Estimating the 2008 Quetame (Colombia) earthquake source parameters from seismic data and InSAR measurements

Gabriel Dicelis a,*, Marcelo Assumpção a, James Kellogg b, Patricia Pedraza c, Fábio Dias a

a Institute of Astronomy, Geophysics and Atmospheric Sciences (IAG), University of Sao Paulo, Brazil
b University of South Carolina (USC), USA
c Colombian Geological Survey (SGC), Colombia

A B S T R A C T

Seismic waveforms and geodetic measurements (InSAR) were used to determine the location, focal mechanism and coseismic surface displacements of the Mw 5.9 earthquake which struck the center of Colombia on May 24, 2008. We determined the focal mechanism of the main event using teleseismic P wave arrivals and regional waveform inversion for the moment tensor. We relocated the best set of aftershocks (30 events) with magnitudes larger than 2.0 recorded from May to June 2008 by a temporary local network as well as by stations of the Colombia national network. We successfully estimated coseismic deformation using SAR interferometry, despite distortion in some areas of the interferogram by atmospheric noise. The deformation was compared to synthetic data for rectangular dislocations in an elastic half-space. Nine source parameters (strike, dip, length, width, strike-slip deformation, dip-slip deformation, latitude shift, longitude shift, and minimum depth) were inverted to fit the observed changes in line-of-sight (LOS) toward the satellite four derived parameters were also estimated (rake, average slip, maximum depth and seismic moment). The aftershock relocation, the focal mechanism and the coseismic dislocation model agree with a right-lateral strike-slip fault with nodal planes oriented NE-SW and NW-SE. We use the results of the waveform inversion, radar interferometry and aftershock relocations to identify the high-angle NE-SW nodal plane as the primary fault. The inferred subsurface rupture length is roughly 11 km, which is consistent with the 12 km long distribution of aftershocks. This coseismic model can provide insights on earthquake mechanisms and seismic hazard assessments for the area, including the 8 million residents of Colombia’s nearby capital city Bogota. The 2008 Quetame earthquake appears to be associated with the northeastward “escape” of the North Andean block, and it may help to illuminate how margin-parallel shear slip is partitioned in the Eastern Cordillera.

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1. Introduction

On May 24, 2008, 19:20 UTC a Mw 5.9 earthquake struck the center of Colombia. The Geological Survey of Colombia (SGC) located the earthquake epicenter southeast of the capital Bogotá at a depth of 0.3 km (Erazo and Tovar, 2010). The earthquake affected five municipalities in the department of Cundinamarca: Quetame, Fosca, Guayabetal, Fomeque, and El Calvario in Meta, resulting in six deaths, 65 people injured, and the destruction of the houses of 9000 people. The permanent national network of Colombia (RSNC) is sparse in this area, only one station (CHI) was closer than 70 km to the main event (Fig. 1). In this case determination of precise hypocenters is very difficult. Locations by RSNC and other agencies (NEIC, ISC, GCMT and IDC) differ up to 20 km horizontally, and depth estimates vary from 0 to 35 km (Fig. 1 and Table 1). Two days after the mainshock, SGC deployed a temporary 9-station, 3-component broadband network (Fig. 1). Between May 2008 and February 2009 the mainshock was followed by a sequence of ~1000 events.

The 2008 earthquake occurred in a seismically active region in the mountains of Colombia's Eastern Cordillera. A combination of geological and climatic factors is responsible for the topography of this region, including severe erosion along the southeastern flank of the mountain belt (Mora et al., 2010). Elevations in the Eastern Cordillera vary between 1500 and 4500 m. Bedrock consists mainly of Paleozoic and Cenozoic sedimentary rocks in the highlands, and Mesozoic sedimentary and Cenozoic rocks partially covered by Quaternary sediments on the southeastern mountain flank (Cortés et al., 2005) as shown in Fig. 1.

Three major fault trends have been recognized in the Eastern Cordillera: 1) NW-SE: Rio Negro Fault and El Tabor, 2) NE-SW: parallel to the southeastern mountain front including the Blanca, Naranjal, Servita, Quetame and Santa Rosa reverse fault systems, and 3) E-W: Río Blanco, Quebrada Honda, and Pescado...
faults. Fig. 1 shows the main faults in our study area. The NE-SW fault system is parallel to the tectonic boundary between the Andean block and the stable South American plate (Fig. 1) (Pennington, 1981; Aggarwal, 1983; Freymuller et al., 1993; Kellogg and Vega, 1995; Trenkamp et al., 2002; Colmenares and Zoback, 2003).

Earthquake focal mechanisms in the northern Andes block are generally bimodal: 1) reverse faults with NW-SE P axes causing crustal thickening that result in permanent deformation, shortening and mountain building, and 2) Right lateral strike-slip faulting (ENE-WSW oriented P axes) that result in rigid block translation of the NA block and “escape” to the northeast (Fig. 2) (Ego et al., 1996; Corredor, 2003; Audemard, 2003; Cortés and Angelier, 2005; Egbue and Kellogg, 2010; Egbue et al., 2014; Audemard, 2014). It is not clear how present-day margin-normal and margin-parallel slip is partitioned in the Eastern Cordillera of Colombia. The 2008 Quetame earthquake discussed in this paper is an “escape” earthquake, and provides a unique opportunity to study margin-parallel “escape” slip-partitioning in the Cordillera. Margin-parallel strike-slip faulting is well documented to the southwest in southern Colombia and Ecuador and to the northeast in Venezuela, but is obscured by over 100 km of margin-

Table 1
Published locations and source parameters for the 2008 Quetame earthquake (2008/05/24 19:20:43.85).

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- Unpublished report from SGC (Erazo and Tovar, 2010).
- Fixed parameter.
- Derived from P-wave focal mechanism, Mw derived from ISOLA waveform inversion.

Fig. 2. Earthquake focal mechanisms in the northern Andes block are generally bimodal: 1) reverse faults with NW-SE P axes and 2) Right lateral strike-slip faulting (ENE-WSW oriented P axes). The red box indicates the study area of the Fig. 1. Inset: The black box indicates the extent area of the map. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)
normal shortening in the Eastern Cordillera in the last 10 Ma (summarized in Egbue and Kellogg, 2010). The preliminary study of the main shock and aftershocks by RSNC suggested a NE-SW oriented distribution, with a shallow depth of 0.3 km, for the ML 5.7 mainshock and epicenter at 4.40°N and 73.81°W. The Global Centroid Moment Tensor Catalogue (GCMT) estimated a Mw 5.9 magnitude for the main event (Table 1), with two nodal planes NP1: strike 193°, dip 82° and NP2: strike 89°, dip 84°. Based on the results of the tomography inversion and polarities of first P-wave arrivals to better constrain the focal mechanism, and we use after-shock relative locations to identify the fault plane as well as the depth of the main event. In this paper we show the relocation of the main shock and a set of aftershocks from the period 24 May to June 6, 2008, recorded by the temporary local network as well as by some stations of the national network. The stations of the temporary network were installed at distances of 15–30 km from the initial epicentre.

We used three SAR images, acquired on ascending tracks provided by PALSAR (Phase Array L-band Synthetic Aperture Radar) aboard the ALOS satellite. The study area has rugged topography, high rainfall and dense vegetation, which make L-band SAR imagery more adequate for interferometric analysis, in comparison with C-band, because it penetrates deeper into the canopy, and produces a more coherent signal even in a forest area (Rosenqvist et al., 2007; Tong et al., 2010). Despite the limitations in the existing InSAR dataset and the challenging climatic and geological conditions, the two complementary data types (seismological and geodetic) were useful to constrain the earthquake focal mechanism and location, as well as to distinguish between the two potential conjugate fault planes. Similar conditions were described by Lohman et al. (2002), where distinguishing the fault and auxiliary plane was difficult due to the symmetry in the deformation fields and radiation patterns (Biggs et al., 2006). Seismic waveform inversion and P-wave first motions constrained the fault plane orientation but not the location. InSAR improved the location, especially the centroid depth, and the estimation of the area of the fault plane. The aftershock relocations helped to conclusively identify the orientation and size of the fault plane.

2. Mainshock and aftershock relocation

We relocated the main event (Fig. 1) using the RSTT method (Regional Seismic Travel Times) from Myers et al. (2010). This method uses a 3D model parameterized by nodes spaced 1°. Assigning a crustal velocity profile and a simple velocity gradient in the upper mantle for each node, the interpolation of the velocity profiles generates a 3D model of the crust and upper mantle. In the Andean region of southern Colombia, the crustal thicknesses seems to be close to the global continental average (~40 km). We used an updated RSTT model of South America that adjusted the average crustal velocity, mantle velocity at the Moho, and the mantle velocity gradient at each node and incorporates the crustal thicknesses of Assumpção et al. (2013a, b). In the Northern Andes (Ecuador, Colombia and Venezuela), crustal thicknesses vary from 30 km for stations near the coast to 50 km in the high Andean ranges, which is an improvement over the 1D, 35 km thick model used by the RSNC.

We relocated the aftershocks based on the idea that waveform similarity is caused by proximity in hypocenter location and similarity in focal mechanism between different events. Then the cross correlation of these waveforms provides a waveform similarity measurement and an accurate time shift (Waldhauser and Ellsworth, 2000; Schaff and Waldhauser, 2005; Cleveland and Ammon, 2013). We used the waveforms of the 60 clear aftershocks recorded at 22 stations (Fig. 3) (both temporary and permanent, with minimum distance ranging from 6 to 10 km) with the program CORR of the SEISAN software package (Havskov and Ottemøller, 1999) to compute the cross correlation at the same stations of all pairs combining them in a cross correlation matrix (Got et al., 1994). With this procedure we got the most consistent set of 31 events with high precision cross-correlated travel times.

Assuming that the hypocentral separations between the aftershocks of the Quetame 2008 event are small compared to the event-station distance and the scale length of the velocity heterogeneity, then the effects of errors in structure can also be effectively minimized by relative earthquake location methods (Poupine et al., 1984; Frémont and Malone, 1987; Got et al., 1994). We determined relative relocations of the events using the double difference method (Waldhauser and Ellsworth, 2000; Waldhauser, 2001) using HypoDD code. This program allows optimal event relocation when there are measurement errors and earth model uncertainties, and avoids the need for station corrections. This HypoDD code also accepts the use of ordinary phase picks catalog, high precision cross correlated travel times or a combination of both. We used the 31 events with high precision cross-correlated arrivals and the catalog of P- and S-wave picks arrivals for 61 events (60 aftershocks and the RSTT starting location of the mainshock). The mainshock was not used in the cross correlation process because it was not recorded by the temporal network and its waveform (recorded by the permanent network) was very different from the aftershocks. 30 events were identified as clustered including the mainshock. The average uncertainty of the cluster was 0.6 km in depth, 1.0 km in the north-south direction, and 0.6 km in the east-west direction. 31 events were identified as isolated (hypoDD eliminates events that lose linkage as a result of the reweighting, due to outlier removal or distance cutoff). Fig. 4 shows the double difference locations based on catalog and cross correlation data.

Locations with regional sparse network (RSNC) or teleseismic stations (NEIC, ISC) have errors of about 10–20 km (see Fig. 1). Absolute locations of the aftershocks using the closer temporary stations have much lower errors (a few km). Using hypoDD to relocate the mainshock together with the aftershocks (using both the local and regional network) can significantly improve the absolute location of the mainshock. Most of the aftershocks are aligned and clustered in depth in a relatively narrow plane striking NNE-SSW (020°) and dipping steeply (85°) to the northwest. The alignment of the epicenters along a direction coincident with the strike of the mainshock’s fault mechanism and the strike of the mainshock towards the cluster confirms the improvement of the mainshock relocation (Fig. 4).

3. Seismological determination of source geometry

Table 1 shows various location estimates for this earthquake including the GCMT focal mechanism solution (Ekström et al., 2012) consistent with dextral strike slip on a NNE-SSW plane. In order to estimate and constrain a fault plane solution we read P first arrivals at 81 stations from IRIS-USGS (II and IU), Mednet (MN), USArrays reference network stations (US-REF) and Caribbean (CU) up to a distance of 40°. We also used phases from 22 stations of the RSNC, up to a distance of 1000 km. The single event mechanism of the May 24, 2008 earthquakes is a well-constrained strike-slip fault. The nodal plane with strike 193°, dip 84° to the NW and rake −176° (Fig. 5) agrees well with the plane estimated by the relocation of aftershocks (strike 200°, dip 85°, given by fitting a plane to the relocated hypocenters).

In order to further confirm this focal mechanism we used regional waveform inversion with ISOLA code (ISOLated Asperities; Sokos and Zahradník, 2008), with two broadband 3-component
stations within 300 km of the earthquake (RREF and HEL); other
closer stations were saturated. We assessed the solution quality by
quantifying the variability of the focal mechanism solutions in
source position and time, thereby not relying solely on the wave-
form fits to assess the quality of the solutions. This assessment was
done using two quality indicators: Variance Reduction (VR) and
Condition Number (CN). A VR of 80% indicates a good
fit for the solution. However, a high VR can be obtained using a few stations
and may not guarantee a reliable solution. On the other hand, a low
CN = 2.1 indicates a well conditioned problem, as defined by Sokos
and Zahradník (2008) (See Appendix A).

Following the methodology presented by Dias et al. (2016), for
each station, a velocity model was determined using the Rayleigh
and Love surface wave dispersion. These specific velocity models
were used to calculate the Green’s Functions using the RSTT loca-
tion (see Table 1) and depths ranging between 4 and 25 km. To
evaluate the waveform inversion uncertainty, we used the Fre-
quency Range Test (Dias et al., 2016). This test consists of inverting
for several frequency ranges, and checking the variability of the
solutions (Fig. A1).

Our preferred solution from the waveform inversion is the one
using the frequency band of 0.01–0.04 Hz, because it has the best
fit. The waveform fit and information about the best focal solution
considering both the variance reduction and the polarity fit are in

Fig. 3. Example of aligned P-wave arrivals (vertical component) at station CHI after cross-correlation.

Fig. 4. Left panel: map view of the double difference locations, showing a clear linear trend of the aftershocks oriented NNE-SSW, the red box indicates the orientation of the cross
sections. Top right panel shows the cross section along profile AA’ (N110° E direction) across the fault, with a steep dip angle of 85° to the northwest. Lower right panel shows the
cross section along the NE-SW profile BB’ (N20° E direction). Open star is the mainshock RSTT location, yellow star is double difference relocation of the mainshock. The solid black
triangles are the near stations of the temporary network. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Fig. 5. Fault plane solution of the main event. Xs represent compressional and open circles dilatation of P-wave motion. One nodal plane has strike 193° and dip 84° to NW, corresponding to a strike-slip mechanism with rake -176°. This mechanism agrees with 92.8% of the 152 polarities.

Fig. 6. Solution of the waveform inversion for the RREF and HEL stations. Black traces are the observed displacement waveforms (Z-vertical, N-North, E-East), and the red traces are the synthetics. The P- and S-wave arrival times, station azimuth and distance, as well as the amplitude scale are indicated. The gray area in the beachball shows the range of possible solutions. Legend gives the information about focal mechanism: Decomposition into the DC, CLVD and VOL components, moment magnitude Mw, centroid depth, centroid time shift with respect to origin time (CT), nodal planes (NP1 and NP2), condition number (CN), variance reduction (VR), maximum Kagan angle among the solutions, frequency range of the inversion (FR). The small condition number (CN = 2.1) indicates a well-constrained problem. Variance reduction (VR = 80%) indicates a good fit of the solution. This solution corresponds to a strike-slip fault. One nodal plane strikes 12°, dip 82°, rake 165°, and is consistent with the fault plane solution for P-wave polarities (Fig. 5), and it has a maximum k-angle of 16°. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

The best-fitting solution using ISOLA shows a right-lateral strike-slip fault, striking 12°, dip 82° and rake 165° (Fig. 6). The range of possible solutions (shown in gray) was estimated using a threshold of 0.95 for the VR and polarity fit. Thus, the acceptable solutions have VR between 0.75 and 0.79 and polarity fit between
The cycle of fringes with a half wavelength of ~12 cm for L-band of displacement of the surface toward the satellite is represented by Rosen et al., 2004; Fialko, 2004; Parsons et al., 2006. Werner 1998; Massonnet and Feigl, 1998; Wang et al., 2003; centimeters and even millimeters due to seismic movement.

The interferogram that gives the surface LOS displacement, at ranges of taken before and after an earthquake were used to make an inter- satellite, acquired by the Japanese space agency JAXA, from 2 weeks before to 13 months after the earthquake (Table 2). We stacked the two unwrapped interferograms to obtain an average interferogram to reduce noise artifacts (Sandwell and Price, 1998; Lohman et al., 2002). We used GMTSAR InSAR soft- ware (Sandwell et al., 2011) to produce LOS displacement. All maps and figures in this study were generated using GMT code (Wessel and Smith, 1998).

We removed the contribution to the interferometric phase from the topography and from the curvature of the Earth. However, it is necessary to remove spurious contributions to the radar phase before interpreting them as coseismic surface displacements, because artifact fringes can be introduced in the interferogram by DEM topography errors and large values of perpendicular baselines (component of the orbital separation perpendicular to the LOS). To quantify this effect Massonnet and Rabaud (1993) defined the altitude of ambiguity (Ha) as the shift in altitude needed to produce one topographic fringe, which can perturb the coseismic signature or even cause complete decorrelation in highland areas. We detrended and removed the topographic effects of the interferogram using 30-m SRTM-3, which has an expected vertical accuracy of less than 15 m (Kellendörfer et al., 2004; Smith and Sandwell, 2003). However, the vertical accuracy can vary by as much as 50 m outside North America in high elevations and rugged topography. Thus, to establish the vertical accuracy of the SRTM-3 topographic data in our study area, we used two GPS permanent stations of the Geologic Service of Colombia (BOG and VILL) as control points. These stations did not show significant displacement due to the earthquake (Mora et al., 2009). We did not find significant differences between the SRTM-3 and the control points; for station BOG we found a difference in height of ~1.58 m and for station VILL a difference of 1.27 m. The altitudes of ambiguity of the coseismic interferograms 1 and 2 are 154 m and 534 m, respectively (Table 2). At this range, a 15 topographic error the DEM produces errors in LOS of 0.6 cm and 0.2 cm, respectively, or ~0.05 fraction of a fringe in phase.

Separating the tropospheric noise and the deformation signal is a challenge, particularly when the signal is small (Rigo and Massonnet, 1999; Feigl, 2002). No good atmospheric model is available to correct for the propagation delays in this area. Then we corrected the remaining errors associated with the orbital error and phase delay through the troposphere by estimating the delay to elevation relationship observed in the interferogram. Following Rosen et al. (2004) and Liu and Jung (2014) we fit a two degree polynomial function of elevation (Béjar-Pizarro et al., 2013). The interferogram is mainly smooth (Fig. 7a) except for two concentric artifacts at ~4.1° latitude that we interpret as large tropospheric and ionospheric perturbations (Fig. 7b). We constrain the region to

4. InSAR determination of source geometry

InSAR is a useful tool to detect and analyze moderate coseismic events in areas difficult to access with traditional means, although radar is not as sensitive or comprehensive as seismic data (Mellors and Magistrale, 2004a, 2004b; Lohman and Simons, 2005b). Very few earthquakes with magnitude less than 6 have been studied using InSAR, but examples include: M5 (depth 2 km) 1992 Fawskin earthquake (Feigl and Thurber, 2009), USA; M6.1 (depth 7–9 km) Eureka Valley, USA (Massonnet and Feigl, 1995a,b; Peltzer and Rosen, 1995); M5.4 (depth 2.6 km) 1992 Landers aftershock (Feigl et al., 1995; Massonnet et al., 1993); M5.4 and 5.3 events (depth 5 km and 4.2 km) the Zagros Mountains, Iran (Lohman and Simons, 2005a); the M5.6 (depth 9.4 km) Little Skull mountain (Lohman et al., 2002), M5.0 (depth 8 km) St. Paul de Fenouillet, France (Rigo and Massonnet, 1999); M4.7 (depth 1.2 km) Katanning, Australia (Dawson et al., 2008).

Quetame is located in the central region of Colombia, with highly undulating topography and dense vegetation covering at least 57% of the area and grass and crops covering 31%. The average rainfall is ~2,600 mm/yr. This kind of cover and land use often reduces the InSAR observation quality (Funning et al., 2005a; Engdahl and Magistrale, 2005). In this study we used three SAR images, ac- cording on ascending tracks provided by PALSAR aboard the ALOS satellite, acquired by the Japanese space agency JAXA, from 2 weeks before to 13 months after the earthquake (Table 2).

The phase differences in complex SAR images of the surface taken before and after an earthquake were used to make an inter- ferogram that gives the surface LOS displacement, at ranges of centimeters and even millimeters due to seismic movement (Zebker et al., 1994; Massonnet and Feigl, 1995a,b; Goldstein and Werner 1998; Massonnet and Feigl, 1998; Wang et al., 2003; Rosen et al., 2004; Fialko, 2004; Parsons et al., 2006). The displacement of the surface toward the satellite is represented by a cycle of fringes with a half wavelength of ~12 cm for L-band of ALOS-PALSAR. A complex radar interferogram is created by multiply- ing the reference image by the complex conjugate of the repeat image, producing a phase difference map. This phase difference is affected by many factors such as atmospheric and ionospheric noise, the shape of the earth, topography, orbit error and surface displacement (Bürgmann et al., 2000). One of the limitations of deformation measurements made with InSAR is that interferograms are only sensitive to surface movements in the LOS direction towards or away from the satellite (Wright et al., 2004a, 2004b; Feigl, 2002; Fialko et al., 2001). Given these limitations we retrieved the component of the coseismic deformation in the LOS direction for the available interferograms.

We combined the SAR pre-seismic image with two post- earthquake images to create two different interferograms (Table 2). We stacked the two unwrapped interferograms to obtain an average interferogram to reduce noise artifacts (Sandwell and Price, 1998; Lohman et al., 2002). We used GMTSAR InSAR soft- ware (Sandwell et al., 2011) to produce LOS displacement. All maps and figures in this study were generated using GMT code (Wessel and Smith, 1998).

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Separating the tropospheric noise and the deformation signal is a challenge, particularly when the signal is small (Rigo and Massonnet, 1999; Feigl, 2002). No good atmospheric model is available to correct for the propagation delays in this area. Then we corrected the remaining errors associated with the orbital error and phase delay through the troposphere by estimating the delay to elevation relationship observed in the interferogram. Following Rosen et al. (2004) and Liu and Jung (2014) we fit a two degree polynomial function of elevation (Béjar-Pizarro et al., 2013). The interferogram is mainly smooth (Fig. 7a) except for two concentric artifacts at ~4.1° latitude that we interpret as large tropospheric and ionospheric perturbations (Fig. 7b). We constrain the region to

Table 2

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<td>A 18235</td>
<td>2009/06/27</td>
<td>367</td>
<td>−448</td>
</tr>
</tbody>
</table>
unwrap the phase (red box in Fig. 7b) in order to avoid the gaps with no phase signal (correlation gaps) and the two large tropospheric and ionospheric perturbations. We used as common master the 2008/05/09 image (Table 2) with LOS vector 34.3°/C14° from vertical. In this convention LOS displacement away from the spacecraft indicates subsidence. The distance measured along the LOS vector between the satellite and the ground surface is given by \( r = u \cdot s \)

where \( s \) is the unit vector which points from ground surface to the spacecraft, \( r \) is the change in range and \( u \) is the deformation; the sign convention used in this study is such that an upward movement will produce a positive value. The phase in the interferogram reaches a maximum of \(-0.42\pi\) (Fig. 7) that is equivalent to \(-2.5\) cm that we interpret as subsidence (Fig. 8c).

We subsampled the interferogram using the quadtree method (Jónsson et al., 2002) to reduce the number of data values to be used from \(-1.44 \times 10^6\) to 7735 samples. For the forward model of the coseismic elastic deformation we used the expressions for finite displacements given by Okada (1985). The inversion problem using finite displacements seeks to minimize the difference between the synthetic model of the three dimensional displacement field and the InSAR LOS observations. We know the fault plane as a result of the aftershock distribution. Thus, for a first approximation of the coseismic displacements, we created a large number of random displacement models, fixing the strike and dip of the fault by the aftershock relocations (strike 200° and dip 85°) and the rupture length and width at 10 km, using the relations of Wells and

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**Fig. 7.** a): Coseismic interferogram before unwrapping, in this convention LOS change far from satellite indicates subsidence b): Coseismic interferogram corrected by estimating the elevation-related delay of the interferogram. The red box indicates the region to unwrap the phase to avoid the gaps with no phase signal (correlation gaps) and the tropospheric and ionospheric perturbations. The yellow star indicates the relocation of the main event obtained in this study. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

**Fig. 8.** a: Synthetic model displacement field; the displacement vectors show the direction and amplitude of the ground displacements. b: LOS wrapped displacements model. c: LOS deformation model in cm, in this convention, displacements away from the satellite (blue) are represent subsidence, the maximum displacement is 2.44 cm. The satellite-target direction is shown by the black arrow, inc is the look angle for ALOS (34.3°) and ald is the azimuth look direction (ALD) perpendicular to satellite heading direction. The white rectangle indicates the position and size of the fault plane. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)
Coppersmith (1994) for an earthquake of ~6Mw. We varied the epicenter ±10 km (with respect to a preliminary epicenter at 73.795W, 4.397S), depth between 0 and 20 km, and slip (U1, U2) taking the depth as the minimum depth of the fault rectangle, and U1 and U2 as the strike- and dip-components of the slip in the Okada formulation (Feigl and Dupré, 1999); this plane was constrained to be below the surface (depth > 0) (e.g., Dawson et al., 2008; Segall, 2010). We also used the seismic moment Mo as a nonlinear constraint to avoid unrealistic fault area and depth estimates. We constrained the moment magnitude in the range (5.6 < Mw < 6.3), Mw = 2/3 log10(M0) – 10.7, where M0 = μAU, μ is the shear modulus of the faulted rock, A is the area of the fault, and U is the average displacement or slip on the fault (Fig. 8).

Our preferred model (Fig. 8a and Table 3) is the NNE-SSW strike-slip solution with a small normal dip-slip component, the rupture plane extends from a depth of 4.8 ± 0.3 km to 14.3 ± 0.4 km, and a width of 9.5 ± 0.4 km (Fig. 8) (quoted formal uncertainty is at the 1σ level). The model has a slip-to-length ratio of 2.8 × 10^{-5}, in agreement with empirical estimates for intraplate earthquakes which range from 2 × 10^{-5} to 1.0 × 10^{-4} (e.g., Scholz, 1990; Funning, 2005b). Observed and model LOS are shown in Fig. 9a and b, residuals are shown in Fig. 9c.

We used the elastic dislocation formulation of Okada (1985) to model the fault dislocation, assuming a homogeneous elastic half-space to relate surface displacement observations to slip on the fault at depth. The estimated LOS displacement model indicates that failure propagates northeastward. The transition between the extensional and contractional quadrants (Fig. 9a) is associated with a clear dextral motion. We conclude that the Quetame earthquake is a right-lateral strike-slip event, aligned with the east Andean frontal fault system.

In Fig. 9a the darkest blue band that corresponds to the maximum negative displacement seems to be contaminated by an artifact from incomplete removal of the large tropospheric and ionospheric perturbations seen in Fig. 7a. This can also be noticed by the abrupt change in RMS sign in the same region in Fig. 9c. The compressional SE lobe also seems to be partially contaminated by the same artifact because the synthetic model (Fig. 9b) has a weak signal in this area compared to the dark blue area in the observed LOS signal (Fig. 9a). The lack of independent InSAR data (only ascending data were available) limits measurements to one component of displacement in the LOS direction and hinders a good correction of atmospheric and/or ionospheric perturbations.

### Table 3

InSAR derived fault parameters for a uniform slip model; easting and northing are in km and represent the center point of the projection of the bottom of the fault rectangle at the surface. The uncertainties are 1σ (See Appendix B).

<table>
<thead>
<tr>
<th>Strike-slip solution with dip-slip component</th>
<th>Pure strike-slip solution</th>
</tr>
</thead>
<tbody>
<tr>
<td>Strike (°)</td>
<td>195 ± 2.5</td>
</tr>
<tr>
<td>Dip (°)</td>
<td>83 ± 3.5</td>
</tr>
<tr>
<td>Rake (°)</td>
<td>−171 ± 2</td>
</tr>
<tr>
<td>U1 (cm)</td>
<td>31 ± 3</td>
</tr>
<tr>
<td>U2 (cm)</td>
<td>−31 ± 5</td>
</tr>
<tr>
<td>Length (km)</td>
<td>10.7 ± 0.8</td>
</tr>
<tr>
<td>Min Depth (km)</td>
<td>4.7 ± 0.5</td>
</tr>
<tr>
<td>Max Depth (km)</td>
<td>14.6 ± 1</td>
</tr>
<tr>
<td>Easting (km)</td>
<td>−5.6 ± 0.5</td>
</tr>
<tr>
<td>Northing (km)</td>
<td>1.7 ± 0.5</td>
</tr>
<tr>
<td>Mw (10^17 N.m)</td>
<td>9.9 ± 1</td>
</tr>
<tr>
<td>rms displacement residual (mm)</td>
<td>6.0 ± 0.1</td>
</tr>
</tbody>
</table>

a) U is the average slip (Aki and Richards, 2002).
b) U1 is the left lateral transverse component of the strike-slip.
c) U2 is the thrusting dip-slip component.
d) The depth refers to the bottom of the fault rectangle according to the convention of Okada (1985).
e) Mw was computed using the relation Mw = μAU, where μ = 3 × 10^{11} is the shear modulus of the upper crust.

Fig. 9. a) Observed LOS b) Modeled LOS using the best fitting model (Table 3) c) residual LOS (observed minus model). The black box indicates the position of the fault plane rectangle and the yellow star indicates the centroid location of this rectangle for the mainshock given by the InSAR inversion. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)
5. Discussion and conclusions

This study integrates different sources of information (seismological and geodetic) at varied spatial scales (local, regional, teleseismic) to estimate the source parameters for the 2008 Quetame earthquake.

Despite the lack of independent InSAR data, we were able to use geodetic measurements (InSAR), to precisely determine the location, focal mechanism and coseismic surface displacements. We used seismological data (P-wave polarities and waveforms) to constrain the fault plane orientation and aftershock relocations to conclusively identify the orientation of the fault plane. Although the strike-slip mechanism was already known, we improved the hypocentral location, identified the fault plane, and estimated the rupture geometry. An interesting contribution from InSAR analysis was the estimation of the depth, rupture size and co-seismic displacement.

Our preferred model is the NNE-SSW right-lateral strike-slip solution (striking 195° and steeply dipping 82° to the NW) with a small normal component (Table 3). The rupture nucleated at a depth of ~10 km with an average slip of 0.3 m over a 9 km × 11 km fault area. The seismic moment and rupture area give an estimate of stress drop of 1 MPa, which is lower than the global average of 3 MPa, but within the observed range (Allmann and Shearer, 2009).

One of the most interesting aspects of the Quetame earthquake is the fact that it was a right-lateral strike-slip event, aligned with the east Andean frontal fault system that does not take advantage of any of the predicted low-angle reverse fault planes, and may even cut through the Servita thrust fault (Fig. 10). However, the 2008 events appear to be confined to the hanging wall of the deepest thrust fault, a blind ramp at 18 km depth with 9 km of displacement. The hanging-wall location of the shocks leaves open the possibility that some of the deeper margin-parallel “escape” of the North Andean block is transferred to the northwest down the basal ramp to a deeper strike-slip fault system in the lower crust and upper mantle. GPS measurements by the continuous Geored network show that margin-parallel “escape” (10 mm/yr) now exceeds margin-normal compression (3 e 4 mm/yr) in the Eastern Cordillera (Kellogg and Mora, 2016).

Margin-parallel strike-slip faulting is well documented to the southwest in Ecuador and to the northeast in Venezuela, but was obscured by over 100 km of Miocene-Pliocene margin-normal shortening in the Eastern Cordillera (Egbue and Kellogg, 2010). The 2008 Quetame earthquake is an “escape” earthquake, and it provides a unique example of how margin-parallel shear slip is partitioned. Understanding this seismic event is important to provide insights on the mechanisms of earthquakes in the Eastern Cordillera and for the assessment of seismic hazard in the area near Bogota, the densely populated (~8 million inhabitants) capital city of Colombia.
Acknowledgements

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Appendix A. Waveform inversion quality indicators

Fig. A1 shows the result for the test. The Condition Number (CN) is derived from the Green’s function matrix (G) which relates data (observed waveforms) and model parameter. This number is the ratio of the maximum and minimum singular values of G for the best fitting source position; a small value of CN indicates a well conditioned problem. The Variance Reduction (VR) represents fitness of the solution.

The horizontal bars indicate the frequency used during the waveform inversion. The beach balls are color-coded according to their P-wave first polarity fit (denoted as PF in the figure). For example, PF = 1 means that 100% of the polarities were satisfied. The polarities used are identical to the Fig. 5. Note that, although most of the mechanisms are color-coded as red (PF < 0.60), the solutions do not present drastic variation among them, i.e., the mechanism is stable. The low polarity fit is due to the presence of many polarities close to the nodal planes, therefore a small rotation of the solution causes the low PF. Same effect was showed by Dias et al. (2016) in their Fig. A1, panels C and D.

Fig. A2 illustrates the Kagan angle (K-angle) with respect to the reference solution from Global CMT (strike/dip/rake = 196°/82°/-179°). K-angle expresses the minimum rotation between two focal mechanisms (Kagan, 1991) and, according to Zahradník and Custódio (2012), similar focal mechanisms have K-angles less than 20°–30° and they are very dissimilar if Kagan angle is greater than 40°.

Fig. A1. Frequency range test for the Quetame event using velocity models derived from the surface-wave dispersion. The horizontal bars (with the double-couple beachballs in the middle) denote the frequency ranges used. The variance reduction (VR) of the seismograms is indicated in the vertical axis. Panels A and B show the frequency range test for the stations RREF and HEL used in the waveform inversion. Panel C shows the test using the two-station inversion. The beachballs are color-coded according to the fitted first motion polarities, e.g., PF = 0.90 means that 90% of the polarities were satisfied (see the colorbar denoted PF).
Appendix B. Optimization and insar error analysis

The nonlinear Monte Carlo inversion methods are based on generating a large number of models in a uniform random fashion between pairs of upper and lower bounds, which were chosen a priori (Sambridge, 2002). Each generated model is tested for its fit to the observed data and then accepted or rejected. This approach allows combining prior information about the parameter search space, like the direction of the fault plane inferred by relocation of aftershocks, with other observed information. The final set of accepted models was used for interpretation (Press, 1968, 1970). Using this method we define probability distributions for each parameter for the solutions that satisfy the nonlinear inversion process (Fig. B1 to B3). The model space, and consequently we tune the parameter set and refine the bound constraints, taking ±3σ of probability distributions as lower and upper bounds for next iterations, assuming N(0, σ²) distribution of the error.
Fig. B1. Monte Carlo diagram of the preferred solution (see Table 3), modeling Strike-Slip solution with dip-slip component. The histograms show the distributions of the solutions satisfying the nonlinear inversion process, the mean or expected value indicates which value collected as many valid solutions.

Fig. B2. a) Observed Interferogram b) Modeled interferogram with uniform slip on a single fault using the best fitting model (see Fig. B1) c) residual phase between observed minus model. The interferogram in the figure b shows a bigger lobe of displacement elongated NW–SW with a fraction of fringe trough LOS deformation of ~3 cm (0.25 fringes).
To assess the error caused by InSAR noise (spatial correlations of atmospheric and orbital errors) in the linear set of parameters, we used the method described in Wright et al. (2004a and 1999); Parsons et al. (2006); Dawson et al. (2008). We analyzed a region with no significant coseismic displacement in the western part of the interferogram (Fig. 7a). We used a simple exponential function that represents the 1D covariance function (Hanssen, 2001)

\[
C(r) = \sigma^2 e^{-r/l}
\]

\(\sigma^2\) is the variance (6.6 mm), \(r\) is the separation of the observations in km (0.1 km) and \(l\) corresponds to the e-folding correlation length scale, determined by the resample process of the interferogram, maximum quadtree block size of ~10 km. To simulate the correlated noise we build a variance-covariance-matrix (vcm) for the sample data, assuming isotropic noise throughout the interferogram subset. We calculated a spatial correlated noise vector \(y\), by

\[y = Lx\]

Where \(x\) is a Gaussian uncorrelated noise, \(L\) is the Cholesky decomposition matrix, often used to create correlations among random variables, \(\Sigma LL^T\) (Rubinstein, 1981). The vector correlated noise \(y\) is added to a synthetic model in order to get a new noise data set, this procedure is repeated 100 times and used in the inversion procedure without constrains to give a set of 100 model solutions of the parameters, the distribution of these model parameters is used to estimated the error of each parameter. We find a large trade-off between slip and minimum depth and slip and rake, whereas the seismic moment is well-constrained.

Fig. B3. Model parameter tradeoffs for Monte Carlo inversions of correlated noise, the parameter scatterplots show 100 best-fit solutions for one dataset perturbed with correlated noise, some trade-offs are evident as slip and MinDepth. The red line indicates the mean value of the parameter.
References


