Extension and Dynamics of the Andes inferred from the 2016 Parina (Huarichancara) Earthquake

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Abstract

The M_w 6.1 2016 Parina earthquake led to extension of the south Peruvian Andes along a normal fault with evidence of Holocene slip. We use InSAR, seismology and field mapping to determine a source model for this event and show that extension at Parina is oriented NE-SW, which is parallel to the shortening direction in the adjacent sub-Andean lowlands. In addition, we use earthquake source models and GPS data to demonstrate that shortening within the sub-Andes is parallel to topographic gradients. Both observations imply that forces resulting from spatial variations in gravitational potential energy are important in controlling the geometry of the deformation in the Andes. We calculate the horizontal forces per unit length acting between the Andes and South America due to these potential energy contrasts to be $4 - 8 \times 10^{12}$ N per metre along-strike of the mountain range. Normal faulting at Parina implies that the Andes in south Peru have reached the maximum elevation that can be supported by the forces transmitted across the adjacent foreland, which requires that the foreland faults have an effective coefficient of friction $\lesssim 0.2$. Additionally, the onset of extension in parts of the central Andes following orogen-wide compression in the late Miocene suggests there has been a change in the force balance within the mountains. We propose that shortening on weak detachment faults within the Andean foreland since ~5-9 Ma reduced the shear tractions acting along the base of the upper crust in the eastern Andes, leading to extension in the highest parts of the range.

Key Points:

- The Parina earthquake led to NE-SW extension in the high Andes of south Peru

- Faults in the South American forelands have an effective coefficient of friction less than 0.2

21   • The strength of faults within the forelands modulates the stress state in the high Andes

1   Introduction

23 On the 1st December 2016 (22:40:26 UTC) a M_w 6.1 normal-faulting earthquake occurred $\sim$70 km

24 west of Lake Titicaca in the south Peruvian Andes (Fig. 1a). This event marks the first $M_w > 6$

25 extensional earthquake to occur within the Andes for 70 years, and provides a rare opportunity to

26 study crustal deformation, lithosphere rheology and the forces driving deformation in this region.

27   Observations of crustal deformation within the high Andes are sparse, but are fundamental to

28 testing models of orogen dynamics [e.g. England and Houseman, 1989; Liu et al., 2002; Flesch and

29 Kreemer, 2010]. One particular area of contention is whether extension in the high Andes is parallel

30 [e.g. Dalmayrac and Molnar, 1981] or perpendicular [e.g. Sébrier et al., 1985; Mercier et al., 1992] to

31 shortening in the adjacent sub-Andean lowlands. Extension in the Andes parallel to shortening in the

32 sub-Andes can be explained by the balance between the forces arising from differences in gravitational

33 potential energy between the mountains and lowlands, and the forces acting through the foreland

34 lithosphere [Dalmayrac and Molnar, 1981; Froidevaux and Isacks, 1984]. However, if extension in the

35 Andes is perpendicular to sub-Andean shortening, shear tractions acting on the base of the Andean

36 lithosphere are required to influence the stress state in the mountains [e.g. Liu et al., 2002].

37   Current observations of normal-faulting earthquakes in the Andes [Doser, 1987; Cabrera and Se-

38 brier, 1998; Devlin et al., 2012; Jay et al., 2015] provide a limited constraint on the geometry of the

39 extension, as the majority of the events are too small ($M_w < 5.5$) to break the full thickness of the seis-

40 mogenic layer, so may not be representative of the dominant crustal strain [e.g. Brune, 1968; Jackson

41 and White, 1989; Jackson and Blenkinsop, 1997; Scholz, 2002]. To supplement the sparse earthquake

42 data, a number of studies have inferred the direction of extension throughout the Andes to be N-S

43 based on slip vectors along minor fault planes [e.g. Sébrier et al., 1985; Mercier et al., 1992; Cladouhos

44 et al., 1994; Schoenbohm and Strecker, 2009; Daxberger and Riller, 2015; Tibaldi and Bonali, 2018].

45 However slip on minor faults also represents a small component of the total crustal strain, and may

46 not accurately record the bulk orientation of extension [Riller et al., 2017].

47   There also exists competing views regarding the orientation of shortening across the sub-Andean

48 lowlands. Studies of earthquakes suggest shortening in the sub-Andes is parallel to gradients in the

49 topography [Assumpção, 1992; Assumpção and Araujo, 1993; Lamb, 2000]. Alternatively, some studies

50 have argued that the GPS velocity field in the sub-Andes can be explained by a combination of elastic

51 strain accumulation on the locked Peru-Chile subduction interface and slip along a detachment fault

52 beneath the sub-Andes. These elastic models assume motion on the detachment fault beneath the sub-

Andes has a slip vector parallel to relative plate motion between the Nazca plate and South America, rather than parallel to gradients in topography [Bevis et al., 2001; Chlieh et al., 2011].

In this study we use geomorphology, seismology and interferometric synthetic aperture radar (In-SAR) to determine a detailed source model of the 2016 Parina earthquake, and relate the pattern of co- and post-seismic fault slip to the surface geomorphology. We then combine our source model for the Parina earthquake with additional seismological source models, and existing GPS data, to re-assess the pattern of crustal deformation in the high Andes, and its relationship with shortening in the adjacent sub-Andean lowlands. We use the pattern of deformation to test models of the forces driving deformation in the Andes, and investigate the rheology of the South American lithosphere. Finally, we discuss possible causes for the onset of normal faulting in the high Andes at ∼5-9 Ma, and suggest a general model for the support and evolution of mountain ranges based upon the strength of faults in their forelands.

2   1st December 2016 Parina Earthquake

The M_w 6.1 Parina earthquake occurred in a region of south Peru that has experienced a number of recent M_w 5-5.8 normal-faulting earthquakes [Dziewonski et al., 1981; Cabrera and Sebrier, 1998; Ekström et al., 2012; Devlin et al., 2012; Jay et al., 2015] (Fig. 1a). Geomorphic evidence of recent normal faulting in the epicentral region is scarce [e.g. Benavente et al., 2013], but has been documented further north near Cuzco [Suarez et al., 1983; Sébrier et al., 1985; Mercier et al., 1992; Benavente et al., 2013], and to the south near Arequipa [Lavenu et al., 2000]. The limited geomorphic expression of normal faults in south Peru probably reflects the small amount of finite extensional strain in the high mountains (< 1%; Sébrier et al. [1985]).

2.1   Local Geomorphology and Surface Ruptures

Surface ruptures associated with the Parina earthquake were mapped within three days of the event (Fig. 1b), and consist of two NW-SE trending sections with an along-strike length of ∼12 km, a maximum height of ∼30 cm (downthrown to the SW), and maximum tensional opening of ∼30 cm (Fig. 2). The ruptures coincide with a 150 m-high escarpment that extends an additional 10 km NW of Parina, and bounds a fluvial basin on its SW side (Fig. 1c). Tensional opening across the ruptures was largest in areas where vertical offsets were also highest, which is a common feature of normal faults that steepen in the near-surface [e.g. Jackson et al., 1982].

The northern section of ruptures consist of discontinuous, overlapping splays that map onto NW-SE trending, metre-high scarps visible in pre-event satellite imagery (Fig. 2b). Both the tensional

84 opening and vertical offsets across the ruptures in the northern section were typically <10 cm (Fig.

85 2d). In the southern section the ruptures have both opening and vertical components up to 30 cm

86 (Fig. 2c), and form a semi-continuous trace that also coincides with a metre-high scarp (Fig. 2a).

87 Dammed ponds on the downthrown (SW) side of the scarps pre-date the Parina earthquake (Fig.

88 2a,b), and suggest there has probably been ≲10-20 m of slip on the Parina Fault since the last major

89 glacial epoch reset the landscape in the region at ~10-30 ka [Clapperton, 1983; Smith et al., 2005],

90 implying a fault slip rate ≲1 mm/yr.

2.2 Teleseismic Body Waveform Modelling

92 We determined the strike, dip, rake, centroid depth, source-time function and moment release of the

93 Parina earthquake by modelling the event as a finite-duration rupture at a point source, and performing

94 a joint inversion of long-period P and SH seismic waveforms recorded at teleseismic distances using

95 the MT5 program of Zwick et al. [1994] (based on the algorithm of McCaffrey and Abers [1988]

96 and McCaffrey et al. [1991]). The methodology behind this procedure has been described in detail

97 elsewhere [see Molnar and Lyon-Caen, 1989; Taymaz et al., 1990], hence only a brief summary will be

98 provided here.

99   Broadband seismograms recorded at stations within 30-90° of the earthquake epicentre were se-

100 lected from the IRIS data management centre, and were filtered to reproduce the response of a long-

101 period WWSSN instrument (15-100 s). Long-period waveforms are insensitive to small-scale hetero-

102 geneity in the source region velocity structure [Taymaz et al., 1990], therefore we use a simple 1-D

103 crustal velocity model with $V_p = 6.5$ km s^{-1}, $V_s = 3.7$ km s^{-1} and density 2800 kg m^{-3} [Dorbath,

104 1996; Schurr et al., 1999]. Inaccuracies in the velocity model typically lead to uncertainties in the

105 centroid depth estimate of ±4 km for shallow (<30 km) crustal events [Taymaz et al., 1990]. Accurate

106 arrival times of the P and SH phases were picked from the broadband seismograms. The starting

107 model for the inversion was taken as the best double-couple gCMT solution [Ekström et al., 2012],

108 and the final model is constrained to have a double-couple moment tensor.

109   The minimum-misfit solution consists of a normal-fault source with a strike/dip/rake of 144/39/276,

110 a centroid depth of 12 km, and a seismic moment of 1.1×10^{18} Nm (M_w 6.0) (Fig. 3). The mapped

111 surface ruptures trend parallel to the minimum-misfit strike estimate, and the sense of motion across

112 them indicates the SW-dipping nodal plane is the fault plane.

113   Uncertainties in the source model were estimated by fixing each parameter of interest at values

114 away from their best fit, and performing inversions in which all other parameters were free to vary.

115 We estimate the strike to be constrained between 110 and 160° by the deterioration in fit between SH

116 waveforms recorded at stations DBIC, COYC and EFI outside this range (Fig. 4). Within this range

117 of possible strikes, the dip cannot be lower than 35° otherwise the stations MPOM and SJG would

118 lie near the P wave nodal plane, and the amplitude of the predicted P arrivals would be too low to

119 fit the observations. Mechanisms with dips $>55^o$ do not fit the observed waveforms as they predict

120 the incorrect SH first motion at stations FFC and SSPA. The rake must be in the range 250-300° to

121 fit the SH first motions and amplitudes at stations DBIC, COYC, EFI and MESJ (Fig. 4). All the

122 models that fit the observed waveforms have a slip vector azimuth between 211° and 245° (NE-SW).

123 Centroid depths greater than 20 km cannot match the observed SH waveforms as the direct S and

124 reflected sS phases separate, which is not observed (e.g. see clear waveform misfits at stations EFI

125 and COYC in Fig. 4). For all inversion results with a good fit to the observed waveforms, the seismic

126 moment is $0.9 - 1.4 \times 10^{18}$ Nm (M_w 5.9-6.1).

2.3   Co-seismic InSAR: Observations

128 We formed Sentinel-1 ascending- and descending-track interferograms covering the co-seismic period

129 of the Parina earthquake using the European Space Agency's SNAP software (`http://step.esa.`

130 `int/main/`). The effect of topography on phase in the interferograms was removed using an SRTM

131 3-arcsecond elevation model [Farr et al., 2007], following which we phase filtered the interferograms

132 [Goldstein et al., 1988], and unwrapped using the statistical-cost network-flow algorithm (SNAPHU)

133 of Chen and Zebker [2001]. Additional information about the interferograms is provided in Table 3 in

134 the Supplementary Information.

135   The wrapped interferograms (Fig. 5) show 5-6 concentric, elongate fringes (corresponding to

136 a maximum of ∼15 cm of line-of-sight displacement) oriented ∼NW-SE in both the ascending and

137 descending track, which is consistent with ground motion resulting from a fault with the same geometry

138 as our body-waveform solution. In addition, the similar pattern of displacement in the ascending and

139 descending tracks implies the majority of the deformation was vertical, which is consistent with motion

140 on a dip-slip fault [e.g. Copley et al., 2015]. The smooth fringe pattern, located to the SW of the

141 surface ruptures, and the higher fringe density in the NE than the SW of the displacement patch,

142 implies the majority of slip remained buried on a fault that dips SW, and has an up-dip projection

143 coincident with the surface ruptures. The along-strike width of the fringes is ∼15 km, and faults that

144 are ∼15 km long typically produce M_w 6.0-6.5 earthquakes [Wells and Coppersmith, 1994].

2.4 Co-seismic InSAR: Modelling

We determine the source parameters of the fault responsible for the 2016 Parina earthquake by inverting the InSAR measurements of surface displacement for the best fitting constant-slip, rectangular, elastic dislocation [Okada, 1985] using a simulated annealing algorithm that minimizes the root-mean-square (RMS) misfit between model and observations [e.g. Copley et al., 2015]. We solve for the location of the fault plane, its length, width, strike, dip, rake, the amount of slip and a constant offset and linear ramp in line-of-sight (LOS) across each interferogram to account for long-wavelength atmospheric and orbital artefacts. The interferograms are down-sampled uniformly over the inversion window to 200×200 m pixels to reduce the number of data points in the inversion to $\sim$70,000. The starting model for the inversion is taken to be the SW-dipping nodal plane of our minimum-misfit body-waveform solution.

To assess the fit between the models and the observations, we estimate the noise levels in the interferograms using the magnitude of phase variations within non-deforming regions. Noise in the interferograms leads to apparent LOS variations between 1.0 and 1.1 cm in the ascending interferogram, and between 0.8 and 0.9 cm in the descending interferogram.

The inversion results suggest the InSAR data are best-fit by $\sim$0.5 m of slip buried on a 13 km long normal fault with a strike/dip/rake of 135/40/250 that mostly ruptured between 3 and 10 km depth (Fig. 6). The surface projection of the best-fit fault plane is coincident with the surface ruptures. RMS residuals between the best-fit model and the InSAR observations are 0.8 cm for the ascending track and 0.9 cm for the descending track, therefore the data are fit to within the noise levels, and a more complex model of distributed fault slip is not required or justified by the data.

To test the sensitivity of the InSAR measurements to changes in the fault geometry, we performed a grid search of inversions in which each fault parameter was independently fixed at a range of values, whilst all others could vary, and measured how the misfit evolved (Fig. 7). We find that fault planes with strike $= 120 - 145^o$, dip $= 35 - 55^o$, rake $= 250 - 270^o$, top depth $= 1.5 - 4.5$ km and bottom depth $= 6 - 14$ km can match the InSAR observations to within the noise levels. Together the range of acceptable solutions allows the slip vector azimuth to be between 230° and 280° (Fig. 7). The teleseismic waveform modelling constrains the slip vector azimuth to be between 211° and 245°, therefore the range of slip vectors consistent with all of the InSAR, seismology and surface rupture constraints have an azimuth of 230-245° (NE-SW; see Fig. 7).

2.5   Post-seismic Deformation

Models of co-seismic fault motion indicate the majority of slip in the Parina earthquake remained buried below 3 km depth (Figs. 6, 7). However the 10-30 cm high surface ruptures measured three days after the earthquake suggest that some co-seismic slip reached the surface. Decreasing co-seismic slip on faults at shallow depths is common [e.g. Fialko et al., 2005] and has previously been observed for normal faults [e.g. Copley, 2012; Bie et al., 2014]. Limited shallow co-seismic slip is likely to reflect either inelastic deformation of the near-surface [Kaneko and Fialko, 2011], rate-strengthening frictional properties of the shallow fault zone [Marone and Scholz, 1988; Scholz, 1998], or both. In the latter case, the fault may creep post-seismically ('afterslip'), causing transient, short-wavelength surface deformation near the co-seismic rupture.

We formed interferograms covering the first 8 months following the Parina earthquake (Fig. 8a). In addition, we also generated a time-series of post-seismic deformation by extracting the relative LOS motions to either side of the surface ruptures from a series of interferograms that were formed with a common post-earthquake acquisition (collected within 19 days of the earthquake) as the reference image in each pair.

The descending-track interferograms reveal a sharp offset in the LOS displacement field across the central section of surface ruptures, with the size of the offset increasing with time after the earthquake (Fig. 8a,f). The LOS offset decays with distance perpendicular to the surface ruptures over a length scale of ∼3-5 km (Fig. 8f), indicating the majority of post-seismic deformation is within the top 5 km of the crust. Post-seismic signals are less clear in the ascending-track interferograms due to the larger component of atmospheric noise (Supp. Fig. 14). The temporal evolution of the post-seismic deformation follows a typical logarithmic transient decay [e.g. Smith and Wyss, 1968; Ingleby and Wright, 2017] with a relaxation time of 12 days (Fig. 8g), which is similar to estimates from a number of other studies [e.g. Savage et al., 2005; Fielding et al., 2009].

We interpret the post-seismic InSAR observations to reflect shallow afterslip on the up-dip extension of the co-seismic fault plane, and not shallow poro-elastic deformation [e.g. Peltzer et al., 1998; Fielding et al., 2009], because the sharp step in LOS directly correlates with the location of the co-seismic surface ruptures, and the polarity of the LOS motions are the same as those in the earthquake. We inverted the observed surface displacements 6 months after the Parina earthquake for the best fit, constant-slip fault (as in Section 2.4) with the rake fixed to that of the co-seismic fault model. The post-seismic InSAR measurements are consistent with ∼6-7 cm of afterslip on a fault with equivalent strike, dip and length to the co-seismic fault plane, but with afterslip focused between 0 and 4 km depth (Fig. 8b,d). However the constant-slip model cannot match both the smooth LOS displacement

pattern south of the surface ruptures, and the sharp change in LOS across the central section of the ruptures (Fig. 8a,f), indicating that afterslip did not extend to the surface along the whole fault. In addition, a small region of symmetrical footwall uplift and hanginwall subsidence across the southern section of surface ruptures requires a locally steeper-dipping fault plane of $\sim70^o$ to fit the observed LOS motion in that area (Fig. 8a,f).

2.6   Parina Earthquake: Summary

We find that the Parina earthquake ruptured a 13 km long, NW-SE trending, SW-dipping, shallow-crustal normal fault with a NE-SW ($230\text{-}245^o$) oriented slip vector azimuth (source models summarised in Table 1). The majority of slip in the earthquake (~0.5 m) was buried between 3-10 km depth, however a small component of slip did extend to the surface (0.1-0.3 m). Within 6 months of the earthquake around 15-35% of the shallow co-seismic slip deficit was released through post-seismic afterslip up-dip of the co-seismic rupture.

Only by combining geodetic, seismological and field observations have we been able to properly quantify the uncertainty in the earthquake slip vector, and show that the direction of extension at Parina is NE-SW. Seismicity surrounding Parina is typically confined to the top 10-15 km of the crust (Fig. 1a), suggesting the Parina earthquake ruptured most, if not all, of the seismogenic layer. As a result, the NE-SW extension is likely to be representative of the dominant strain within this region.

Shortening in the adjacent sub-Andean lowlands recorded in earthquake slip vectors and the orientation of fold axes and thrust fault scarps is parallel to the direction of extension at Parina (Fig. 1a), implying the mountains are locally deforming in plane strain. Smaller normal-faulting earthquakes with poorly-constrained source parameters show a similar pattern, but with significant variability, in which the direction of extension appears to be roughly oriented (within $\pm45^o$) with shortening in the adjacent sub-Andes [Cabrera and Sebrier, 1998; Devlin et al., 2012; Jay et al., 2015]. This is the pattern of fault slip expected if deformation within a mountain range is controlled by a contrast in gravitational potential energy between the mountains and its surrounding lowlands [e.g. Dalmayrac and Molnar, 1981; Copley et al., 2009].

In the next section we assess the extent to which potential energy contrasts control the deformation patterns in the Andes and the South American forelands, using well-constrained depths and mechanisms of recent moderate-to-large magnitude earthquakes and existing GPS measurements.

3   Dynamics of Deformation in the Andes

The forces controlling deformation in mountain ranges derive predominantly from (1) relative motions of the bounding plates, (2) gravity acting on density contrasts within the lithosphere, and (3) shear tractions on the base of the overriding lithosphere from the underthrusting of rigid foreland material [Artyushkov, 1973; England and Houseman, 1989; Wdowinski et al., 1989; Copley et al., 2011b]. These forces are in a quasi-static balance with each other, viscous resistance within the ductile lithosphere, and shear resistance on faults within the brittle crust. Different contributions of these forces to the overall balance will be reflected in the pattern of active faulting within the mountain range and its forelands [e.g. Molnar and Lyon-Caen, 1988].

In order to investigate the pattern of active faulting in the Andes and South America, we compiled a catalogue of earthquakes with $M_w > 5$ that have been modelled using body waveforms or P wave first motions from literature sources [Suarez et al., 1983; Chinn and Isacks, 1983; Kadinsky-Cade et al., 1985; Doser, 1987; Assumpção and Suarez, 1988; Vega and Buforn, 1991; Assumpção, 1992; Assumpção and Araujo, 1993; Tavera and Buforn, 2001; Alvarado and Beck, 2006; Emmerson, 2007; Devlin et al., 2012; Scott et al., 2014] and 13 new solutions of our own (shown in Table 2). In addition, we included well-constrained gCMT catalogue events with >80% double-couple moment tensors (as defined in Jackson et al. [2002]) from Dziewonski et al. [1981] and Ekström et al. [2012], with hypocentral depths taken from the catalogue of Engdahl et al. [1998] where available. We have included waveform-modelled earthquakes with M_w 5-5.5 in this compilation, as small earthquakes provide information on the depth extent of faulting in South America [e.g. Assumpção and Suarez, 1988; Assumpção, 1992; Devlin et al., 2012]. Below we describe the pattern of earthquakes, in conjunction with GPS and geomorphological information, and discuss the implications for the forces acting on the Andes and the South American foreland.

3.1   Faulting in the high Andes

Moderate-magnitude earthquakes in the crust of the high Andes are infrequent, but reveal a pattern of predominantly shallow (<10-15 km) normal and strike-slip faulting events [Doser, 1987; Cabrera and Sebrier, 1998; Holtkamp et al., 2011; Devlin et al., 2012; Jay et al., 2015, This study] (Fig. 1a). One exception to this pattern is seen adjacent to the Shira Uplift in south Peru, where oblique reverse faulting in the high Andes was observed in the 1969 Pariahuanca earthquakes [Philip and Megard, 1977; Suarez et al., 1983; Sébrier et al., 1988] (see PH1969; Fig. 1a). Otherwise the majority of the reverse-faulting earthquakes are confined to regions <3000 m elevation [Suarez et al., 1983]. The dependence of faulting mechanism on elevation is typical of mountain ranges in which gravitational

269 potential energy contrasts are an important factor in the forces controlling deformation [Dalmayrac

270 and Molnar, 1981; Copley et al., 2009].

271   Nearly all of the moment release from recent normal-faulting seismicity in the high Andes has been

272 focused in south Peru, whilst the Altiplano and Puna plateau have remained essentially aseismic over

273 the same time period (Fig. 1a). Geomorphological evidence of recent normal faulting is also concen-

274 trated in south Peru, with a number studies identifying metre-high Holocene fault scarps bounding

275 footwall uplifts with a few hundred metres of relief (e.g., Sangararra Fault [Suarez et al., 1983]; Tam-

276 bomachay Fault [Mercier et al., 1992]; Langui-Layo Fault [Benavente et al., 2013]; Parina Fault [This

277 study]). GPS measurements in south Peru limit the cumulative NE-SW extension rate across the

278 high Andes to <5 mm/yr [Kendrick et al., 2001; Villegas-Lanza et al., 2016], which is equivalent to

279 an extensional strain rate $< 2 \times 10^{-8}$ yr^{-1} over the 250 km wide plateau. In contrast, there is no

280 evidence of recent fault-controlled relief in the Bolivian Altiplano, and the shallow crust is inferred to

281 be undeforming [Lamb and Hoke, 1997; Lamb, 2000; Weiss et al., 2016]. Schoenbohm and Strecker

282 [2009] and Zhou et al. [2013] identified a number of putative normal faults within the Puna plateau,

283 however few of those mapped were associated with Holocene surface ruptures, and their estimated slip

284 rates are ≲0.1 mm/yr over the last 0.5 Myrs. Therefore both the recent seismicity and Holocene fault

285 activity imply that the high Andes in south Peru is extending faster than the Altiplano, and possibly

286 faster than the Puna plateau.

287   Differences in strain rate within an isostatically-compensated mountain belt deforming in response

288 to variations in gravitational potential energy are related to differences in elevation [England and

289 Houseman, 1989; D'Agostino et al., 2014]. Crustal thickness estimates [Assumpção et al., 2013] and

290 small (<50 mGal) free-air gravity anomalies in the central Andean plateau both suggest that eleva-

291 tion contrasts on length-scales much larger than the elastic thickness of the Andean crust ($T_e \approx 10$

292 km; Rodríguez Tribaldos et al. [2017]) are predominantly isostatically supported. We isolated the

293 long-wavelength, isostatically-compensated component of the topography by filtering the SRTM 3-

294 arcsecond elevation model [Farr et al., 2007] using a Gaussian filter with a width of 100 km (i.e.

295 $>> T_e$). We find that the recent normal-faulting seismicity in south Peru occurs in a region with the

296 highest smoothed modal elevation in the whole Andes at ∼4500 m, whilst the Altiplano and Puna

297 have lower modal elevations of ∼3800 m and ∼4200 m, respectively (Supp. Fig. 15). Therefore

298 the distribution of recent normal-faulting earthquakes may reflect higher extensional strain rates in

299 the south Peruvian Andes compared to the surrounding mountains due to the region's high relative

300 elevation [e.g. England and Molnar, 1997].

301   The current elevation contrasts influencing the extension rates within the Andes will decay expo-

nentially via flow of the lithosphere on a timescale that is dependent its viscosity. Using the unde-forming Bolivian Altiplano as a reference, we can combine the current maximum extensional strain rate estimate in south Peru, and the current elevation contrast between south Peru and Bolivia, to compute a lower bound on the vertically-averaged Newtonian viscosity of the lithosphere ($\eta > 3 \times 10^{21}$ Pa s; see Appendix A). Assuming the forces acting on the Andes have remained constant in the late Cenozoic, and given our lower bound on the lithosphere viscosity, elevation contrasts between south Peru and the surrounding high Andes will decay with a time constant of at least $\sim$3 Myrs, and are unlikely to have existed much earlier than 10 Ma (see Appendix A).

Strike-slip faulting in the high Andes SW of Cuzco occurs where there are significant along-strike changes in the geometry of the mountain belt and the amount of E-W shortening across the sub-Andes [Kley and Monaldi, 1998] (Fig. 1a). In these isolated regions the Andes does not behave in a two-dimensional manner, as there is a component of strain along-strike, and therefore balancing forces on cross-sections perpendicular to the range front is not appropriate. In this study we focus on regions where extension in the high Andes is parallel to shortening in the adjacent lowlands, so the forces acting on the mountain belt can be treated in two dimensions [Dalmayrac and Molnar, 1981].

3.2 Faulting in the Eastern Cordillera

The Eastern Cordillera defines the area between the low relief, high elevation Andean plateau, and the sub-Andean lowlands (Fig. 1a). Seismicity in this region consists mainly of moderate magnitude reverse and strike-slip faulting, such as the 1998 M_w 6.6 Aiquile earthquake (Funning et al. [2005]; AQ1998 on Fig. 1a). The N-S trending strike-slip rupture of the Aiquile earthquake cut across pre-existing fold-thrust belt structures that were active at $\sim$10 Ma, which indicates the intermediate principal stress has switched from being horizontal at the time in which the fold-thrust belt was active, to vertical [Funning et al., 2005]. This transition in stress state can be accounted for by recent uplift of the Eastern Cordillera [Gubbels et al., 1993; Lamb and Hoke, 1997], during which the forces due to contrasts in gravitational potential energy between the Eastern Cordillera and lowlands increased, such that they now balance the compressional forces transmitted across the sub-Andean ranges into the orogen interior.

North of the Aiquile earthquake epicentre are the Cochabamba and Punata basins (Fig. 1a). These asymmetrical half grabens are filled with post-Miocene sediments and are bounded by large ($\sim$2 km basement relief), NE-SW to E-W striking normal faults [Renner and Velasco, 2000], which are roughly perpendicular to the trend of the range front in the adjacent sub-Andes. Kennan [1993] proposed the basins formed predominantly through range-parallel (NW-SE) extension. Faults with

a similar geometry relative to the orogen front are seen in southern Tibet [e.g. Armijo et al., 1986], and have been attributed to the influence of potential energy contrasts driving radial spreading of the mountains over the adjacent rigid foreland [e.g. Armijo et al., 1986; Copley et al., 2011b] - a feature reflected in the earthquake slip vectors and GPS data from the Andes discussed in the next section. A similar process may account for the range-perpendicular normal faults at Cochabamba and Punata.

3.3   Faulting in the sub-Andes and Peru-Chile Trench

Faulting on the western edge of the Andes is dominated by large-magnitude (M_w >7.5) megathrust earthquakes along the subduction interface between the Andean forearc and the subducting Nazca plate [e.g. Pritchard et al., 2007; Sladen et al., 2010; Lin et al., 2013]. Slip in these megathrust earthquakes is consistently parallel to independent estimates of the orientation of motion between the Nazca plate and the South American shield [Norabuena et al., 1999; DeMets et al., 2010] (Fig. 9a). Although there is evidence of recent reverse faulting within the Andean forearc [e.g. Hall et al., 2012; Benavente et al., 2017], motion on these faults is small compared to slip on the megathrust.

Unlike slip on the megathrust, the slip vectors of earthquakes on the eastern side of the Andes vary significantly along-strike of the mountain belt, with the most conspicuous changes occurring between south Peru and northern Argentina. North of Cuzco a thrust-faulting earthquake in 1998 (Fig. 9a; CZ1998) ruptured the shallow detachment beneath the sub-Andes with a NNE-directed slip vector, which is oriented $\sim 70^o$ anti-clockwise to the Nazca-South America plate motion. Moving south between southern Peru and northern Bolivia the earthquake slip vectors rotate clockwise from NNE to NE (Fig. 9a). Around the Santa Cruz bend the slip vectors rotate rapidly clockwise by a further $\sim$30-40° to point $\sim$E [Chinn and Isacks, 1983; Assumpção and Araujo, 1993; Devlin et al., 2012] (Fig. 9a). Finally, within the sub-Andes of southern Bolivia and northern Argentina the slip vectors remain pointing $\sim$E, which is oriented $\sim$20-30° clockwise relative to Nazca-South America plate motion. The azimuth of the GPS velocity field in the eastern Andes relative to stable South America also rotates clockwise between south Peru and northern Argentina, mirroring the pattern in the earthquake slip vectors [Lamb, 2000; Kendrick et al., 2001; Weiss et al., 2016] (Fig. 9b).

Elastic strain accumulation within the Andes from oblique subduction of the Nazca plate beneath the Andean forearc has been proposed to account for the rotation in the GPS velocities in the eastern Andes [Bevis and Martel, 2001; Bevis et al., 2001]. Although the elastic model of Bevis et al. [2001] fits the azimuth of GPS velocities south of the Santa Cruz bend, in southern Peru and northern Bolivia it consistently underpredicts the northward component of the GPS velocities and earthquake slip vectors (Fig. 9b,e-f; see Table 3). These spatially coherent misfits imply the elastic models are

366   not accurately capturing the orientation of the shortening direction in the sub-Andes. In addition, the

367   assumption in Bevis et al. [2001] that shortening in the sub-Andes can be modelled as slip parallel to

368   Nazca-South America plate motion on a detachment horizon beneath the sub-Andes is not consistent

369   with the observed earthquake slip vectors (Fig. 9a).

370   Alternatively, if the eastern Andes are deforming predominantly in response to horizontal contrasts

371   in gravitational potential energy, then the direction of shortening will be parallel to topographic

372   gradients [e.g. Lamb, 2000; Copley and McKenzie, 2007; Flesch and Kreemer, 2010]. We find that

373   earthquake slip vectors in the sub-Andes are consistently parallel to gradients in the topography

374   (Fig. 9b,d; Table 3). The topographic gradients also match the general pattern of clockwise rotating

375   GPS azimuths between southern Peru and northern Argentina, the northward component of the

376   GPS velocity field near Cuzco, and the rotation of the GPS velocities around the Santa Cruz bend

377   (Fig. 9b). In addition, the azimuth of shortening inferred from the GPS and earthquake slip vectors

378   are statistically better fit by the direction of topographic gradients than an elastic model of strain

379   accumulation on the Peru-Chile subduction zone (see statistical comparison in Table 3).

380   We therefore conclude that the forces resulting from relative plate motion control the orientation

381   of slip on the subduction interface on the western side of the Andes, whilst contrasts in gravitational

382   potential energy between the mountains and forelands control the orientation of slip on faults in the

383   sub-Andean belt in the east [e.g. Assumpção and Araujo, 1993; Lamb, 2000].

3.4   Faulting and Flexure in the South American Forelands

385   Reverse-faulting earthquakes in regions >300 km from the Andes range front (Fig. 9a) indicate

386   the South American foreland crust is breaking in response to compressional forces acting through

387   the lithosphere [Assumpção, 1992]. However, within 300 km of the Andes range front, asymmetrical

388   sedimentary basins [McGroder et al., 2015], long-wavelength, negative free-air gravity anomalies [Lyon-

389   Caen et al., 1985], and shallow extensional faulting overlying compressional faulting at the base of the

390   crust (Fig. 10, 11a,c) imply that the forelands are bending in response to the load of the Andes [e.g.

391   Lyon-Caen et al., 1985; Watts et al., 1995].

392   Evidence for extension in the Andes foreland is limited to a single M_w 6.4 normal-faulting earth-

393   quake beneath the Llanos Basin (Fig. 9a, Table 2). Long-period waveform modelling suggests this

394   earthquake had a centroid depth of 18^{+5}_{-5} km (Fig. 10). The Moho beneath the Llanos Basin is at ∼40-

395   45 km depth [Assumpção et al., 2013], which implies the normal-faulting earthquake reflects extension

396   in the shallow basement of the bending Andean foreland.

397   A M_w 5.3 reverse-faulting earthquake has also been recorded beneath the Llanos Basin (Fig. 11a,c,

Table 2). We determined the centroid depth of this earthquake by comparing synthetic waveforms to stacked vertical-component seismograms recorded at a small-aperture array in Spitsbergen using the method of Craig et al. [2012]. The observed seismograms can be matched by a reverse-faulting earthquake at a centroid depth of 28^{+2}_{-2} km (Supp. Fig. 12). Reverse faulting has also been recorded at 42 km depth beneath the Madre de Dios Basin in central Peru [Assumpção, 1992] (Fig. 11a), indicating that the bending portion of the South American lithosphere probably remains in compression from 26-30 km to at least the base of the crust (Fig. 11c).

Observations of shallow extensional earthquakes overlying deeper compressional earthquakes within the continental lithosphere are rare, and have only been documented previously in the forelands of Tibet [Molnar et al., 1977; Ni and Barazangi, 1984; Maggi et al., 2000b], and possibly in small ($M_w < 4.5$) earthquakes recorded in the forelands of the Alborz mountains [Nemati et al., 2013]. The pattern of earthquakes in both the Tibetan and Andean forelands implies that flexural stresses are large enough to break faults in underthrusting continental lithosphere.

If we treat the flexed South American forelands as a thin bending plate, the magnitude of the flexural stresses can be related to the curvature of the plate, and a factor that is governed by the assumed rheology of the plate [Turcotte and Schubert, 2002]:

$$\sigma_f = \frac{ET}{2(1-\nu^2)} \frac{d^2w}{dx^2}, \tag{1}$$

where σ_f is the maximum flexural stress, E is Young's modulus, ν is Poisson's ratio, T is either the effective elastic thickness of the plate (in a purely elastic model), or the thickness of the aseismic elastic core (in an elastic-plastic model with a constant yield stress), and $\frac{d^2w}{dx^2}$ is the curvature of the plate.

To estimate the curvature and effective elastic thickness (T_e) of the South American foreland beneath the Llanos Basin we followed the method of McKenzie and Fairhead [1997] and modelled the free-air gravity anomalies produced by flexure of an elastic plate overlying an inviscid mantle half-space, with the plate overlain by low-density sediments. We find the gravity anomalies in the Llanos Basin can be fit by a plate model with a T_e of at least 20 km (Fig. 11b,d). However the misfit between the modelled and observed gravity field at large T_e values increases by only a fraction of the minimum misfit, suggesting our estimate represents only a lower bound (Fig. 11d). Our results are consistent with those of Pérez-Gussinyé et al. [2007] and McKenzie et al. [2014], who find that T_e in the Andean forelands is between 15 and 25 km, which is significantly smaller than the seismogenic thickness (40-45 km; Assumpção and Suarez [1988]; Emmerson [2007]). Stewart and Watts [1997] found a larger T_e of 50 km in the Llanos Basin, however their study relied on modelling individual

429 Bouguer gravity profiles, and fixed the location of the plate break, which has typically been shown to

430 overestimate T_e (see discussion in Jackson et al. [2008]).

431   The maximum curvature of the plate in the best-fit model is $\sim 4 \times 10^{-7}$ m^{-1} (Fig. 11b). Assuming

432 the plate has a Young's modulus of 10^{11} Pa and Poisson's ratio of 0.25, Eq. 1 would give a maximum

433 extensional flexural stress of $\sim$530 MPa. This stress will subject normal faults in the top half of the

434 bending plate, which dip at $\sim$35° relative to the plate surface (e.g. Llanos Basin earthquake; Fig. 10),

435 to average shear stresses $\lesssim$130 MPa. For a simple elastic-plastic plate with a constant yield stress

436 [Turcotte and Schubert, 2002], and an elastic core <17 km thick (based on the depth distribution of

437 earthquakes; Fig. 11c), the maximum flexural stresses are reduced to $\lesssim$360 MPa, and the average

438 shear stresses on faults to $\lesssim$90 MPa. Further earthquakes could reduce our estimate of the elastic core

439 thickness, and therefore reduce our stress estimates.

440   In contrast to the bending portion of the South American forelands, the normal faulting in the

441 high Andes and reverse faulting throughout the foreland crust >300 km from the Andes range front

442 appears to be primarily controlled by the forces per unit length exerted between the mountains and

443 South America due to lateral contrasts in gravitational potential energy. In the next section we present

444 calculations to estimate the forces acting through the South American forelands, and place an upper

445 bound on the frictional strength of faults within the foreland crust.

446 4   Strength of the South American Forelands

447 4.1   Forces acting between the Andes and the South American Forelands

448 The forces per unit length due to potential energy contrasts between a mountain belt and its adjacent

449 foreland can be calculated from lateral contrasts in the density structure between the two regions [e.g.

450 Artyushkov, 1973; Dalmayrac and Molnar, 1981]. We build on the original calculations of Dalmayrac

451 and Molnar [1981] by using new estimates of the temperature structure, chemical composition and

452 thickness of the crust and mantle lithosphere to refine the estimates of the forces per unit length acting

453 between the Andes and South America, following a method similar to that of Copley and Woodcock

454 [2016]. A summary of the parameters used in the calculations is provided in Table 4, the choice of

455 parameters and the methodology behind the calculations are described in detail in the Supplementary

456 Information, and an overview of the calculations is provided below.

457   We compute the horizontal force per unit length exerted between the Andes and South America by

458 integrating the differences in vertical normal stress between the mountains and forelands ($\Delta\sigma_{zz}$) from

459 the surface to the depth of isostatic compensation ($F_{GPE} = \int \Delta\sigma_{zz} dz$) [Dalmayrac and Molnar, 1981].

460   By varying the temperature structure, composition and thickness of the crust and mantle lithosphere

461   we find that the horizontal force per unit length acting between the Andes and the South American

462   foreland is $\sim 4 - 8 \times 10^{12}$ N per metre along-strike of the mountain range (Fig. 12a). This result is

463   consistent with previous studies [Dalmayrac and Molnar, 1981; Froidevaux and Isacks, 1984; Molnar

464   and Lyon-Caen, 1988; Meijer et al., 1997; Husson and Ricard, 2004; Oncken et al., 2012], but includes

465   a larger range of uncertainty that is related to the range in possible lithospheric mantle thickness

466   and the poorly-constrained temperature structure beneath the Andes that had previously not been

467   considered (e.g. Molnar and Stock [2009]; see Supplementary Information).

4.2   Fault Friction in the South American Forelands

469   The forces per unit length due to contrasts in gravitational potential energy acting between the Andes

470   and South America must be balanced by the forces acting through the South American foreland

471   lithosphere. Reverse-faulting earthquakes throughout the non-bending regions of the South American

472   forelands, outboard of the region of normal faulting in the high Andes, implies that these forces are

473   large enough to exceed the static frictional strength of faults in the forelands (as is also seen in the

474   Tibetan foreland in India [Copley et al., 2011a]).

475   We construct one-dimensional yield stress profiles that reflect the stress state as a function of

476   depth to the east of the sub-Andean belt in the bending and non-bending part of the South American

477   foreland, and integrate the yield stress to estimate the force the lithosphere can support in these

478   different regions ($F_{fl} = \int \Delta\sigma_{xx} dz$) [e.g. Goetze and Evans, 1979; Brace and Kohlstedt, 1980]. We

479   then compare these estimates of foreland strength (F_{fl}) with our estimate of the forces per unit length

480   acting through the foreland lithosphere (F_{GPE}) to place an upper bound on the frictional strength of

481   faults in the South American crust.

482   Faults modulate the strength of the brittle portion of the lithosphere, which can be expressed as

483   a function of the effective coefficient of friction [Turcotte and Schubert, 2002]:

$$\Delta\sigma_{xx} = \frac{2\mu'\rho g z}{\pm \sin 2\theta - \mu'(1 + \cos 2\theta)}, \tag{2}$$

484   where μ' is the effective coefficient of friction, ρ is the average crustal density, g is the acceleration

485   due to gravity, z is the depth, θ is the angle of the fault from the vertical, and $\Delta\sigma_{xx}$ is the horizontal

486   differential stress the faults can support. The $\pm \sin 2\theta$ term is negative if the faults are extensional

487   and positive if they are compressional.

488   Within 300 km of the Andes range front the top $\sim$5-8 km of the crust consists of sedimentary

489   basins that contain no earthquakes [McGroder et al., 2015] (Fig. 11c), and that we assume support

negligible force. Outside of the sedimentary basins, earthquakes occur throughout the thickness of the crust [Assumpção and Suarez, 1988; Assumpção, 1992, This study] (Fig. 11c), and therefore we assume that stresses in the basement crust are supported by frictional resistance on faults given by Eq. 2.

Earthquakes are absent from the foreland mantle lithosphere beneath South America (Fig. 11c). A similar pattern of seismicity has been observed in other continental shield regions (e.g., India [Maggi et al., 2000a; Craig et al., 2012]; Africa [Craig et al., 2011]; Sibera [Sloan et al., 2011]), and has been attributed to the temperature of the continental lithospheric mantle generally exceeding ∼600 ^{o}C [McKenzie et al., 2005; Jackson et al., 2008], above which mantle rocks deform predominantly by ductile creep mechanisms. We use a dry olivine dislocation creep law to estimate an upper bound on the ductile strength of the lithospheric mantle [Karato and Wu, 1993]:

$$\Delta\sigma_{xx} = S\dot{\varepsilon}_r^{\frac{1}{n}} A^{-\frac{1}{n}} \exp\left(\frac{E + pV}{nRT}\right),\tag{3}$$

where S is the shear modulus, $\dot{\varepsilon}_r$ is the reference strain rate which we calculate by imposing stress continuity at the brittle-ductile transition, A is a constant, n is the power-law exponent, E is the active energy, p is the pressure, V is the activation volume, R is the gas constant and T is the temperature (parameters given in Table 4). It has been suggested that upper mantle rocks at temperatures less than 750 ^{o}C may deform by low temperature plasticity (Peierl's creep) [e.g. Mei et al., 2010; England and Molnar, 2015], which would support smaller differential stresses for a given strain rate than dislocation creep. If so, the forces supported by faults in the lithosphere would be slightly higher than our estimates. However, we also perform calculations in which the lithospheric mantle supports no stress, providing an absolute upper bound on the possible forces acting on faults within the crust.

By performing a parameter sweep through the range of variables that control the shape of the yield stress envelope with depth (effective coefficient of friction, seismogenic thickness, fault dips, neutral fibre depth, elastic core thickness and sediment thickness; parameter range given in Table 4), we can define an envelope of the force that the lithosphere can support as a function of the effective coefficient of friction on crustal faults (Fig. 12b-d). We find that, in order to break in response to the $4 - 8 \times 10^{12}$ Nm^{-1} that we estimate above to be acting through the forelands, faults in the bending and non-bending regions of South America must have an effective coefficient of friction $\lesssim 0.15$. If the lithospheric mantle supported no stress at all, which would resolve all the force acting through the foreland lithosphere onto faults in the crust, the maximum possible effective coefficient of friction would be 0.2 (Fig. 12d). If the faults in the South American crust were stronger, then they would not be breaking in earthquakes as a result of the forces per unit length generated by gravitational

521 potential energy contrasts between the Andes its bounding lowlands.

522 Our low estimate of the effective coefficient of friction on faults in the South American crust is

523 consistent with studies that infer elevated pore-fluid pressures [Sibson, 2004] or foliated, phyllosilicate-

524 rich fault gouges [Imber et al., 2008; Collettini et al., 2009] may reduce the frictional strength of active

525 fault zones relative to laboratory estimates of bare rock friction [Byerlee, 1978]. Notably, faults in the

526 forelands with an effective coefficient of friction $\lesssim$0.2, that dip between 30-50° (Fig. 10), will support

527 depth-averaged shear stresses $\lesssim$110 MPa, which is similar to our estimate derived from the analysis

528 of flexure in Section 3.4.

529 Previous studies of the dynamics in South America based on a 'thin-viscous sheet' approximation

530 [England and McKenzie, 1982] imply the deviatoric stresses acting to deform the lithosphere are rel-

531 atively small ($\sigma'_{xx} \approx 10 - 30$ MPa; Dalmayrac and Molnar [1981]; Lamb [2000]; Flesch and Kreemer

532 [2010]). Our estimates of the horizontal forces acting through the $\sim$100-150 km thick foreland litho-

533 sphere imply deviatoric stresses on the order of a few 10's of megapascal, which is consistent with the

534 aforementioned studies.

5   Discussion

5.1   Late Miocene Change in the Dynamics of the Andes

537 The Parina earthquake and the surrounding seismicity demonstrates that the high Andes in south

538 Peru are currently undergoing extension. A M_w 6.2-6.8 normal-faulting earthquake recorded in 1946

539 at Ancash in central Peru [Heim, 1949; Doser, 1987], and a M_w 5.4 earthquake in the Bolivian Altiplano

540 (Fig. 9a), suggest that a wider region of the high central Andes is also currently extending.

541 Geological evidence of recent normal faulting in the high Andes is widespread, with examples of

542 extensional structures in regions at elevations >3000 m from central and southern Peru [Dalmayrac

543 and Molnar, 1981; Sébrier et al., 1985, 1988; Mercier et al., 1992; Veloza et al., 2012; Benavente et al.,

544 2013; Kar et al., 2016], northern Chile [Tibaldi et al., 2009; Tibaldi and Bonali, 2018], southern and

545 northern Bolivia [Mercier, 1981; Lavenu et al., 2000; Lamb, 2000] and northern Argentina [Cladouhos

546 et al., 1994; Marrett et al., 1994; Lavenu et al., 2000; Schoenbohm and Strecker, 2009] (Figs. 9a,

547 13). The timing of motion on these normal faults can be bracketed by cross-cutting relationships with

548 the extensive volcanics erupted continually throughout the Andean orogeny. The earliest evidence

549 of extension in the high Andes varies from $\sim$7-9 Ma in southern Bolivia and northern Argentina

550 [Cladouhos et al., 1994; Montero Lopez et al., 2010], to $\sim$5-7 Ma in central and southern Peru [Mercier,

551 1981; Cabrera et al., 1991; Giovanni et al., 2010; Benavente et al., 2013; Kar et al., 2016], and follows

an earlier period of widespread ∼E-W compression [Megard et al., 1984; Sébrier et al., 1985; Ellison et al., 1989; Mercier et al., 1992; Elger et al., 2005; Tibaldi et al., 2009]. Within the central Bolivian Altiplano folded Pliocene strata in the Corque-Corococo basin imply the shallow crust experienced waning shortening between ∼10 Ma and 3 Ma [Hoke and Lamb, 2007; Lamb, 2011], and has recently been undeforming [Elger et al., 2005; Lamb, 2011] (Fig. 13). The transition in stress state within parts of the high Andes of Peru and Argentina from dominantly compression to extension since the late Miocene, and the slowing of shortening in the lower elevation Bolivian Altiplano, is indicative of an orogen-wide change in the force balance.

Over the same time interval (∼5-16 Ma) paleoelevation proxies infer there has been 1-3 km of surface uplift in the high Andes of southern Peru and Bolivia [e.g. Garzione et al., 2006; Barke and Lamb, 2006; Garzione et al., 2008; Saylor and Horton, 2014; Kar et al., 2016; Garzione et al., 2017]. In addition, structural reconstructions and proxies for rock exhumation have recorded a distinct shift in the locus, rate and style of deformation on the eastern edge of the Andes. Prior to ∼10 Ma deformation in the eastern Andes consisted of slow, long-wavelength folding and bivergent reverse faulting of lower Paleozoic strata and metamorphic basement focused in the Eastern Cordillera [Kley, 1996; Oncken et al., 2006; Espurt et al., 2008; Perez et al., 2016]. Since ∼10 Ma shortening rates in the eastern Andes have increased significantly [Oncken et al., 2006, 2012], and the locus of deformation has propagated rapidly eastwards into the sub-Andean lowlands [Gubbels et al., 1993; Echavarria et al., 2003; Gautheron et al., 2013; Perez et al., 2016], where Mesozoic foreland basin fill is exposed in short-wavelength (∼5-10 km) anticlinal ridges bound by closely-spaced splay faults above shallow-dipping detachments [e.g. Kley, 1996].

Any attempt to explain the onset of normal faulting in the high Andes ∼5-9 Ma must also account for the observations described above (summarised in Fig. 13).

5.2   Possible causes of Normal Faulting in the High Andes

Removal of dense lower crust and mantle lithosphere (delamination) could increase the gravitational potential energy contrast between the mountains and forelands, leading to extension in the highest parts of the Andes [e.g. England and Houseman, 1989; Garzione et al., 2006]. Delamination has been proposed to account for the composition, timing and volume of ignimbrite and mafic volcanism [e.g. Kay and Mahlburg Kay, 1993], the Helium isotope ratios of hydrothermal fluids [Hoke and Lamb, 2007], and the possible rapid Miocene-recent uplift of the central Andes [e.g. Garzione et al., 2006, 2008, 2017]. However the relationship between crustal thickening and uplift rates in the Bolivian Altiplano [Lamb, 2011, 2016], and inconsistent seismic models that independently infer thick, thin

584 and variable thickness lithospheric mantle beneath the Andes [e.g. Whitman et al., 1992; Beck and

585 Zandt, 2002; Phillips et al., 2012; Priestley and McKenzie, 2013; Ward et al., 2016], call into question

586 whether delamination beneath the central Andes coeval with extension actually occurred.

587   Another mechanism that has been suggested to induce extension in the high Andes is dynamic

588 uplift of central Peru due to the subduction of the buoyant Nazca Ridge [Mercier et al., 1992; McNulty

589 and Farber, 2002; Margirier et al., 2017]. Our observations of normal-faulting earthquakes 400 km

590 south of the Nazca Ridge (Fig. 13), in addition to evidence of extensional faulting within elevated

591 regions across the Andes, irrespective of the nature of the underlying subducting plate (e.g. flat slab

592 segments; Gutscher et al. [2000]), indicates that the pattern of extension is probably unrelated to

593 subduction processes. Similarly, the timing and distribution of normal-faulting earthquakes appears

594 unrelated to the subduction zone earthquake cycle.

595   Changes in the horizontal forces supported by the lithosphere bounding the Andes could also cause

596 the onset of extension in the high mountains. The switch in shortening style in the eastern Andes

597 ∼10 Ma from steeply-dipping reverse faulting to shallow-dipping, weak detachment faulting within

598 the foreland basin sediments could have reduced the shear stresses acting on the base of the overriding

599 Andean lithosphere, leading to the onset of extension in the high mountains, a reduction in shortening

600 rates in the Bolivia Altiplano, and enhanced shortening rates within the lowlands of the eastern Andes

601 [Babeyko et al., 2006; Oncken et al., 2012]. We explore this possibility further below.

602   The spatial relationship between normal faulting and foreland deformation style is suggestive of

603 a causal link. Late Miocene to recent normal faulting in the high Andes correlates along-strike with

604 regions of wide thin-skinned fold-thrust belts [Kley et al., 1999], the locations of low-angle thrust-

605 faulting earthquakes [Devlin et al., 2012], and thick Paleozoic and Mesozoic sediments in the adjacent

606 foreland basin [McGroder et al., 2015] (Supp. Fig. 17). In contrast, in regions where the active

607 foreland deformation style comprises predominantly reverse faulting emerging at the front of steep

608 topography (e.g., Shira Uplift of Peru [Kley et al., 1999]; Sierra Pampeanas of Argentina [Alvarado

609 and Beck, 2006]; Oriente of Ecuador [Kley et al., 1999]) the style of faulting in the adjacent high

610 Andes is dominantly compressional (Fig. 13). A particularly clear example is seen in the high Andes

611 adjacent the Shira Uplift in Peru, where faulting in the 1969 M_w 6.1 and 6.6 Pariahuanca earthquakes

612 was compressional [Philip and Megard, 1977; Suarez et al., 1983; Sébrier et al., 1988] (see PH1969; Fig.

613 1a). However, both north and south of Shira, where the foreland has wide and active thin-skinned

614 fold-thrust belts [Kley et al., 1999], the high Andes are undergoing extension [Heim, 1949; Doser, 1987;

615 Devlin et al., 2012, This study] (Figs. 1a, 13).

616   It is possible that once the Andes reached the maximum elevation that could be supported by

the steep reverse faults bounding the eastern edge of the range ∼5-9 Ma (Fig. 14a), the plateau widened as faulting propagated eastward into the thick foreland basin sediments that had accumulated concomitant with uplift [e.g. Molnar and Lyon-Caen, 1989]. Foreland deformation then became focused along weak detachment layers above the sediment-basement interface, and the rigid Brazilian Shield was underthrust further beneath the mountains [Kley et al., 1999; Lamb, 2006, 2011; Phillips et al., 2012] (Fig. 14b). As the force balance in the eastern Andes is controlled by the horizontal shear stresses acting on the base of the overriding lithosphere [e.g. Huppert, 1982; McKenzie et al., 2000], reducing the frictional shear stresses along the nose of the foreland fold-thrust belt would be balanced by extensional viscous strain and normal-faulting within the overriding lithosphere (see the calculations presented in Craig and Copley [2018]). In addition to the onset of extension within the high Andes, the ductile lower crust in front of the indenting foreland lithosphere would shorten and thicken [Barke and Lamb, 2006; Lamb, 2011], balancing the flux of material out over the South American foreland. At the surface this would be visible as the mountains spreading eastwards over the forelands coeval with extension in the highest parts of the range - a feature that is reflected in the current pattern of crustal deformation (Fig. 14b).

Previous models of the dynamics in the Andes have typically treated the mountain belt in terms of vertically-averaged stress and strain (i.e. a thin-viscous sheet) [Lamb, 2000; Liu et al., 2002; Flesch and Kreemer, 2010]. Under the thin-sheet assumptions shear stresses on the base of the lithosphere are negligible, and weakening the eastern side of the Andes would simply enhance shortening strain rates in that region to maintain the horizontal force balance. Our insight is to include the influence of the underthrust South American foreland, which will partition the force balance with depth beneath the mountains (i.e. F_u and F_p in Fig. 14b), and accommodate extension of the upper crust in the Andes.

5.3   Implications for the Evolution of Orogenic Belts

A generalisation of the model presented above, in which mountain belts weaken their forelands through syn-orogenic sedimentation, is likely to be broadly applicable to orogen evolution [e.g. Lamb, 2006; Copley et al., 2009; Reynolds et al., 2015]. For example, the deformation style within the forelands of the Tien Shan is dependent on the thickness of the sedimentary basins bounding the basement uplifts [Kober et al., 2013]. Along range fronts adjacent to thick (10-15 km) sedimentary basins, anticlinal folding above buried reverse faults extends ∼30 km into the piedmont, suggesting shortening is beginning to step out from the basement-cutting faults onto low-angle detachments in the foreland stratigraphy [Avouac and Tapponnier, 1993]. However in regions where the basin sediments are thin

649 (<5 km), faulting has remained focused along steep reverse faults that outcrop at the range front

650 [Avouac and Tapponnier, 1993]. Quaternary-Holocene fault scarps [Thompson et al., 2002], and

651 recent seismicity [Sloan et al., 2011], indicate shortening is still active throughout the interior of the

652 Tien Shan, which probably reflects the fact that the steep, basement-cutting reverse faults can still

653 transmit enough force into the mountain range to overcome the gravitational body forces acting to

654 drive extension.

655   In contrast, within the southern forelands of Tibet, deformation has focused onto detachment hori-

656 zons that cut through the foreland stratigraphy, and emerge ∼80-100 km south of the high Himalaya

657 [Lavé and Avouac, 2000]. Geomorphic evidence for uplift [Lavé and Avouac, 2000], and the pattern of

658 shortening reflected in thrust-faulting earthquakes [Molnar and Lyon-Caen, 1989], is confined to the

659 edges of the mountain belt, whilst extension occurs simultaneously in the high plateau [Armijo et al.,

660 1986]. Tibet therefore represents a more advanced stage of orogen evolution, similar to the Altiplano,

661 in which shortening has localised into the sediments of the foreland basin.

662   In summary, we suggest the mechanical properties and nature of the stratigraphy within foreland

663 basins may be important in modulating the forces transmitted across the forelands, the stress field

664 within orogenic belts, and the elevation that mountain belts can attain.

6   Conclusions

666 We have used InSAR, seismology and field observations to determine a source model for the 1st

667 December 2016 M_w 6.1 Parina earthquake in the high Andes of south Peru. Our models suggest

668 the earthquake was generated by slip on a shallow-crustal normal fault that accommodated NE-SW

669 extension. Slip in the earthquake was partially buried, and around 15-35% of the shallow co-seismic

670 slip deficit was accommodated by post-seismic afterslip within 6 months of the event.

671   We find that extension in the Parina earthquake was parallel to the direction of shortening across

672 the adjacent sub-Andean lowlands, and that the pattern of shortening in the sub-Andes is controlled

673 by gradients in the topography. In addition, we find that normal faulting earthquakes and geomorpho-

674 logical evidence for recent extensional deformation within the Andes concentrates in regions with the

675 highest elevations. All of these observations imply forces generated by lateral contrasts in gravitational

676 potential energy are important in controlling deformation in the high Andes and its eastern lowlands.

677   We calculated the force per unit length exerted between the Andes and its forelands due to varia-

678 tions in gravitational potential energy to be $4 - 8 \times 10^{12}$ N per metre along-strike. Earthquakes within

679 the South American forelands > 300 km from the Andes range front suggest the crust is breaking

680 in response to these forces, which requires the foreland faults have an effective coefficient of friction

$\lesssim 0.2$, and support average shear stresses $\lesssim 110$ MPa. Within 300 km of the Andes range front the extensional stresses in the shallow, bending portion of the foreland lithosphere are large enough to break faults. Using estimates of the curvature of the South American foreland and the rheology of the lithosphere, we can place similar upper bounds on the magnitude of the shear stresses acting on faults in the bending South American forelands ($\lesssim$90-130 MPa).

Normal faulting in the Andes began in the late Miocene following a period of orogen-wide compression, which reflects a change in the force balance across the mountain range. The coeval onset of low-angle detachment faulting in the sub-Andes and extension in the high Andes $\sim$5-9 Ma, in addition to the spatial correlation between extension in the high mountains and wide thin-skinned fold-thrust belts in the adjacent sub-Andes, suggests weak detachment horizons in the foreland stratigraphy could have reduced the forces transmitted into the mountain belt by shear stresses along the base of the overriding lithosphere, leading to the onset of extension in the high mountains. The dynamics and evolution of the Andes, and we suggest most mountain ranges, may therefore be influenced by the mechanical strength of the stratigraphy in its forelands.

7 Acknowledgements

A   Derivation of Relaxation Time-scale for Elevation Contrasts

Consider two uniform columns of isostatically-compensated lithosphere with constant viscosity η, thickness L_1 and L_2 separated by a height difference h. A simple 1-D horizontal force balance between body forces acting on the columns and viscous resistance, assuming shear stresses on horizontal planes are negligible, is given by [D'Agostino et al., 2012; England and Molnar, 2015]:

$$2L\frac{\partial \tau_{xx}}{\partial x} = \frac{\partial \Gamma}{\partial x}, \tag{A.1}$$

where τ_{xx} is the vertically-averaged deviatoric stress acting on the columns and Γ is the gravitational potential energy per unit area of the column. Assuming the columns are Newtonian (i.e. $\tau_{xx} = 2\eta\dot{\varepsilon}_{xx}$) and integrating Eq. A.1 yields:

$$\dot{\varepsilon}_{xx} = \frac{\Delta\Gamma}{4\eta L}, \tag{A.2}$$

where $\Delta\Gamma$ is the difference in gravitational potential energy between a column with zero strain rate (i.e. undeforming) and the column in question. For Airy isostatic compensation $\Delta\Gamma$ can be expressed approximately as γh, where γ is some constant scaling elevation contrasts between the deforming and undeforming area (h) to the gravitational potential energy contrast [D'Agostino et al., 2014]. In our case, one of the columns (i.e. the Bolivian Altiplano) is assumed to be undeforming, therefore the strain rate in the deforming column (i.e. south Peru) is proportional to the elevation of that column above the undeforming column.

Assuming incompressibility of each column of lithosphere, and the conservation of area in cross section, horizontal extension of the column at strain rate $\dot{\varepsilon}_{xx}$ in response to horizontal contrasts in gravitational potential energy would lead to thinning given by:

$$\frac{dL}{dt} = -\dot{\varepsilon}_{xx}L = -\frac{\gamma h}{4\eta} \tag{A.3}$$

The difference in the thickness of the higher, deforming column and the undeforming column $L_1 - L_2$ is related to the elevation contrast through an isostatic relationship $L_1 - L_2 = h\frac{\rho_m}{\Delta\rho}$, where $\Delta\rho$ is the density difference between mantle and crust. Substituting the isostatic relationship into Eq. A.3 gives the following first order ODE:

$$\frac{dh}{dt} + \frac{\gamma\Delta\rho}{4\eta\rho_m}h = 0 \tag{A.4}$$

This equation has a particular solution of the form $h(t) = h_0 \exp(-t/t_r)$, where $t_r = \frac{4\eta\rho_m}{\gamma\Delta\rho}$ and h_0

is the starting elevation of the deforming column above the undeforming column. Therefore the elevation contrasts between deforming and undeforming regions within a mountain belt will decay exponentially at a rate dependent on the density contrasts between crust and mantle and the viscosity of the lithosphere.

Given the maximum possible extensional strain rate in south Peru as 2×10^{-8} yr^{-1}, γ as 1.5×10^{9} N m^{-1} per metre of elevation (see calculations in Section 4.1), h as 700 m and L as 150 km, we can use Eq. A.2 to compute a lower bound on the viscosity, which is $\eta > 3 \times 10^{21}$ Pa s. Our lower bound on the viscosity of the lithosphere is consistent with Lamb [2000], who infer the average viscosity of the lithosphere in the Andes is $5-10 \times 10^{21}$ Pa s. Taking ρ_m and ρ_c as 3300 and 2800 kg m^{-3}, respectively, the relaxation time-scale of elevation contrasts would be $t_r > 3$ Myrs. Under constant force boundary conditions, after >9 Myrs around 95% of the elevation contrasts between south Peru and Bolivia will have been removed via extension. We therefore infer that the elevation contrasts between south Peru and Bolivia are unlikely to have existed for >>10 Myrs.

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Tables

Method	Strike	Dip	Rake	L, km	W, km	$\bar{u}$, m	M_0, Nm
gCMT	148°	43°	274°	-	-	$0.4^{\ddagger}$	2.2×10^{18}
USGS	134°	35°	263°	-	-	$0.5^{\ddagger}$	2.3×10^{18}
Ruptures	$130\text{-}145^o$	-	-	12	-	0.1-0.3	$0.4\text{-}1.3 \times 10^{18}$ †
BWF	144° ($110\text{-}160^o$)	39° ($30\text{-}55^o$)	276° ($250\text{-}300^o$)	-	-	$0.2^{\ddagger}$	$0.8\text{-}1.4 \times 10^{18}$
InSAR	135° ($120\text{-}145^o$)	40° ($35\text{-}55^o$)	250° ($250\text{-}270^o$)	13	12	0.5	2.4×10^{18}

Table 1: Comparison of the source parameters determined from the surface ruptures, seismology and InSAR observations of the Parina earthquake. BWF indicates the long-period body-waveform solution. L and W are the along-strike length and down-dip width of the rupture patch, respectively. Moment is calculated using $M_0 = \mu A \bar{u}$, where μ is the shear modulus (30 GPa), A is the fault area and $\bar{u}$ is the average fault slip. Values marked (†) have been calculated assuming the fault rupture is square (i.e. $L \approx W$), and values marked (‡) have been calculated assuming the fault is 13 km long and square. For the body-waveform and InSAR model the range of acceptable fault geometry parameters is given in brackets next to the best-fit value.

Date	Time (GMT)	Lon, ^{o}	Lat, ^{o}	s/d/r	M_w	z, km	% DC	Mech
1993 07 22	04:57:15	-71.18	6.49	024/35/-93	6.4	18	-	, b
1994 05 31	17:42:04	-72.04	7.39	138/38/78	6.1	6	-	, b
1995 01 19	15:05:11	-72.94	5.04	214/57/97	6.5	21	-	, b
1995 10 03	01:51:37	-77.53	-2.55	216/40/103	6.8	19	-	, b
2001 06 29	22:30:21	-70.35	-15.37	110/58/-107	5.4	6	82	, a
2001 08 12	00:16:32	-70.03	-16.76	130/45/-108	5.1	6	87	, a
2006 08 09	22:36:11	-70.75	-14.45	082/51/-93	5.2	8	87	, a
2009 11 06	08:50:02	-64.46	-23.44	224/30/131	5.4	20	92	, a
2010 02 27	15:45:43	-65.76	-24.85	162/25/97	6.2	9	99	, B
2011 10 06	11:12:37	-64.22	-24.18	176/56/91	5.9	15	-	, b
2014 02 17	09:41:40	-71.58	6.64	008/67/117	5.3	28	70	, a
2015 11 29	18:52:59	-64.65	-23.56	222/29/129	5.8	4	98	, B
2017 04 18	17:49:59	-75.31	-2.73	028/42/88	5.9	15	88	, B

Table 2: Waveform modelling results for earthquakes in South America (not including Parina, see Table 1). % DC is the percentage double-couple of the gCMT moment tensor (as defined in Jackson et al. [2002]). Mechanisms with reference *a* are modelled by fitting vertical-component seismograms with synthetic waveforms following the methods of Maggi et al. [2000b] and Craig et al. [2012] based on Chapman [1978] and Kennett et al. [1995] (see Supplementary Information), whilst those with reference *b* are modelled using long-period body-waveform inversion. For events labelled *B* we fixed the earthquake mechanism and inverted for depth, source-time function and moment only.

	Test	Test 1: all		Test 2: N. of Santa Cruz	
		Megathrust	**Topography**	**Megathrust**	**Topography**
SV	RMS, deg	29	9	42	7
GPS	RMS, deg	33	25	42	31

Table 3: Results of a statistical comparison between the azimuth of the GPS velocity field and slip vectors (SV) in the eastern Andes, with the elastic model of strain accumulation on the Peru-Chile megathrust from Bevis et al. [2001] and the azimuth of topographic gradients [e.g. Copley and McKenzie, 2007]. A map view of the measurement sites and their azimuths is shown in Fig. 9b. Topographic gradients were calculated from a 300 km Gaussian filtered version of the SRTM 3-arcsecond elevation model [Farr et al., 2007]. Filter widths between 100-400 km give similar results. RMS is the root-mean square misfit between the observed and modelled azimuth. Two tests are shown, one in which all GPS and slip vector measurements shown in Fig. 9b are used to compute the misfits (Test 1), and one in which only measurements from north of Santa Cruz are used to compute the misfits (Test 2). In all cases the azimuth of the topographic gradients fit the observations better than the model of elastic strain accumulation of Bevis et al. [2001].

Variable	Value	Source
Crustal thickness, lowlands	35-45 km	[Assumpção et al., 2013]
Crustal thickness, Andes	65-80 km	[Assumpção et al., 2013]
Lithosphere thickness, lowlands	120-140 km	[McKenzie and Priestley, 2008]
Lithosphere thickness, Andes	180-220 km or 0 km	[see Supplementary Information]
Moho temperature, lowlands	600-700 ^oC	[see Supplementary Information]
Moho temperature, Andes	700-1000 ^oC	[Weber et al., 2002]
Mantle potential temperature	1315 ^oC	[McKenzie et al., 2005]
Thermal expansivity†, crust	3×10^{-5} K^{-1}	[Turcotte and Schubert, 2002]
Thermal expansivity†, mantle	$3 - 4.5 \times 10^{-5}$ K^{-1}	[Bouhifd et al., 1996]
Asthenosphere density (ρ_a, 0 ^oC)	3330 kg m^{-3}	[Turcotte and Schubert, 2002]
Lithospheric Mantle density (0 ^oC)	$\rho_a - 50$ kg m^{-3}	[Lucassen et al., 2005; McKenzie and Priestley, 2016]
Crustal density (0 ^oC)	2800 kg m^{-3}	[Lucassen et al., 1999]
Foreland fault dips	30-50°	[This study]
Activation energy (E)	540 kJ mol^{-1}	[Karato and Wu, 1993]
Activation volume (V)	20 cm^3 mol^{-1}	[Karato and Wu, 1993]
Stress exponent (n)	3	[Karato and Wu, 1993]
Seismogenic thickness	40-45 km	[Assumpção and Suarez, 1988]
Foreland sediment thickness	5-8 km	[McGroder et al., 2015]
Neutral fibre depth	13-28 km	[see Section 3.5]
Elastic core thickness‡	1-20 km	[see caption]

Table 4: Parameters used in the calculations for the force exerted between the mountain range and forelands, and for estimating the frictional strength of faults in the South American forelands. (†) This is the linear coefficient of thermal expansion. (‡) The elastic core thickness is the extent of the region either side of the neutral fibre in the bending South American foreland over which the stress field changes from being extensional to compressional.
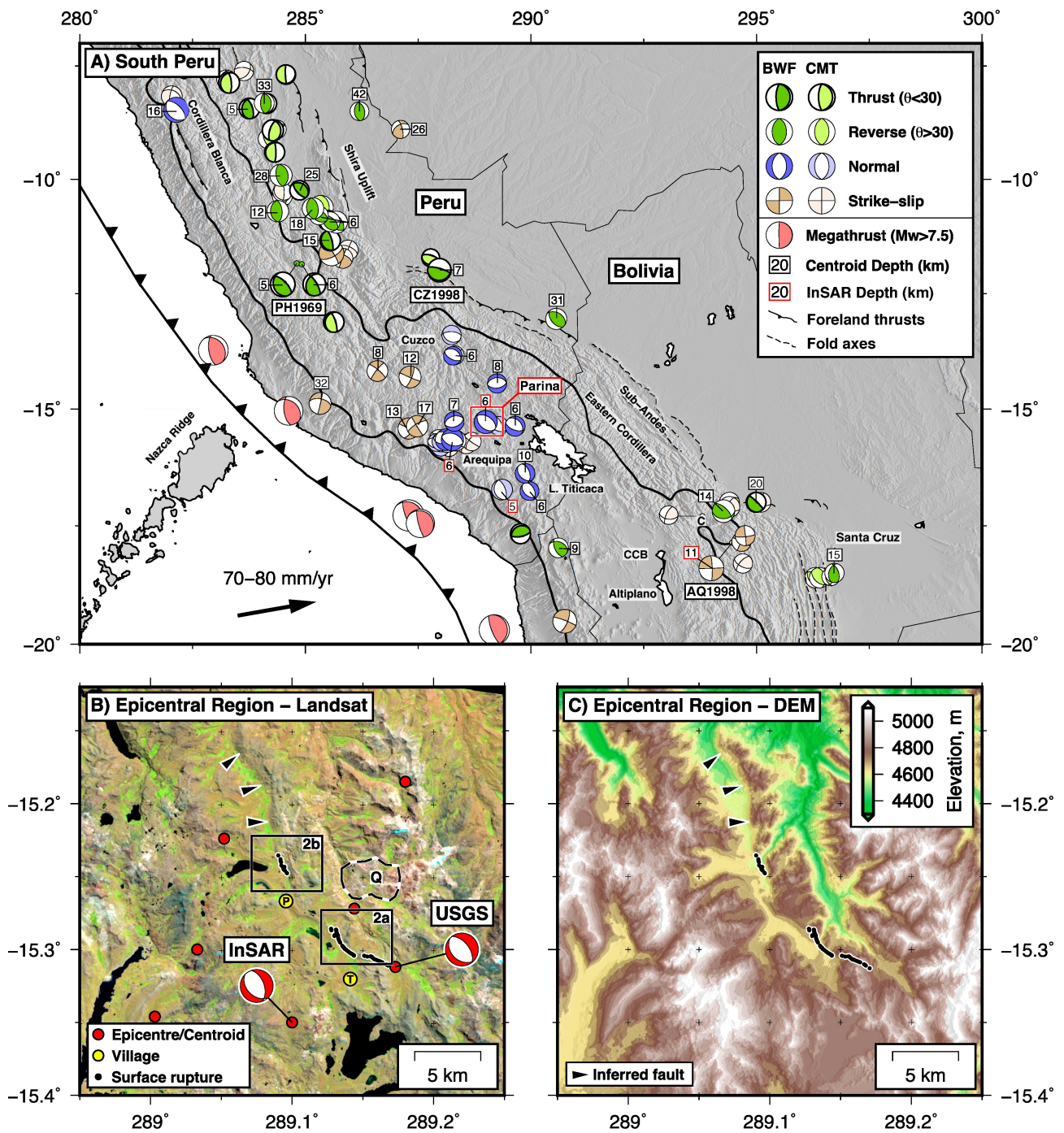
Figures

Figure 1: Overview of the Parina earthquake epicentral region. Waveform modelled (BWF) and global catalogue earthquakes (gCMT; Ekström et al. [2012]) with $M_w > 5$ in the overriding plate are shown in (a) (see Supplementary Table 1 for sources). We do not include earthquakes offshore in the forearc. PH1969 are the 1969 Pariahuanca earthquakes [Philip and Megard, 1977; Suarez et al., 1983], AQ1998 is the 1998 Aiquile earthquake [Funning et al., 2005] and CZ1998 is the 1998 Cuzco foreland earthquake [Emmerson, 2007]. C marks the Cochabamba and Punata basins, and CCB marks the Corque-Corococo Basin. The thick black line is the 3000 m contour from the 100 km Gaussian filtered topography. The red box marks the location of (b) and (c). (b) is a Landsat 8 (bands 4,3,2) image of the epicentral region. Black dots are the co-seismic surface ruptures, and black triangles mark the inferred extension of the Parina Fault north of the surface ruptures. P and T marks the villages of Parina and Togra. The centroid and mechanism of the best-fit InSAR fault plane, and the epicentre of the USGS source model, are shown together with the epicentres of $M_w > 4.0$ NEIC aftershocks (location error >5 km). Black boxes outline the locations of the satellite imagery in Figure 2, and the region marked 'Q' is an open cast quarry. (c) is the SRTM 1-arcsecond topography covering the same region as (b), highlighting the NW-SE trending step in topography associated with the Parina Fault.

48
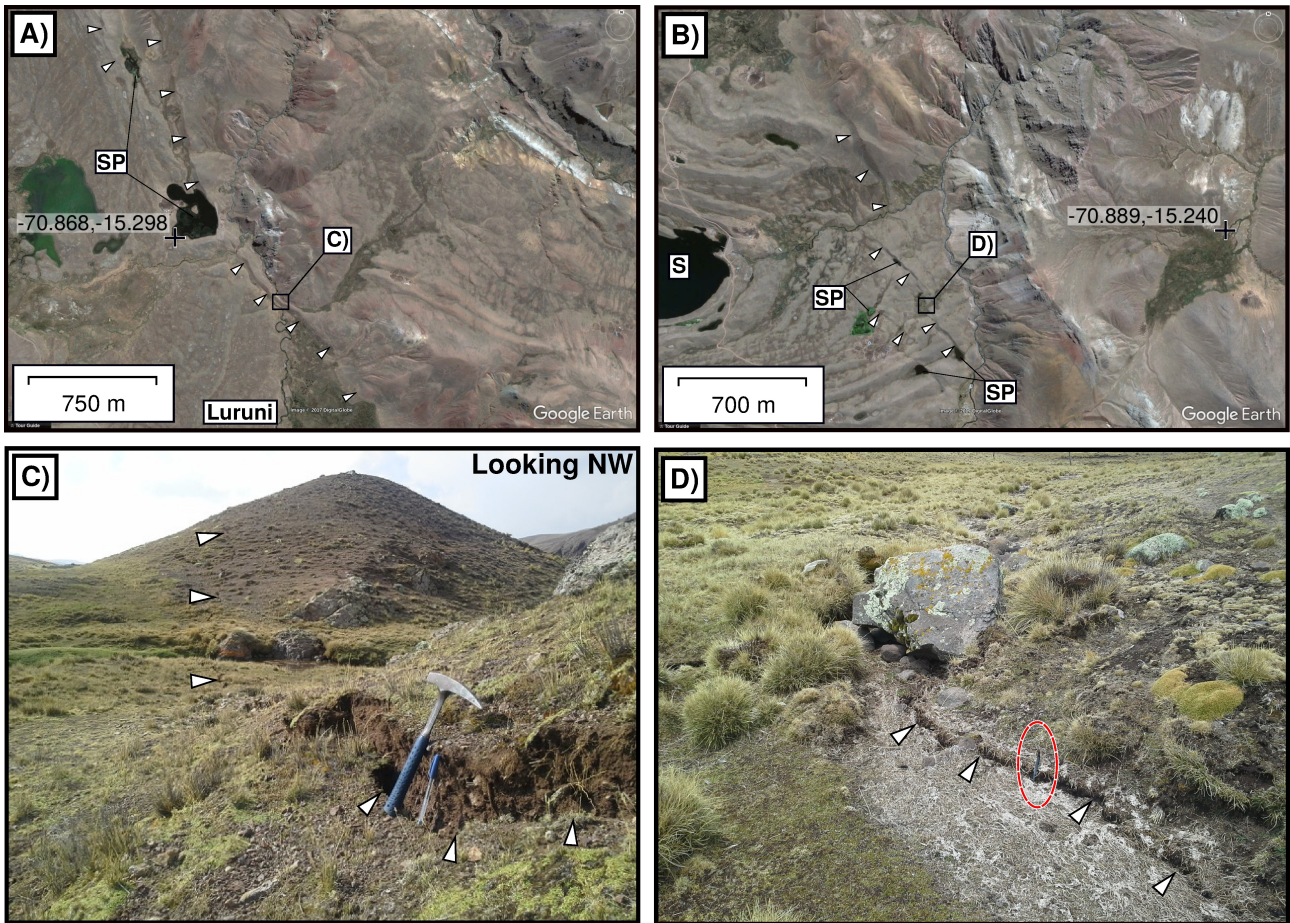
Figure 2: Geomorphology and surface ruptures of the 1st December 2016 Parina earthquake. (a) shows a GoogleEarth satellite image (from 2015) of the southern end of the co-seismic ruptures near Togra (see Fig. 1b). White triangles pick out small scarps in the topography that are associated with the mapped co-seismic surface ruptures, and bodies of water marked SP are small ponds dammed against the footwall scarp of the Parina Fault. (b) shows a similar GoogleEarth image (from 2015) over the northern end of the ruptures near Saguanani Lake (marked S). The vertical offsets across the surface ruptures in the southern section are up to 30 cm in height (c, geological hammer for scale) and have a small opening component, whilst at the northern end of the rupture, most of the features are hairline ruptures and have <10 cm offset (d, pen circled in red for scale).
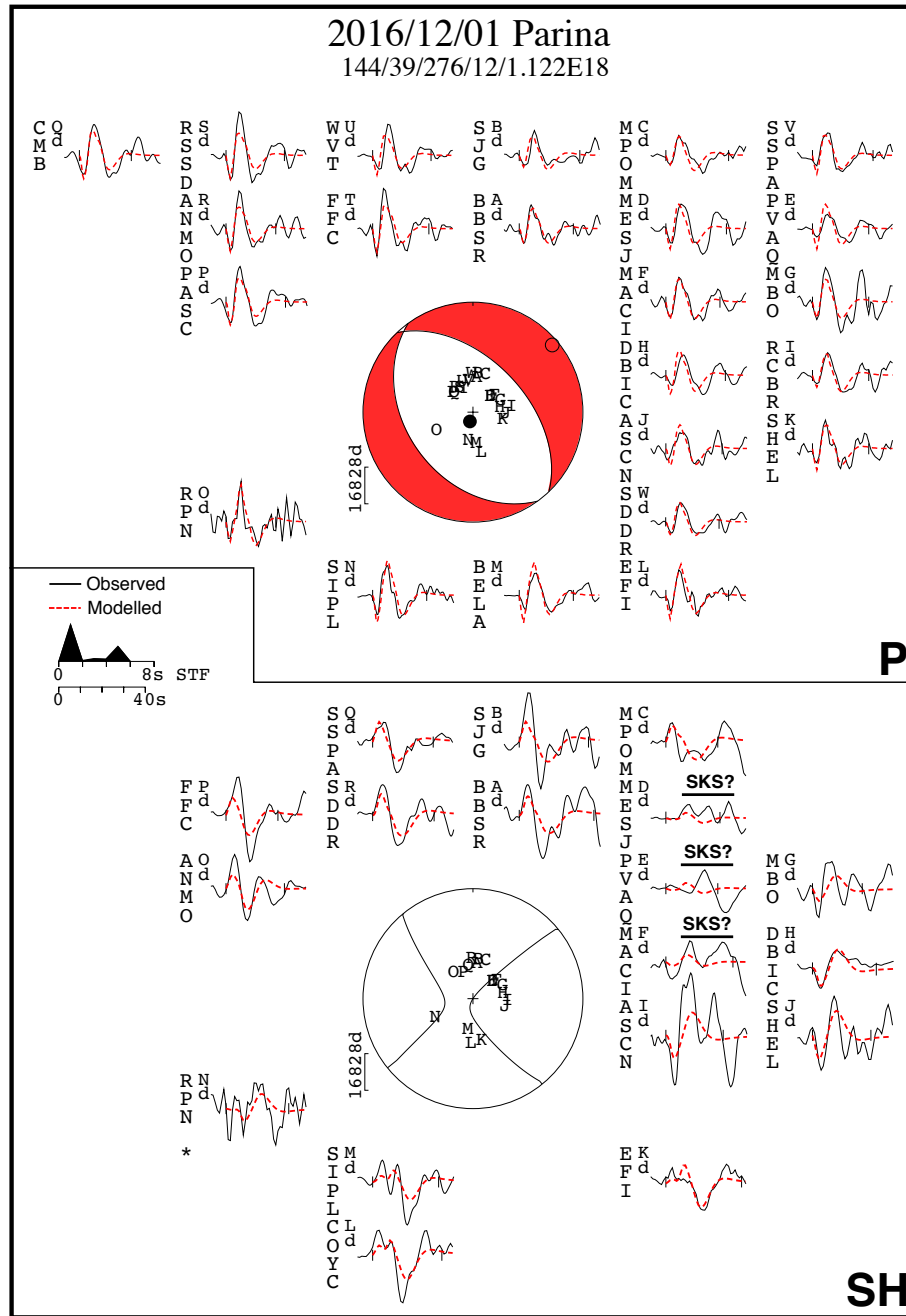
Figure 3: Minimum-misfit teleseismic body-waveform solution for the 1st December 2016 Parina earthquake. Details of the inversion result are shown below the title with the form strike/dip/rake/depth/moment, where depth is in kilometres and moment is in Nm. The top panel shows a lower hemisphere projection of the *P* wave nodal planes, with the station distribution used in the best-fit inversion shown as capital letters. Observed and modelled seismograms are 40 second time-series and are shown as black and red dashed lines, respectively. The seismogram station code is shown to the left of the corresponding seismogram, and the inversion time window is demarcated by vertical black ticks. P and T axes are projected onto the focal sphere as a black filled circle and an open circle, respectively. The lower panel shows the equivalent for the *SH* waves. Seismograms that are labelled with SKS occur within the 80-85° range from the epicentre, therefore may contain some signal from phases that have interacted with the core. Stations with an asterisk are weighted to 0 in the inversion as they contain a large component of noise, but are included for comparison with the model predictions. The source-time function is shown as black triangles.
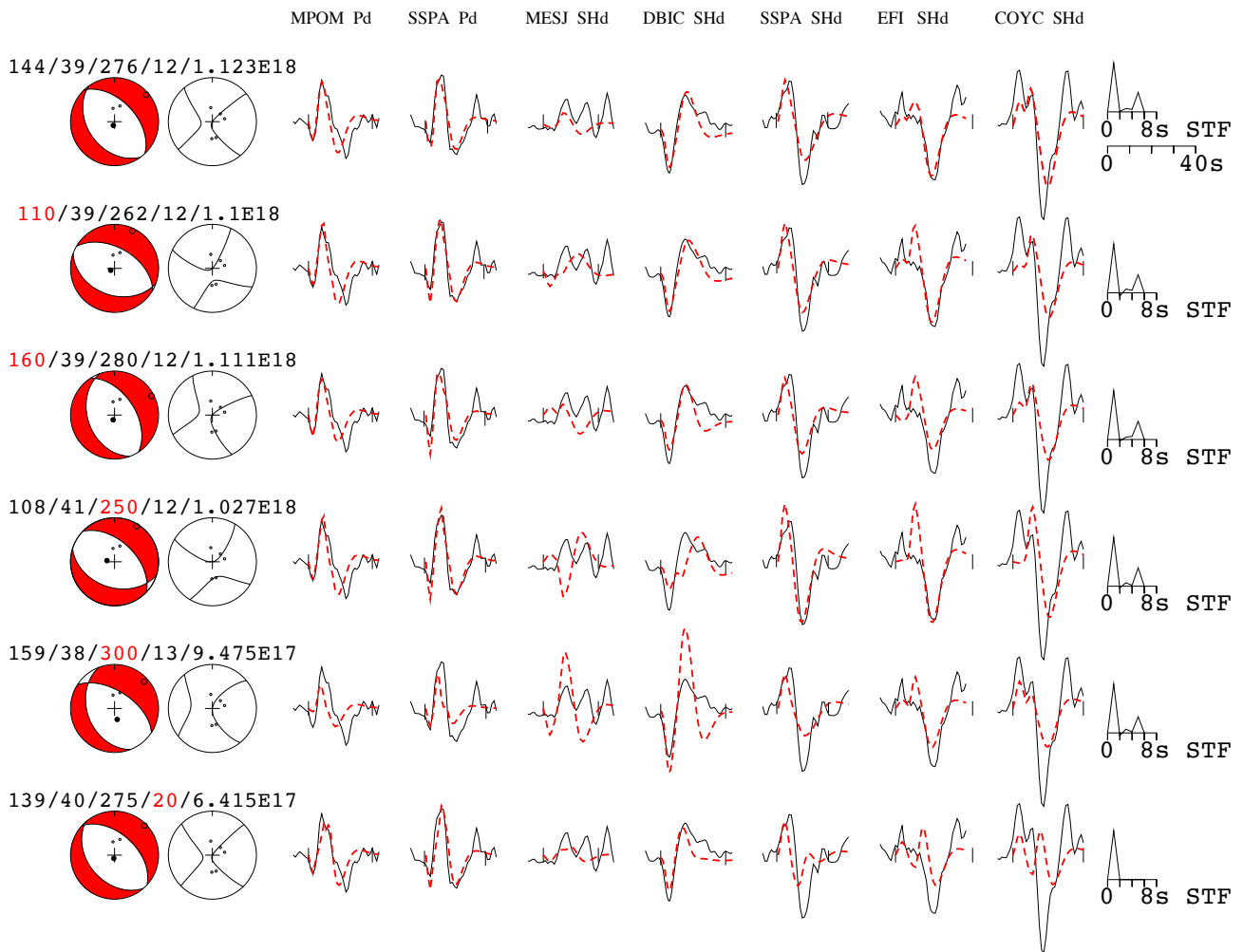
Figure 4: Sensitivity tests showing fits between selected waveforms of the 2016 Parina earthquake given variations in the source parameters. Each line shows *P* and *SH* focal spheres, the fault parameters in strike/dip/rake/centroid depth/moment formation, the fit between modelled (red dashed) and observed (black solid) waveforms at a set of stations, and the source-time function for that particular model. The station codes are given at the top in each column, where Pd are vertical-component waveforms and SHd are transverse waveforms. The best-fit model presented in Fig. 3 is shown on the top row, and each subsequent row shows different inversion results in which the parameter in red is held fixed, whilst all other parameters can vary.
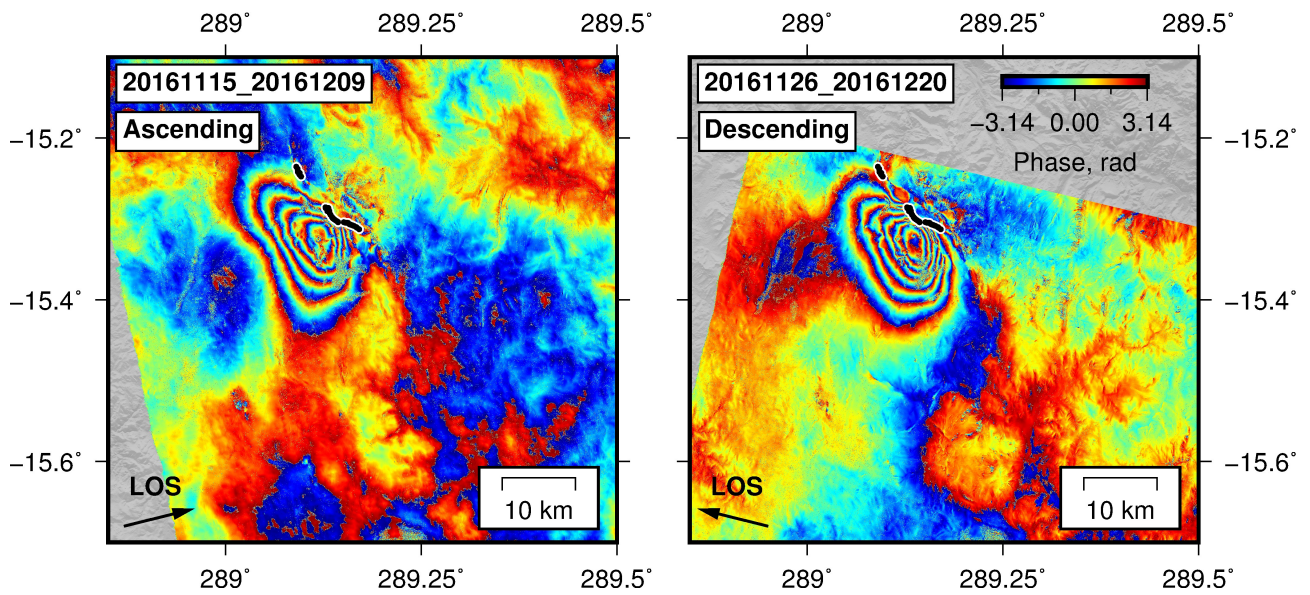
Figure 5: Wrapped co-seismic interferograms covering the 1st December 2016 Parina earthquake. The dates of the pre- and post-event acquisitions are given in the form YYYYMMDD, and the line-of-sight vector is shown in the bottom left. The incidence angle of this vector relative to the Earth's surface is between 30-43°. Black dots outline the locations of the co-seismic surface ruptures.
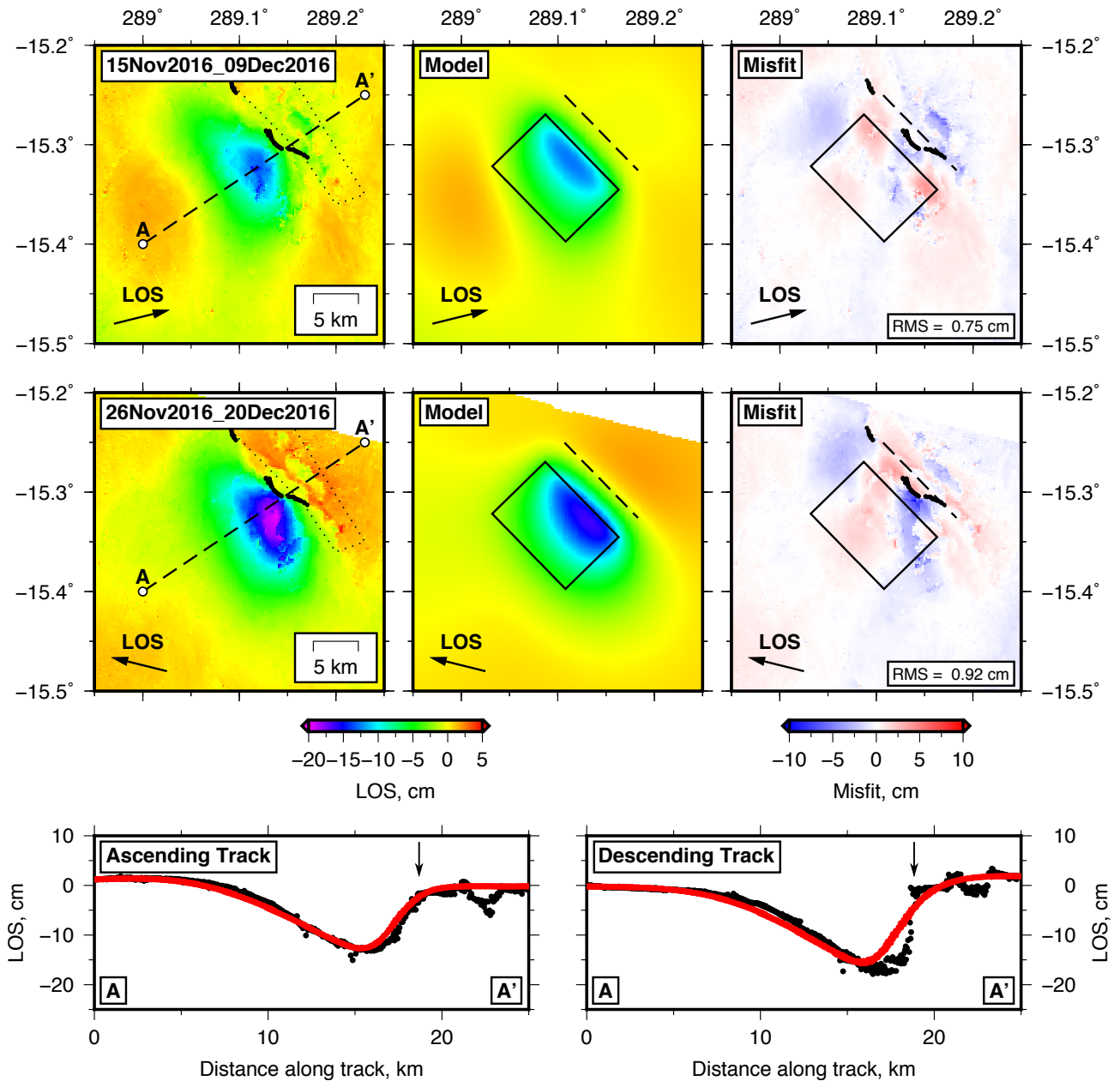
Figure 6: Results of inversions for the best-fitting elastic dislocation model to the observed InSAR surface displacements. The left panels show the unwrapped and downsampled interferograms covering the co-seismic period. We use the convention that +ve LOS is motion towards the satellite. The black polygon delineates a region with short wavelength, spatially variable LOS displacements that are likely to represent offsets at the surface generated by local slope failure in the steep valley NE of Parina, and open cast quarrying activity (Fig. 1b). The black dots are the mapped co-seismic surface ruptures. The best-fitting model is shown in the middle panels, with the map view of the fault plane shown as a black rectangle, and its surface projection as a black dashed line. The misfit between the observations and models is shown in the panels on the right. Profiles between A-A' through the observations (black dots) and models (red dots) are shown below. Steep gradients of LOS displacement near the surface ruptures (shown by the vertical arrows) are clear in the descending-track profile, however are not present in the smoother ascending track data. These differences between the surface displacement may reflect post-seismic afterslip, as the descending track acquisition was collected 19 days after the earthquake, whilst the ascending track data only includes 8 days of post-seismic deformation. We present models and a discussion of the post-seismic deformation in Section 2.5.
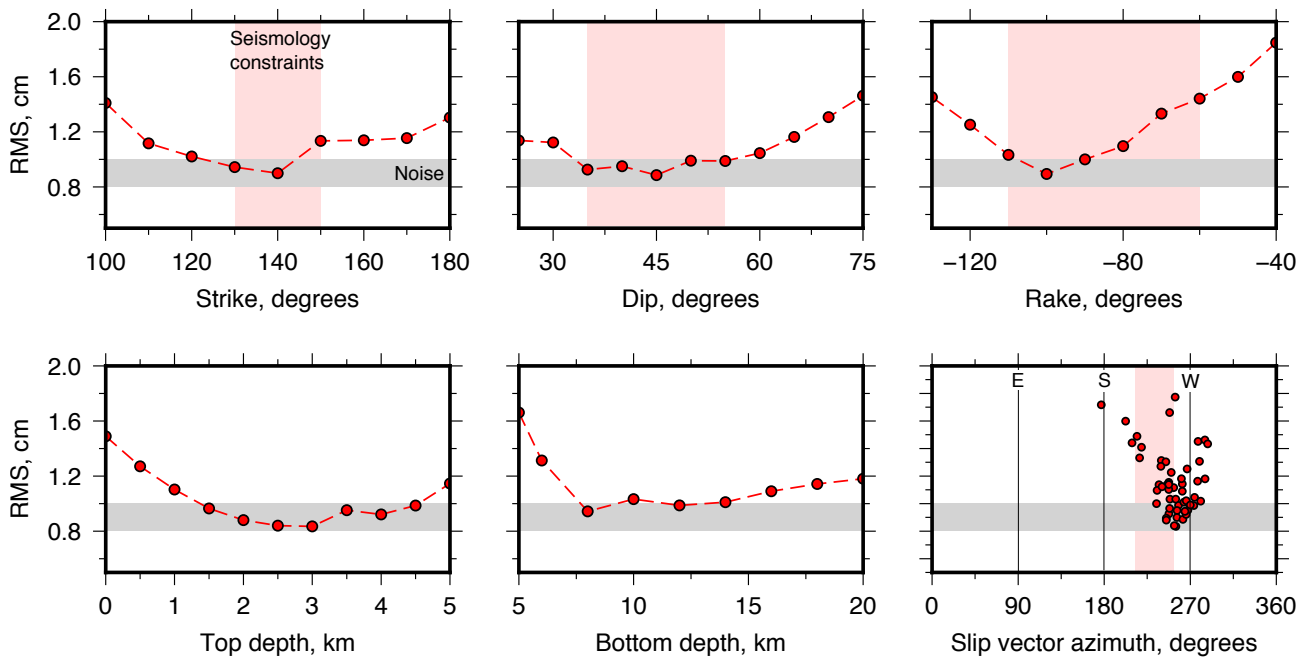
Figure 7: Sensitivity analysis of the InSAR inversions to the fault parameters. The light red bars mark the possible range of fault parameters determined from the seismological body-waveform modelling (Figs. 3,4) and the orientation of the surface ruptures. The grey bar shows the noise levels in the interferograms. The slip vector azimuth is given as the motion of the footwall relative to the hangingwall in degrees clockwise from north.
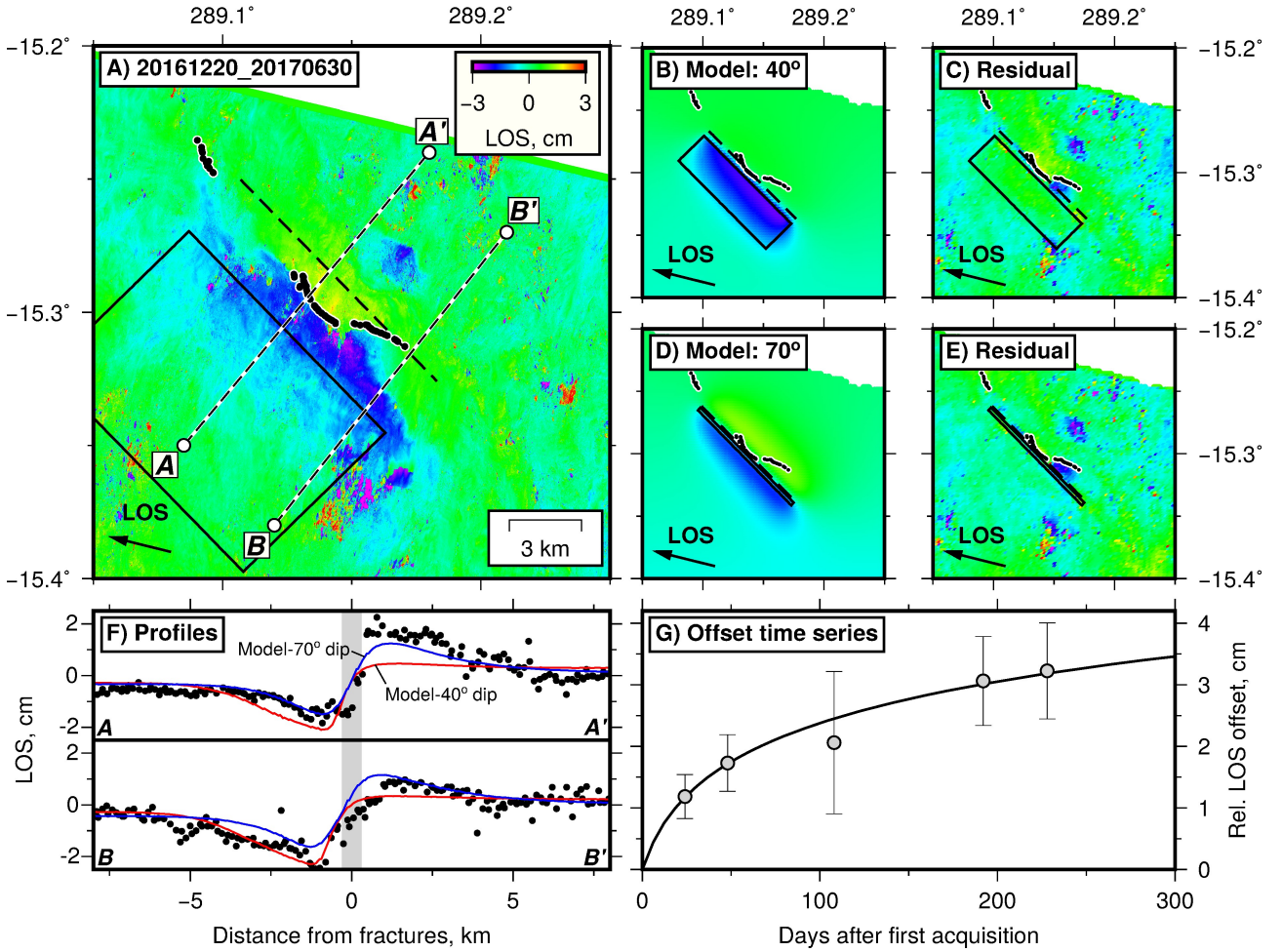
Figure 8: Post-seismic deformation following the 2016 Parina earthquake. (a) shows a descending-track interferogram covering 6 months of post-seismic deformation with a linear, long-wavelength ramp removed. The pre- and post-event acquisition dates are given in YYYYMMDD format in the top left, and the black dots are the co-seismic surface ruptures. The map view of the best-fit co-seismic fault plane is shown as a black solid line, and its surface projection as a black dashed line. (b) shows the best-fit fault model to the post-seismic surface displacement when the fault strike, dip and rake are fixed to the co-seismic fault plane. (c) shows the residual between the model prediction and the observed surface displacement. (d) and (e) show the best-fit fault model when the strike and rake are fixed to their co-seismic values, and the dip is fixed to 70°. Profiles through the deformation field between A-A' and B-B' are shown in (f), where the black dots are the observed LOS changes, the red lines are the model in (b), the blue lines are the model in (d), and the grey bar denotes the location of the surface ruptures. The temporal evolution of deformation in (g) records the change in relative LOS between the immediate hangingwall and footwall of the Parina Fault derived from a series of interferograms with a common reference. Error bars are the 1 standard deviation bounds on the difference in LOS offset between the immediate hangingwall and footwall of the fault. The black solid line is a best-fit curve of the form $a \ln (1 + t/t_r)$, showing that the post-seismic deformation decays with the logarithm of elapsed time [e.g. Smith and Wyss, 1968; Savage et al., 2005; Fielding et al., 2009].
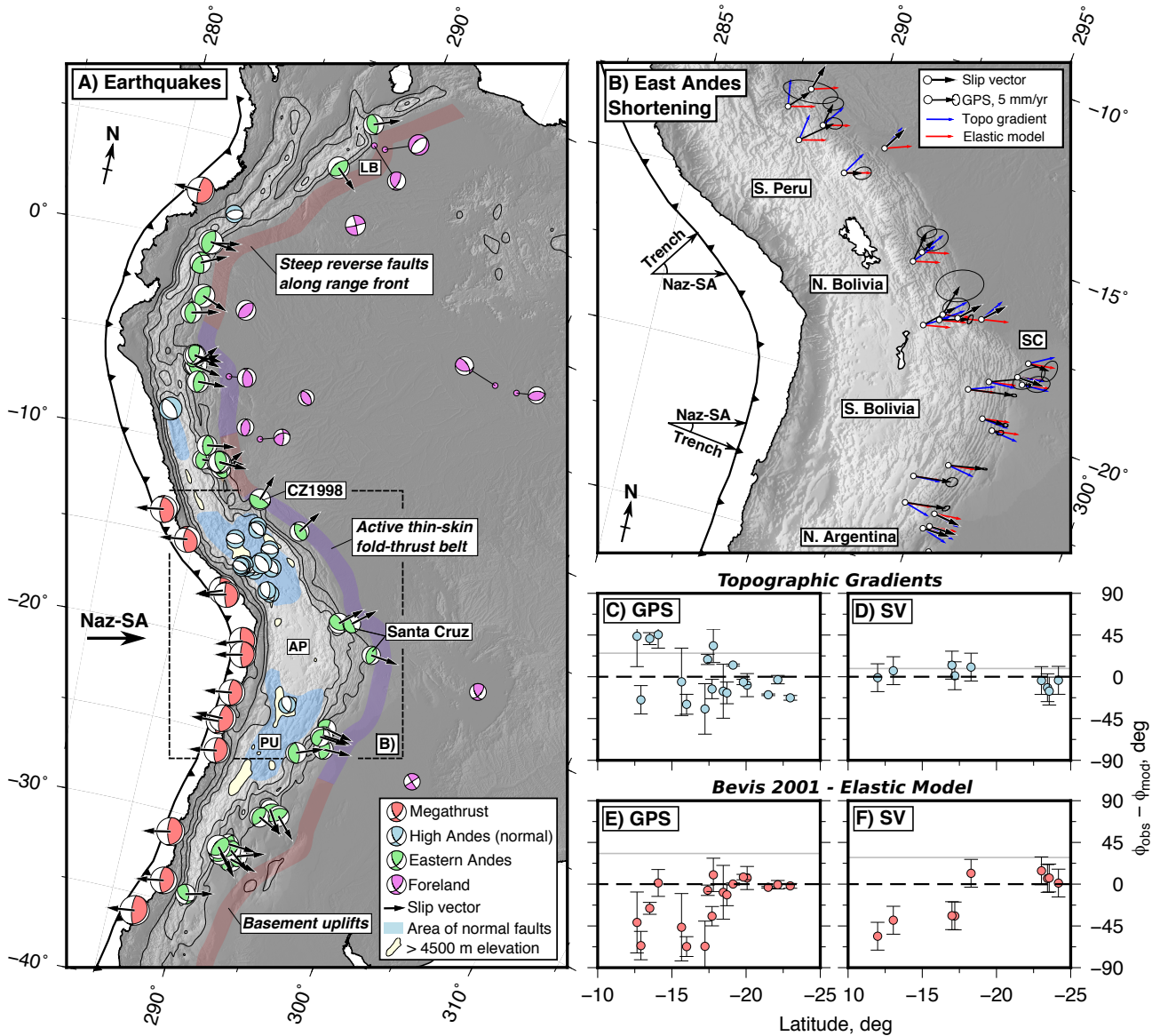
Figure 9: Earthquake and GPS data demonstrating the pattern of deformation in the regions bounding the Andes. (a) shows selected waveform-modelled earthquakes and well-constrained gCMT events compiled from the literature and this study (see Supplementary Table 1 and 2 for a list of sources). Thin lines are the 1, 2, 3 and 4.5 km elevation contours from the Gaussian filtered topography (300 km filter width). The light blue region is the area in which evidence for recent extension has been documented in the high Andes (Supp. Fig. 17). LB is the Llanos Basin, AP is the Altiplano, PU is the Puna plateau and SC is Santa Cruz. Both (a) and (b) are in oblique Mercator projection about the Euler pole of the Nazca-South America convergence taken from DeMets et al. [2010], therefore slip vectors parallel to relative plate motion will be horizontal [McKenzie and Parker, 1967]. (b) shows the azimuth of shortening in the eastern Andes inferred from selected GPS sites [Kendrick et al., 2001; Weiss et al., 2016] and earthquake slip vectors, compared with gradients in the filtered topography (300 km Gaussian filter) and the elastic block model of Bevis et al. [2001]. GPS velocities in (b) are shown relative to stable South America with 1σ uncertainty ellipses. (c-d) compare the azimuth of the topographic gradients with the azimuth of the GPS and earthquake slip vectors, respectively. (e-f) compare the predicted azimuths for the block model of Bevis et al. [2001] with the GPS and earthquake slip vectors. The grey lines show the the root-mean square misfit (always positive), see also Table 3.
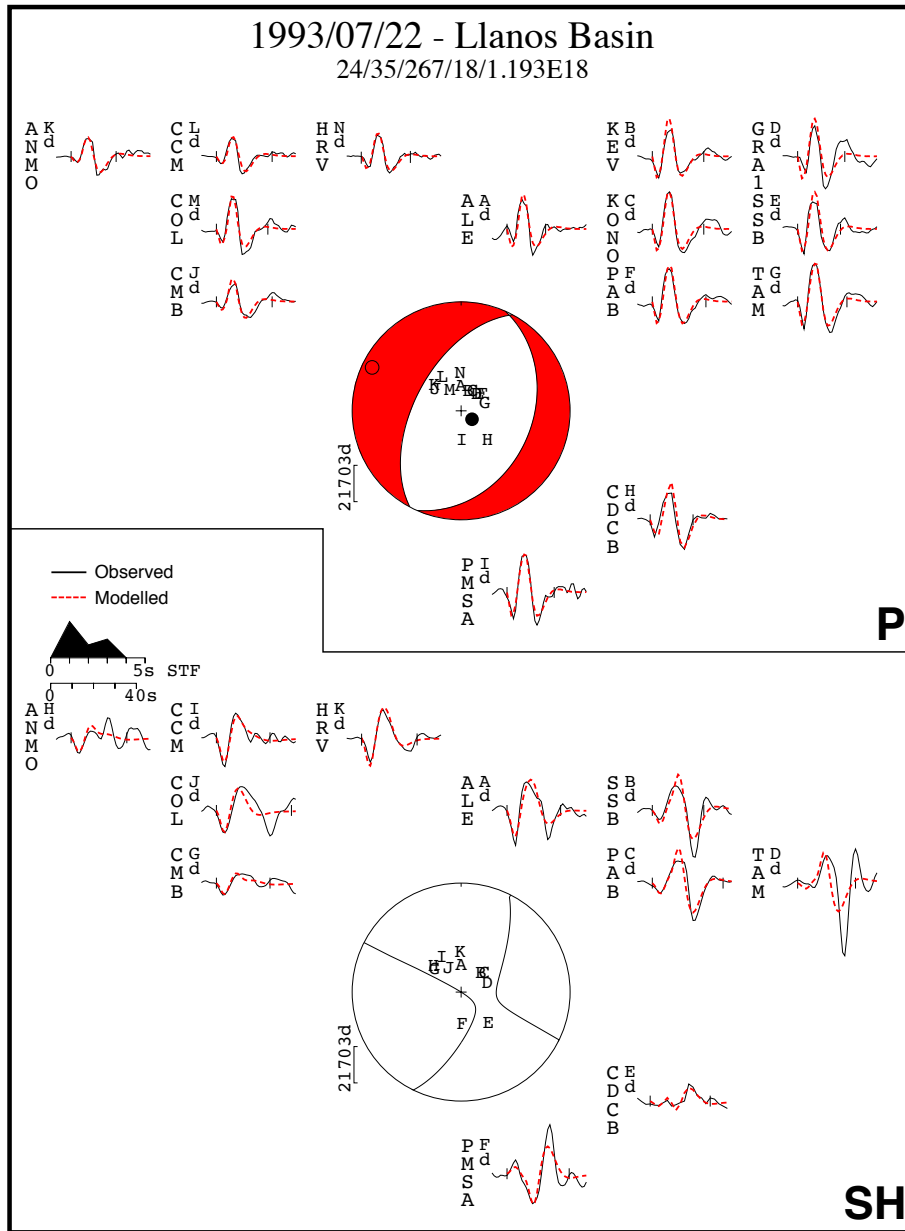
Figure 10: Minimum-misfit teleseismic body-waveform solution for the 22nd July 1993 Llanos Basin earthquake. Figure layout is the same as for Fig. 3. The uncertainties in the fault parameters were estimated using the method described in Section 2.2 and are strike = $\pm 25^o$, dip = $\pm 5^o$, rake = $\pm 25^o$ and depth ± 5 km.
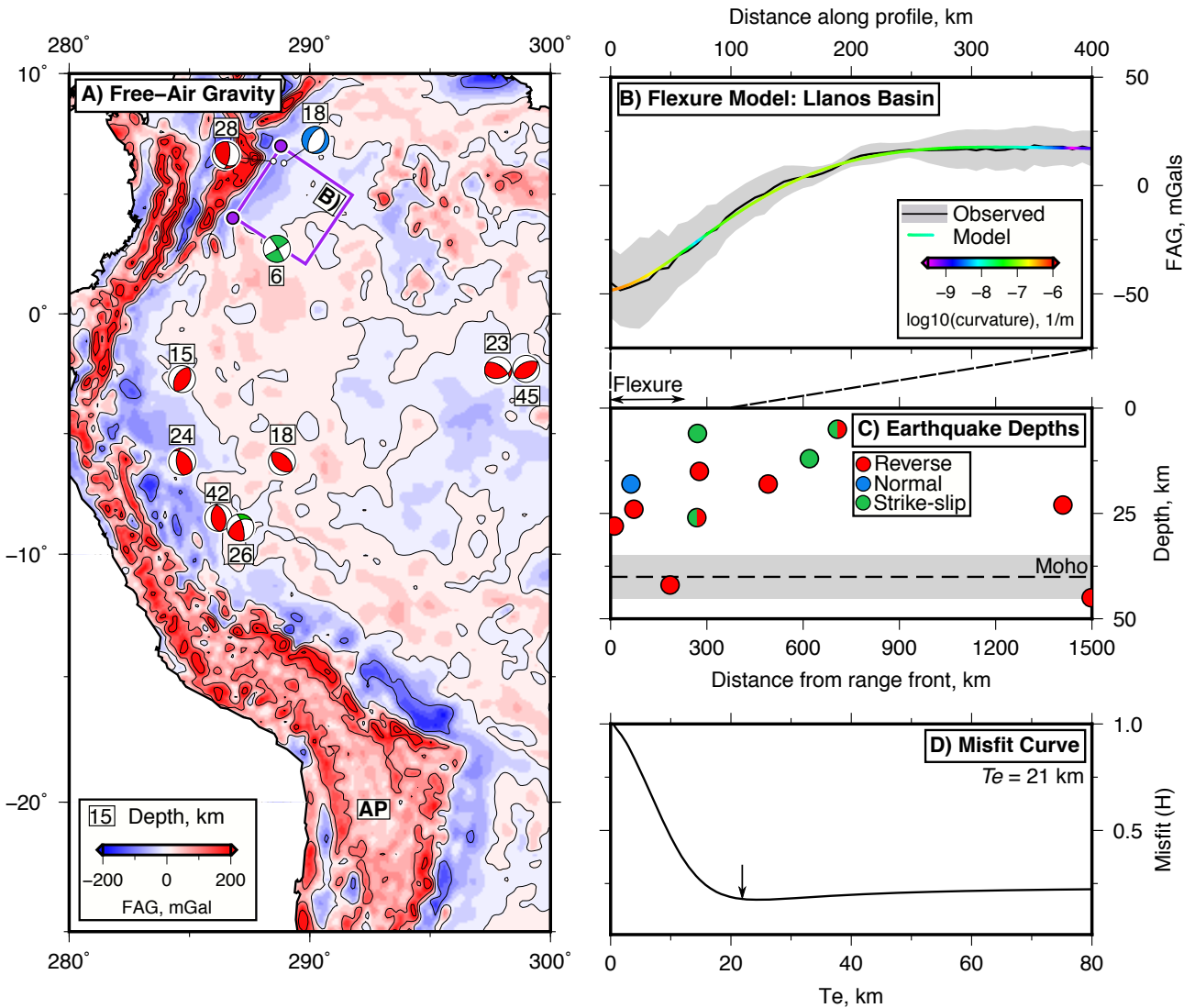
Figure 11: Free-air gravity data, earthquakes and plate models used to study the flexure of the South American forelands. (a) shows the EIGEN-6c free-air gravity map of South America [Shako et al., 2014] low-pass filtered to exclude signals with wavelength <50 km that are unrelated to flexure [e.g. Bry and White, 2007]. The purple box is the region within which profiles through the gravity field perpendicular to the Andes range front have been extracted and stacked. AP marks the Altiplano. (b) shows the stacked free-air gravity anomalies with 1 standard deviation (σ) bounds, and the best-fit model to the observed gravity profile of an elastic plate overlying an inviscid half-space bending in response to a line load and bending moment [e.g. Turcotte and Schubert, 2002]. We assume that the sediments in the foreland basin are 500 kg m^{-3} less dense than the mantle beneath the plate, and Young's modulus and Poisson's ratio of the plate are 10^{11} Pa and 0.25, respectively [Turcotte and Schubert, 2002]. The colour of the line shows the absolute curvature of the plate model. (c) shows the mechanisms of waveform-modelled foreland earthquakes as a function of distance from the Andes range front. The Moho depth and uncertainty (1σ) is taken from the catalogue of Assumpção et al. [2013]. (d) is the weighted misfit between the model and observations computed for a range of elastic thicknesses, defined as $H = \frac{1}{N}\Sigma\left[\frac{g_m - g_o}{\sigma}\right]^2$, where g_m and g_o are the modelled and observed free-air gravity, respectively, and N is the number of measurements along the profile.
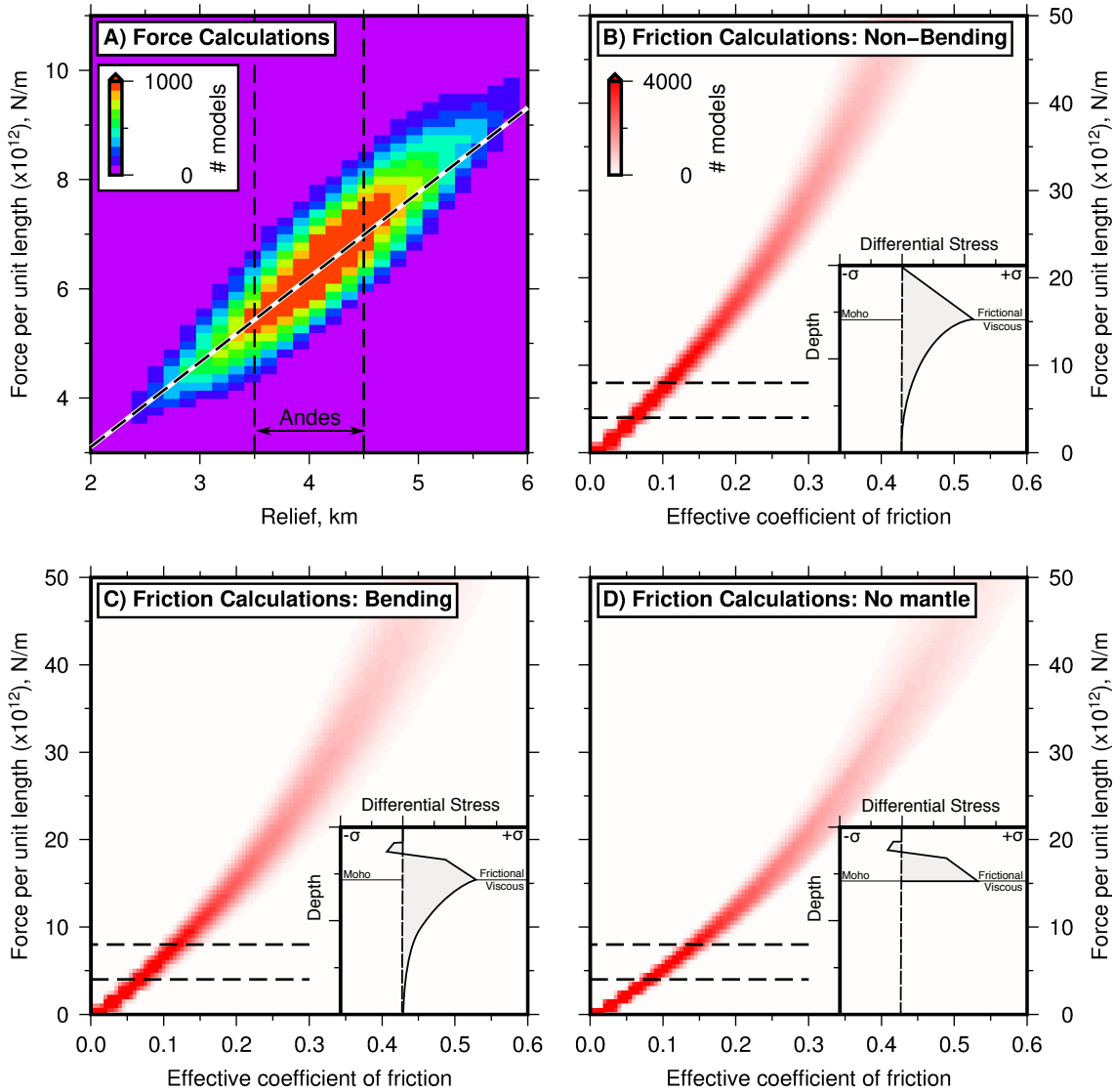
Figure 12: Results of calculations for the forces per unit length acting through the South American lithosphere, due to differences in gravitational potential energy between the Andes and the South American forelands. (a) shows the force calculations as a function of relief between the mountains and forelands (described in Section 4.1). Vertical black dashed lines delineate the range of appropriate solutions for the Andes. The dashed black and white line defines the relationship between relief and force assuming crustal isostasy, with a crustal thickness of 70 km and 40 km in the mountains and lowlands, respectively. (b,c,d) show the results of calculations for the maximum force that can be supported by the foreland lithosphere as a function of the effective coefficient of friction along faults, as described in the Section 4.2. The three different cases shown are; (b) when the whole crust is in compression and there is no sediment layer (i.e. the foreland outside the bending region); (c) when the top 5-8 km of crust is sediment that supports no force, the top 5-13 km of basement is in extension, and the bottom ∼20 km is in compression (i.e. the foreland in the bending region); and (d) the end-member scenario when the lithosphere is bending, and the mantle supports no stress at all. Schematic yield stress profiles are shown inset. Horizontal black dashed lines are the bounds on the horizontal force acting through the South American lithosphere from the calculations in (a). Where the envelope of models plots above these horizontal dashed lines, the lithosphere would be be too strong to break in earthquakes in response to the forces generated by contrasts in gravitational potential energy acting between the mountains and forelands.
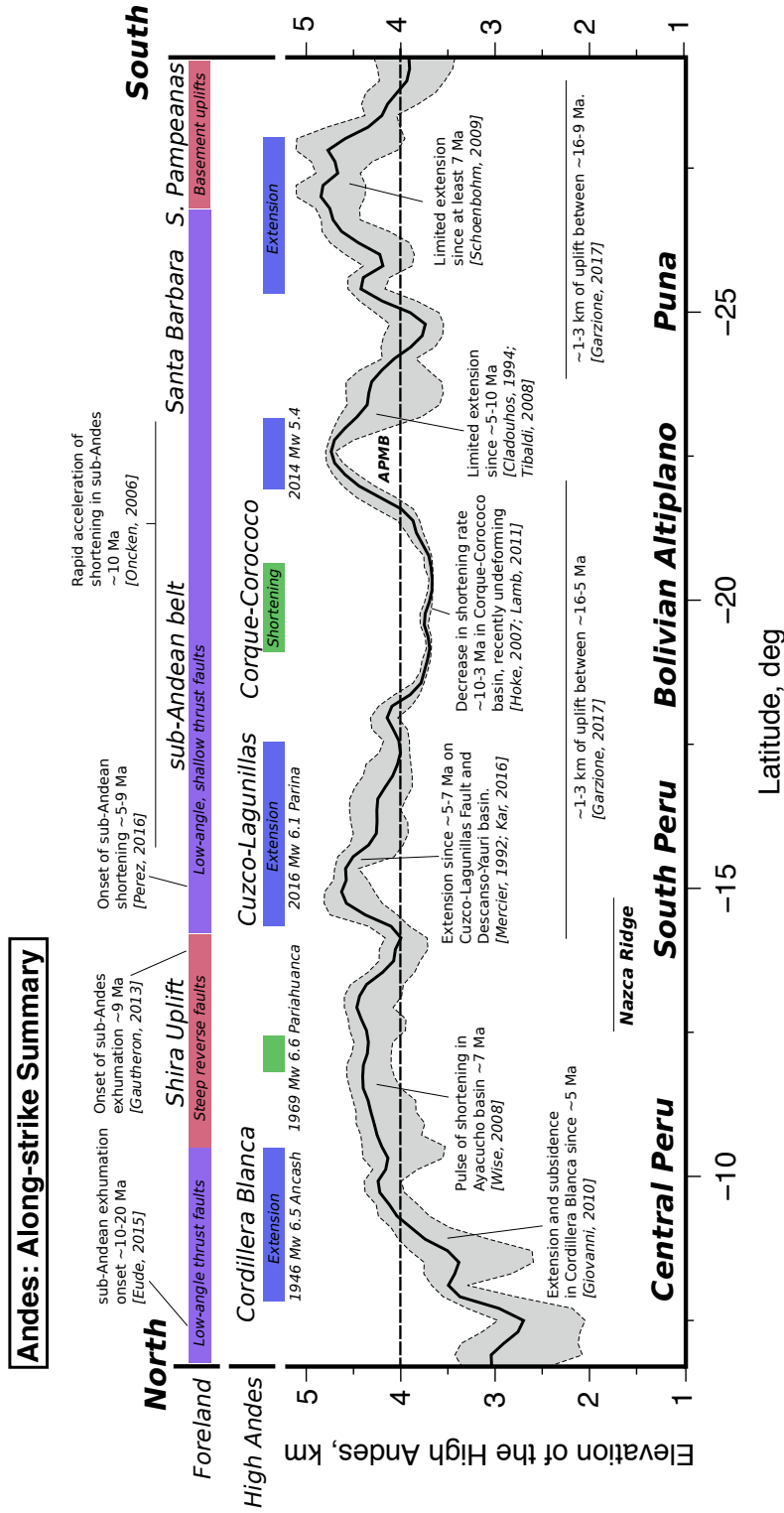
Figure 13: Summary of the topography and recent deformation history in the Andes between central Peru and northern Argentina. Topography is extracted in a 30 km swath down the spine of the range from a 100 km Gaussian filtered version of the SRTM 3-arcsecond elevation model. The maximum, mean and minimum elevations are shown. Evidence for the timing and pattern of deformation in the high Andes and the adjacent foreland is summarised above and below the elevation curve. APMB is the Altiplano-Puna magma body.
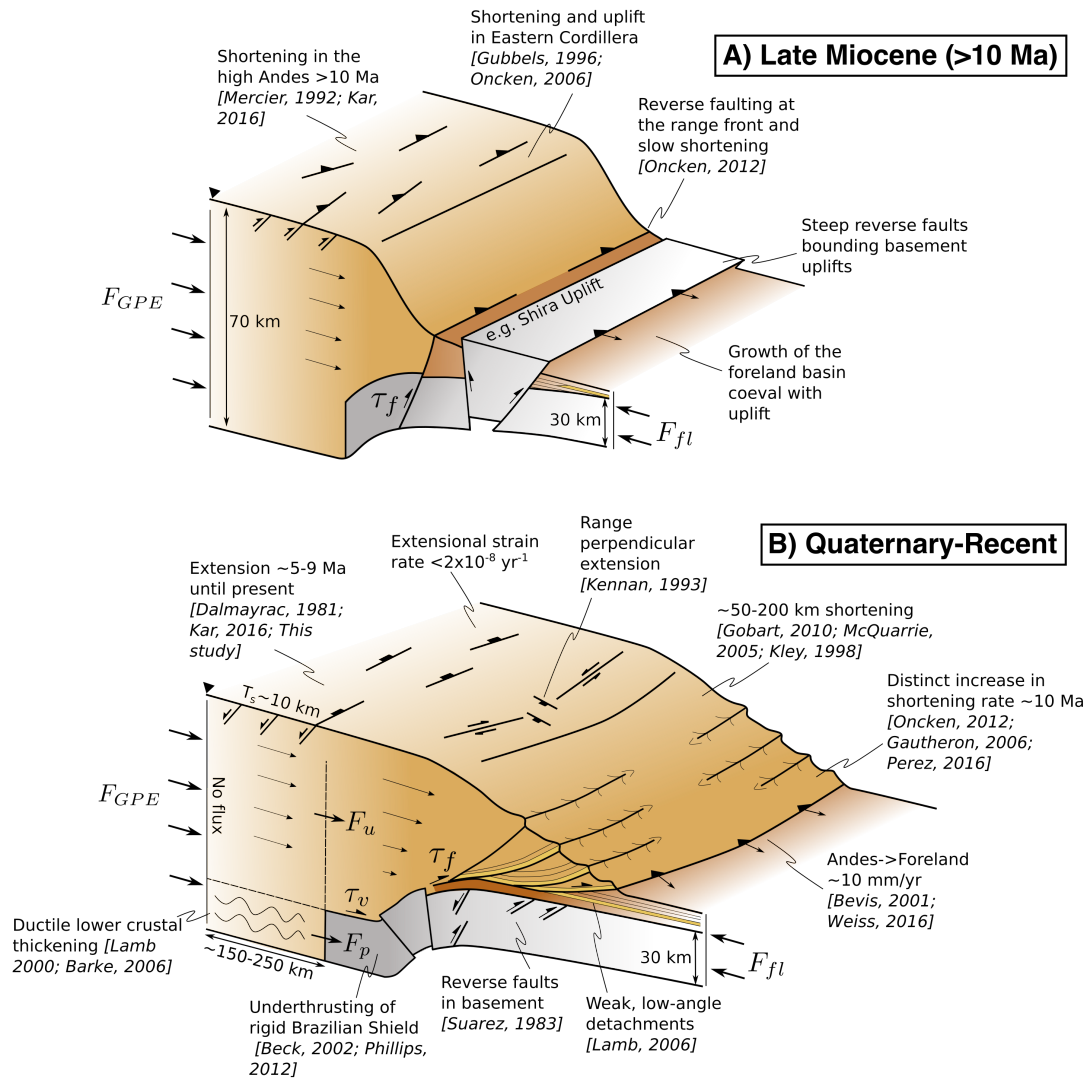
Figure 14: Sketch of the evolution of faulting in the Andes since the late Miocene. A reduction in the frictional shear stresses (τ_f) along the base of the eastern Andes ∼5-9 Ma would have reduced the horizontal force acting through upper crust overriding the rigid foreland (F_u). A reduction in τ_f would be balanced ($F_{GPE} = F_u + F_p$) by an increase in the viscous resistance to shortening in the ductile lower crust (F_p), extensional viscous strain within the centre of the mountain belt, and an increase in the rate of propagation of the mountains over the rigid foreland.