain recovery occurs elastically during the seismic event. As a matter of fact, measured crustal displacements associated with megathrust earthquakes that originated at various subduction zones are consistent with such elastic rebound models, the region of maximum coseismic surface displacement being commonly located close to the trench (and offshore in most of the cases) [74].

The uplift pattern depicted by our FEM results (Figure 8), not taking into account coseismic elastic strain recovery, is consistent with the long-term uplift observed along the coastal area of the active margin of Peru. Macharé and Ortlieb [10] determined a maximum uplift rate of around 0.7 mm/yr, while Zeumann and Hampel [11] suggested that long-term surface maximum uplift in the studied sector of the Andes during the Quaternary is of ca. 1 mm/yr. These values agree with the values estimated by Sillard et al. [75], who, using the technique of cosmogenic ¹⁰Be exposure dating, estimated that cumulative uplift rates have progressively increased since approximately the middle Pleistocene from about 0.4 to 0.9 mm/yr. Moreover, these authors suggested that the increasing uplift rate for 800,000 years is related to the subduction of the Nazca Ridge.

However, such surface uplift is affected by erosion, which is not considered in our rock uplift model. As erosion rates obtained by Wipf et al. [39] for the study area range from 0.06 mm/yr to 4 mm/yr, our model satisfactorily compares with measured uplift rates. Due to the locked plate interface, modelled interseismic deformation produces an uplift pattern that extends inboard to the Coastal Cordillera and, to a lesser extent, even to the Western Cordillera. On the other hand, coseismic deformation is characterised by a distinct maximum of focussed ground motion in the offshore forearc basin domain. This latter uplift component is consistent with the Pliocene–Quaternary vertical motion and emergence of the eastern sector of the forearc basin system (i.e., the East Pisco basin), which is currently located at a maximum elevation of about 650 m above sea level. Within this framework, the uncertainties involved in earthquake recurrence intervals and erosion rates

in the extrapolation of the coupled coseismic–interseismic deformation do not represent a major issue, since our aim is to obtain information on the location of zones of larger uplift and not on the absolute amounts of surface motion.

FEM results, besides showing predictable stress and strain accumulation at the upper tip of the unlocked (i.e., free to slip) plate interface, also show zones of stress concentration (Figure 9) and strain accumulation at deeper levels along the plate boundary (Figure 10a). This is consistent with transient seismic behaviour at much greater depths, which is testified by the seismicity (Figure 10b).

In conclusion, integrated thermal and mechanical modelling conducted in this study provide useful insights into the active tectonic setting and plate-scale geodynamics of the Peruvian margin along the axis of the subducting aseismic Nazca Ridge.

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