Originally published as:


DOI: 10.1029/2009GL039276
Impact of megathrust geometry on inversion of coseismic slip from geodetic data: Application to the 1960 Chile earthquake

M. S. Moreno,\textsuperscript{1} J. Bolte,\textsuperscript{2} J. Klotz,\textsuperscript{1} and D. Melnick\textsuperscript{3}

\textsuperscript{1}Department of Geodesy and Remote Sensing, Helmholtz-Zentrum Potsdam, Germany

\textsuperscript{2}Zentrum für Astronomie und Astrophysik, Technische Universität Berlin, Germany

\textsuperscript{3}Institut für Geowissenschaften, Universität Potsdam, Germany
We analyze the role of megathrust geometry on slip estimation using the 1960 Chile earthquake ($M_W=9.5$) as an example. A variable slip distribution for this earthquake has been derived by Barrientos and Ward [1990] applying an elastic dislocation model with a planar fault geometry. Their model shows slip patches at 80-110 km depth, isolated from the seimogenic zone, interpreted as aseismic slip. We invert the same geodetic data set using a finite element model (FEM) with precise geometry derived from geophysical data. Isoparametric FEM is implemented to constrain the slip distribution of curve-shaped elements. Slip resolved by our precise geometry model is limited to the shallow region of the plate interface suggesting that the deep patches of moment were most likely an artifact of the planar geometry. Our study emphasizes the importance of fault geometry on slip estimation of large earthquakes.
1. Introduction

Spatial and temporal distribution of slip along subduction zones provides primary information for understanding the earthquake process, and consequently to forecast the possible locus of future earthquakes. Inversions of geodetic data have been broadly used to infer coseismic slip on faults. It has been well recognized that inversion procedures may affect the slip distribution solution [Beresnev, 2003]. Incorrect assumptions about fault geometries [Lee et al., 2006], crustal structure [Savage, 1987; Masterlark, 2003], and smoothing constraints [Du et al., 1992] may significantly affect the predicted slip distribution. Moreover, the spatial heterogeneity of the model resolution may lead to artifacts in poorly resolved areas of the fault plane [Page et al., 2009]. Slip artifacts may appear as robust features and thus may be wrongly interpreted as seismic asperities.

The great Chile earthquake of 22 May 1960 ($M_W=9.5$) constitutes the largest recorded seismic event [Engdahl and Villaseñor, 2002]. The earthquake ruptured about 1000 km of the interface between the subducting Nazca plate and the South American continent (Figure 1). Coseismic data determined from relative sea-level changes, leveling lines, and triangulations [Plafker and Savage, 1970] indicated maximum uplift of 5 m in the outer shelf and subsidence of up to 2.7 m in the mainland. Barrientos and Ward [1990] used a dislocation model with planar slab geometry to invert the available geodetic data into a variable slip planar distribution (VSP model). Their results show a main body of slip along the shallow part of the megathrust with up to 41 m of displacement (Figure 2a). In addition, three patches appear isolated from the main body at 80-110 km depth below
the volcanic arc. These deep patches were interpreted as aseismic slip at a depth of more than 40 km beneath the down-dip limits of the rupture.

In recent years, the crustal structure of the south-central Chile margin has been derived from local seismological networks [Bohm et al., 2002; Haberland et al., 2006; Lange et al., 2007], deep seismic reflection and refraction profiles [Lüth et al., 2003; Groß et al., 2007], and forward modeling of the gravity field [Tassara et al., 2006; Tašárová, 2007]. The precise geometry of the slab obtained from combining these geophysical data sets differs notably from Barrientos and Ward’s assumption of a planar slab; in the present image the dip angle has significant changes along the strike of the margin (Figure S1).

In order to understand the deformation processes related to this event, we attempt to prove if these deep slip zones are needed to explain the data or if this feature may be an artifact of the inversion. We invert the geodetic data employed by Barrientos and Ward [1990] using a 3D finite element model (FEM) with precise fault geometry. The complex geometry leads to non-uniform distance of the nodes and non-rectangular elements on the fault. This requires the application of isoparametric finite elements with quadratic shape functions.

2. FEM inversion

We constructed a 3D FEM that incorporates the precise geometry of the slab and continental Moho, as well as topography and bathymetry (Figure S2). To test the possible occurrence of deep slip, we used in our inversion the fault nodes localized above a depth of ∼240 km. This selection consisted of 773 pair of nodes. Only thrust slip components were considered based on seismological [Cifuentes and Silver, 1989; Kanamori and Cipar,
and geodetic [Plafker and Savage, 1970] studies, which have suggested that the earthquake mechanism was almost entirely dip-slip. Details of the model design can be found in the Auxiliary Material.

The relation between surface deformation and slip along the fault is expressed by the linear system:

\[ Gx ≈ d, \]  \hspace{1cm} (1)

where \( G \) is the synthetic Green's function matrix, \( x \) is the unknown slip vector and \( d \) is the surface observation vector. The matrix \( G \) integrates geometry, material properties and meshing of the FEM model, constructed following the technique of Masterlark [2003]. In doing so, a forward model runs as many times as the number of nodes on the fault. In each run a unity dislocation is applied to a pair of nodes and null dislocation for all other nodes on the fault.

Constraints that minimize the differences between neighboring nodes and weight the observations [Wald and Heaton, 1994] are added to the system of equations:

\[
\begin{pmatrix}
 w^{-1}G \\
 \lambda S
\end{pmatrix} x \approx \begin{pmatrix}
 w^{-1}d \\
 0
\end{pmatrix}.
\]  \hspace{1cm} (2)

Here, \( S \) is the matrix of smoothing constraints, \( \lambda \) is the linear smoothing coefficient and \( w \) is the diagonal matrix, which normalizes and weights the data. We applied isoparametric finite elements with quadratic shape functions to construct the matrix \( S \). The constraint matrix is the stiffness matrix for the 2D Laplace problem, which forms the linear system of equations in a finite element analysis (See Auxiliary Material for details).
The MATLAB routine `lsqlin`, a subspace trust-region method based on the interior-reflective Newton method described in Coleman and Li [1996], was used to solve the inversion. Minimum and maximum slip constraints were applied to avoid models with unreasonable slip patterns and to improve the model resolution [Du et al., 1992; Harris and Segall, 1987]. No additional constraints were imposed at the upper, lower, and lateral fault borders. The smoothing parameter was estimated from the trade-off curve between misfit and slip roughness. The selected value is obtained in the inflection of the curve and gives an optimal balance between data fit and model roughness [Du et al., 1992; Bürgmann et al., 2005]. To calculate a geodetic moment, we introduced the effective shear modulus [Wu and Chen, 2003]. Error propagation from the data to the model is described in the Auxiliary Material.

We inverted the geodetic data compiled by Plafker and Savage [1970], which were used by Barrientos and Ward [1990] (Figure 1). The data consist of: 1) Vertical displacements of 166 sites along the coast, observed in 1968. These observations may include postseismic effects, although Plafker and Savage reported that their data are consistent with about 20 measurements obtained shortly after the earthquake. 2) Inland elevation differences obtained from a leveling line surveyed in 1957-1959 and 1963-1964. Unfortunately, the original data are not available, therefore we digitalized the leveling line from Plafker and Savage [1970] in the form of 150 points. 3) Surface shear strain of eight sites calculated from the angle changes along a triangulation arc, observed in 1950-1952 and 1966-1968.
In order to gain insight into the influence of megathrust geometry on slip inversion, we constructed a second FEM using a planar fault that replicates the parameters of the VSP model (Figure S2b).

3. Results

Inversion of synthetically generated data indicates that our curved fault model successfully recover slip above ~150 km depth (Figure S3). Inversions over a wide range of $\lambda$ values were performed to obtain a trade-off curve (Figure S4a). The optimal value for the smoothing coefficient was found at the inflection point of the curve, $\lambda = 7.5 \times 10^{-6}$. This inflection was also observed in the curve of $\lambda$ versus earthquake magnitude (Figure S4b). The preferred slip distribution produced a geodetic moment of $9.64 \pm 0.52 \times 10^{22}$ N m, equivalent to a moment magnitude of $9.26 \pm 0.02$. The model fits the vertical data with a root mean square (RMS) of 0.63 m.

Independent of the degree of smoothing chosen, the resolved slip consists of a compact but heterogeneous slip body located exclusively under the continental slope, shelf, and Coastal Range (Figure S5). The earthquake moment of the preferred slip model (Figure 2b) is mainly released along the northern segment of the fault (38-42° S). In this region, the rupture propagated to the trench and its down-dip limit reached to a depth of ~55 km. Contrastingly, in the southern segment of the rupture (42-46° S) slip propagation stopped before the trench and its down-dip limit progressively shallowed to a depth of ~30 km. The slip magnitude increased from 20 m in the hypocentral area to a peak of 44 m at 40.5° S. Farther south the dislocation gradually decreased to 10 m at 42-43° S.
exhibiting a southernmost peak of 32 m at 44.7° S. No slip appears deeper than ~55 km along the entire rupture zone.

The slip distribution resulted from the FEM with a planar fault (Figure 2c) generates a geodetic moment of $9.72 \pm 0.62 \times 10^{22}$ N m, equivalent to an earthquake magnitude of $9.26 \pm 0.02$. Vertical observations are simulated with a RMS of 0.78 m, which is 24% higher than the RMS obtained from the precise geometry model. The spatial distribution of the slip shows a main body parallel to the coast and three patches of deep slip isolated from the main body at ~80-150 km depth below the volcanic arc. The main slip body is similar in magnitude and shape to the VSP model (Figure 2a). The deep patches of slip are comparable in magnitude to the isolated patches of the VSP model, but not equally distributed. These differences may arise from the inversion methodologies; Barrientos and Ward [1990] have solved an ill-posed inverse problem using an iterative gradient technique. Conversely, we have introduced constraints to restore the well-posed character of the solution.

Surface vertical displacements and horizontal strains predicted by the model with precise geometry are shown in Figure 3 together with the geodetic data. The synthetic vertical changes are well correlated in magnitude with most of the observations except for a few isolated points located east of the main rupture in the southernmost region (Figure S7 and S8). These sites show anomalous uplift, not reproduced by the coseismic simulation. On average, the entire continental shelf was uplifted over 2 m. The largest predicted vertical deformation is located at the northern segment of the rupture, with maximum uplift and subsidence of ~8.3 m and 2.5 m, respectively.
4. Discussion

Our FEM modeling has shown that patterns of coseismic slip differ significantly between models with precise and simplified geometries (Figure 2). The precise 3D FEM model shows slip exclusively along the seismogenic zone, as expected by the thermal control on the seismic behavior of the megathrust [Hyndman and Wang, 1995]. In turn, isolated patches of slip at 80-110 km depth, below the volcanic arc, result only from the planar models. Coseismic frictional slip in this deep and hot sector of the plate interface is not expected to be triggered by a shallow megathrust earthquake, as continuous viscous flow is expected to dominate [Bürgmann and Dresen, 2008]. Thus, we suggest that the patches of deep slip are artifacts of the inversion arising from the simplified, planar slab geometry.

The slip distribution of the precise 3D FEM model is similar to the main body of slip derived from the VSP model, but provides detailed information about the rupture distribution. The fault dislocation is spatially distributed into adjacent asperities, releasing most of the earthquake moment along the northern part and southern edge of the fault (Figure 2b). A slip patch of over 20 m in the hypocentral area and an asperity with slip of more than 25 m at 41.5° S are observed in our results, but not predicted by the VSP model. These patches of higher slip may represent long-term asperities controlled by the inherited geological structure of the margin and heterogeneous density distribution [Wells et al., 2003; Song and Simons, 2003].

Our proposed slip distribution of the 1960 Chile earthquake indicates that the rupture zone narrowed southward. This is likely a result of the gradual variation in age of the
incoming plate along the trench. The slab age decreases southward resulting in an increase in heat-flow, and consequently a narrower seismogenic zone [Hyndman and Wang, 1995].

Stress perturbation on the continental mantle induced by this earthquake has produced a prolonged postseismic viscoelastic deformation distributed over a broad region [Khazaradze et al., 2002; Hu et al., 2004]. The anomalous uplift localized on the south-eastern sector of the rupture may result from this postseismic deformation and/or from afterslip. Alternatively, shallow faults in the overriding plate may have been triggered by the megathrust earthquake producing localized uplift [Moreno et al., 2008; Melnick et al., 2009].

Most inversion studies have adopted simple planar faults or several planar segments to simulate a complex curved fault. Two main reasons may justify this assumption: (1) to avoid the problems of the bending of the fault, e.g. singular buildup of stress, curve-shaped elements, and non-orthogonal array of nodes, and (2) to reduce the computational costs needed for a high-resolution fault geometry. Oglesby and Archuleta [2003] investigated the dynamics of a non-planar fault (two planar segments). They affirmed that dip changes at depth do not affect the predicted fault slip. Konca et al. [2008] suggested that the derived slip is not affected by introducing a curved fault. In contrast, Lee et al. [2006] and [Maerten et al., 2005] introduced a 3D curved fault surface and showed that inversions with simple fault models may lead to inconsistencies and artificial slip distributions. Our study reveals that the slip inversion is very sensitive to the fault geometry, especially in zones outside the observation network. Errors in the fault geometry assumption are nonlinear and may change the slip distribution significantly. By incorporating precise fault
geometry, the Green’s function better represents the fault system, resulting in a more robust and thus realistic slip distribution.

5. Conclusions

Our slip distribution of the 1960 Chile earthquake indicates that most if not all of the moment was released exclusively along the shallow part of the plate interface. We suggest that the patches of deep slip found by Barrientos and Ward [1990] interpreted as aseismic slip are likely an artifact of their simplified, planar fault geometry.

The calculated geodetic moment is $9.64 \pm 0.52 \times 10^{22} \text{ N m}$, which is about 1/5 of the seismic moment obtained for this earthquake [Engdahl and Villaseñor, 2002]. This finding confirms a deficit of the geodetic moment originally reported by Barrientos and Ward [1990].

A stable solution that fits geodetic data to a high degree does not necessarily represent the correct earthquake slip distribution. Our isoparametric FEM confines the slip distribution on complex fault geometries removing artifacts in poorly resolved areas. The flexibility of our method permits the extension to interseismic and postseismic surface deformation scenarios, and to different tectonic settings.

Acknowledgments. Our special thanks go to: Jessica Papke, Klaus Bataille, Matt Miller and Dietrich Lange for fruitful comments and discussions. We also appreciate the valuable comments and suggestions of two anonymous reviewers. Marcos Moreno gratefully acknowledges a scholarship granted by the German Academic Exchange Service (DAAD).
References


Figure 1. Tectonic setting of the south-central Chile margin. Red dotted line indicates approximately the rupture zone of the 1960 Chile earthquake. Geodetic data used in the slip inversion are shown [Plafker and Savage, 1970].
Figure 2. Slip distributions of the 1960 event depicted by contour lines every 5 m. a) Slip predicted by the VSP model of Barrientos and Ward [1990]. b) Slip predicted by the precise 3D FEM model. c) Slip predicted by the FEM model with a planar fault.
Figure 3. Coseismic surface deformation predicted by the precise 3D FEM model and geodetic observations are colored by the vertical displacements.
Impact of megathrust geometry on inversion of coseismic slip from geodetic data: Application to the 1960 Chile earthquake

M. S. Moreno\textsuperscript{1} J. Bolte,\textsuperscript{2} J. Klotz,\textsuperscript{1} and D. Melnick,\textsuperscript{3}

 Auxiliary Material

Introduction

Supplementary material for this article contains a model set up description, a model resolution test, description of the smoothing constraints, Gaussian error propagation from the data to the model, and Figures S1-S8.

\textsuperscript{1}Helmholtz-Zentrum Potsdam, Germany

\textsuperscript{2}Zentrum für Astronomie und Astrophysik, Technische Universität Berlin, Germany

\textsuperscript{3}Institut für Geowissenschaften, Universität Potsdam, Germany
1. Model set up

All the FEM calculations were done using the ANSYS® Academic Research, v. 11.0 software. The surface deformation was predicted kinematically from fault-slip regardless of the loading mechanism and friction properties. Fault-slip was implemented using the split-node technique [Melosh and Raefsky, 1981] applying linear constraint equations. This method allows to introduce fault displacements into FEM simulating the double-couple acting on the fault. After applying the constraint equation, nodes at both sides of the fault are equally displaced but in opposite directions. Nodes are forced to remain on the fault and consequently can only slide along the interface. The upper surface of the model was assumed to be stress-free allowing freely deformation, whereas lateral and base boundaries were fixed from orthogonal displacements.

A 3D-spherical FEM incorporated topography and bathymetry [Smith and Sandwell, 1997], as well a precise geometry of the slab and continental Moho, which were derived from combining local seismological networks [Bohm et al., 2002; Haberland et al., 2006; Lange et al., 2007], seismic reflection and refraction profiles [Lüth et al., 2003; Groß et al., 2007], and forward models of the gravity field [Tassara et al., 2006; Tašárová, 2007] (Figure S1).

Our model structure extents from 78-60° W to 34-48° S and to a depth of 500 km. The deformation in our study area varies insignificantly by expanding the lateral and bottom boundaries of the mesh. Therefore, the study area is most likely far enough from the boundary constraint effects. The fault plane is a deformable contact surface consisting of eight-node elements, whereas the rest of the mesh is composed of 10-node tetrahedral-
shaped elements. These are quadratic elements with mid-side nodes that follow curved geometric boundaries, well suited to model irregular meshes. The element size is between 10 and 20 km near the fault zone, and 25 km in the rest of the upper crust, whereas in the oceanic crust and mantle it is 30 and 50 km, respectively. The node spacing is \( \sim 5 \) km in the upper surface. Our FEM consists of an elastic upper plate, an elastic subducting plate, a viscoelastic continental mantle, and a viscoelastic oceanic mantle (Figure S2). Though coseismic static slip is independent of viscoelasticity, we have designed a model that can be used to model different phases of the seismic cycle. The thickness of the elastic oceanic plate was set to 30 km [Watt and Zhong, 2000], while the bottom limit of the elastic upper plate was defined by the continental Moho (Figure S2). The modeled continental crust is on average 40-45 km thick, with local extremes of 53-55 km (maximum), and \( \sim 25 \) km (minimum). We specified a Young’s modulus of 100, 120 and 160 GPa, for the continental, oceanic, and mantle layers, respectively, following Hu et al. [2004] and Moreno et al. [2008]. The Poisson’s ratio was set to 0.265 and 0.30 for continental, and oceanic crust, respectively [Christensen, 1996]. The viscosity was set to \( 2.5 \times 10^{19} \) and \( 1 \times 10^{19} \) Pa s for the continental and oceanic mantle, respectively [Hu et al., 2004]. Density values of 2700 and 3300 kg/m\(^3\) were used for the continental and oceanic layers, respectively.

2. Model resolution

We tested the resolving power of the geodetic network and our inversion method by attempting to recover an assumed slip distribution between 80-200 km depth. Synthetically generated fault-slip patterns that mimic the deep slip zones obtained by Barrientos and Ward [1990] (Figure S3a) were used to calculate vertical and horizontal displacements at
the geodetic sites. Observation uncertainties were added to the surface displacements to make a set of synthetic data. Subsequently, we inverted for the distribution of slip using separately both vertical and horizontal displacements. The test results indicate that the corresponding best-fitting models (Figure S3 b-c) closely recover the location, geometry, magnitude and gradient amplitude of the synthetic slip patches. The models based on vertical and horizontal data fit the synthetic displacements with a RMS of 1.61 and 1.50 m, respectively. Dislocation resolution is poorest in both models at depths greater than 150 km and improves directly beneath the geodetic network. We conclude that our curved fault model is able to recover slip above ~150 km depth. Furthermore, lower amount of horizontal than vertical data may not affect the distribution of slip at depth.

3. Smoothing Constraints

The complex geometry leads to non-uniform distance of the nodes and non-rectangular elements on the fault. This requires the application of isoparametric finite elements with quadratic shape functions to construct the matrix of smoothing constraints, $S$. This method makes use of curved quadrilaterals with parabolic boundaries. This approach allows for a quadratic displacement field in every curved quadrilateral with inter-element continuity.

The matrix $S$ is the stiffness matrix for the 2D Laplace problem which forms the linear system of equations in a finite element analysis. With $\mathcal{T}$ the domain of the fault, $T \in \mathcal{T}$ the finite element domains with curved boundaries, and $\mathcal{N}$ the set of the nodes on the fault, the matrix $S$ of the smoothing constraints is given by Bartels et al. [2006] as:

$$S = (S_{z,z'})_{z,z' \in \mathcal{N}} = \left( \int_{\Omega} \nabla N_z \cdot \nabla N_{z'} \; dx \right)_{z,z' \in \mathcal{N}}.$$

(1)
Where
\[
N_z|_T = \begin{cases} 
\varphi_j \circ \Phi_T^{-1} & \text{if } z \in T, \\
0 & \text{if } s \notin T
\end{cases}
\] (2)
is the isoparametric quadratic shape function for the node \( z \) and
\[
\varphi_1(\xi, \eta) = (1 - \xi)(1 - \eta)/4,
\]
\[
\varphi_2(\xi, \eta) = (1 + \xi)(1 - \eta)/4,
\]
\[
\varphi_3(\xi, \eta) = (1 + \xi)(1 + \eta)/4,
\]
\[
\varphi_4(\xi, \eta) = (1 - \xi)(1 + \eta)/4, \text{ if } z \text{ one of the four corner nodes or }
\]
\[
\varphi_5(\xi, \eta) = (1 - \xi^2)(1 - \eta)/2,
\]
\[
\varphi_6(\xi, \eta) = (1 + \xi)(1 - \eta^2)/2,
\]
\[
\varphi_7(\xi, \eta) = (1 - \xi^2)(1 + \eta)/2,
\]
\[
\varphi_8(\xi, \eta) = (1 - \xi)(1 - \eta^2)/2
\]
if \( z \) one of the four midpoints of the line segments connecting the corner nodes. The
diffeomorphismus \( \Phi_T \) maps the curved quadrilateral \( T \) into the reference quadrilateral
\( Q_{\text{ref}} = [-1, 1]^2 \).

4. Gaussian error propagation

To calculate the uncertainties of the slip and earthquake magnitude, we compute the
propagation of errors as follows:

For \( j = 1, \ldots, m \) we calculate \( \epsilon^{(j)} \) as the \texttt{lsqlin} (MATLAB function) solution of
\[
\begin{pmatrix}
w^{-1}G \\
\lambda S
\end{pmatrix} \epsilon^{(j)} = \begin{pmatrix}
w^{-1} \epsilon^{(j)} \\
0
\end{pmatrix},
\] (3)
where \( \epsilon^{(j)} \) is the canonical unit vector with \( \epsilon^{(j)}_i = \delta_{i,j} \) and \( \delta_{i,j} \) the Kronecker delta.
Let us now define

\[ y^{(j)} = (\Delta d_j)^2 \begin{pmatrix} (\epsilon_1^{(j)})^2 \\ (\epsilon_2^{(j)})^2 \\ \vdots \\ (\epsilon_n^{(j)})^2 \end{pmatrix}. \] (4)

For the uncertainties of the slip \( \Delta x_k, k = 1, \ldots, n \); we have

\[ \Delta x_k = \sqrt{y_k^{(1)} + y_k^{(2)} + \cdots + y_k^{(m)}}. \] (5)

If we call \( u_\ell, \ell = 1, \ldots, L \) the average slip in the element \( \ell \), i.e.

\[ u_\ell = \left( \sum_{k \text{ node in element } \ell} x_k \right) / \text{(no. of nodes in element } \ell), \] (6)

then there holds for the uncertainty of the average slip

\[ \Delta u_\ell = \left( \sqrt{\sum_{k \text{ node in element } \ell} (\Delta x_k)^2} \right) / \text{(no. of nodes in element } \ell), \] (7)

and for the uncertainty of the magnitude

\[ \Delta M = \mu \cdot \sqrt{\sum_{\ell=1}^{L} (A_\ell)^2 \cdot (\Delta u_\ell)^2}, \] (8)

where \( \mu \) is the effective shear modulus [Wu and Chen, 2003].

The resulting fault-slip errors mapped on the fault plane are shown in Figure S6.

References


Figure captions

Figure S1. Geophysical data used to construct the 3D geometry of the south-central Chile margin. Contour lines indicate the depth of the upper surface of the slab.

Figure S2. a) Geometry of the south-central Chile margin incorporated in our precise 3D FEM model. b) Planar fault geometry that replicated fault parameters of the VSP model and incorporated in our planar fault model. c) 3-D finite element model configuration.

Figure S3. Resolution ability of the network and inversion method to recover synthetically introduced slip patches at 80-200 km depth. a) Initial synthetic fault-slip patterns that mimic the deep slip patches found by Barrientos and Ward [1990]. b) Result from inverting the network vertical velocities due to the synthetic fault-slip pattern. c) Result from inverting the network horizontal velocities due to the synthetic fault-slip pattern. d-e) Tradeoff curves between the roughness and misfit of the slip distribution for the models that use vertical (d) and horizontal (e) data.

Figure S4. a) Tradeoff curve between model roughness and misfit of the slip distribution for a broad range of smoothing coefficient ($\lambda$). The misfit is expressed by the residual norm ($\| \sum_d (d - Gx) \|^2$). The optimal value for $\lambda$ is $7.5 \times 10^{-6}$ located on the center of the bend of the curve. b) Residual norm as a function of $\lambda$ (green curve) and earthquake magnitude as a function of $\lambda$ (blue curve).

Figure S5. Distributed slip model using alternative smoothing coefficients, $\lambda$, ranging from oversmoothed (a-b) to undersmoothed (d-f). Optimal model found from a tradeoff curve (Figure S4) is shown in c.
**Figure S6.** Fault-slip errors mapped on the fault plane. Black contour lines indicate the slip distribution error every 1 m. Estimated coseismic slip is depicted by orange contour lines every 20 m.

**Figure S7.** Margin normal profiles showing observed (blue dots) and synthetic (red lines) vertical displacements. Below is plotted the slab geometry from the curved (red lines), and planar models (green lines).

**Figure S8.** Margin parallel profile showing the vertical displacements of the leveling line (blue line) and the synthetic vertical displacements (red line).
Figure S2
Figure S3
Figure S4
Figure S5
Figure S6
Figure S7