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Block modeling of crustal deformation in Tierra del Fuego from GNSS velocities

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ABSTRACT

The Tierra del Fuego (TDF) main island is divided by a major transform boundary between the South America and Scotia tectonic plates. Using a block model, we infer slip rates, locking depths and inclinations of active faults in TDF from inversion of site velocities derived from Global Navigation Satellite System observations. We use interseismic velocities from 48 sites, obtained from field measurements spanning 20 years. Euler vectors consistent with a simple seismic cycle are estimated for each block. In addition, we introduce far-field information into the modeling by applying constraints on Euler vectors of major tectonic plates. The difference between model and observed surface deformation near the Magallanes Fagnano Fault System (MFS) is reduced by considering finite dip in the forward model. For this tectonic boundary global plate circuits models predict relative movements between 7 and 9 mm yr^{-1}, while our regional model indicates that a strike-slip rate of 5.9 ± 0.2 mm yr^{-1} is accommodated across the MFS. Our results indicate faults dipping 66^{+6}_{-4}° southward, locked to a depth of 11^{+5}_{-5} km, which are consistent with geological models for the MFS. However, normal slip also dominates the fault perpendicular motion throughout the eastern MFS, with a maximum rate along the Fagnano Lake.

1. Introduction

The Tierra del Fuego (TDF) main island, southernmost South America, is bisected by the active transform boundary between the South America (SAM) and Scotia (SCO) tectonic plates (Pelayo and Wiens, 1989). In the TDF region, the plate boundary is represented by the Magallanes Fagnano Fault System (MFS, Lodolo et al., 2003). The MFS extends from the Atlantic offshore to the western arm of the Strait of Magellan and splits the TDF island into two continental blocks. The E–W strike and relative movement of the MFS through the southern part of the island produce robust topography and geomorphology, expressed as a series of lineaments and depressions (Irigoyen River valley, Turbio River valley, Fagnano Lake) (e.g., Lodolo et al., 2003; Menichetti et al., 2008). The main trace of the MFS probably provides the trans-extensional forces that produce the 105 km-long Fagnano Lake. This trace exits the lake toward the northwest.

In 1993, geodetic GPS observations were commenced in the Argentine part of the TDF region with the aim to determine the recent relative crustal movements along the MFS. These observations allowed the first detection of the relative horizontal displacement between the northern and southern part of the island (Del Cogliano et al., 2000). Based on an independent set of GPS data, of limited spatial resolution and observation time-span, a first attempt was made to characterize the tectonic deformation along the MFS by a simple kinematic model (Smalley et al., 2003). However, this model assumed an infinitely extended, vertical fault plane, neglecting the actual geometry of the faults' surface traces, and was not able to reliably resolve a locking depth nor a fault inclination. A substantial densification of our regional GPS network lead to a detailed quantitative description of the horizontal surface deformation by means of a strain analysis (Mendoza et al., 2011). These results were based on episodic GPS-only observations carried out until 2008. However, no model for the deformation was proposed.

In this paper we use high quality GNSS data that have enough spatial coverage to begin probing the deep structure of fault slip in this major continental transform tectonic system. Our analysis derived from these geodetic observations aims to complement the knowledge on the present-day dynamics of the MFS. Understanding the many subtle aspects of physical structure of such systems (e.g., Moody and Hill, 1956; Sylvester, 1988) may aid in illuminating the physics controlling aseismic and seismic partitioning of deformation. More generally, this is also a problem of great societal importance since the earthquake potential of this strike-slip system in TDF could endanger population centers and fundamental infrastructure (e.g. natural gas pipelines).

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2. Data and methods

2.1. Geodetic observations

We collected campaign-type GPS observations in the TDF area since late 1993. Moreover, the number of observed sites has been steadily growing, and during 2010 to 2013 many semi-permanent GNSS stations were installed, recording continuously for several months. In addition, selected tracking stations from the International GNSS Service (IGS) were included in the analysis, some of which have GPS + GLONASS capabilities since 2006. In total, the regional measurements set spans 20 years, and has sufficient precision and distribution to reliably infer deformation at near-fault spatial scale. This densification also allows for the estimation of a second invariant of the strain rate tensor, possibly with improved signal to noise ratio (e.g., Kreemer et al., 2014).

2.2. GNSS data analysis

The observations were processed with the *Bernese Software* version 5.1 (Dach et al., 2007), and models recommended by the International Earth Rotation Service (IERS) were used (Petit and Luzum, 2010). In addition, ocean tidal loading corrections, according to Savcenko and Bosch (2012), and absolute phase-center corrections for satellites and receivers, as issued by the IGS (file *IGS08.atx*), were applied. The tropospheric delay was modeled with the Global Mapping Function (GMF, Boehm et al., 2006), including 2-hourly zenith delay estimates. First order ionospheric delays were eliminated by means of the ionosphere-free linear combination, and higher-order terms were modeled according to Fritsche et al. (2005). The IGS08 reference frame (Rebischung et al., 2012) was introduced by means of constraints on coordinates of selected IGS tracking sites. To assure a homogeneous set of GPS + GLONASS precise orbits and clocks, and consistent Earth Orientation Parameters (EOP), the reprocessed products computed by Fritsche et al. (2014) were used. These reprocessing products extend from late 1993 to early 2013 and were supplemented with operational IGS products covering the time period from 2013 to early 2014.

Too optimistic formal uncertainties for the site velocities are typically obtained from the GNSS data analysis (e.g., Mao et al., 1999). This is due to the fact that a simplified correlation dependency is assumed (non differenced observations are considered to be uncorrelated) and also due to the huge number of observations adjusted (in our case 3.4×10^8 double differenced ionosphere-free linear combinations were simultaneously inverted). Therefore, and in order to obtain realistic uncertainties for the velocity estimates, the position time series were analyzed. For every site, trends and trends' uncertainties for the north and east components, together with white and flicker noise parameters, were estimated by means of the software tool *CATS* (Williams, 2003). On average, the realistic uncertainties were 20 times larger than the formal uncertainties obtained from the least squares adjustment of GNSS observations. Therefore, a scaling factor of $400 = 20^2$ was applied to the whole variance–covariance matrix of the velocity estimates as obtained from the GNSS analysis. In our modeling the velocities resulting from the multi-year GNSS cumulative solution, together with their rescaled uncertainties, were used as input observations.

2.3. Strain analysis

In order to quantify the surface deformation associated to the MFS main deformation zone, a locally uniform strain rate field was computed by inverting the observed velocities (Shen et al., 2007). The three components of the strain rate tensor $(\dot{\epsilon}_{ee}, \dot{\epsilon}_{nn}, \dot{\epsilon}_{en})$, together with a rigid rotation rate, were estimated for each point of a regular grid, every 5 km, near the plate boundary.

This method does not require an optimal Delaunay triangulation nor assume uniform deformation within polygons. The inversions were carried out by least squares adjustments, where the observed velocities were reweighted by a factor Ae^{-d^2/σ^2}. Here, A weights the contribution of each site according to the area of its corresponding cell in a Voronoi tessellation of the sites' locations, whereas d is the distance between the site and the point being evaluated and σ is a smoothing factor, in this case equal to 30 km. Note that only the area of each Voronoi cell was employed in the adjustments. The strain rate field was modeled as a smooth and continuous function, regardless of the cell's boundaries. Finally, the second invariant of the strain rate, defined as $\sqrt{\dot{\epsilon}_{ee}^2 + \dot{\epsilon}_{nn}^2 + 2\dot{\epsilon}_{en}^2}$ (Kreemer et al., 2003), was computed for every point of the grid.

2.4. Block modeling

In this work, the methodology described by Meade and Hager (2005) was applied. The model relates block motions and fault slip rates to observed interseismic deformation, and makes use of the analytic solutions given by Okada (1985) for the surface deformation due to an arbitrarily inclined and finite dislocation in a homogeneous elastic half-space.

Given a block partition (bounded by faults), fault segments $\left(\vec{x}_F\right)$ and site locations $\left(\vec{x}_S\right)$, the interseismic velocity is interpreted as the difference between the block velocity and the coseismic slip deficit (CSD) velocity

$$\vec{v}_I = \vec{v}_B\left(\vec{x}_S\right) - \vec{v}_{\mathrm{CSD}}\left(\vec{x}_S, \vec{x}_F\right). \tag{1}$$

Both the block and the CSD velocities are expressed as linear transformations of Euler poles and rotation rates

$$\vec{v}_B = \mathbf{R}_E\mathbf{R}_B\vec{\Omega} = \mathbf{R}_1\vec{\Omega} \tag{2}$$

$$\vec{v}_{\mathrm{CSD}} = \mathbf{R}_M\mathbf{R}_O\underbrace{\mathbf{R}_F\mathbf{R}_{\Delta\vec{v}}\vec{\Omega}}_{\vec{s}} = \mathbf{R}_2\vec{\Omega} \tag{3}$$

where $\mathbf{R}_B\vec{\Omega} = \vec{\Omega} \times \vec{x}_S$ account for the relative block rotation, $\mathbf{R}_E$ transforms from a geocentric $\{x, y, z\}$ into a local $\{e, n\}$ system, $\mathbf{R}_{\Delta\vec{v}}$ gives the fault-parallel and fault-normal relative velocity vector $\Delta\vec{v} = (v_\parallel, v_\perp)$, $\mathbf{R}_F$ projects this velocity vector into Okada's coordinate system, $\mathbf{R}_O$ contains the partial derivatives of Okada's elastic Green's functions with respect to each component of the slip rate vector $\vec{s}$ and $\mathbf{R}_M$ computes the inverse projection from Oblique Mercator (oriented along each fault segment) into a local $\{e, n\}$ system. Finally, if a priori information from geodetic or geophysical studies is available, either as absolute $\left(\vec{\Omega}\right)$ or relative $\left(\vec{\Omega}_{ij}\right)$ Euler poles and rotation rates, the forward model can be written as

$$\begin{pmatrix} \vec{v}_I \\ \vec{\Omega} \\ \vec{\Omega}_{ij} \end{pmatrix} = \begin{pmatrix} \mathbf{R}_1 - \mathbf{R}_2 \\ \mathbf{I} \\ \mathbf{I}_{ij} \end{pmatrix} \vec{\Omega} \tag{4}$$

where $\mathbf{I}$ is the identity matrix and $\mathbf{I}_{ij}$ accounts for the relative rotation of block i with respect to block j. Rewriting this model as $\vec{d} = \mathbf{R}\vec{\Omega}$, the Euler pole and rotation rate of each block can be estimated by linear least squares

$$\hat{\vec{\Omega}} = \left(\mathbf{R}^T\mathbf{W}\mathbf{R}\right)^{-1}\mathbf{R}^T\mathbf{W}\vec{d} \tag{5}$$

where $\mathbf{W}$ contains weights for the observations and appropriate constraints for the pseudo observations (i.e., a priori Euler poles and rotation rates). In turn, the slip vector accommodated by each fault segment can