Modeling of stress transfer in the Coquimbo region of central Chile

A. Gardi,[1,2] A. Lemoine,[1,3] R. Madariaga,[1] and J. Campos[4]

Received 17 September 2004; revised 30 November 2005; accepted 29 December 2005; published 28 April 2006.

[1] We study the seismicity and stress transfer in the Coquimbo region of central Chile, where an exceptional series of more than 12 earthquakes of magnitudes from 6 to 7.6 has occurred since July 1997. In this area, the oceanic Nazca plate is subducted under the continental lithosphere of South America. Below 50 km, the downgoing slab slips aseismically with respect to the South American plate at a rate close to 6.5 cm/yr. The Coquimbo region was the site of major earthquakes of $M > 8$ in 1880 and 1943. After many years of quiescence, the seismic activity of the plate interface suddenly increased in mid-1997 and continued at least until 2004. The first group of events occurred in July 1997 in the middle of the locked plate interface. In October 1997, the activity moved inland to the Punitaqui-Ovalle area, just above the transition from the seismogenic zone to that of aseismic slip. The main event of the series was the $M = 7.6$ 15 October 1997 Punitaqui earthquake. This is an intraslab compressional earthquake that occurred at $\sim$60 km depth, on a subvertical plane located very close to the downdip edge of the seismogenic coupled interface. We performed simulations of Coulomb stress transfer for earthquakes near the bottom of the seismogenic zone. We found that a simple model of stress transfer from the aseismic slip at depths greater than 50 km can explain the triggering not only of the Punitaqui earthquake but also of the July 1997 sequence. Additional simulations show that the seismicity that followed the 1997 event for almost 7 years can also be simply explained as a result of increased Coulomb stresses on the seismogenic plate interface as a result of the 1997 event and aseismic slip.

Citation: Gardi, A., A. Lemoine, R. Madariaga, and J. Campos (2006), Modeling of stress transfer in the Coquimbo region of central Chile, *J. Geophys. Res.*, *111*, B04307, doi:10.1029/2004JB003440.

1. Introduction

[2] We study the recent seismicity of the Coquimbo region in Central Chile where seismic activity increased very rapidly in the late 90s after a series of events that occurred from July 1997 to January 1998, and has continued at a high rate since then. Earthquakes in this region are due to the subduction of the Nazca plate under North Central Chile (Figure 1). The Coquimbo region is located in the middle of the shallow dipping segment of the Chilean subduction zone identified by *Barazangi and Isacks* [1976]. In this area the downgoing slab plunges initially with a dip angle of 25° to 30°, then it unbends and plunges under the Andes with a dip angle of nearly 10°. The depth of the transition from the seismogenic zone to aseismic slip in Coquimbo was estimated as 40–50 km by *Tichelaar and Ruff* [1991] from seismic observations, and confirmed by *Klotz et al.* [2001] from geodesy. According to *Oleskevich et al.* [1999] the Nazca plate here is relatively cold so that the temperature at the bottom of the locked zone is only 200°C and the location of the thermally activated creep isotherm, 350°C, is undetermined but deeper than 100 km.

[3] Earthquakes in the Coquimbo region are of several types: outer rise events, thrust-faulting earthquakes located at the interface between the Nazca and the South American plates, intraplate events occurring inside the downgoing slab, and continental events [see, e.g., *Lomnitz*, 1971; *Madariaga*, 1998]. In this paper we are interested in some relatively large events that occur between 50–60 km depth inside the downgoing slab, near the downdip edge of the strongly coupled seismogenic interface. These events are quite common and can produce significant destruction as discussed by *Malgrange et al.* [1981], *Beck et al.* [1998], and *Kausel and Campos* [1992]. This kind of seismicity is found also in subduction zones of central Mexico [*Cocco et al.*, 1997; *Mikumo et al.*, 1999; *Gardi et al.*, 2000; *Lemoine et al.*, 2001] and central Peru [*Lemoine et al.*, 2001, 2002]. Intraplate earthquakes near the bottom of the seismogenic zone are of two types: along-slab tensional events, also called slab pull events, and along-slab compressional (slab push) earthquakes [see, e.g., *Astiz and Kanamori*, 1986; *Lay et al.*, 1989]. Our goal is to study the origin of the events that occur just below the bottom of the seismogenic zone, not to model the intermediate depth seismicity that occurs

[1] Laboratoire de Géologie, Ecole Normale Supérieure, Paris, France.
[2] Now at Institut de Physique du Globe, Paris, France.
[3] Now at Risques Sismiques, Aménagement et Risques Naturels, Bureau de Recherche Géologique et Minière, Orleans, France.
[4] Departamento de Geofísica, Universidad de Chile, Santiago, Chile.

0148-0227/06/2004JB003440$09.00

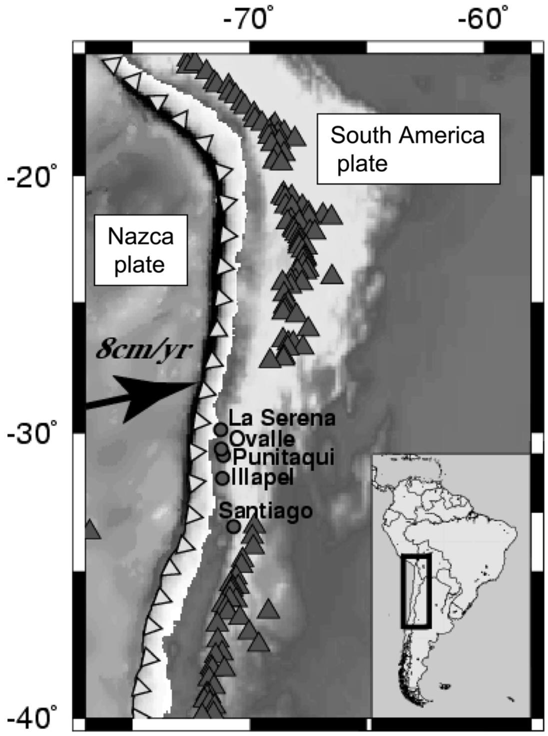

Figure 1. Tectonic setting of central Chile. The line with open triangles shows the position of the trench. Solid triangles indicate Quaternary volcanism. The black arrow indicates direction and rate of plate convergence.

inside the Nazca plate at depths greater than 80 km. This kind of seismicity is more likely to be due to thermal stresses and dehydration inside the downgoing slab. It is remarkable that intermediate depth seismicity at depth greater than 70 km has been almost absent in the last 15 years in the Coquimbo region.

[4] The last large subduction zone earthquake in the Coquimbo region occurred in April 1943 [*Lomnitz*, 1971]. *Beck et al.* [1998] studied far field body wave records of this earthquake and found that it ruptured a 100 km stretch of the subduction zone centered around 31°. Seismicity in the area suddenly increased in early July 1997 when several shallow thrust events of magnitude $M_w > 6$ occurred on the plate interface in a lapse of three weeks. The July 1997 events followed a cascade pattern extending southward over a distance of about 40 km toward the future site of the 15 October 1997 $M_w = 7.1$ Punitaqui earthquake [*Lemoine et al.*, 2001; *Pardo et al.*, 2002b]. This earthquake occurred inland, inside the downgoing Nazca plate, just below the downdip end of the strongly coupled interplate zone located around 50 km depths [*Tichelaar and Ruff*, 1991]. The Punitaqui earthquake had an along-slab compressional mechanism and was followed by several $M_w > 6$ interplate earthquakes that occurred just above it. The enhanced seismicity continued in the Coquimbo region at least until early 2004. A medium-sized thrust event of $M_w = 6.8$ (HRV) occurred there on 20 June 2003 and an along-slab tensional event on 10 January 2004 ($M_w = 5.7$).

[5] It has been proposed by many authors that intraslab earthquakes near the bottom of the seismogenic zone are due to stress transfer during the seismic cycle of subduction zones. *Astiz and Kanamori* [1986] proposed that tensional (slab pull) events occur before large thrust events in the coupled plate interface, while compressional (slab push) events occur after the main thrust event. The 1997 Punitaqui earthquake is an along-slab compressional event that took place almost 50 years after the 1943 earthquake. We examine in this paper an alternative view to that of *Astiz and Kanamori* [1986], that along-slab compressional and tensional events are due to stresses generated inside the Nazca plate by aseismic slip below the seismogenic zone. Aseismic slip below the bottom of the seismogenic zone in Coquimbo has been estimated as 6.5 cm/yr by *Klotz et al.* [2001] and *Khazaradze and Klotz* [2003]. This rate is lower than the secular convergence rate between the Nazca and South American plates which is about 8 cm/yr along a N78.1° strike [*De Mets et al.*, 1990]. As discussed by *Norabuena et al.* [1999] this is probably due to internal deformation of the South American plate. We propose that the stresses produced by aseismic slip at depth trigger the intraslab seismic activity. A similar model was proposed for the triggering of a slab pull event in another downgoing slab, the Weber event of 1990 in New Zealand, by *Robinson* [2003].

2. Recent Seismicity in the Coquimbo Region

[6] The northern part of the Coquimbo region of central Chile, situated roughly between 29.5°S and 31°S, was the site of a sudden increase in seismicity that started in July 1997, and was still in progress early in 2004. Unfortunately, only a handful of seismic stations operate continuously in the area so that the only sources of seismicity data available to us are the National Earthquake Information Center (NEIC) and temporary networks deployed by *Pardo et al.* [2002a]. Figure 2a shows all $3 < M < 8$ events in the time interval 1992–2004 reported in the Earthquake Data Reports (EDR).

[7] The seismicity distribution of the Coquimbo area from 1992 to early 2004 is shown in Figure 2a. Although we cannot simply distinguish the background seismicity from the events of 1997 to 2004, Figure 2a clearly shows a bow shaped zone of strong seismicity in the area from 30°–31°S. From 30° to 30.5°S, the seismicity describes a longitudinal zone offshore oriented roughly N-S, while from 30.6 to 31°S there is a large E-W seismicity cluster that connects the offshore seismic area to that of the 15 October Punitaqui earthquake and its aftershocks. Intermediate depth seismicity, below the transition from seismic to aseismic slip, is practically absent in the period of time examined in Figure 2a.

[8] The curve in Figure 2b shows the cumulative number of earthquakes versus time for the period 1992–2004 in the area from 30°S to 31.5°S. Seismicity before July 1997 occurred at a rate indicated by the dotted line in Figure 2b. The beginning of the Coquimbo sequence in July 1997 is indicated in the curve, as well as a new abrupt increase in the seismicity associated with the 15 October 1997 Punitaqui earthquake. After this sequence, around mid-1998, seismicity stabilized to a rate indicated by the dashed line. The seismicity rate in the entire area from 29.5°S to 31°S increased significantly after October 1997 and became clearly larger than what it was before the July 1997 events.

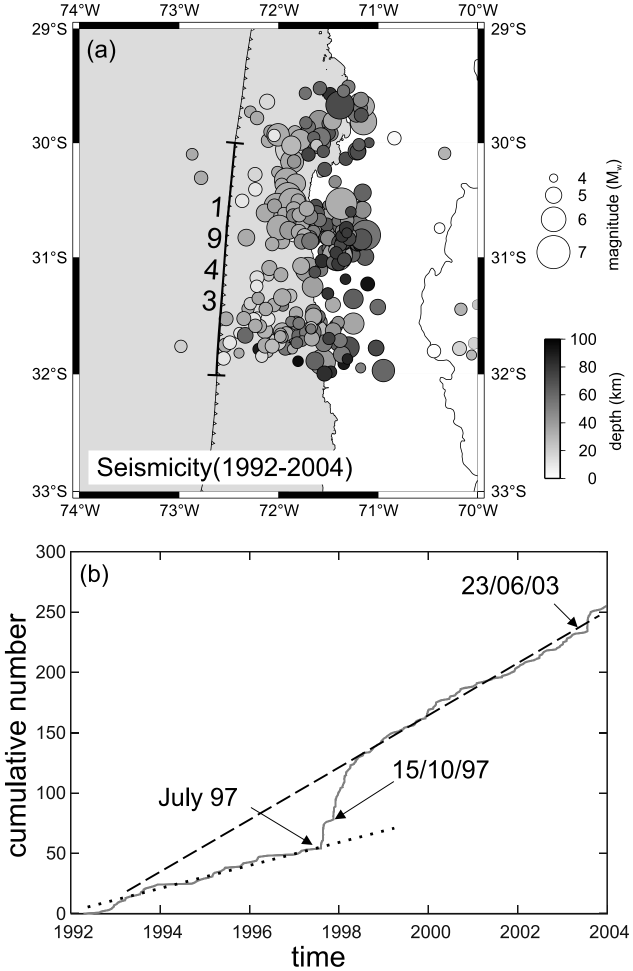

Figure 2. (a) Map view with National Earthquake Information Center seismicity for the period 1992–2004. The line with triangles shows the position of the trench, while the solid trace gives the rupture extension of the large shallow 1943 earthquake. (b) Cumulative seismicity versus time. The slope of the curve is proportional to the rate of seismicity. The dotted and dashed lines highlight the slope of the curve before July 1997 and after mid-1998, respectively.

[9] In order to explore in greater detail the seismicity, we relocated events in the study area for the period from July 1997 to September 2003, using a master event relative location technique [*Fitch*, 1975]. We assumed that travel time residuals could be adequately fitted by first-order perturbations since the relocated earthquakes were less than 50 km from the master event whose location was fixed. As we could not get accurate depths estimates from the relocation method (because of the trade-off between the depth and origin time), we used EDR (Earthquake Data Reports) depths. We selected events for which at least 20 P wave readings were available at teleseismic and local stations. The 106 events relocated in this way are shown in Figure 3a. The seismic activity started offshore in the northwestern area in July 1997, and then it moved south-east, inland, toward the source area of the Punitaqui earthquake region, where it was still in progress in early 2004.

[10] Eleven $M_w > 5.8$ events from the period July 1997 to June 2003 were modeled as point double couple sources using very broad band, P and SH waveform digital data from the IRIS network (Figure 3b). In order to avoid upper mantle triplications and core arrivals, we used teleseismic stations in the range of $30° < \Delta < 90°$. For each inversion, we selected a set of stations that represent the best azimuthal coverage as possible, deconvolved the records to a common

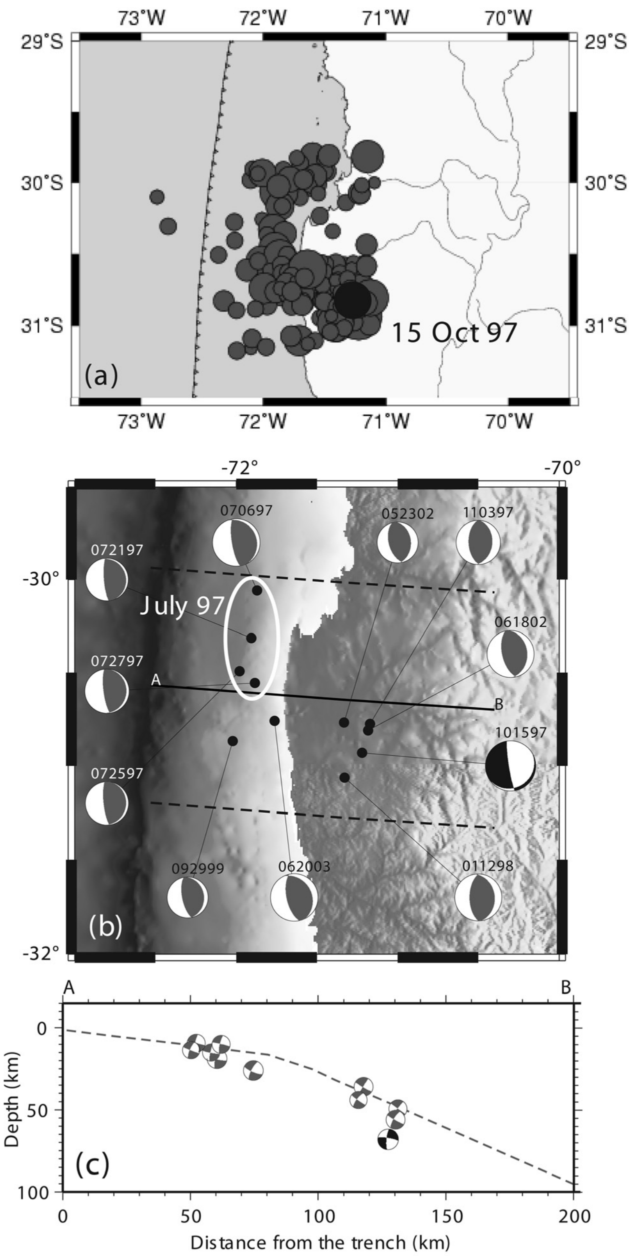

Figure 3. (a) Map view with relocated seismicity. The black circle refers to the 15 October 1997 Punitaqui event. (b) Map view with modeled events ($M > 5.8$) from July 1997 to June 2003. The black focal mechanism refers to the 15 October 1997 Punitaqui earthquake. The trace AB gives the orientation of the vertical profile shown in Figure 3c. (c) Cross section with seismicity of the area delimited by the two black dashed lines in Figure 3b. The dashed line gives the plate boundary as deduced by *Pardo et al.* [2002a].

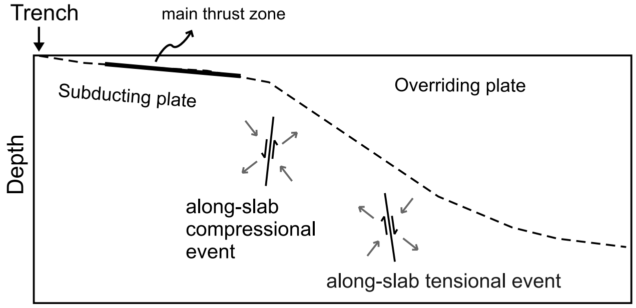

Figure 4. Along-slab compressional and tensional mechanisms: Directions of maximum compression or of the maximum extension are oriented subparallel to the slab.

station response and integrated them to displacement. In order to avoid problems with low- and high-frequency noise, we filtered the displacement records with band-pass Butterworth filters whose corner frequency depend on the earthquake studied. In modeling the far field body waves we approximated the velocity structure near the source and beneath the stations by a half-space with velocities obtained from regional studies [*Pardo et al.*, 2002b]. The main contributions to the seismograms come from the direct waves (P and S) and the reflected phases from the free surface (pP, sP, pS and sS). We used a maximum likelihood criterion to obtain the source parameters that provide the best fit between observed and synthetics waveform [*Nábělek*, 1984]. During the inversion, we solved simultaneously for the focal mechanism, focal depth and source time function using the CMT-Harvard solutions as a priori models.

[11] Figure 3b depicts the focal mechanisms of the modeled events. All of them show a thrust faulting mechanism, except the 15 October 1997 Punitaqui earthquake. This event had a normal faulting mechanism with P axes subparallel to the subducting plate (see also Figure 3c). The AB line in Figure 3b shows the orientation and the lateral extension of the vertical profile of Figure 3c, on which we projected the seismicity of the area delimited by the two black dashed lines of Figure 3b. The plate boundary drawn on the vertical section of Figure 3c was computed from the distribution of seismicity recorded by a local network installed by *Pardo et al.* [2002b] in the area of the Punitaqui earthquake between December 1999 and March 2000. All the thrust faulting earthquakes apparently occurred in the coupled zone at the interface between the Nazca and South American plates and have fault planes almost parallel to the subduction interface. The 15 October 1997 Punitaqui earthquake, on the other hand, was clearly an intraslab event with an along-slab compressional mechanism. *Lemoine et al.* [2001] found that the S waves from the Punitaqui earthquake presented a clear downward directivity, from which they concluded that this event took place on the subvertical fault plane. This plane will be considered as the fault in the following models.

3. Possible Origin of the Punitaqui Slab-Push Earthquake

[12] As observed in Figure 3b, the 15 October 1997 Punitaqui earthquake is an intraplate event with a very different mechanism from other large intraslab events that occur frequently in the transition from the coupled zone to that of aseismic slip at depth. Those events, shown in Figure 4, are usually along-slab tensional (or *slab pull*) (see, for example, the 1965 slab pull earthquake about 100 km south of Punitaqui studied by *Malgrange et al.* [1981]). Compressional (*slab push*) earthquakes, see also Figure 4, are much rarer than extensional events and, until recently, only a few had been reported in studies of Chilean seismicity by *Astiz and Kanamori* [1986] and in the worldwide survey by *Lay et al.* [1989]. *Lemoine et al.* [2001] found that along-slab compressional events do occur, but they are usually relatively small and difficult to model. They reported a number of these events at several places along the Eastern Pacific subduction zones from Chile, Peru and Mexico.

[13] Several models proposed in the literature for the occurrence of the large compressional (slab push) events were discussed by *Lemoine et al.* [2001]. The best known of these models, proposed by *Astiz and Kanamori* [1986] and further developed by *Dmowska et al.* [1996], is shown in the cartoon of Figure 5a. According to this model, right after a large subduction zone earthquake, once the main thrust zone has slipped, slab-oriented compressional stresses are transferred toward intermediate depths near the bottom of the seismogenic zone. These stresses tend to promote slab compressional events at intermediate depths, between 50 and 70 km. The 15 October 1997 earthquake, however, occurred more than 50 years after the last major event in the region, the M = 8 Illapel earthquake of April 1943 [*Beck et al.*, 1998]. It is thus very unlikely that this event can be explained by the stress transfer model of Astiz and Kanamori.

[14] We propose an alternative hypothesis, shown in Figure 5b. The solid bold line highlights the deepest portion

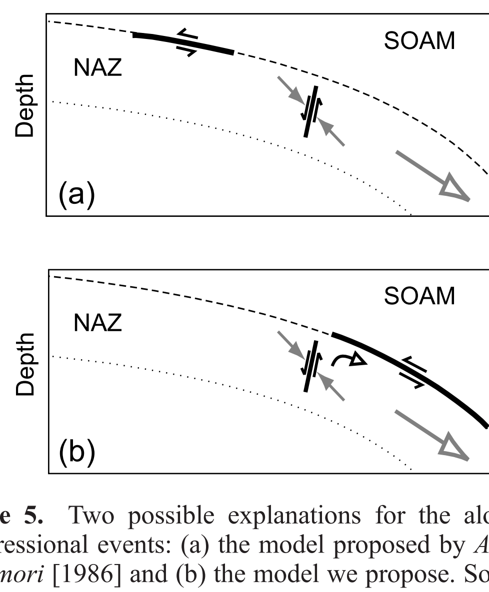

Figure 5. Two possible explanations for the along-slab compressional events: (a) the model proposed by *Astiz and Kanamori* [1986] and (b) the model we propose. Solid gray arrows give the axis of maximum compression. Open gray and black arrows indicate the movement of the subducting plate and the potential rotation of the block located inside the Nazca plate and delimited by the subvertical fault plane and the aseismic interface, respectively. NAZ, Nazca plate; SOAM, South America plate.

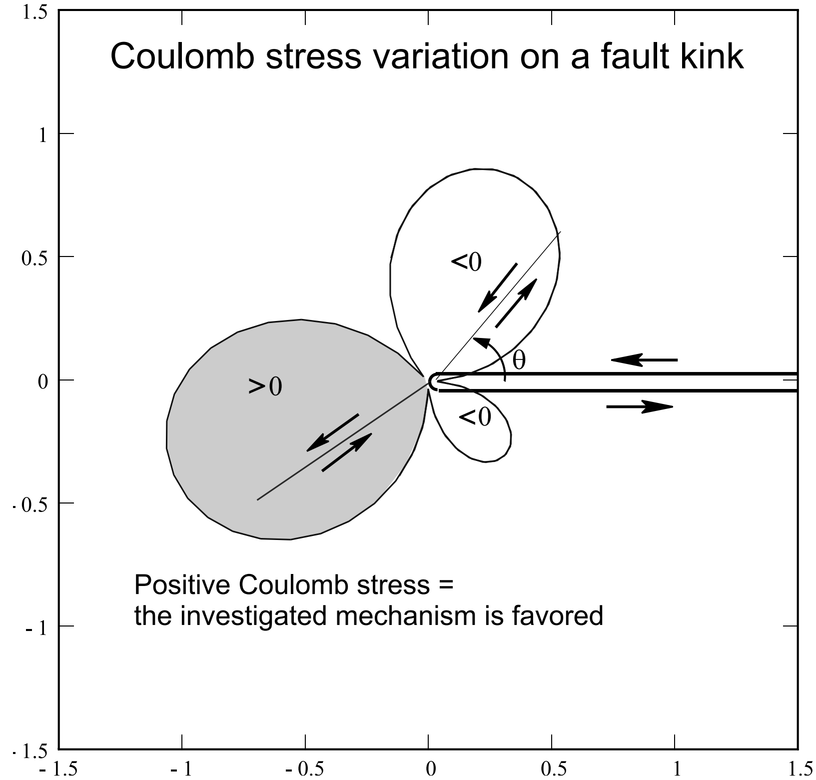

Figure 6. Coulomb stress change resolved onto planes emanating from a mode II crack as a function of the kink angle. For different angles we show the Coulomb stress variation for the same mechanism as the one acting on the primary fault.

of the plate interface along which aseismic slip occurs. Since the shallow plate interface is locked, the loading due to aseismic slip at the contact between the Nazca and South American plates should favor the rupture of the Nazca plate along potential fault planes that make a steep angle with respect to the plate interface. The main reason is that aseismic slip at depth reduces normal stresses and unclamps faults inside the downgoing slab right ahead of the transition from the seismogenic and aseismic slip zones. Recently, several authors including *Aochi et al.* [2002] and *Poliakov et al.* [2002] have studied branching of a shear crack. From their results, we can compute the Coulomb stresses transferred to faults that make an angle θ with the aseismic slip zone. In Figure 6 the stress field is that produced before rupture occurs on the branched fault. In this computation we assumed that slip along the aseismic interface increases as the square root of the distance from the bottom of the seismogenic zone. We use this crack-like model instead of a constant slip in the aseismic zone as used by *Savage* [1983] in order to simulate the transition zone between aseismic slip and the locked interface. Slip in the transition zone should be smoothed by viscous deformation and other mechanical processes like stress corrosion, etc. If more detailed models of the transition zone became available we could make better estimations of stress transfer, but we believe that our model captures the essential features of stress transfer in the transition from aseismic to locked slip. We take as positive the Coulomb stress that would favor downdip motion similar to that of the Punitaqui earthquake. This Coulomb stress is positive in the grey area of Figure 6 so that Punitaqui-like earthquakes are favored by aseismic slip in the contact between the Nazca and South American plates. This is a rough model because obviously the Punitaqui earthquake was not a direct branch from the aseismic interface, but it explains the main motivation for the following study with more usual stress transfer techniques [*Lin and Stein*, 2004; *Robinson*, 2003].

[15] In the region where the Punitaqui earthquake occurred, inside the Nazca plate ahead of the aseismic slip zone, slab compressional events are promoted. These events have a mechanism that is compatible with a rotation of the block situated inside the Nazca plate and delimited by the Punitaqui fault plane and the aseismic interface, as shown in Figure 5b by the open black arrow. According to the model we propose, slab compressional earthquakes may occur at any time, not only at the beginning of the seismic cycle, as proposed by *Astiz and Kanamori* [1986]. In order to test the two scenarii of Figure 5, we performed Coulomb stress change computations, as presented in the following section.

4. Stress Transfer in the Subduction Zone

[16] We developed a program to compute stress transfer in subduction zones based on *Okada*'s [1985, 1992] static Green functions for a homogeneous half-space. We used a Boundary Element Method derived from *Bonafede and Neri* [2000] in order to compute the stress field due to a given slip distribution on a planar fault surface by subdividing the fault in rectangular elements. Stresses due to each boundary element were computed using Okada's static Green functions, and summed at each point of the half-space. Finally, the stress field was resolved onto the secondary fault where

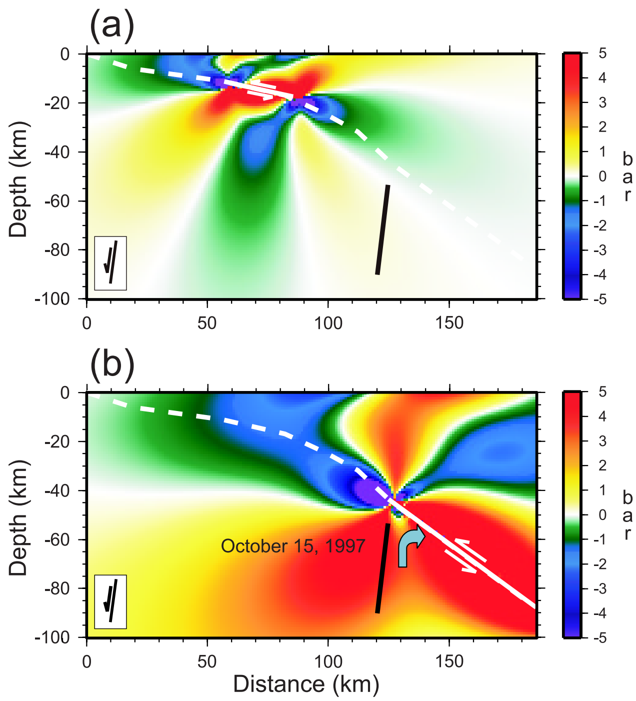

Figure 7. Coulomb stress changes on the Punitaqui fault due to (a) July swarm and (b) deep creep. The horizontal axis represents the distance from the trench. Solid white lines and arrows give position and mechanism of the fault that has ruptured, while black solid lines and arrows refer to the receiver faults. The white dashed line represents the interplate interface, as deduced from work by *Pardo et al.* [2002a].

we want to evaluate the stresses caused by an earthquake on the primary fault. All our computations were done in 3D although for the present application we only plotted 2D vertical cross sections perpendicular to the strike of the subduction one in Central Chile.

[17] Using Coulomb failure assumptions [e.g., *Jaeger and Cook*, 1979], we computed the Coulomb stress change, $\Delta\sigma_f$, as follows [*King et al.*, 1994; *Nostro et al.*, 1997]:

$$\Delta\sigma_f = \Delta\tau + \mu'\Delta\sigma_n$$

where $\Delta\tau$ is the shear stress change on the fault in the direction of slip, $\Delta\sigma_n$ is the normal stress change on the fault, and μ' is the effective friction coefficient which includes the influences of both mechanical friction and pore fluid pressure. The fault is brought closer to Coulomb failure when either the shear stress rises (load increases) or the normal stress decreases (unclamping).

[18] In our computations we did not include the determination of the optimally oriented fault planes. These are the planes where the maximum changes in the Coulomb failure stress are produced, they represent the planes where aftershocks are expected to occur [*Stein et al.*, 1992; *King et al.*, 1994]. Since our purpose was to test whether Coulomb stress transfer favored the occurrence of the different earthquakes in the Punitaqui sequence we computed the stress transferred on well identified "receiver faults". In our computations we assumed that the regional stress field is dominated by stresses produced by aseismic slip between the Nazca and South American plates below the seismogenic zone. There is no doubt that there are additional regional stresses in the source area of the Punitaqui earthquake but we have no way of estimating them. We think, however, that late in the seismic cycle, the dominant source of regional stress in the plate interface is aseismic slip, the direct cause of elastic rebound.

[19] In order to compute the stress changes produced by the different earthquakes in the sequence, we estimated their stress drop from wave form inversion. Assuming a simple fault geometry for each event we computed the corresponding slip distribution. The simulations presented in this work were all done imposing crack-like slip distributions. This is a very important point because we were interested in stress changes in the vicinity of the aseismic slip zone at depth. Coulomb stress computations with uniform slip distributions would produce unphysical stress concentrations near the fault edge.

[20] The effective coefficient of friction μ' used for the computations of Coulomb stress changes presented in this section is equal to 0.4, as often proposed in literature for subduction zones [*Stein et al.*, 1992; *King et al.*, 1994; *Lin and Stein*, 2004]. The elastic half-space is characterized by Poisson's ratio 0.27, Young's modulus of 173 GPa and rigidity μ = 68 MPa, derived for the average lithosphere of the Preliminary Reference Earth Model (PREM) [*Dziewonski and Anderson*, 1981]. In order to test the validity of the code, we performed computations on vertical left- and right-lateral faults and we compared our results to those computed by *Lin and Stein* [2004]. In case of a dislocation (uniform slip along the primary fault), these results agree with those of *King et al.* [1994] and *Nostro et al.* [1997].

4.1. Stress Transfer Between the July 1997 Events and the Punitaqui Earthquake

[21] In order to test the scenario of Figure 5a, we simulated the occurrence of the July 1997 swarm (Figure 7a) and computed the Coulomb stress transferred on the 1997 Punitaqui earthquake. A similar computation was already reported by *Lemoine et al.* [2001] using a different software. We imposed a slip of 0.6 m on the fault whose dimensions correspond to the total area which ruptured during July 1997. This value of slip was computed summing the displacements estimated for the four major earthquakes (those with $M_w \geq 6$) of the July 1997 swarm, previously modeled by *Lemoine et al.* [2001]. The resulting stress perturbation tensor is resolved onto planes parallel to the fault of the 15 October 1997 Punitaqui earthquake, the subvertical fault depicted by the symbol in the lower left corner of Figure 7a. The solid black line gives the orientation and the position of the Punitaqui fault plane, which is the receiver fault in this test. Positive Coulomb stress changes mean that the Punitaqui event is enhanced, while negative stress perturbations inhibit it. As we can see in Figure 7a, the 1997 July swarm causes positive stress changes on the Punitaqui fault plane but its maximum value is very low, less than 0.01 MPa (0.1 bar). Several authors have proposed that static Coulomb stress changes as low as 0.01 MPa can affect the locations of aftershocks [e.g., *King et al.*, 1994]. Whether static stress changes smaller than 0.01 MPa can also trigger large earthquakes like Punitaqui still remains an unresolved issue. In some instances even

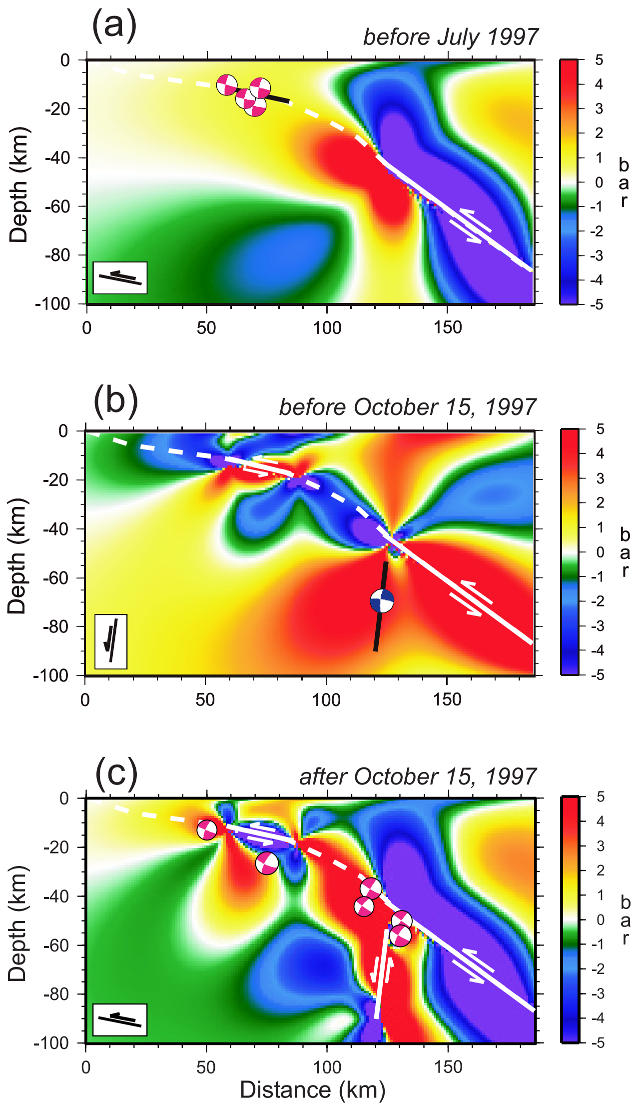

Figure 8. Coulomb stress changes (a) before the July 1997 events, (b) before the 15 October 1997 Punitaqui earthquake, and (c) after the 15 October 1997 Punitaqui earthquake. Representation is the same as in Figure 7.

smaller stress changes appear to influence the occurrence of subsequent seismicity [e.g., *Harris and Simpson*, 1996], but there are rarely enough events to do a rigorous statistical test [*Harris*, 2003]. Therefore it is difficult to decide whether the July swarm triggered the Punitaqui event, as already concluded by *Lemoine et al.* [2001].

4.2. Stress Transfer Between Aseismic Slip and the Punitaqui Earthquake

[22] As we have already discussed, the Punitaqui earthquake occurred in the downgoing slab, very close to the transition from the seismogenic to the aseismic slip zone. The bottom of the seismogenic zone was located at a depth of 50 km by *Oleskevich et al.* [1999] on the basis of previous seismological studies by *Tichelaar and Ruff* [1991]. *Klotz et al.* [2001] and *Khazaradze and Klotz* [2003] reported the results of campaign mode GPS measurements made in the Coquimbo region before the 1997 event. From the elastic deformation curves shown in Figure 8 of *Khazaradze and Klotz* [2003], we determined that the aseismic slip rate along the plate contact below 60 km was roughly 6.5 cm/yr. The exact value depends on the dip angle, but as discussed by *Khazaradze and Klotz* [2003], the dip angle does not change the slip rate very much. As in Southern Peru, this rate is substantially less than the secular convergence rate between the Nazca plate and South America which is close to 8 cm/yr according to *De Mets et al.* [1990] and *Norabuena et al.* [1999].

[23] We consider now at the possible triggering of the Punitaqui earthquake by this aseismic slip. In order to simulate aseismic slip along the plate interface at depths greater than 60 km, we assumed a maximum aseismic slip of 3.5 m. This amount is the slip deficit produced by a plate subducting at the rate of 6.5 cm/yr during a time interval of 50 years, the time elapsed since the last great, shallow thrust earthquake of April 1943. Because the Punitaqui earthquake was situated very close to the tip of the aseismic slip zone, we tapered off the slip distribution in order to simulate the effects of stress relaxation in the transition zone from aseismic slip and the locked plate interface. As a simple model for the transition zone we assumed that slip had a crack-like parabolic taper as it approached the locked zone. Using a constant slip along the aseismic slip zone would produce very large unphysical stresses because of the localized dislocation [see *Freund and Barnett*, 1976]. Then we computed the stress change produced by the tapered aseismic slip using our boundary integral method based on Okada's formulas. Finally, the resulting stress perturbation tensor is resolved onto planes parallel to that of the 1997 Punitaqui earthquake.

[24] Figure 7b depicts the distribution of Coulomb stress change produced by aseismic slip along the deepest portion of the plate interface on planes parallel to the fault of the Punitaqui earthquake. Such a scenario corresponds to that imaged by the cartoon of Figure 5b. As shown in Figure 7b we find that 3.5 m of maximum slip deficit produce Coulomb stress changes of the order of 5 bars along the aseismic slip zone and its immediate vicinity. In contrast to the Coulomb stresses transmitted to the Punitaqui earthquake from the July 1997 events in Figure 7a, the positive Coulomb stress change caused by aseismic slip at depth onto the Punitaqui fault is quite large, from 2 to 5 bar. This value is in excellent agreement to that computed for a similar situation in New Zealand by *Robinson* [2003]. Deep aseismic slip can thus favor slab push earthquakes and this seems to be independent from the earthquake cycle of the main seismogenic thrust fault, since no great shallow earthquakes have occurred in that region after April 1943.

[25] On the basis of this result, we studied whether or not aseismic slip at depth had the potential to play a role also on the occurrence of the July 1997 earthquakes. We verified this hypothesis by computing static stress changes due to deep aseismic slip onto the faults responsible for the July 1997 swarm. The result of this simulation is shown in Figure 8a, where the July 1997 rupture area is represented by the solid black line, on which we superposed the focal mechanisms of the four major events of the swarm. All of them occurred in a region dominated by small but positive Coulomb stress change. Therefore aseismic slip at depth also favored the occurrence of the July 1997 earthquakes that occurred 30 km away from the aseismic slip transition.

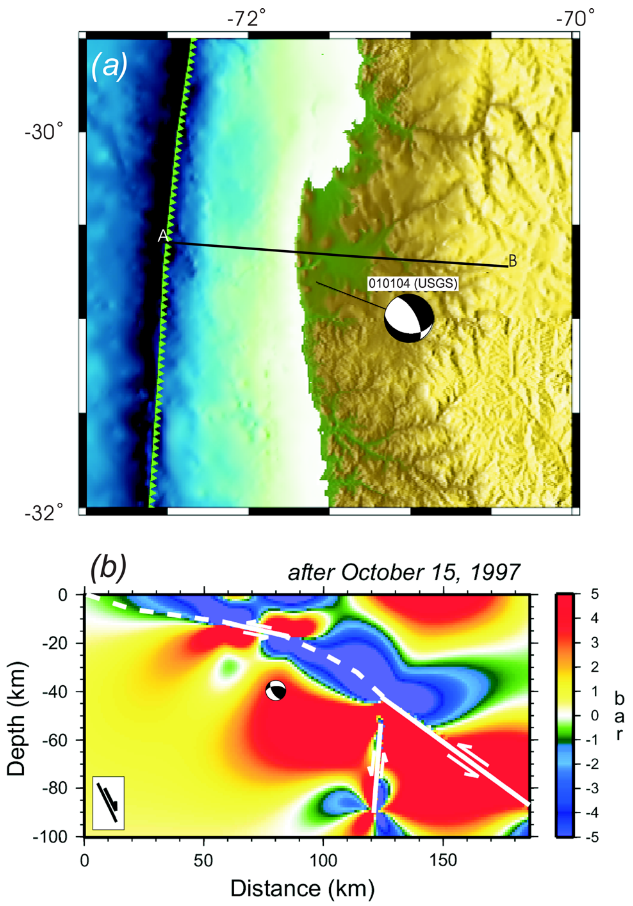

Figure 9. (a) Map view with location and focal mechanism of the 10 January 2004 earthquake, as computed by the U.S. Geological Survey. (b) Coulomb stress changes due to the same rupture sequence of Figure 8c but resolved onto the fault responsible for the 10 January 2004 event. Representation is the same as in Figure 7.

This is not surprising, because it is in agreement with the elastic rebound model for subduction zones.

[26] Summing the contributions of the aseismic slip and of the July 1997 swarm and resolving onto the Punitaqui fault plane for a normal mechanism, we obtain the stress change cross section shown in Figure 8b. This plot shows the Coulomb stress change just before the 15 October 1997 Punitaqui earthquake. Our computation predicts a Coulomb stress increase on the Punitaqui fault plane, suggesting that the Punitaqui fault was brought closer to failure by the stress perturbations because of the preceding ruptures and the aseismic slip.

[27] Moving ahead in the rupture sequence, we added the Coulomb stress change field produced by the occurrence of the Punitaqui event. Figure 8c depicts the cumulative Coulomb stress changes due to the July interplate swarm, slip in the aseismic interplate zone, plus the effect of the 15 October 1997 earthquake. Coulomb stress in Figure 8c is resolved onto planes parallel to that of the plate interface earthquake of 20 June 2003. The maximum slip for the Punitaqui earthquake is close to 2 m as shown by *Lemoine et al.* [2001]. The cumulative Coulomb stress change distribution is compared with the distribution of the thrust earthquakes occurred between January 1998 and August 2003, all of which were thrust events located near or on the plate interface. Again, we can see that the spatial distribution of these events fits the prediction quite accurately.

[28] On 10 January 2004, an earthquake of M_w = 5.7 occurred in central Chile. Its epicenter and focal mechanism are shown in Figure 9a. This earthquake occurred at 57 km depth, inside the subducted plate and had an along-slab tensional (slab pull) mechanism. Figure 9b shows the Coulomb stress change distribution obtained after the same sequence of Figure 8c (i.e., by summing the contributions of the deep aseismic slip, the July 1997 swarm and the Punitaqui earthquake) but resolved for the fault plane and rake of the 10 January 2004 event, as indicated by the symbol in the lower left corner of Figure 9b. The predicted distribution is consistent with the mechanism of the 10 January 2004 earthquake, suggesting that this event was also brought closer to failure by the preceding rupture sequence.

5. Discussion and Conclusions

[29] The Coquimbo region in central Chile had a sudden increase in seismic activity that started in mid-1997 and was still in progress in early 2004. The sequence of events started in July 1997 with a set of thrust events that preceded by about 3 months the M = 7.1 15 October 1997 Punitaqui earthquake. This is a rare along-slab compressional earthquake that occurred inside the Nazca plate near the bottom of the seismogenic zone. This event was followed by an intense aftershock activity located in the locked plate interface, trenchward from the Punitaqui event.

[30] In order to study in detail the seismicity of the Coquimbo region and to investigate a possible fault interaction between the different events that occurred in the area, we relocated events for the period July 1997 to September 2003, we modeled eleven M_w > 5.8 events and we performed stress transfer simulations for the well located events taking into account stresses produced by slip in the deep aseismic contact between the Nazca and South American plates. In order to account for possible inelastic relaxation processes near the top of the aseismic zone we applied a parabolic (crack-like) taper to the aseismic slip distribution. We found that Coulomb stress changes explain well the current seismicity in the Coquimbo region. Our study shows that the compressional (slab push) Punitaqui earthquake could have been triggered by aseismic slip at depth. Actually, the triggering of most of the earthquakes that occurred in the Coquimbo since 1997 could be produced by stress transfer from the aseismic slip zone. This is not surprising since GPS measurements reported by *Klotz et al.* [2001] show that aseismic slip was of the order of 6.5 cm/yr below 50 km in the Coquimbo area before 1997. The aseismically slipping plate interface loaded the main thrust zone bringing this region closer to failure in July 1997. The contributions from aseismic slip and the July 1997 ruptures promoted the Punitaqui earthquake, which, in its turn triggered abundant plate interface events that lasted more than 7 years.

[31] The usual explanation for the occurrence of compressional and tensional earthquakes near the bottom of the

seismogenic zone is stress transfer during large thrust events [*Astiz and Kanamori*, 1986; *Dmowska et al.*, 1996]. In this model, slab compressional earthquakes like Punitaqui should occur early in the seismic cycle after a large thrust event has ruptured the locked plate interface. This implies that earthquakes like the 1997 event should occur only at the beginning of the earthquake cycle of the main thrust fault. The last great shallow earthquake in the Coquimbo area occurred in April 1943 and we demonstrated that the sequence of thrust events of July 1997 was not large enough to produce significant stress changes on the Punitaqui fault. On the other hand, stress transfer due to continuous slip in the aseismic zone at depth clearly favors the rupture of planes oriented as the 1997 Punitaqui earthquake fault. Such events may occur at any time during the seismic cycle, once the stresses near the bottom of the seismogenic zone produced by aseismic slip are large enough.

[32] Independently from our study, GPS data from *Klotz et al.* [2001] showed that before the sequence of July 1997 the Tongoy station located in the Northern edge of the 1997 rupture zone was lagging behind the other stations along the coast, indicating perhaps a slip event at depth. This is an interesting possibility because similar aseismic displacements have been revealed, by continuous GPS observations, in 1998 and 2002 in the Guerrero seismic gap, Mexico [*Lowry et al.*, 2001; *Kostoglodov et al.*, 2003]. Another silent thrust slip event has been reported by *Hirose et al.* [1999] in southwest Japan on the basis of GPS observations. Closer to Coquimbo, a large aseismic slip event was detected by *Kanamori and Cipar* [1974] before the main shock of the 1960 earthquake (M = 9.5). The authors suggested that the signal was generated by slow slip on a part of the plate interface deeper than the coseismic rupture plane, which was confirmed by vertical displacement data of *Linde and Silver* [1989]. We hope that the ongoing GPS campaigns in the Coquimbo region, conducted with a very dense network of permanent stations, will soon help to recognize the presence of silent slip events along the deeper part of the plate boundary. As suggested by the simulations presented in this paper, we think that continuous aseismic slip is an essential element of stress transfer in subduction zones and that it is one of the main sources of regional stresses in the area.

[33] **Acknowledgments.** This work was partially done while A.G. worked at Institut de Radioprotection et Sûreté Nucléaire, Fontenay-aux-Roses, France. A.G. wishes to thank Oona Scotti for fruitful discussions on Coulomb stress change simulations and Alon Ziv for encouragement and constructive and critical comments. Maps and plots were drawn using GMT 3.0 [*Wessel and Smith*, 1991]. This research was partially funded by the program ECOS-Conicyt of cooperation between Chile and France under contract C01U02, by program PICS 2473 of the DRI/CNRS, and by a contract with the Programme National ACI Risques Naturels of INSU/CNRS. We thank an anonymous reviewer and the associate editor for their help in improving our manuscript.

References

Aochi, H., R. Madariaga, and E. Fukuyama (2002), Effect of normal stress during rupture propagation along nonplanar faults, *J. Geophys. Res.*, *107*(B2), 2038, doi:10.1029/2001JB000500.

Astiz, L., and H. Kanamori (1986), Interplate coupling and temporal variations of mechanisms of intermediate-depth earthquakes in Chile, *Bull. Seismol. Soc. Am.*, *76*, 1614–1622.

Barazangi, M., and B. L. Isacks (1976), Spatial distribution of earthquakes and subduction of the Nazca plate beneath South America, *Geology*, *4*, 686–692.

Beck, S., S. Barrientos, E. Kausel, and M. Reyes (1998), Source characteristics of historic earthquakes along the central Chile subduction zone, *J. S. Am. Earth Sci.*, *11*, 115–129.

Bonafede, M., and A. Neri (2000), Effects induced by an earthquake on its fault plane: A boundary element study, *Geophys. J. Int.*, *141*, 43–56.

Cocco, M., J. Pacheco, S. K. Singh, and F. Courboulex (1997), The Zihuatanejo, Mexico earthquake of 1994 December 10 (M = 6.6): Source characteristics and tectonic implications, *Geophys. J. Int.*, *131*, 135–145.

De Mets, C., R. G. Gordon, D. F. Argus, and S. Stein (1990), Current plate motions, *Geophys. J. Int.*, *101*, 425–478.

Dmowska, R., J. R. Rice, L. Lovinson, and D. Josell (1996), Stress transfer and seismic phenomena in coupled subduction zones during the earthquake cycle, *J. Geophys. Res.*, *101*, 3015–3029.

Dziewonski, A. M., and D. L. Anderson (1981), Preliminary Reference Earth Model, *Phys. Earth Planet. Inter.*, *25*, 297–356.

Fitch, T. J. (1975), Compressional velocities in source regions of deep earthquakes: An application of the master earthquake technique, *Earth Planet. Sci. Lett.*, *26*, 156–166.

Freund, L. B., and D. M. Barnett (1976), A two-dimensional analysis of surface deformation due to dip-slip faulting, *Bull. Seismol. Soc. Am.*, *66*, 667–675.

Gardi, A., M. Cocco, A. M. Negredo, R. Sabadini, and S. K. Singh (2000), Dynamic modelling of the subduction zone of central Mexico, *Geophys. J. Int.*, *143*, 809–820.

Harris, R. A. (2003), Stress triggers, stress shadows and seismic hazard, in *International Handbook of Earthquake and Engineering Seismology, Part B*, chap. 73, edited by W. H. K. Lee et al., Elsevier, New York.

Harris, R. A., and R. W. Simpson (1996), In the shadow of 1857: Effect of the great Ft. Tejon earthquake on subsequent earthquakes in southern California, *Geophys. Res. Lett.*, *23*, 229–232.

Hirose, H., K. Hirakara, F. Kimata, N. Fujii, and S. Miyazaki (1999), A slow thrust slip event following the two 1996 Hyuganada earthquakes beneath the Bungo Channel, southwest Japan, *Geophys. Res. Lett.*, *26*, 3237–3240.

Kanamori, H., and B. Cipar (1974), Focal process of the great Chilean earthquake of May 22 (1960), *Phys. Earth Planet. Inter.*, *9*, 128–136.

Kausel, E., and J. Campos (1992), The M_s = 8 tensional earthquake of 9 December 1950 of northern Chile and its relation to the seismic potential of the region, *Phys. Earth Planet. Inter.*, *72*, 220–235.

Khazaradze, G., and J. Klotz (2003), Short- and long-term effects of GPS measured crustal deformation rates along the south central Andes, *J. Geophys. Res.*, *108*(B6), 2289, doi:10.1029/2002JB001879.

King, G. C. P., R. S. Stein, and J. Lin (1994), Static stress change and the triggering of earthquakes, *Bull. Seismol. Soc. Am.*, *84*, 935–953.

Klotz, J., G. Khazaradze, D. Angermann, C. Reigber, R. Perdomo, and O. Cifuentes (2001), Earthquake cycle dominates contemporary crustal deformation in central and southern Andes, *Earth Planet. Sci. Lett.*, *193*, 437–446.

Kostoglodov, V., S. K. Singh, J. A. Santiago, S. I. Franco, K. M. Larson, A. R. Lowry, and R. Bilham (2003), A large silent earthquake in the Guerrero seismic gap, Mexico, *Geophys. Res. Lett.*, *30*(15), 1807, doi:10.1029/2003GL017219.

Jaeger, J. C., and N. G. W. Cook (1979), *Fundamentals of Rock Mechanics*, 3rd ed., 593 pp., CRC Press, Boca Raton, Fla.

Lay, T., L. Astiz, H. Kanamori, and D. H. Christensen (1989), Temporal variation of large intraplate earthquakes in coupled subduction zones, *Phys. Earth Planet. Inter.*, *54*, 258–312.

Lemoine, A., R. Madariaga, and J. Campos (2001), Evidence for earthquake interaction in central Chile: The July 1997–September 1998 sequence, *Geophys. Res. Lett.*, *28*, 2743–2746.

Lemoine, A., R. Madariaga, and J. Campos (2002), Slab-pull and slab-push earthquakes in the Mexican, Chilean and Peruvian subduction zones, *Phys. Earth Planet. Inter.*, *132*, 157–175.

Lin, J., and R. S. Stein (2004), Stress triggering in thrust and subduction earthquakes between the southern San Andreas and nearby thrust and strike-slip faults, *J. Geophys. Res.*, *109*, B02303, doi:10.1029/2003JB002607.

Linde, A. T., and P. G. Silver (1989), Elevation changes and the great 1960 Chilean earthquake: Support for aseismic slip, *Geophys. Res. Lett.*, *16*, 1305–1308.

Lomnitz, C. (1971), Grandes terremotos y tsunamis en Chile durante el período 1535–1955, *Geofis. Panam.*, *1*, 151–178.

Lowry, A. R., K. M. Larson, V. Kostoglodov, and R. Bilham (2001), Transient slip on the subduction interface in Guerrero, southern Mexico, *Geophys. Res. Lett.*, *28*, 3753–3756.

Madariaga, R. (1998), Sismicidad de Chile, *Fis. Tierra*, *10*, 221–258.

Malgrange, M., A. Deschamps, and R. Madariaga (1981), Thrust and extensional faulting under the Chilean coast: 1965 and 1971 Aconcagua earthquakes, *Geophys. J. R. Astron. Soc.*, *66*, 313–332.

Mikumo, T., S. H. Singh, and M. A. Santoyo (1999), A possible stress interaction between large thrust and normal faulting earthquakes in the Mexican subduction zone, *Bull. Seismol. Soc. Am.*, *89*, 1418–1427.
Nábělek, J. (1984), Determination of earthquake source parameters from inversion of body waves, Ph.D. thesis, Mass. Inst. of Technol., Cambridge, Mass.
Norabuena, E. O., T. H. Dixon, S. Stein, and C. G. A. Harrison (1999), Decelerating Nazca–South America and Nazca-Pacific plate motions, *Geophys. Res. Lett.*, *26*, 3405–3408.
Nostro, C., M. Cocco, and M. E. Belardinelli (1997), Static stress changes in extensional regimes: An application to southern Apennines (Italy), *Bull. Seismol. Soc. Am.*, *87*, 234–248.
Okada, Y. (1985), Surface deformation due to shear and tensile faults in a half-space, *Bull. Seismol. Soc. Am.*, *75*, 1135–1154.
Okada, Y. (1992), Internal deformation due to shear and tensil faults in a half-space, *Bull. Seismol. Soc. Am.*, *82*, 1018–1040.
Oleskevich, D., R. Hyndman, and K. Wang (1999), The updip and downdip limits to great subduction zone earthquakes: Thermal and structural models of Cascadia, South Alaska, SW Japan, and Chile, *J. Geophys. Res.*, *104*, 14,956–14,991.
Pardo, M., D. Compte, and T. Monfret (2002a), Seismotectonic and stress distribution in the central Chile subduction zone, *J. S. Am. Earth Sci.*, *15*, 11–22.
Pardo, M., D. Compte, T. Monfret, R. Boroschek, and M. Astroza (2002b), The October 15, 1997 Punitaqui earthquake (M_w = 7.1): A destructive event within the subducting Nazca plate in central Chile, *Tectonophysics*, *345*, 199–210.
Poliakov, A., R. Dmowska, and J. R. Rice (2002), Dynamic shear rupture interactions with fault bends and off-axis secondary faulting, *J. Geophys. Res.*, *107*(B11), 2295, doi:10.1029/2001JB000572.
Robinson, R. (2003), Fault interactions and subduction tectonics: A reexamination of the Weber, New Zealand, earthquake sequence of 1990, *Geophys. J. Int.*, *154*, 745–753.
Savage, J. C. (1983), A dislocation model of strain accumulation and release at a subduction zone, *J. Geophys. Res.*, *88*, 4984–4996.
Stein, R. S., G. C. P. King, and J. Lin (1992), Change in failure stress on the southern San Andreas fault system caused by the 1992 magnitude = 7.4 Landers earthquake, *Science*, *258*, 1328–1332.
Tichelaar, B. W., and L. J. Ruff (1991), Seismic coupling along the Chilean subduction zone, *J. Geophys. Res.*, *96*, 11,997–12,022.
Wessel, P., and W. F. Smith (1991), Free software helps map and display data, *Eos Trans. AGU*, *72*, 441.

J. Campos, Departamento de Geofísica, Universidad de Chile, Casilla 2777, Santiago, Chile.

A. Gardi, Institut de Physique du Globe, Paris F-75252 Cedex 05, France. (gardi@ipgp.jussieu.fr)

A. Lemoine, Risques Sismiques, ARN, BRGM, 3, avenue Claude Guillemin, BP 36009, F-45060 Orléans Cedex 02, France.

R. Madariaga, Laboratoire de Géologie, Ecole Normale Supérieure, Paris F-75231 Paris Cedex 05, France.