Enhancement of long period components of recorded and synthetic ground motions using InSAR

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A B S T R A C T

Tall buildings and flexible structures require a better characterization of long period ground motion spectra than the one provided by current seismic building codes. Motivated by that, a methodology is proposed and tested to improve recorded and synthetic ground motions which are consistent with the observed co-seismic displacement field obtained from interferometric synthetic aperture radar (InSAR) analysis of image data for the Tocopilla 2007 earthquake (Mw = 7.7) in Northern Chile. A methodology is proposed to correct the observed motions such that, after double integration, they are coherent with the local value of the residual displacement. Synthetic records are generated by using a stochastic finite-fault model coupled with a long period pulse to capture the long period fling effect.

It is observed that the proposed co-seismic correction yields records with more accurate long-period spectral components as compared with regular correction schemes such as acausal filtering. These signals provide an estimate for the velocity and displacement spectra, which are essential for tall-building design. Furthermore, hints are provided as to the shape of long-period spectra for seismic zones prone to large co-seismic displacements such as the Nazca-South American zone.

1. Introduction

Design of flexible structures in seismic regions requires a more consistent characterization of low-frequency ground motion components which are, historically, not well represented by design spectra in building codes. Recorded acceleration data used to derive the response spectra have built-in low-frequency noise that is filtered during baseline correction and signal processing, which in turn annihilates also the true low-frequency contents of the record [1]. Thus, acceleration records usually integrate to near-zero residual ground displacement, or at best yield an inaccurate estimate of the co-seismic displacement, distorting the peak ground displacement (PGD) value and the displacement controlled region of the response spectra. This effect is greatly amplified in places where the crustal deformation may reach from a few centimeters to meters, the seismological near-field. Recent advances in building codes have included to some extent provisions which tend to enhance the long-period portion of the design spectrum by considering different effects. An example of this is the Eurocode 8 [2]. Nevertheless, methods are needed which improve our confidence in the records used to calibrate design spectra for the long period range.

Several methodologies have been proposed for acceleration record processing aiming to recover co-seismic displacement [3]. In a recent publication [4], strong-motion data were corrected by using measurements of continuous GPS stations (1–30 Hz sampling rate). These data showed the low-frequency behavior of the seismic motions and enabled a guided baseline correction of the available acceleration records. These studies show that fling-pulse type recorded motions can be attributed to co-seismic displacement reached gradually from the onset of strong motion. Therefore, knowledge of the static co-seismic displacement field is a tool to improve consistency of ground motion records and spectra at low-frequencies.

A promising alternative to determine the co-seismic displacement field is the use of interferometric SAR synthetic aperture radar (InSAR) [5]. In this technique the complex phase of two interfered satellite radar images of the region affected by an earthquake is the basis to derive the change in surface geometry during the earthquake in the satellite line of sight [6–8] (LOS). The displacements inferred by the procedure have different uses, such as inverse determination of a finite-fault earthquake slip distribution [9–12], the identification of local site effects and activation of secondary faults, and morphology studies (inflation and deflation) at regional level [13–15], to name a few. Moreover, the wide
coverage of satellite images and the sub-centimeter accuracy [16] of the differential InSAR technique (D-InSAR) make it a powerful tool for global analysis of seismic regions.

Additionally, knowing the earthquake geometry, slip distribution, and rupture propagation at a fault the current state-of-the-art strong-motion seismology provides different methods for the generation of consistent synthetic ground motions. For instance, the ‘stochastic method’ described elsewhere [17], considers the earthquake-source Fourier spectrum of small to intermediate size earthquakes described by a $\alpha^2$-shaped spectrum which can be used to assemble a larger fault. The complexities observed in strong motion records, in the 0.5–100 Hz frequency range, motivate the use of a stochastic framework to model these records since deterministic models fail to reproduce the features observed. The complexities, which are seemingly random phase and amplitude values, proceed from the anelastic behavior of the crust together with frequency dependent effects, geometric spreading, and refraction and reflection of seismic waves in the heterogeneous media. The stochastic framework is capable of modeling the random nature of the process while keeping its physical basis [18–25]. Although the method seems to work reasonably well for frequencies above 1 Hz, the low-frequency components are better represented by deterministic models based on elastic wave propagation theory. Thus, broadband approaches tend to use a hybrid stochastic-deterministic approach (e.g. [26–28]). A more simplistic approach consists in using a deterministic wave [29] calibrated and incorporated into the model to account for effects below 1 Hz which proves to be a low computational-cost alternative to the hybrid method. This last approach will be tested herein with the Tocopilla 2007 ($M_w=7.7$) event, which was covered by SAR imaging and by more than 11 broadband seismic stations.

In summary, the purpose of this work is to show how InSAR can be used to improve recorded and synthetic records based on the idea of meeting the observed co-seismic displacement field at the recording station. The proposed methodology is illustrated in three steps. First, to determine the interferograms and co-seismic displacement fields for the 2007 Tocopilla (Chile) earthquake. That information has a value per se, which goes beyond the other objectives of this work due to the multiple uses that such interferograms could be put to. Second, to identify finite-fault models corresponding to these events, by using slip inversions from InSAR measurements, which will yield a model of the 3D co-seismic displacement field. And third, use this field to enhance the quality of long-period spectral predictions of recorded ground motions and also any synthetic ground motions generated for arbitrary locations.

A flowchart summarizing the methodology used in this research is presented in Fig. 1. Step 1 involves InSAR image analysis from raw satellite data to determine co-seismic displacements. Step 2 is the inversion of the co-seismic displacement data to provide a plausible fault geometry and slip distribution. And Step 3 leads to the correction of existing strong ground motion records and synthesis of new ground motion-data consistent with the identified surface displacements.

2. **InSAR analysis**

Synthetic Aperture Radar (SAR) processing yields an image of a portion or ‘swath’ of Earth’s surface called a Single Look Complex image (SLC image), which are obtained by an imaging radar mounted on a satellite or other moving platforms. SAR enabled satellites, which are the most common source of SAR data for seismic applications, usually orbit Earth at altitudes of around 800 km and produce images approximately 100 km wide and several hundred kilometers long along the satellite track. An SLC image consists of an array of complex-valued pixels which represent a specific point in the target scene. Mathematically the image is represented as \( I(\phi, \lambda) = \sigma(\phi, \lambda) \exp[j\phi(\phi, \lambda)] \) where the pair \((\phi, \lambda)\) is geographical coordinates, longitude and latitude respectively. The magnitude \(\sigma(\phi, \lambda)\) is called the terrain reflectivity, which is just a measure of how visible the pixel is to the radar, and the phase \(\phi(\phi, \lambda)\) is related to the average time taken for the signal to travel from the radar to the objects within the pixel and back to the radar. This time delay can be related to the average distance of all objects within the pixel. Pixel sizes range from a few meters to tens of meters depending on various factors. Due to the periodicity of harmonic functions, the measured phase \(\phi\) can only be recovered in the interval \((0, 2\pi)\), which is called phase wrapping. For a given pixel, the phase is proportional to the number of wavelengths that fit in twice the distance traveled from the antenna to the target, i.e.

\[
\phi(\phi, \lambda) = 2\pi R_{ave}(\phi, \lambda) / \lambda \mod 2\pi
\]

(1)

where \(R_{ave}\) is the average range of all objects within a pixel. As shown, objects separated in range by one wavelength \(\lambda\) will present the same phase value. Since the wavelength \(\lambda\) is small (~5 cm) compared to the pixel dimensions and sizes of objects inside, several cycles of the \((0, 2\pi)\) interval occur for the phase within any given pixel. This phase wrapping within the pixel results in a very random-looking phase image which resembles static on a television set.
A thorough presentation of SAR processing goes beyond the scope of this article. Interested readers are referred to the excellent literature on the topic (e.g., [30,31]).

It is now possible to introduce InSAR analysis using these basic SAR concepts. The complete sequence of steps for InSAR and D-InSAR analysis used in this study is described in Fig. 3, please refer to this figure for the rest of this section. As with SAR, literature in the topic is vast (e.g., [32,5,33]). Consider two Single Look Complex (SLC) images $I_1$ and $I_2$ of a certain geographic location on Earth obtained at different times to allow for surface deformation to occur. The images are rectangular grids with a complex scalar value defined per pixel, i.e.

$$I_1(\Phi, A) = \sigma_1(\Phi, A) \cdot \exp(j \cdot \phi_1(\Phi, A))$$

$$I_2(\Phi, A) = \sigma_2(\Phi, A) \cdot \exp(j \cdot \phi_2(\Phi, A))$$

where $\sigma_i$ and $\phi_i$ are the terrain reflectivity and complex phase of the $i$-th image. An interferogram is formed by complex multiplication of the first image (reference or master image) with the conjugate of the second (slave image):

$$I_{12}(\Phi, A) = I_1(\Phi, A) \cdot I_2(\Phi, A) = \sigma_1 \sigma_2 \exp(j \cdot (\phi_1 - \phi_2))$$

$$= \sigma_1 \sigma_2 \exp(j \cdot (\Delta \phi))$$

where $\Delta \phi$ is called the interferometric phase. With reference to Fig. 2, if the radar sends pulses with wavelength $\lambda$, i.e. deformation to occur, the images are rectangular grids with a distance traveled by the pulse, $2D$, twice the distance traveled by the pulse, $2D$. The term $\phi_1 - \phi_2$ is called the 'flat phase unwrapping term 2$\pi n$. As indicated before, phase is wrapped in the interval $[0,2\pi]$ during SAR and InSAR processing. The process of recovering this term is called 'phase unwrapping' and is an area of active research in InSAR.

**SAR interferometry cannot be done on any two SLCs as there are technical limitations to when InSAR possible. Limitations on InSAR come from the value of the perpendicular baseline, $B_p$, which should be as small as possible for deformation mapping. Typical values of the perpendicular baseline range from a few meters in specialized missions to several hundreds of meters in regular missions. Depending on the system characteristics there are theoretical limits for $B_p$. Another limitation is the so-called temporal baseline, which is the time interval between satellite passes. The repeat orbit period of most SAR satellites is 28–35 days, but can be as low as 11 days as is the case of the German TerraSAR-X satellite. As a rule of thumb, the greater the time-lapse between passes the greater the de-correlation of the SAR images, which implies more noise and makes interpretation of the sources of deformation more difficult. There are other more subtle limitations which require more insight into SAR technology, yet the ones presented here are the most relevant.**

### 3. Earthquake interferograms

By using the procedure outlined in Fig. 3, LC images for the Tocopilla earthquake (2007, $M_w=7.7$) were produced and interfered. Table 1 shows the macroseismic data associated with the main event and some of the following aftershocks. This event was chosen because of two reasons. First, it is a medium sized event for the local seismic setting, which translates in a reduced co-seismic displacement field, requiring fewer interferograms to cover the area. And second, because its location in Northern Chile's desertic landscape makes it an ideal earthquake to study using InSAR, as it is very unlikely that vegetation or human activity will obscure the deformation signal. Other candidate earthquakes for this study were Pisco (2007), Antofagasta (1995), and Maule (2010) earthquakes. These were explored but were discarded because of lack of adequate acceleration data, low-quality interferometric results, and extension of the displacement field, respectively. Interferograms were also generated for the Pisco and Antofagasta earthquake and can be found elsewhere [34]. The interferograms used herein have also been processed and explored in other recent studies [46,47].

The raw radar scenes were obtained from the European Space Agency (ESA) and satellites ERS-1/2 and ENVISAT. ALOS satellite data was obtained from the Japanese Space Agency (JAXA) through Alaska Satellite Facility (ASF) which is the data node for the Americas. The scenes and data used in this study are shown in Table 2. The resulting unwrapped interferograms for the different paths are shown in Fig. 4.

Interferograms show fringe patterns associated with the LOS displacement component using colors which cycle according to a scale of 5 cm indicating relative displacement between equal fringe colors. Positive values indicate movement towards the sensor which, in general for this zone, are some combination of uplift and westward motion (see Fig. 5 for a general sense of the involved motions). Fringes represent iso-LOS-displacement contours, which can be used to find relative LOS-displacements.
between any points within the map. For instance, Path 096 that 
neat the coast the crust moved towards the sensor during 
the event while 368 confirms this and also shows a zone in 
the central valley that presented movement away from the 
radar. Northward there is a zone of little deformation that can 
be used as a reference point for measuring displacements (zero 

displacements). By counting fringes from this point toward the 
central valley, it is possible to identify almost four fringes, i.e., 
-nearly 20 cm of negative LOS displacement. Instead, by moving 
toward the coast the fringe count is nearly 6, i.e., 30 cm of relative 
LOS displacement towards the sensor. The different images 
presented in this figure, (a), (b) and (c) correspond to the same 
displacement field, but seen from a different LOS corresponding 
to the satellite pass. The different LOS vectors explain why the 
fringe patterns look different for the different interferograms 
and the same earthquake event. All this will become apparent 
as the complete theoretical co-seismic displacement fields are 
determined.

4. Inversion for fault slip using InSAR

The problem of predicting the co-seismic displacement field 
occuring on the surface of an elastic half-space due to a buried 

Table 1
Macro-seismic information for Tocopilla 2007 mainshock and some aftershocks.

<table>
<thead>
<tr>
<th>Lon</th>
<th>Lat</th>
<th>Strike</th>
<th>Dip</th>
<th>Mw</th>
<th>Date</th>
<th>Time</th>
<th>Depth [km]</th>
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<tbody>
<tr>
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<td>22.64</td>
<td>358</td>
<td>20</td>
<td>7.7</td>
<td>2007/11/14</td>
<td>15:40</td>
<td>37.6</td>
</tr>
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<td>71.12</td>
<td>23.03</td>
<td>339</td>
<td>22</td>
<td>5.3</td>
<td>2007/11/15</td>
<td>11:14</td>
<td>19.8</td>
</tr>
<tr>
<td>71.01</td>
<td>22.99</td>
<td>5</td>
<td>21</td>
<td>6.3</td>
<td>2007/11/15</td>
<td>15:03</td>
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</tr>
<tr>
<td>70.94</td>
<td>22.98</td>
<td>360</td>
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<td>6.8</td>
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<td>15:06</td>
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<td>7</td>
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<td>08:42</td>
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<td>70.88</td>
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<td>353</td>
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<td>2007/11/17</td>
<td>18:13</td>
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<tr>
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<td>19.12</td>
<td>246</td>
<td>44</td>
<td>5.8</td>
<td>2007/11/18</td>
<td>07:02</td>
<td>98.4</td>
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<td>5.0</td>
<td>2007/11/26</td>
<td>06:14</td>
<td>36.9</td>
</tr>
</tbody>
</table>

Table 2
SAR scene data used in this study (‘D’ and ‘A’ denote a descending and ascending passes).

<table>
<thead>
<tr>
<th>Satellite</th>
<th>Master date</th>
<th>Slave date</th>
<th>Path</th>
<th>Frames</th>
<th>B1 (m)</th>
<th>Pass</th>
</tr>
</thead>
<tbody>
<tr>
<td>ALOS</td>
<td>10/14/2007</td>
<td>29/11/2007</td>
<td>103</td>
<td>6720, 6730, 6740</td>
<td>538</td>
<td>A</td>
</tr>
<tr>
<td>ENVISAT</td>
<td>5/11/2007</td>
<td>10/12/2007</td>
<td>096</td>
<td>4041, 4059, 4077</td>
<td>178</td>
<td>D</td>
</tr>
<tr>
<td>ENVISAT</td>
<td>20/10/2007</td>
<td>24/11/2007</td>
<td>368</td>
<td>4041, 4059, 4077</td>
<td>271</td>
<td>D</td>
</tr>
</tbody>
</table>
Shear dislocation is called, herein, the forward problem. The inverse problem is to infer the fault geometry and slip distribution from measurements of the static displacements. Closed-form solutions exist for the co-seismic displacement field due to a buried rectangular fault with constant uniform slip [35]. These are used to discretize and assemble larger and more complex faults with variable slip distribution. Once the large fault has been divided into subfaults, the inverse problem, i.e., to identify fault parameters (strike, dip, fault slip distribution and dimensions) at each sub-fault from the InSAR co-seismic displacement field can be stated as

\[ u(x, g, m, v) = G(x, g, m) \cdot m = d(x) \]

where \( d(x) \) is the vector of known LOS co-seismic displacements at points \( x \); \( G \) is the static Green function matrix containing in each column the static displacement at \( x \) for a unit slip in each patch element of the fault projected on the LOS vector; \( g \) represents geometric parameters of the model, namely fault location, orientation (strike and dip), dimensions and discretization; and \( v \) the crustal Poisson’s ratio obtained from a 1D model of the crust in this region (Crust 2.0 [36]) together with other elastic parameters; and \( m \) collects the unknown slip parameters associated with the chosen patch discretization to be identified. In other words, “find \( g \) and \( m \) such that \( u \) is as close as possible to the measurements \( d \)”.

In order to simplify the problem numerically it is common to reduce the number of unknowns by using some additional physical assumptions or known information. For example, it is usual to have a centroid moment tensor solution (CMT) for the earthquake being studied, which provides information of the fault location (hypocenter and depth), the fault plane orientation (strike and dip), and expected seismic moment. By using this information, the geometric parameters \( g \) are constrained and the remaining linear inverse problem is solved for the slip parameters. Details of this methodology have been presented earlier in the literature [11].

In this study, strike and dip angles along with a hypocenter location were obtained from previous studies which used wavelet-domain inversion [37] of teleseismic waves due to each event.
Faulting extent in strike and dip direction are adjusted manually until the identified slip distribution is completely contained within the identified fault. By discretizing the slip and observed co-seismic displacement field, and under the assumptions described, the linear inversion problem is \( Gm = d \), where \( G \) is constant; \( d \) is given by the observed LOS co-seismic displacements; and \( m \) is the vector of unknown model parameters defining the strike and dip components of slip for each fault patch.

Eq. (6) is over determined, even after the additional simplifications made, since model parameters are in the few hundreds whereas InSAR displacements are in the millions of observation pixels. Data reduction techniques are, therefore, employed to down-sample the InSAR image by selecting points in areas of the image which have greater impact on the resolution of the slip distribution, such as areas with high deformation gradients and discarding other points. Spatial averages of the LOS displacement are taken around the chosen points to reduce the error associated with the measurement. The averaging window sizes are selected based on the local displacement field gradients, small windows are selected for areas with high gradient and high data quality. On the other hand, areas with low quality interferometric signal (low coherence) are usually masked out of the scene.

Solutions to this problem should provide smooth 2D slip distributions as well as account for modeling errors produced by incorrect InSAR baseline estimations and post-seismic creep among other factors [9,38]. Thus, the inversion model is expanded to include these considerations. The final inverse problem to be solved which may be written as [11]

\[
\begin{bmatrix}
G & Q \\
\kappa^2 D & 0
\end{bmatrix}
\begin{bmatrix}
m \\
m_Q
\end{bmatrix}
= \begin{bmatrix}
d \\
0
\end{bmatrix}
\tag{7}
\]

where \( D \) is a smoothing discretized Laplacian matrix acting on the slip distribution with smoothing weight \( \kappa^2 \); \( Q \) is a matrix which models the propagation of baseline errors into InSAR displacements, and is usually assumed to be a quadratic function of the space coordinates with coefficients and \( m_Q \). Models with different \( \kappa^2 \) values are inverted to assess the effect of this parameter. The final value is chosen to minimize the tradeoff between smoothing and solution error [11].

Eq. (7) was solved using a constrained linear least-squares solver [39], which reinforces expected slip directions. In this subsection zone, the slip was constrained to reverse faulting and to left-lateral strike slip motion (positive strike and dip-slip).

Despite other authors using a curved fault geometry, a plane geometry was chosen for the fault model for two reasons. First, it was found that results of the model fit for this earthquake were more sensitive to the \( \kappa^2 \) parameter than the geometry of the fault and, second, it is much easier to translate this geometry to the geometry needed by the stochastic modeling program mentioned later.

The identified fault mechanism is shown in Fig. 5 together with the resulting co-seismic displacement field. Arrows reflect horizontal displacements while contours reflect vertical displacements. The Tocopilla event was an underthrusting earthquake that ruptured the Nazca-South American plate interface with an epicenter roughly 40 km south east of the coastal town of Tocopilla, Chile; the rupture is a northward continuation of the 1995 Antofagasta earthquake. The identified slip distribution shows slip concentration in two zones which is consistent with other studies [40,41]. The seismic moment obtained from the inversion was the same to one decimal place as the one reported in these studies (\( M_w=7.7 \)). Please note that the selection of the fault strike is consistent with the reported value and with the overall observed deformation pattern.

Shown in Fig. 6 is the comparison between the real and the synthetic co-seismic interferograms for one satellite path. Also shown at the rightmost column is residual image between the real and synthetic interferograms. This image shows an accuracy of \( \pm 2.5 \) cm within this interferogram which can be attributed to misfit, atmospheric effects not accounted for, noise, subsidiary faulting, localized effects, etc. Note that the high gradient and high deformation areas present in the image are predicted with a higher accuracy. Furthermore, the larger residuals occur in areas far from the main deformation and so have little impact on the resolution of the fault slip. Similar remarks apply to the other frames involved in the study. The presence of post-seismic creep is minimized by using interferograms which span at most 25 days (after the main event) and by inverting simultaneously the quadratic ramp mentioned earlier. Thus it is deemed that the inversion results are adequate for the purposes of this article which is to show the enabling technology with a simple case. A model which is more geophysically correct is deemed of high value but not further sought.

5. Recorded accelerations and displacements

Acceleration data for the Tocopilla 2007 event were obtained from the Integrated Plate-Boundary Observatory in Chile (IPOC), a network of broadband seismometers, accelerometers, and GPS stations coordinated by GFZ Potsdam and the Institut de Physique
du Globe de Paris (IPGP) in collaboration with local institutions, Universidad Católica del Norte and Universidad de Chile. Records from 11 stations located in bedrock, denoted as PB01-02, PB04-08 and HMB, PSG, MNM and PAT, are corrected for co-seismic displacement.

Instrumental response was first removed from the measured data by using the instrument zero-pole-gain parameters provided by the manufacturers (Guralp CMG5 for the GFZ stations, and Kinematics ETNA for the IPGP stations). The instrument corrected accelerations present low-frequency noise which alters the baseline of the records. The usual procedure to eliminate this noise is to apply a high-pass filter of order 4. The cutoff period was chosen to account for the known co-seismic displacement field.

A summary of the peak ground values and co-seismic displacements inferred from InSAR (values in parenthesis are those obtained by high-pass filtering).

The intermediate interval times $t_1$ and $t_2$ are free parameters to be chosen so that the final displacement provides the observed co-seismic displacement. Possible bounds for $t_1$ range from $t_0$ to the PGA time of the record, whereas for $t_2$ range from the PGA time till the end of the record, $t_0$. A baseline correction is determined for each time pair $(t_1, t_2)$ and the co-seismic displacement (average for the final seconds of the record) is compared with the InSAR value. Many $t_1, t_2$ pairs meet the criterion of producing a residual displacement value near the co-seismic displacement, thus an additional criterion is needed to find a feasible pair. Extreme cases include very near or very far apart $t_1$ and $t_2$ values, with both extremes not being physically meaningful. Choosing the pair which minimizes the record RMS velocity was found to be an adequate criterion, yielding $t_1$ and $t_2$ values which tended to bracket the strongest portions of the record, which is where the co-seismic displacement is expected to occur, in other cases the interval contained portions of the record which exhibited strong pulses. Within the $t_2$ interval the acceleration baseline exhibits in this interval a complicated path that in average is represented by $a_2$ [43]. A summary of the $t_1$ and $t_2$ pairs found is given in Table 4.

The filtering procedure was performed using a standard acausal Butterworth filter of order 4. The cutoff period was chosen at $t_c=10$ s for all records. The chosen period cutoff is consistent with the maximum usable period found for digital instruments for this magnitude of events [42] and was, thus, chosen as the best possible filter for this event.

The baseline correction procedure chosen subdivides the record into three intervals, namely $I_1 = [t_1, t_c]$, $I_2 = [t_c, t_2]$ and $I_3 = [t_2, t_e]$, where $t_1$ and $t_e$ are start and end times of the record defined by its cumulative 1% and 99% Arias intensity (also called the Husid plot). If $a_0(t)$ is the instrument corrected data and $a_1(t)$ the acceleration baseline, the corrected acceleration is $a(t)=a_0(t)-a_1(t)$. The proposed baseline $a_1(t)$ is a piecewise constant function for each interval, $a_1$ for $t \in [t_1, t_c]$ with $l=1,2,3$.

In summary the procedure to compute $a_1$, $a_2$, and $a_3$ uses parabolic fits to the uncorrected displacements within certain subintervals of $I_1$ and $I_2$ (leading to $a_1$ and $a_2$) and velocity continuity for the displacement fits (leading to $a_2$). More details on this can be found elsewhere [43].

Table 3
Summary of peak ground values and co-seismic displacements inferred from InSAR (values in parenthesis are those obtained by high-pass filtering).

<table>
<thead>
<tr>
<th>Station</th>
<th>E–W</th>
<th>N–S</th>
<th>Up-down</th>
<th>E–W</th>
<th>N–S</th>
<th>Up–down</th>
<th>Static offset</th>
</tr>
</thead>
<tbody>
<tr>
<td>PB01</td>
<td>0.294</td>
<td>0.120</td>
<td>0.114</td>
<td>9.7</td>
<td>5.6</td>
<td>4.3</td>
<td>(1.3)</td>
</tr>
<tr>
<td>PB02</td>
<td>0.533</td>
<td>0.409</td>
<td>0.390</td>
<td>24.4</td>
<td>14.9</td>
<td>14.0</td>
<td>(1.5)</td>
</tr>
<tr>
<td>PB03</td>
<td>0.372</td>
<td>0.482</td>
<td>0.295</td>
<td>28.4</td>
<td>13.2</td>
<td>10.5</td>
<td>(3.5)</td>
</tr>
<tr>
<td>PB05</td>
<td>0.369</td>
<td>0.555</td>
<td>0.193</td>
<td>26.5</td>
<td>13.2</td>
<td>9.4</td>
<td>(6.6)</td>
</tr>
<tr>
<td>PB06</td>
<td>0.285</td>
<td>0.299</td>
<td>0.145</td>
<td>12.7</td>
<td>13.3</td>
<td>7.3</td>
<td>(5.0)</td>
</tr>
<tr>
<td>PB07</td>
<td>0.377</td>
<td>0.457</td>
<td>0.335</td>
<td>20.3</td>
<td>16.0</td>
<td>14.6</td>
<td>(27.8)</td>
</tr>
<tr>
<td>PB08</td>
<td>0.032</td>
<td>0.041</td>
<td>0.019</td>
<td>2.4</td>
<td>2.8</td>
<td>2.8</td>
<td>(2.8)</td>
</tr>
<tr>
<td>HMB</td>
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<td>0.072</td>
<td>0.039</td>
<td>3.6</td>
<td>4.3</td>
<td>2.8</td>
<td>(1.7)</td>
</tr>
<tr>
<td>PAT</td>
<td>0.089</td>
<td>0.098</td>
<td>0.052</td>
<td>8.4</td>
<td>5.9</td>
<td>5.1</td>
<td>(2.9)</td>
</tr>
<tr>
<td>PSG</td>
<td>0.023</td>
<td>0.020</td>
<td>0.014</td>
<td>2.2</td>
<td>1.4</td>
<td>1.0</td>
<td>(2.8)</td>
</tr>
</tbody>
</table>

Table 4
$t_1$ and $t_2$ values for the proposed correction of the Tocopilla 2007 ground motion database. Times are given in seconds relative to the beginning of the record.

<table>
<thead>
<tr>
<th>Station</th>
<th>East–west</th>
<th>North–south</th>
<th>Up–down</th>
</tr>
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<tr>
<td>PB01</td>
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<td>14.75</td>
<td>19.8</td>
</tr>
<tr>
<td>PB02</td>
<td>27.21</td>
<td>62.8</td>
<td>35</td>
</tr>
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<td>PB04</td>
<td>16.96</td>
<td>74.95</td>
<td>13.04</td>
</tr>
<tr>
<td>PB05</td>
<td>23.32</td>
<td>94.72</td>
<td>16.2</td>
</tr>
<tr>
<td>PB06</td>
<td>17.28</td>
<td>39.88</td>
<td>11.18</td>
</tr>
<tr>
<td>PB07</td>
<td>27.54</td>
<td>70.28</td>
<td>29.08</td>
</tr>
<tr>
<td>PB08</td>
<td>87.41</td>
<td>38.81</td>
<td>102.8</td>
</tr>
<tr>
<td>PB06</td>
<td>94.62</td>
<td>27.59</td>
<td>51.96</td>
</tr>
<tr>
<td>PATCX</td>
<td>6.39</td>
<td>24.01</td>
<td>18.62</td>
</tr>
<tr>
<td>PSGCX</td>
<td>158.94</td>
<td>36.45</td>
<td>38.87</td>
</tr>
</tbody>
</table>

The filtering procedure was performed using a standard acausal Butterworth filter of order 4. The cutoff period was chosen at $t_c=10$ s for all records. The chosen period cutoff is consistent with the maximum usable period found for digital instruments for this magnitude of events [42] and was, thus, chosen as the best possible filter for this event.

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In summary the procedure to compute $a_1$, $a_2$, and $a_3$ uses parabolic fits to the uncorrected displacements within certain subintervals of $I_1$ and $I_2$ (leading to $a_1$ and $a_2$) and velocity continuity for the displacement fits (leading to $a_2$). More details on this can be found elsewhere [43].
Shown in Fig. 7 is a typical displacement response spectra for one arbitrary station, PB06, comparing the proposed method with regular filtering. Discrepancies in spectral displacement ordinates become apparent starting at about 3–4 s and may be significant at longer periods. One salient aspect about the proposed correction scheme is that it leads to spectra which do not converge to the PGD except at very long periods. This occurs when the co-seismic displacement is larger than the PGD value or when it is associated to a very long period pulse linked to the co-seismic buildup of the residual displacement. As structural periods get longer the displacement demand will increase unbounded until the co-seismic value is reached. This can have profound implications for design spectra used in future versions of Chile’s seismic codes.

Another alternative representation of the corrected normalized response spectra for the different Tocopilla stations in rock is shown in Figs. 8(a) and (b). The east–west and up–down components have been normalized relative to their corresponding PGA and PGD. The average spectral value is indicated in the darker line and the rectangles reflect the standard deviation associated with the mean estimations. Values outside the unit square imply acceleration and displacement amplifications. The 45° lines indicate normalized period $T = T_0$ where $T_0 = 2\pi \sqrt{\text{PGD/PGA}}$ takes different values for the spectrum at each station. This figure shows that, in average for horizontal components, it cannot be assumed that the maximum spectral displacement is larger than the PGD as is the case of spectral accelerations which are usually higher than the PGA. The average line for different events could be used to separate places where the displacement spectrum presents amplifications of the PGD from those where this does not occur. Also noteworthy is that the up–down components do, in average, present spectral displacement amplifications over the PGD. On the other hand, filtering leads to an amplification of the elastic displacements and then a convergence to a lower PGD value, which is similar to the classical shape assumed for the design displacement spectra.

Table 3 shows that the PGD values are significantly modified by the correction procedure, while the PGA and PGV values are less affected. Values in Table 3 within parenthesis correspond to direct filtering of the record without accounting for the co-seismic displacement.

Finally, Fig. 9(a) shows the east–west corrected components of rock acceleration and displacement recorded during the Tocopilla (2007) earthquake. Only stations with significant ground motion are presented. The decay of the co-seismic displacement with distance is apparent. The residual displacement was always met within a 1% error, which was the tolerance set to the algorithm.

### 6. Generation of synthetic records

Next, a slight modification of an existing methodology [19] is tested and used to generate synthetic ground motions compatible with the observed co-seismic displacement field. The analysis splits the high-frequency (> 1 Hz) and low-frequency (< 1 Hz) components of the signals. The high-frequency component will be generated by an existing finite-fault formulation of the stochastic method with an underlying $\nu^2$ source spectrum and dynamic corner frequency [18,22,21]. The finite-fault model geometry and slip distribution (converted to seismic moment distribution) are directly imported from InSAR inversion results, and the attenuation and duration models calibrated to reproduce as best as possible the recorded motions for the Tocopilla 2007 earthquake; some model parameters are obtained from previous research [40,44,41]. The
low-frequency signal component is modified by adding a one-sided velocity pulse [29], which represents the near-field co-seismic displacement buildup. Because the detail of the stochastic method may be found in previous research [23,17,18,22,19,24], the presentation next only provides some essential concepts of the technique as relevant to this study.

The low-frequency component of the synthetic acceleration record was modeled using a deterministic velocity pulse wavelet [29]

\[
v(t) = \frac{A}{2} \left[ 1 + \cos \left( \frac{2\pi f_p}{2} (t - t_0) \right) \right] \cos(2\pi f_p (t - t_0) + \psi) \\
t_0 - \frac{\gamma}{2f_p} \leq t \leq t_0 + \frac{\gamma}{2f_p}
\]

and zero elsewhere. This wavelet corresponds to an amplitude modulated cosine with a prevailing frequency parameter \(f_p\). \(\gamma\) controls the peak velocity of the pulse (PGV); \(\gamma\) determines the oscillatory nature of the pulse; \(t_0\) is the time of occurrence of the peak of the wavelet envelope; and \(\psi\) is the phase shift of the modulated signal. By integrating this velocity pulse with zero initial displacement, the final value of the static displacement is

\[
A_{\text{static}} = \frac{A}{2\pi f_p} \frac{\sin(\pi \gamma)}{1 - \frac{2\pi f_p}{\gamma}} \cos(\psi)
\]

which is taken equal to the final co-seismic InSAR displacement. The main hypothesis behind this model is that the PGD is very sensitive to the correct characterization of this pulse.

The estimation of the five wavelet parameters \(A, f_p, \gamma, t_0\) and \(\psi\) in Eq. (9), should be done ideally using a suite of recorded velocity traces which incorporate the effect of near-fault directivity and co-seismic displacement by using the proposed baseline correction criterion. Unfortunately, a complete database of records with these characteristics is not yet available for the region, and estimations of these parameters need to be based on a mixture of available information, recorded data, some reasonable physical assumptions, and also some judgement. As an alternative, physical wave-propagation models could be used to capture the scaling relation of these parameters with event magnitude, distance and site to source geometry, but this is beyond the scope of this article. Instead, note that the \(f_p\) parameter is related to the pulse interval length, which implies that using this wavelet to model co-seismic fling relates this parameter to the rise time of the displacement. Thus, \(f_p\) is chosen such that the interval length is the same as the duration of the strongest motion, since it had already been noted that this buildup occurs within this interval. This interval starts when the first S-waves arrive from the event nucleation and continues until all of the main asperities have slipped and the corresponding S-waves have reached the site. Intuitively if the rupture speed of the event is large (as in super-shear events) the co-seismic fling step pulse should resemble the step response of the crust or, if the rupture speed is very slow, then the co-seismic pulse should resemble the slip function convolved over the entire rupturing fault with the crustal impulse response. In general the observed co-seismic pulse should be something between these two extremes.

To obtain this time interval a characteristic dimension of an equivalent fault with uniform slip equal to the average identified slip which outputs the same amount of total seismic moment as the real event is calculated. By using this characteristic length and the rupture speed, the pulse interval can be estimated, which is the time taken by a rupture with speed \(v_{\text{rup}} = \gamma f\) to transverse this length. The characteristic length \(a\) that preserves the total released seismic moment and average slip is given by \(a = \frac{\gamma}{2\pi f_p} M_0 / \mu \Delta\) where \(\Delta\) is the average slip on the extended fault model, \(\mu\) the crustal shear modulus, \(M_0\) the released seismic moment and \(\alpha\) a shape factor. The chosen value for the Tocopilla event is \(\alpha \approx 3\), which is

![Fig. 9. Baseline-corrected ground motions for the Tocopilla (2007) earthquake in the east–west direction: (a) acceleration and (b) displacement.](image-url)
equivalent to a rectangular fault with a 1:3 side aspect ratio. With the characteristic dimension, the model average rupture velocity (in the case of Tocopilla this is $V_{rup}\approx3.09\text{ km/s}$), the $f_s$ parameter is estimated with good correlation to observed motions.

Furthermore, as noted by Mavroiedis and Papageorgiou [29], the $\gamma$ parameter was close to unity when there was residual displacement. The chosen value here is $\gamma=1.1$ since unity is mathematically forbidden. $\psi$ is chosen to vary from $30^\circ$ to $90^\circ$ to model the change of the pulse waveform observed in the far-field.

The following is a summary of the parameters chosen to model the present case. The parameters chosen and, especially, the attenuation and subevent duration models were calibrated to match recorded motions during the Tocopilla 2007 earthquake. They are $Q_0=1500$, $\eta=0.5$, $R_1=45\text{ km}$, $p_1=1$, $R_2=100\text{ km}$, $p_2=-0.1$, and $p_3=2$. The $f_{\text{max}}$ filter was used with a 1 Hz cutoff frequency, and the parameters used in the subevent duration model were $R_{\text{min}}=45\text{ km}$, $T_{\text{min}}=0.5\text{ s}$, $R_{\text{max}}=\infty$, and $b_1=0.04$. More details on the meaning of these parameters and a succinct explanation of the SFFM can be found in Appendix A and more in the references.

The superposition of the low and high-frequency models yields the final simulated records. Because as presented, the SFFM is unable to distinguish different ground motion components, the only difference in synthetic record components will be found in the long-period range which must integrate to a different co-seismic displacement value.

Shown in Fig. 10 is a typical comparison of one realization of a synthetic acceleration record along with the envelope of all realizations. In the near field stations (within the co-seismic field) the PGA of the E–W component tended to agree with the simulated results whereas the other components were overestimated. The synthetic record begins after the real record because the stochastic method only models S-Wave propagation and the first motions shown are due to P-Waves primarily. The duration of the synthetic records is somewhat longer than the real record, due to misfit in the duration and attenuation model, but the effect of the two asperities is visible in the synthetics. In the far-field stations (PB02 and beyond) the duration tends to become similar to the real recorded duration. Also shown is the resulting mean response spectrum for 20 synthetic record realizations plotted against the recorded signal and spectrum for site PB04 (Tocopilla). This site is located in the near field above the identified faulting plane. Twenty realizations per component and per station were deemed enough to capture the variability of the stochastic method. The point to point differences in time series are apparent, yet the co-seismic displacement is equal as expected. The misfit in the average PGA shows how the model is unable to account for what occurred at all stations despite the effort to fit the model. Results for other near field stations show similar trends and are presented elsewhere [34]. They show that the model tends to lead to a better approximation in average sense for near field stations than for far-field ones favoring east–west components rather than north–south or up–down components which are overestimated.

7. Conclusions

This paper proposes InSAR as a useful tool which is complementary to current strong-motion acceleration networks, allowing a
correction of the recorded motions which approximately preserves the long period components of these. Furthermore, generation of artificial records for a particular earthquake also benefits from InSAR since these records can now be consistent with the co-seismic displacement field of the event.

The analysis was performed for the Tocopilla 2007 earthquake which featured adequate SAR coverage and ground instrumentation. Based on the interferograms produced, the co-seismic displacement field was obtained for the two regions affected, and used to identify the fault parameters and slip distribution. The modeled displacement field was used to correct existing ground motion data and compared with the same records corrected by regular acausal filtering. The results show that the filtering underestimates the PGD, the long-period spectral values, and the residual displacement is eliminated, whereas the proposed methodology shows residual displacement and larger values of spectral displacements at long periods and PGD. When the PGD is lower than the co-seismic displacement the response spectrum grows monotonically until reaching the PGD, showing no amplification for lower frequencies for the east–west components. This is not observed for the north–south or up–down components. This has great impact on the shape adopted for the code design spectrum and more studies should be performed to assess the true shape of the spectrum in sites prone to large co-seismic displacements.

Additionally, the data made available by the InSAR inversion was used to the extent possible to test a stochastic methodology to compute synthetic ground motions which are compatible with the measured data. Corrected ground motion records were qualitatively compared with synthetic accelerograms for the same locations showing the goodness of fit of the stochastic model. Notable was the fit of the model for east–west components located near the fault when compared to stations in the far-field. Additionally, north–south and up–down components were overestimated due to the incapability of the model to distinguish horizontal components or produce results for the vertical component.

More analysis of similar data and testing of other ground motion synthesis models are probably required before stating certain general trends, we have intentionally avoided proposing herein a design spectrum compatible with co-seismic displacement for the region under study. However, such result would be straightforward, and though approximate, much closer to reality than current design spectra derived from ground motion records with this co-seismic displacement filtered or processed using methods which do not account for this effect.

InSAR results are extraordinary in terms of their simplicity and resolution quality. There is a promising future for this technique in connection with practical seismic engineering. Interferograms also allow to monitor uninstrumented areas, and show peculiarities attributed to local site effects, such as the activation of secondary faults, local subsidence or inflation, topographic amplification that are difficult to capture by localized instrumentation. A shortcoming, however, is the inability of the method to resolve the time effects on the faulting process, which needs to be assumed. Although not presented herein, the Antofagasta (1995) and Tocopilla (2007) and Pisco (2007) [34], and the three form a good database to develop interesting regional studies of seismic hazard. Please note that although the results herein were derived by the authors, they were possible thanks to the application of ideas, developments, and procedures devised by many other researchers from the area of geophysics. The goal was to bring a structural engineering perspective and practical application to all these relevant results, take advantage of InSAR technology, and move forward in the direction of proposing a consistent design spectrum for the design of flexible structures in the period range of 3 s and beyond.

Acknowledgements

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Appendix A. The stochastic finite-fault model

The spectrum of the synthesized ground motion at a site is given by \( Y(f|R,M_0) = E(M_0,f) | R | \mathcal{G}(f) \mathcal{W}(f) \). For small amplitude events [18], the theoretical functional form for \( E(M_0,f) \) is given by the \( \omega^2 \)-spectrum, or Brune point source, and its mathematical expression is

\[
E(M_0,f) = \frac{M_0}{4\pi \rho \beta^2} \left[ 1 + \left( \frac{\omega}{\omega_c} \right)^{2} \right]^{-1} A^{60} \cdot \hat{n}^{60} \tag{A.1}
\]

where \( A^{60} \) is the near field radiation pattern vector; \( \hat{n}^{60} \) is a unit vector which points from the source to the site; \( \theta \) and \( \phi \) are the spherical angles used to describe this vector; \( M_0 = \mu A R^5 \) is the total released seismic moment; \( \beta \) is the S-wave propagation speed; and \( \rho \) is the crustal average mass density. This spectrum corresponds to that of the observed ground motion at a site without including path-dependent effects like attenuation. The corner frequency, in the inverse of the fault rise time \( \omega_c = 1/\tau \), is given by [17]

\[
f_c = \frac{\nu_s^2}{\beta^2} \frac{1}{\mathcal{I}} \left( \frac{\tau}{t_s} \right)^3 \quad M_0 = \Delta \sigma L^3 \tag{A.2}
\]

where \( f_c \) represents the corner frequency (scaling law); and \( M_0 \) the seismic moment (scaling law); \( L \) is an equivalent fault radius or predominant length; \( \nu_s = v_{rms} / \beta \) is the ratio of rupture speed to shear wave velocity; and \( z_s = \tau_s / t_s \), with \( t_s \) the time needed to reach \( x = u(t_s)/v_{s} \) of the final static slip (typically \( z_s = 0.5 \)). The seismic moment \( M_0 \) is related to an equivalent fault-radius and stress drop parameter \( \Delta \sigma \), which in the SFFM has no attached physical meaning [19] and is considered as a free model parameter. In any case \( \Delta \sigma \) controls the structural amplitudes at frequencies greater than \( f_c \) and is used to calibrate with observed ground motions.

A model used for the path effect including distance and frequency dependent terms is [17]

\[
P(R,f) = Z(R) \exp \left\{ -\frac{R}{Q \Omega f^\eta} \right\} \quad \Omega = R \leq R_1 \quad Z(R) = \left\{ \begin{array}{ll}
R_0 / R, & R \leq R_1, \\
Z(R_1)(R_1 / R)^{\delta}, & R_1 < R \leq R_2, \\
\vdots & \\
Z(R_0)(R_0 / R)^{\delta}, & R_0 < R < \infty
\end{array} \right. \tag{A.4}
\]

where \( Z(R) \) includes the geometric attenuation term; \( R_0 \) is a minimum distance (usually \( R_0 = 1 \)) and the exponential term is a frequency dependent anelastic attenuation with the quality factor \( Q \) of the form [45] \( Q(f) = \max(Q_0 f^\eta, Q_{min}) \), where \( Q_0, Q_{min} \) and \( \eta \) are parameters to fit this quality factor to the ground motions.
Table 5

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
<th>Units</th>
<th>Source</th>
</tr>
</thead>
<tbody>
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<td>bars</td>
<td>Atkinson et al.</td>
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<td>%</td>
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<tr>
<td>$V_{rupture}/\beta$</td>
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<td></td>
<td>Pritchard et al.</td>
</tr>
<tr>
<td>Crust density ($\rho$)</td>
<td>$2.88$</td>
<td>(g/cm$^3$)</td>
<td>This study$^a$</td>
</tr>
</tbody>
</table>

$^a$ Epicenter location coincides with NEIC catalog, depth was chosen to be compatible with the proposed fault model.

References

[44] 7th international symposium on Andean geodynamics. The Mw 7.7 Tocopilla earthquake of November 2007: characteristics of a subduction earthquake that occurred in the brittle-ductile transition zone of the northern Chile seismic gap, ISAG; 2008.

