Decadal volcanic deformation in the Central Andes Volcanic Zone revealed by InSAR time series

S. T. Henderson and M. E. Pritchard
Department of Earth and Atmospheric Sciences, Cornell University 3162 Snee Hall, Ithaca, New York 14850, USA
(stb54@cornell.edu)

Decadal trends of volcanic deformation in the Central Andes Volcanic Zone (CVZ) are identified with Interferometric Synthetic Aperture Radar (InSAR) stacks and time series velocity maps covering an area 19°S–27°S and 66°W–69°W. We combine over 750 ERS and Envisat interferograms from two descending and three ascending tracks. These tracks cover 100,000 km² and span 1992–2011. Our analysis extends observations at Cerro Blanco, Uturuncu, and Lazufre volcanic centers and uncovers two previously undocumented deformation centers: Cerro Overo in Northern Chile and Putana Volcano in Southwest Bolivia. Cerro Overo exhibits a transition from steady ~0.4 cm/yr deflation to 0.5 cm/yr inflation over several years. Putana Volcano underwent a short-lived episode of uplift between 13 September 2009 and 31 January 2010, with a maximum uplift of 4.0 cm. Cerro Blanco continues ~1.0 cm/yr deflation since 1995. Uplift at Lazufre began between 1997 and 2000 and has gradually accelerated to 3.5 cm/yr since 2005. Uturuncu volcano continues 1.0 cm/yr monotonic uplift since 1992 and shows evidence for a broad moat of subsidence surrounding the uplifting region. Four of the nine deformation events in the CVZ are not obviously associated with a particular volcanic edifice. Furthermore, there is significant spatial and temporal variability of these deformation events within a small geographic area.

Components: 8,580 words, 14 figures, 3 tables.

Keywords: InSAR; volcano; Andes; geodesy.

Index Terms: 1243 Space geodetic surveys; 8485 Remote sensing of volcanoes; 8419 Volcano monitoring.

Received 22 October 2012; Revised 17 January 2013; Accepted 22 January 2013; Published 6 May 2013.


1. Introduction

The Central Andes Volcanic Zone (CVZ, 14°S–28°S) is one of four distinct segments of the volcanic arc along the western margin of South America. Subduction of the Nazca Plate beneath the South America Plate drives CVZ volcanism, which terminates to the north and south at the Peruvian and Chilean-Pampean flat slab segments [e.g., Barazangi and Isacks, 1976; Cahill and Isacks, 1992; Ramos et al., 2002]. High elevation, thick crust, and a high concentration of silicic calderas are unique characteristics of this volcanic region [e.g., Isacks, 1988; de Silva, 1989; Beck and Zandt, 2002; Allmendinger et al., 1997], which contains 69 Holocene edifices, at least 13 of which have erupted in the last century [Siebert et al., 2011].

Here we present two decades of geodetic observations of volcanic deformation within the CVZ using Interferometric Synthetic Aperture Radar (InSAR). This survey has two principal goals: first,
Figure 1. Overview of InSAR data with cm/yr volcanic deformation emphasized. Descending tracks 282 (west) and 10 (east) cover the five major deformation centers in this study: Uturuncu, Lazufre, Cerro Blanco, Putana, and Cerro Overo. Red triangles are volcanos with Holocene activity from the Smithsonian Global Volcanism Database [Siebert et al., 2011]. The red outline shows the maximal extent to the Altiplano-Puna Magma Body (APMB) [Zandt et al., 2003]. The green outline approximates the extent of deposits related to the Altiplano-Puna Volcanic Complex (APVC) [de Silva, 1989].

...to better understand the distribution, frequency, and geometry of deformation throughout the arc; and second, to improve individual volcano hazard assessments by determining background deformation rates. Deformation patterns have been shown to sometimes precede volcanism [e.g., Voight et al., 1998], and in other cases, deformation ceases without eruption [e.g., Poland, 2010; Moran et al., 2011]. Therefore, in order to evaluate the implications of deformation at a particular volcano, it is important to rigorously sample surface deformation in both space and time [e.g., Dzurisin, 2003]. Furthermore, whether or not future eruptions occur, these InSAR measurements place important constraints on the volume of intruding material [e.g., Delaney and Metcigue, 1994; Johnson et al., 2000], which are of interest to the debate surrounding the emplacement and eventual fate of plutons [e.g., Glazner et al., 2004; Bachmann et al., 2007; Lipman, 2007; Menand et al., 2011].

[5] Previous petrological studies indicate multiple viable levels of magma storage within the thick crust of the CVZ. Three principal magma storage depths have been deduced by petrologic studies of a concentration of Tertiary ignimbrites in the CVZ known as the Altiplano-Puna Volcanic Complex (APVC, 21°S–24°S) [de Silva, 1989]: Mantle melts initially pond and mix at 60–70 km, evolve at 20–25 km depth, and evolve further at 4–8 km prior to eruption [e.g., de Silva and Gosnold, 2007; Kay et al., 2010]. Interestingly, much of the areal extent of the APVC is underlain by a geophysically imaged layer of partial melt at 19 km below topography, known as the Altiplano-Puna Magma Body (APMB; Figure 1) [Chmielowski et al., 1999; Zandt et al., 2003; Schilling et al., 2006].

[5] It is unclear whether the APMB is an ephemeral feature or if it is a long-lived source of the APVC eruptions. Nevertheless, previous geodetic inverse modeling places the Uturuncu deformation source between 12 and 25 km [Pritchard and Simons, 2004]. More recent numerical forward modeling of geodetic data may indicate that the interconnected partial melts of the APMB are currently feeding a large diapir rising beneath Uturuncu volcano [Fialko and Pearse,
Table 1. InSAR Data Summary

<table>
<thead>
<tr>
<th>Track</th>
<th>Beam</th>
<th># IFGs</th>
<th># Dates</th>
<th>Dates</th>
<th>Orbit</th>
<th>Platform(s)</th>
<th>Volcanos</th>
</tr>
</thead>
<tbody>
<tr>
<td>89</td>
<td>6</td>
<td>73</td>
<td>20</td>
<td>2004–2010</td>
<td>asc</td>
<td>Envisat</td>
<td>U</td>
</tr>
<tr>
<td>3</td>
<td>2</td>
<td>15</td>
<td>22</td>
<td>1997–2010</td>
<td>asc</td>
<td>ERS, Envisat</td>
<td>U</td>
</tr>
<tr>
<td>318</td>
<td>2</td>
<td>133</td>
<td>19</td>
<td>2005–2010</td>
<td>asc</td>
<td>Envisat</td>
<td>L, B</td>
</tr>
</tbody>
</table>

*Features of the five tracks of InSAR data used in this report. For spatial extents refer to Figure 3. Volcano abbreviations are the following: U = Uturuncu, LC = Laguna Colorado, P = Putana, B = Cerro Blanco, O = Cerro Overo, and L = Lazufre. Deformation at Cerro Blanco is also seen in descending track 239, Laguna Colorado in track 53, and Lazufre in track 404.

2012]. Measurement of active diapiric ascent is unprecedented and raises questions about the pervasiveness of diapir-related surface deformation [Brooks, 2012].

[6] In this paper, we present evidence that the deformation signal at Uturuncu is unique among the other deforming volcanos in the CVZ. We first describe our method for surveying large regions of a volcanic arc with InSAR, and we report on the spatial and temporal variation in surface deformation from 1992 to 2011 for a large subregion of the CVZ (19°S–27°S). We also provide order of magnitude estimates of source parameters based on standard inverse models. The five major deformation centers discussed in this paper are shown in Figure 1.

2. Data

[7] We generated 757 interferograms from five different tracks, spanning 1992 through 2011 and covering over 800 km along the volcanic arc (Table 1). All radar scenes are acquired from C-band ERS-1, ERS-2, and Envisat satellites. Full details of our processing steps are covered in the auxiliary material, and in the following sections, we briefly outline the essential methodology.

[8] We use ROI PAC software [Rosen et al., 2004] to process interferograms, pairing individual radar scenes with maximum baselines between 200 and 500 m (depending on the satellite track) to minimize spatial decorrelation. We use precise orbital information when available from the Delft Institute for Earth-Oriented Space Research [Scharroo and Visser, 1998; Doornbos and Scharroo, 2005], otherwise we use precise orbits from the German Space Agency. Many interferograms contain residual ramps resulting from the incomplete removal of orbital phase signals, which we remove by fitting a linear or quadratic surface. Interferograms are unwrapped with the Statistical-cost Network-flow Algorithm for Phase Un-wrapping (SNAPHU) [Chen and Zebker, 2001] and downsampled to 720 m per pixel [Rosen et al., 2000]. Downsampling filters short wavelength signals and decreases computational requirements, which are desirable for surveying large areas for volcanic deformation signals with surface footprints greater than several kilometers. Finally, interferograms are co-registered onto a common grid in radar coordinates.

2.1. Sources of Noise

[9] The dominant source of noise in single interferograms is attributed to phase delays as the radar signal propagates through the atmosphere [e.g., Goldstein, 1995; Zebker et al., 1997]. Owing to the large relief on volcanic edifices, atmospheric phase delays can sometimes be misconstrued as uplift or subsidence signals, and therefore care must be taken to separate these signals [Delacourt et al., 1998].

[10] In the CVZ, both descending and ascending interferograms regularly show correlated noise that varies with atmospheric water vapor concentrations [Fournier et al., 2011]. Ascending tracks, which are acquired at approximately 10 P.M. local time in the CVZ, can have additional large ionospheric signals that are related to variability in electron density [e.g., Gray et al., 2000; Xu et al., 2004]. Ionospheric signals are easily identified as SW-NE trending, high amplitude streaks that are roughly parallel to the Earth’s magnetic field at this latitude [Loveless et al., 2010; Fournier et al., 2010]. We note that for ascending track 318, four dates between March 2008 and March 2010 have strong ionospheric signals, which represents 22% of all acquisition dates for the track (Table 1). In track 3, four dates between April 2002 and November 2009 have strong signals, representing 18% of acquisition dates (Table 1). We choose to omit these dates from our analysis since we have sufficient scenes without ionospheric effects.
3. Interferogram Stacks

[11] By combining many interferograms spanning different dates, temporally variable noise is partially canceled out, improving the accuracy of deformation rate measurements [e.g., Wright et al., 2004; Fialko, 2006; Biggs et al., 2007]. Stacking of rectified interferograms is a standard way to improve signal-to-noise by a factor of \( \sqrt{N} \) where \( N \) is the number of independent interferograms being combined. However, the theoretical gain in signal-to-noise ratio assumes constant velocity deformation over the time span of \( N \) observations and Gaussian noise with a mean of zero. Both of these assumptions are likely false for InSAR data. Nevertheless, we find that this simple approach to stacking does produce smoothly varying spatial deformation patterns that are not correlated with topography (Figure 1).

4. Time Series Inversion

[12] The InSAR time series inversion is another post-processing technique that can enhance the precision of surface deformation measurements to sub-centimeter levels [Casu et al., 2006; Ferretti et al., 2007; Hooper et al., 2011]. In addition, time series analysis extracts a cumulative deformation history for each pixel in a set of interferograms. There are many versions of time series algorithms in use [e.g., Lauknes et al., 2011; Hetland et al., 2011], but most are based on persistent scatter (PS) [Ferretti et al., 2001] or small-baseline subset (SBAS) [Berardino et al., 2002] algorithms. PS techniques are generally favorable in areas of low radar signal coherence, but given the high coherence in the CVZ, we employ an SBAS-based algorithm in our inversion.

[13] All SBAS inversion schemes assume only that velocity is constant between consecutive dates, instead of over the total elapsed time of observations. Therefore, any function may be applied as a best fit regression to the time series output of cumulative deformation versus time, potentially improving the accuracy of ground velocity measurements. Our time series inversion algorithm closely follows procedures outlined in Lundgren et al. [2001] and Berardino et al. [2002], which have been applied successfully in several extended-area surveys [Tizzani et al., 2007; Casu et al., 2008].

[14] For a combined set of unwrapped ERS and Envisat Interferograms, we solve the following linear inverse problem pixel-by-pixel using singular value decomposition (SVD):

\[
Bv = \delta \phi
\]  

[15] where \( B \) is the \( m \times n \) design matrix for a set of \( m \) interferograms containing \( n + 1 \) dates, and the nonzero entries of \( B \) are the time spans between consecutive dates; \( v \) is a column vector of pixel velocities for each date interval; and \( \delta \phi \) is the data vector containing unweighted deformation values extracted from individual interferograms. For most pixels, the system of equations is overdetermined and \( v \) contains the velocities that minimize the \( L_2 \) residual.

[16] We integrate output velocities to produce plots of cumulative deformation versus time. The deformation history can then be fit with another linear regression to produce ground velocity maps over the analyzed time period. We assume a functional form \( v = mx + b \) in our regression, and consequently nonlinear rates will be obscured on this velocity map product. However, we note that maps of the \( L_2 \) residual are useful for identifying regions that deviate from linear deformation.

5. Validation

[17] In order to quantify uncertainties in measured surface deformation rates, previous InSAR time series studies compare results to independent data sets such as leveling data [Pepe et al., 2005; Casu et al., 2006], or GPS measurements [Finnegan et al., 2008]. Unfortunately, there are no GPS measurements in our study region that overlap with InSAR observations. One approach to validate time series results in the absence of ground-based measurements is to compare velocities obtained via an independent processing chain [Casu et al., 2008]. Since the major deformation centers in the CVZ (Uturuncu, Lazufre, and Cerro Blanco) have been observed to be monotonically inflating or deflating over the past two decades, we expect close agreement between time series best fit velocities and average velocities derived from simple stacking. In Figure 2, we compare velocity maps based on these two processing techniques to demonstrate the agreement in amplitude and spatial pattern of deformation signals.

[18] We also follow the procedure outlined by Finnegan et al. [2008] to compare velocities derived for the same ground pixels in overlapping descending tracks 10 and 282 covering the same time period (1995–2010). Direct comparison of pixel velocities is complicated by the fact that
Figure 2. Correlation plots show agreement of pixel velocities in overlapping region of descending data: (a) best fitting Mogi source synthetic, (b) –1 to 1 mm/yr uniform distribution of velocities and Gaussian noise with standard deviation of 4 mm/yr imposed on synthetic Mogi source signal, (c) comparison of velocities obtained by stacking, and (d) comparison velocities derived from time series analysis.
actual displacement values are projected into different line-of-sight (LOS) vectors, whose incidence angles vary both between tracks and up to 9° across a single track. These discrepancies lead to a subtle skewing of values about the 1:1 line, but the general agreement in velocities is clear. The spread of noise can be characterized following Finnegan et al. [2008] as a uniform distribution of velocities between –1 and 1 mm/yr and Gaussian white noise with a standard deviation of 4 mm/yr (Figure 2). While the dominant source of noise in interferograms is not white, this simplifying assumption gives us an approximate bound on the uncertainty in velocities derived through stacking and time series analysis.

Finnegan et al. [2008] make the assumption that InSAR measurements are mostly vertical, but in the case of volcanic deformation signals, radial displacements can be significant. In order to isolate the contribution of geometric distortion in our validation procedure, we took forward models of uplift at Uturuncu from Pritchard and Simons [2004] and projected the noiseless synthetic displacements into LOS for tracks 282 and t10. Plotting the agreement of the projected synthetic data results in a unique distribution about the 1:1 line. We note that for an expansion point source [Mogi, 1958], track 282 shows systematically higher velocities compared to track 10, whereas the stack and best fit velocities derived from actual data show slightly higher velocities in track 10. This effect is more pronounced for the high-amplitude signal of Uturuncu and could be due to different acquisition dates in tracks 282 and track 10. Another possible explanation is that the radial and vertical deformation vectors predicted by elastic half-space models of a spherical point source are not representative of the actual deformation source underlying Uturuncu. Consequently, more complicated source shapes or heterogeneous crustal material should be explored in future modeling studies.

6. Results and Discussion

The Central Andes Volcanic Zone was previously known to contain seven volcanos that have exhibited surface deformation since 1992: Ticsani, Hualca Hualca, Uturuncu, Lazufre, Cerro Blanco, Lastarria, and Lascar (Figure 3 and Table 2). Our analysis adds two volcanos to the list—Putana and Cerro Overo—and we extend previous time series analyses to bring observations through 2011. This extension in observational period illuminates the variety of magnitudes and variability in deformation rates for volcanos in the CVZ. Ticsani and Hualca Hualca volcanos are situated outside the extent of InSAR data used in this study and therefore will not be discussed here.

The diameter of volcanic deformation signals in the CVZ ranges from 5 to 150 km, which implies a large range in source depths based on the classic model of a pressure source embedded in a homogeneous elastic crust [Mogi, 1958]. While the assumptions of this simplified model likely violate the subsurface conditions in the hot, thick crust of the CVZ, it serves as an order of magnitude estimate of source depth and volume. Therefore, in order to provide a direct comparison of source parameters for the centers discussed in this paper and to compare values with previously published studies, we performed a Levenberg-Marquardt inversion for point expansion source parameters with our data set. The results are summarized in Table 3 and show depths ranging from 1 to 24 km below local topography and changes in volume on the order of 0.1–29 m³/yr.
Table 2. Volcanic Deformation in the CVZ\textsuperscript{a}

<table>
<thead>
<tr>
<th>Volcano</th>
<th>$V_{los}$ (cm/duration)</th>
<th>Duration</th>
<th>$dV/dt$</th>
<th>SA (km\textsuperscript{2})</th>
<th>Center (°S, °W)</th>
<th>Track</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>This Study</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Uturuncu</td>
<td>1.1</td>
<td>1992–2011</td>
<td>constant</td>
<td>15,000</td>
<td>(22.27, 67.22)</td>
<td>282</td>
</tr>
<tr>
<td>Lazufre\textsubscript{1}</td>
<td>2.0</td>
<td>1998–2005</td>
<td>acceleration</td>
<td>10,000</td>
<td>(25.25, 68.49)</td>
<td>282</td>
</tr>
<tr>
<td>Lazufre\textsubscript{2}</td>
<td>3.5</td>
<td>2005–2011</td>
<td>constant</td>
<td>13,000</td>
<td>(25.25, 68.49)</td>
<td>282</td>
</tr>
<tr>
<td>Cerro Blanco</td>
<td>–1.0</td>
<td>1992–2010</td>
<td>constant</td>
<td>1200</td>
<td>(26.77, 67.72)</td>
<td>10</td>
</tr>
<tr>
<td>Putana</td>
<td>4.0</td>
<td>09/2009–01/2010</td>
<td>impulse</td>
<td>120</td>
<td>(22.57, 67.87)</td>
<td>282</td>
</tr>
<tr>
<td>Cerro Overo\textsubscript{1}</td>
<td>–0.4</td>
<td>1992–2003</td>
<td>constant</td>
<td>1000</td>
<td>(23.76, 67.41)</td>
<td>10</td>
</tr>
<tr>
<td>Cerro Overo\textsubscript{2}</td>
<td>0.5</td>
<td>2003–2010</td>
<td>constant</td>
<td>1000</td>
<td>(23.76, 67.41)</td>
<td>10</td>
</tr>
<tr>
<td><strong>Previous Studies</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lastarria\textsubscript{b}</td>
<td>1.0</td>
<td>2003–2008</td>
<td>constant</td>
<td>100</td>
<td>(25.17, 68.50)</td>
<td>282</td>
</tr>
<tr>
<td>Lascar\textsubscript{c}</td>
<td>–2.0</td>
<td>07/1995–10/1995</td>
<td>impulse</td>
<td>1</td>
<td>(23.37, 67.73)</td>
<td>282</td>
</tr>
<tr>
<td>Lascar\textsuperscript{d}</td>
<td>–1.0</td>
<td>1994–2009</td>
<td>deceleration</td>
<td>10</td>
<td>(23.37, 67.73)</td>
<td>282</td>
</tr>
<tr>
<td>Hualca Hualca\textsubscript{e}</td>
<td>2.0</td>
<td>1992–1997</td>
<td>constant</td>
<td>11,000</td>
<td>(15.71, 71.89)</td>
<td>454</td>
</tr>
<tr>
<td>Tiscani\textsuperscript{f}</td>
<td>–23.0</td>
<td>06/2005–09/2005</td>
<td>impulse</td>
<td>600</td>
<td>(16.75, 70.60)</td>
<td>411</td>
</tr>
</tbody>
</table>

\textsuperscript{a}Volcanic deformation in the CVZ compiled from this dataset and previous studies. “LOS rate” is the maximum radar line-of-sight deformation rate over the time period listed under “Duration.” All rates are derived from descending interferograms (see “Track” column), which record approximately 90% of vertical displacements and 40% of horizontal displacements. In the case of nonlinear deformation, such as Cerro Overo where constant subsidence is followed by constant uplift, we divide the deformation episode into two table lines. Surface area is calculated based on the formula for an ellipse. Dates are formatted in month/year.

\textsuperscript{b}Froger et al. [2007].
\textsuperscript{c}Pavez et al. [2006].
\textsuperscript{d}Whelley et al. [2011].
\textsuperscript{e}Pritchard and Simons [2004].
\textsuperscript{f}Holtkamp et al. [2011].

In the following sections, we present the InSAR time series results for individual volcanic centers, and we discuss possible source models that explain the observed deformation signals.

6.1. Uturuncu Volcano (22.27°S, 67.22°W, 6008 m)

Uturuncu is a dacitic stratovolcano in southwest Bolivia that has likely not erupted for 270 ka [Sparks et al., 2008]. Pritchard and Simons [2002] identified Uturuncu as a 70 km diameter region uplifting at a maximum rate of 1–2 cm/yr. A follow-up study in 2004 analyzed 12 interferograms between May 1992 and December 2000 to confirm constant uplift rate at 1–2 cm/yr [Pritchard and Simons, 2004]. Sparks et al. [2008] confirmed the uplift continued in the same region through 2006. Recently, it has been suggested that there is evidence for a moat of subsidence surrounding uplift at Uturuncu resulting from lateral transport of melt to a rising diapir [Fialko and Pearse, 2012]. We confirm that uplift at Uturuncu has been relatively constant through 2011 and that the moat of subsidence is a robust signal.

Our time series extends coverage of Uturuncu through 2011 in four separate tracks (descending

Table 3. Mogi Model Inversion Source Parameters\textsuperscript{g}

<table>
<thead>
<tr>
<th>Volcano</th>
<th>Center (°S, °W)</th>
<th>$X$ (km)</th>
<th>$Y$ (km)</th>
<th>$Z$ (km)</th>
<th>$\Delta V$ ($10^6$ m$^3$/yr)</th>
<th>RMSE (cm/yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Uturuncu</td>
<td>(22.27, 67.22)</td>
<td>–2</td>
<td>0</td>
<td>24</td>
<td>29</td>
<td>0.04</td>
</tr>
<tr>
<td>Lazufre</td>
<td>(25.25, 68.49)</td>
<td>–1</td>
<td>–2</td>
<td>13</td>
<td>17</td>
<td>0.08</td>
</tr>
<tr>
<td>Cerro Blanco</td>
<td>(26.77, 67.72)</td>
<td>1</td>
<td>–1</td>
<td>10</td>
<td>–6</td>
<td>0.07</td>
</tr>
<tr>
<td>Laguna Colorado</td>
<td>(22.01, 67.70)</td>
<td>–2</td>
<td>0</td>
<td>15</td>
<td>–3</td>
<td>0.04</td>
</tr>
<tr>
<td>Cerro Overo (uplift)\textsuperscript{b}</td>
<td>(23.76, 67.30)</td>
<td>3</td>
<td>–1</td>
<td>13</td>
<td>20</td>
<td>0.33</td>
</tr>
<tr>
<td>Cerro Overo (sub)\textsuperscript{c}</td>
<td>(23.76, 67.30)</td>
<td>0</td>
<td>–1</td>
<td>9</td>
<td>–10</td>
<td>0.28</td>
</tr>
<tr>
<td>Putana\textsuperscript{d}</td>
<td>(22.25, 67.85)</td>
<td>0</td>
<td>–1</td>
<td>1</td>
<td>0.3</td>
<td>0.18</td>
</tr>
</tbody>
</table>

\textsuperscript{g}Model parameters are based on an implementation of the Matlab\textsuperscript{™} nlinfit routine that uses a Levenberg-Marquardt algorithm weighted by data covariance. Inversions based on descending track 282 stack unless otherwise noted. The inversion assumes Poisson’s ratio $\nu = 0.25$, and depths are with respect to average local topography ($\approx$4–5 km).

\textsuperscript{b}Cerro Overo uplift data from 04 April 2000 to 19 October 2010 (Envisat t10).
\textsuperscript{c}Cerro Overo subsidence data from 02 October 1995 to 21 December 1999 (Envisat t10).
\textsuperscript{d}Putana uplift data from full resolution interferogram 16 January 2011 to 24 August 2008 (Envisat t282).
tracks 282 and 10, and ascending tracks 89 and 3) with hundreds of new interferograms. The velocity maps for each of these tracks clearly show uplift of the same spatial extent (Figure 4), which appears to stay constant through 2011 at a maximum rate of up to 1.1 cm/yr in radar line-of-sight (Figure 5). There is a hint of deceleration in recent years; however, the limited number of interferograms that use recent dates means the apparent deceleration could be due to uncanceled noise.

Figure 4. Profiles through uplift at Uturuncu volcano based on descending stacks of (left) track 282 and (right) track 10. Uplift is axis-symmetric with a diameter of approximately 70 km.

Figure 5. Time series deformation history for pixels near the region of maximum uplift at Uturuncu from four different InSAR tracks. Descending tracks 2010 and 2282 show good agreement for constant 1 cm/yr uplift since 1995. The slightly lower uplift rate in track 3 can be attributed to fewer dates and therefore greater sensitivity to the outlier in 2002. Uplift rate in track 6089 is significantly lower due to the shallower incidence angle of beam 6 compared to beam 2.
Figure 6. The moat of subsidence around uplift at Uturuncu is a persistent feature in both ascending and descending stacks, suggesting it is in fact a real deformation feature. Subtle linear features are associated with the edges of large subsets of interferograms used in the stack. Uplift appears to be broader and of lower magnitude in track 6089 because of the shallow incidence angle of beam 6.

[25] Uturuncu Volcano is the only deforming volcano in the CVZ that appears to have a clear association with the Altiplano-Puna Magma Body. It is located well within the mapped areal extent of this seismically imaged low velocity zone (Figure 1), and geodetic model depths coincide closely with the midcrustal depth of the APMB [Pritchard and Simons, 2004]. An array of 15 seismometers deployed between April 2009 and April 2010 at Uturuncu has shown a high rate of volcanotectonic seismicity compared to other inflating volcanos without accompanying eruptions: This seismicity is concentrated at a depth of 5 km, exhibits frequent swarms, and shows evidence for dynamic triggering by 27 February 2010 Maule earthquake [Jay et al., 2012]. Given these observations, it is possible that the active inflation in the midcrust is causing small earthquakes by perturbing the stresses on overlying faults [e.g., Savage and Clark, 1982; Bonafede et al., 1986; Sparks et al., 2008].

6.1.1. Moat of Subsidence

[26] In order to constrain the full extent of 100 km+ deformation signals, it is essential to stitch multiple frames of InSAR data along a single track, or to use the Envisat SCANSAR wide-swath mode. While Fialko and Pearse [2012] include SCANSAR data over Uturuncu to extend east-west coverage, we use long tracks of InSAR data to better constrain the north-south dimensions of subsidence around Uturuncu.

[27] The moat of subsidence is not as symmetric as the uplift pattern, but this is potentially due to the lower amplitude of subsidence (up to –4 mm/yr), which is on the order of estimated uncertainties in individual pixel velocities. As a result, the continuity of the broad signal is prone to disruption by regions of correlated atmospheric noise. Nevertheless, multiple profiles taken through the center of uplift at Uturuncu show that the subsiding region has a diameter of approximately 150 km (Figure 6), and we have confirmed that measured subsidence is not correlated with topography. Several cartoon models of physical models that can fit these observations are given in Figure 7, and each of these models is discussed in Fialko and Pearse [2012].

[28] We note that stacked Mogi sources can reproduce both the uplift and peripheral subsidence seen in descending data if the inflation source is located at 25 km depth and the deflation source is located near the moho at 75 km depth (Table 3). However, the assumption of elasticity is questionable at such depths. Furthermore, we observe a significant offset in maximum LOS uplift of 6 ± 2 km in descending tracks 282 and 10 compared to ascending track 3 (auxiliary material). The peak deformation offset reflects the ratio of horizontal to surface. This ratio is a function of source geometry assuming expansion sources in an elastic or Maxwell viscoelastic half-space [Fialko and Pearse, 2012]. At Uturuncu, a separation of 6 km could be due
to prolate source geometry [Fialko and Pearse, 2012], crustal anisotropy [Rubin, 1992], or increasing shear modulus with depth [Fialko et al., 2001].

6.1.2. Laguna Colorado (22.01°S, 67.68°W, 5000 m)

A 20 km region centered on the Laguna Colorado ignimbrite appears as a heterogeneity within the moat of subsidence surrounding Uturuncu (Figures 1 and 2). The time series from track 282 shows coherent subsidence at up to −4 mm/yr across a 20 km diameter region. This area is not clearly associated with an active volcanic edifice; however, its perimeter appears associated with the Laguna Colorado deposits named in de Silva and Francis [1991]. This deposit was studied earlier by Baker and Francis [1978], and Salisbury et al. [2011] estimate the age at 1.98 Ma with a dense rock equivalent (DRE) volume of 60 km³. The location of enhanced subsidence within the larger moat of subsidence may indicate heterogeneous vertical or lateral motion of material towards the source of uplift.

6.2. Lazufre (25.25°S, 68.49°W, 4900 m)

Surface deformation at Lazufre was first described by Pritchard and Simons [2002] as a 70 km axisymmetric 2–2.5 cm/yr uplift between the Holocene volcanos Lastarria and Cordon del Azufre. Since then, studies have shown that the areal extent and uplift rate have been increasing in time, and likely perturbing the hydrothermal system underlying Lastarria Volcano [Froger et al., 2007; Ruch et al., 2008, 2009; Ruch and Walter, 2010]. While the broad uplift at Lazufre can be well fit by an inflating sill at 10 km depth, it is difficult to separate the contributions of lateral growth of the sill, increasing pressurization, or other processes which can explain the increasing areal extent of uplift [Anderssohn et al., 2009].

Whereas Ruch et al. [2009] performed separate analyses on ERS and Envisat data, we combine these data sets into a single inversion that extends coverage through 2011 (Figure 8). Surface deformation at Lazufre is comparable in spatial extent and uplift rate to Uturuncu, although it is important to note that the uplift is less symmetrical and profiles through the deformation do not show evidence for a moat of subsidence. Our analysis confirms no deformation between 1995 and 1997, acceleration to cm/yr rates between 1997 and 2005, followed by uplift after 2005 at a constant rate of 3.5 cm/yr (Figure 9). The exact onset time of uplift at Lazufre is difficult to pin down, given the absence of InSAR scenes between 26 October 1997 and 19 March 2000.
By examining the separation of uplift maxima in interferogram stacks from descending track 282 and ascending track 318, we find an offset of 4 ± 2 km (auxiliary material), which is within the maximum range of predicted separation for sills shallower than 12 km in elastic host rock [Fialko et al., 2001]. Both Uturuncu and Lazufre have been modeled as magmatic intrusions; however, additional geophysical data sets acquired in this region will help distinguish the subsurface conditions or temporal evolution of intrusions that lead to larger maximum uplift offsets and subsidence moats versus the deformation pattern observed at Lazufre.

6.3. Cerro Blanco (26.77°S, 67.72°W, 4400 m)

This caldera and rhyolitic dome complex in Northwest Argentina is called Cerro Blanco on some Argentinean maps, but referred to as Robledo in other studies [Siebert et al., 2011]. Pritchard and Simons [2004] reported a subsidence rate of −2.5 cm/yr between 1992 and 1997, and a lesser rate of −1.8 cm/yr from 1996 to 2000. Our time series results indicate that a linear subsidence rate of −1.0 cm/yr throughout the entire period of observation 1992–2011 fits the data well (Figure 10). However, a faster rate would indeed be calculated using only data between 1992 and 1997.
The deflation has a much smaller footprint (20 km diameter) compared to Uturuncu and Lazufre, and preliminary elastic modeling suggests a deflation source at 10 km depth. Subsidence at calderas could be due to several mechanisms including solidification of cooling magma, and removal of hydrothermal fluids followed by compaction [e.g., Newhall and Dzurisin, 1988]. Pritchard and Simons [2004] explored one-dimensional magma chamber cooling models and concluded that an additional source of subsidence must be present to explain the observed subsidence. One possible explanation is the lateral migration of hydrothermal fluids from the section of crust overlaying deeper crystallizing melt.

6.4. Cerro Overo (23.76°S, 67.41°W, 5365 m)

Our analysis shows a previously undocumented 20 km diameter deformation pattern near Cerro Overo on the border of Chile and Argentina that shows a clear transition from −4 mm/yr subsidence to 5 mm/yr uplift between 2003 and 2005 (Figure 11). The similarity in deformation magnitude and spatial extent before and after 2003 is suggestive of a reversible process at the same source depth occurring over a 20 year time scale.

[36] A map in Gonzalez-Ferran [1995] labels Cerro Overo near the SE terminus of the NW trending Cordon Puntas Negras, although no individual description of Cerro Overo is given. Cordon Puntas Negras is a 70 km long collection of lava flows, domes, maars, and andesitic cones erupted throughout the Holocene (Figure 12). Volcan Puntas Negras (−23.73°, −67.53°, 5852 m) lies about 10 km to the west of Cerro Overo and has produced many post-glacial flows. Ten kilometers to the southwest is the El Laco volcanic complex (−23.80°, −67.5°, 5482 m) known for its unique magnetite-bearing lavas [e.g., Gonzalez-Ferran, 1995]. Field evidence at El Laco points to metasomatic alteration of volcanic rocks by circulating hot brines in the shallow crust [Sillitoe and Burrows, 2002]. Thus, although there are no published reports of active fumarole fields or hot springs at Cerro Overo, it is possible that the subsidence and uplift are due to the draining and refilling of a distributed shallow crustal reservoir related to Cordon Puntas Negras.
Figure 12. Outline of deforming region near Cerro Overo shown with a Landsat Thematic Mapper Mosaic background. The deforming zone is clearly offset to the east of the volcanic chain. Black triangles are volcanic edifices from de Silva and Francis [1991], many of them associated with the Holocene Cordon Puntas Negras volcanic chain listed in Siebert et al. [2011]. The location of Cerro Overo is taken from Gonzalez-Ferran [1995, p. 185]. The white triangle marks the location of the El Laco lava flows discussed in the text.

Figure 13. Selection of interferograms showing uplift at Putana Volcano (located within the black circular marker). Black triangles are volcanic edifices from de Silva and Francis [1991], red triangles are Holocene volcanos from Siebert et al. [2011], and purple stars mark the location of geyser fields. The deformation is persistent on various dates and similar amplitude uplift signals are not apparent on neighboring edifices, suggesting a non-atmospheric origin. Based on this collection of interferograms, and local seismic data, we ascertain the deformation was short lived and occurred sometime between 13 September 2009 and 31 January 2010.
Figure 14. Outline of deforming region at Putana volcano shown with a Landsat Thematic Mapper Mosaic background. Red triangles are Holocene volcanoes from Siebert et al. [2011]. Geyser fields are labeled by purple stars. The 5 km diameter region encircled by the red dashed line uplifted up to 4 cm between 13 September 2009 and 31 January 2010.

[37] Preliminary elastic modeling places the sources of inflation and deflation at approximately 10 km depth (Table 3 and the auxiliary material). If the deformation source is due to fluid evacuation followed by re-entry, it is possible that an earthquake caused a change in stress state large enough to reverse flow. From our time series inversion, we know subsidence at Overo occurred through 2003, and sometime between 2003 and 2005 uplift initiated. A candidate earthquake is the Mw 7.8 Tarapaca earthquake of 13 June 2005. However, the epicenter for this earthquake is approximately 450 km from Cerro Overo. Therefore, it is possible the earthquake played a role in the transition from subsidence to uplift, but the large distance could require a dynamic triggering mechanism.

6.5. Putana (22.55°S, 67.85°W, 5890 m)

[38] Putana volcano is part of a 600 km² cone-and-flow complex with no historic eruptions but extensive fumarolic activity [e.g., de Silva and Francis, 1991]. Fumarole measurements taken in 2007 at Putana showed emission temperatures of 82–88°C, source temperatures of 500°C, and relatively high concentrations of magmatic gases [Tassi et al., 2011].

[39] The volcano exhibited a short-lived episode of uplift sometime between 13 September 2009 and 31 January 2010. We note that if volcanic deformation occurs over scales shorter than several months, it can easily be dismissed in a time series plot as an outlier related to atmospheric noise. Consequently, we did not detect deformation at Putana in our time series analysis, but rather from examining independent interferograms from track 282 (Figure 13).

[40] The circular deformation pattern at Putana has a maximum uplift of 4 cm centered on the volcanic edifice and a diameter of 5 km (Figure 14). The USGS PDE catalog shows no magnitude 4 or greater events within 500 km of Putana during the deformation time span, and therefore the uplift does not seem to be triggered or accompanied by large earthquake nearby. However, a reconnaissance vertical component 1–10 Hz seismometer located 4 km northwest of the summit, from August 2009 to February 2011, showed local swarm-like activity on 3 October, 12–13 October, and 9–31 December 2009 [Soler and Amigo, 2012]. We suspect that these swarms could be related to the observed deformation. Finally, we note that the Sol de Mañana geyser field, 20 km northeast of Putana, shows some evidence of broad uplift in several interferograms, although no persistent deformation is found via stacking or time series analysis (Figure 13).

[41] Due to the small spatial footprint of deformation at Putana, the source is likely shallow and therefore a simple Mogi source linear elastic model should provide a good first-order estimate of depth [e.g., Amelung et al., 2000]. We calculate a best-fitting volume addition of $0.3 \times 10^6$ m³ at 1 km depth to explain the deformation at Putana, although there is some hint that the deformation is slightly elongated to the East and West (auxiliary material).

6.6. Selected Non-Detection

[42] Due to the downsampling of our data set to 720 m pixels, previously observed deformation on the scale of 1–10 km² at Lastarria and Lascar was not resolved. Uplift at Lascar Volcano (25.17°S, 68.50°W, 5697 m) on the NW edge of Lazufre deformation was detected with high-resolution InSAR data [Froger et al., 2007]. The uplift had a rate of approximately 1 cm/yr
above background and was detected in Envisat data acquired between April 2003 and January 2008 [Anderssohn et al., 2009; Ruch et al., 2009]. Despite the frequent eruptive activity at Lascar Volcano (22.37°S, 67.73°W, 5592 m) [Matthews et al., 1997], there have been limited detections of ground deformation, most notably co-eruptive subsidence in the crater in 1995 [Pavez et al., 2006] and compaction of pyroclastic flow deposits (Table 2) [Whelley et al., 2011].

7. Conclusion

[43] We have presented results of a decadal (1992–2011) InSAR time series survey of the Central Andes Volcanic zone from a combination of ERS and Envisat interferograms. Such surveys of large volcanic regions are important for identifying trends in background deformation that can inform future hazard assessments. Our analysis has uncovered deformation at two volcanic areas not thought to be deforming (Putana and Cerro Overo) and has detected unique temporal trends at known deformation centers (Uturuncu, Lazufre, and Cerro Blanco).

[44] Interestingly, of the nine known volcanic deformation events in the CVZ, only one is associated with an eruption, and four are not clearly associated with any of the 69 active volcanos in the CVZ. Only the deforming area surrounding Uturuncu volcano shows a broad moat of subsidence, suggesting that if such a signal is characteristic of diapirc ascent, only under the Altiplano-Puna plateau are subsurface conditions currently amenable to observable active diapir formation. Overall, there is significant variation in deformation rates throughout the arc, implying a variety of detectable hydrothermal and magmatic transport mechanisms occur over 20 year time periods.

Acknowledgments

[45] We thank two anonymous reviewers for improving this manuscript. ERS and Envisat data were provided by the European Space Agency (ESA). This work was supported by NASA (National Aeronautics and Space Administration) grant NNX08AT02G issued through the Science Mission Directorate’s Earth Science Division and NSF (National Science Foundation) grant 0908281 which is part of the PLUTONS project (http://plutons.science.oregonstate.edu).

References


Barazangi, M., and B. Isacks (1976), Spatial distribution of earthquakes and subduction of the Nazca plate beneath South America, Geology, 4(11), 686–692.


Fialko, Y. (2006), Interseismic strain accumulation and the earthquake potential on the southern San Andreas fault system, Nature, 441(7096), 968–971.


Ruch, J., and T. R. Walter (2010), Relationship between the InSAR-measured uplift, the structural framework, and the present-day stress field at Lazufrue volcanic area, central Andes, Tectonophysics, 492(1–4), 133–140.
Sillitoe, R., and D. Burrows (2002), New field evidence bearing on the origin of the el laco magnetite deposit, Northern Chile, Econ. Geol., 97, 1101–1109.