Measuring Low Fault Strain Rate with Synthetic Aperture Radar: Application to the Pacific-North America Plate Boundary

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MEASURING LOW FAULT STRAIN RATE WITH SYNTHETIC APERTURE RADAR: APPLICATION TO THE PACIFIC-NORTH AMERICA PLATE BOUNDARY

By

Noel Gourmelen

A DISSERTATION

Submitted to the Faculty of the University of Miami in partial fulfillment of the requirements for the degree of Doctor of Philosophy

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MEASURING LOW FAULT STRAIN RATE WITH SYNTHETIC APERTURE
RADAR: APPLICATION TO THE PACIFIC-NORTH AMERICA PLATE
BOUNDARY

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I use Synthetic Aperture Radar Interferometry (InSAR) to study the present deformation in the Western Basin and Range and Basin and Range – Sierra Nevada transition. I process 350 SAR data over $190\times10^3 \text{ km}^2$ for the period 1992 to 2002. Both stacking and time series processing were applied to produce precise (mm/yr) and high-resolution velocity map for the area. Two new processing techniques have been developed. The first technique solves for the long wavelength ambiguities of the InSAR derived velocity map that arise due to uncertainty in the orbital parameter of the satellite. The technique assimilates continuous GPS data into the InSAR time-series processing. The second technique extracts the horizontal and vertical components of the deformation field from two adjacent radar tracks.
I applied stacking to study the transient deformation across the Central Nevada Seismic Belt and interseismic strain accumulation across the Eastern California Shear Zone.

I show that the current deformation across the Central Nevada Seismic Belt can be explained by a combination of inter-seismic, post-seismic and anthropogenic deformation. The Post-Seismic deformation is associated with visco-elastic relaxation of the Earth's mantle in response to a centennial earthquake sequence of five ~M7 earthquakes along the Central Nevada Seismic Belt. The anthropogenic deformation is a response of the bedrock to water withdrawal in support of mining activity.

A more evolved time-series approach that solves for orbital errors is applied across the Eastern California Shear Zone. The study shows that the Hunter Mountain – Panamint Valley fault system accommodates ~5 mm/yr, a faster rate than geological averages. The region of strain accumulation is a narrow band of ~10 km centered on the Hunter mountain fault, and indicates a very shallow locking depth in agreement with an active low angle normal fault system.
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CHAPTER 1  INTRODUCTION

Low strain rate could be defined as a rate which is comparable or lower than the precision of a given geodetic technique. In earth sciences, the geodetic techniques used to measure ground deformation are mainly Global Positioning System (GPS), and Synthetic Aperture Radar Interferometry (InSAR). GPS and InSAR have distinct precision due to distinct strengths and weaknesses. The strength of the GPS technique lies in its ability to measure absolute, large-scale slow motion. The strength of the InSAR technique is its high measurement resolution and that it is a full remote sensing technique, requiring no fieldwork. The ability of InSAR to measure long wavelength, low strain rate, is poor. It is generally accepted that 0.1mm/yr/km is a typical lower bound. The two techniques are complementary, and combined, can be used to produce high resolution and high precision measurement in region of low strain rate.

The inter-seismic and post-seismic phase of the earthquake cycle are processes associated with low strain rate. In particular, deforming region with fault systems composed of multiple individual and parallel faults, and post-seismic visco-elastic relaxation of the lower crust or upper mantle.

The present study focuses on the deformation across the Eastern California Shear Zone (ECSZ) and Central Nevada Seismic Belt (CNSB), two contiguous regions of interseismic and post-seismic strain accumulation within the Pacific – North America plate boundary.
Both the ECSZ and CNSB are well covered with SAR acquisition from the European Space Agency (ESA) European Remote-sensing Satellites ERS1 and ERS2 and a network of permanent GPS stations.

1.1 **Synthetic Aperture Radar Interferometry (InSAR) limitation in measuring Low Strain Rate**

1.1.1 **InSAR Principles and Sources of Errors**

InSAR is a remote sensing technique that measures changes in electromagnetic wave travel time between the satellite and the ground between two epochs by exploiting the phase difference (usually referred to as interferogram) of two Synthetic Aperture Radar (SAR) images [Gabriel et al., 1989]. The change in the wave travel time is related to ground motion between epochs, the satellite position, and atmospheric water vapor. Applications of ground motion measurement include the measure of deformation associated with fault activity [Massonnet et al., 1993], glacier flow [Goldstein et al., 1993], volcanic activity [Massonnet et al., 1995; Amelung et al., 2000], water level change [Wdowinski et al., 2004], land subsidence [Amelung et al., 1999; Dixon et al., 2006]. For a review of the technique and applications, see [Massonnet and Feigl, 1998; Rosen et al., 2000; Rocca et al., 2000; Burgmann et al., 2000; Hanssen, 2001].

Multiple sources contributes to the measured phase difference between two epochs, the contribution from ground deformation, the contribution from atmospheric water vapor, and the contribution from noise sources such as incomplete removal of the topography, incomplete removal of the orbital
contribution, surface decorrelation, thermal noise and processing noise (such as due to interpolation.

Contribution from atmospheric water vapor, and the contribution from noise sources needs to be mitigated to achieve a high precision on the determination of the ground deformation.

The precision of the InSAR technique improved in recent years with the availability of large amount of SAR data and the development of new methods such as Persistent Scatterer (PS) [Ferretti et al., 2001; Hooper et al., 2004] and Small BAseline Subset (SBAS) [Berardino et al., 2002; Lanari et al., 2007] that produces phase time series.

Contribution from atmospheric water vapor, also referred as Atmospheric Phase Screen (APS), is a spatially correlated but temporally uncorrelated phenomena [Hanssen, 2001], its effect on the measured phase is reduced by a combination of temporal and spatial filtering on the time series [Ferretti et al., 2001; Berardino et al., 2002].

A selection process of coherent ground targets using statistical approaches accounts for the decorrelation term $\phi_{\text{decorr}}(t_2, t_1)$ and selecting coherent targets over time. $\phi_{\text{int}, \epsilon}(t_2, t_1)$ is minimized by performing the processing in radar geometry, thus minimizing the amount of resampling of the initial data.
Orbital phase errors (OPE) are a systematic cause of errors on the measure of $\phi_{def}$, despite the availability of precise SAR satellites orbits and methods to mitigate their effects.

1.1.2 SAR Orbits

The present study uses precise orbits for the ERS1 and ERS2 satellites, provided by the Delft Institute of Technology (http://www.deos.tudelft.nl/ers/precorbs/orbits/). The Delft precise orbits are obtained from a combination of tracking (Satellite Laser Ranging or SLR), gravity model and dynamical models (Atmospheric density, …). Tracking from SLR is only available at a coarse resolution and unevenly around the globe, requiring the use of dynamic models to propagate the orbits at other locations. Delft precise orbits are believed to have a radial precision as low as 5 cm [Scharroo, 2002]. The error on the relative position of two orbits is thought to be lower, as error on the trajectory of the satellite is correlated with the spatial location, two successive estimated orbits will differ from each other by less than they differ from the true position.
A precision of a millimeter on the knowledge of orbit parameters is necessary to remove totally the phase contribution of the orbit in an interferogram [Hanssen, 2001]. Because of the current precision on the determination of orbital parameters, the InSAR estimated ground deformation will contain errors due to incomplete removal of the orbital phase. An extreme case of orbital phase error is shown on Figure 1. Because of the satellite configuration, the OPE is essentially a long wavelength signal that is a critical source of error when measuring deformation with a long wavelength component. Based on comparison of InSAR determined velocity with GPS determined velocity, the

![Interferogram for the big island of Hawaii. Left: high rate of fringes from errors in the knowledge of orbital parameters. Right: After flattening; deformation associated with volcanic centers](image)

Figure 1: Interferogram for the big island of Hawaii. Left: high rate of fringes from errors in the knowledge of orbital parameters. Right: After flattening; deformation associated with volcanic centers
current level of error attributed to incomplete removal of OPE is 0.1 mm/yr/km [Lanari et al., 2004a; Burgmann et al., 2006] but can be larger with SAR data from satellite with poor orbit determination (e.g RADARSAT).

1.1.3  **Current Orbital Correction methods and limitations**

Current methods vary depending of the type of orbital model used to describe the OPE, by the InSAR product from which the orbital model is determined, and by the use of information about the deformation field.

1.1.3.1  **Orbital model**

1.1.3.1.1  **Geometric**

The preferred current method to remove OPE is by fitting and removing a simple plane from the interferometric phase [Ferretti et al., 2001; Berardino et al., 2002; Peltzer et al., 2001a; Fialko, 2006]. Other geometric models include second order polynomial [Wright et al., 2004b] and “twisted” plane [Cavalie et al., 2007]. In general, the model complexity increases with the size of the observed area, the first group of studies applies to $< 10^4 \text{ km}^2$ study areas whereas the second group tend to measure larger region $> 10^4 \text{ km}^2$.

1.1.3.1.2  **Physical**

The Repeat Orbit Interferometry PACkage software (ROI_PAC, http://www.roipac.org/) from the Jet Propulsion Laboratory (JPL) applies a physical model to the orbital correction of interferograms, the model is function of orbital parameters, such as baseline separation between the satellite epochs,
distance between ground and satellite; and takes into account the ground topography.

1.1.3.2 **Interferometric InSAR Product**

OPE can be determined from three interferometric products: 1) a single interferogram, which represents the motion between two epochs, 2) a velocity map which represents the average displacement within a given time period, or 3) a displacement time series which represents the time evolution of the displacement within a given time period.

In the case of the single interferogram, one OPE is determined. It corresponds to the averaged OPE for the two SAR data. In the case of the velocity map, one OPE is determined, it corresponds to the averaged OPE for all the SAR data. In the case of the time-series with N SAR epochs, N-1 OPE are determined; they correspond to the average of each epochs’ OPE with the OPE of the master epoch.

Methods have been established to solve the OPE for each of the three cases, in particular, Kohlhase et al. [2003] and Biggs et al. [2007] established the basis for solving for OPE in the time-series case.

1.1.3.3 **Deformation Field**

Because of the correlation between the phase signal from deformation and the phase signal from orbital uncertainties, $\phi_{def}$ and $\phi_{orb}$ are simultaneously inverted for. Peltzer et al., [2001a] calibrate InSAR data across the Eastern California Shear Zone using a model of long-term plate motion. Fialko [2006]
uses more than 50 GPS velocities to remove a linear ramp from a stack of interferograms for the southern San Andreas Fault. Burgmann [2006], working in the San Francisco Bay Area, use a GPS-constrained tectonic model to remove a ramp from the InSAR velocities obtained using the PS technique.

If a priori information about the displacement field is not available, long-wavelength deformation can be retrieved by simultaneous inversion of a model of deformation and a model of OPE [Wright et al., 2004b; Biggs et al., 2007].

1.2 Low Strain Rate in the Eastern California Shear Zone and Central Nevada Seismic Belt

1.2.1 General statements

The Central Nevada Seismic Belt (CNSB) and the Eastern California Shear Zone (ECSZ) are two finite regions of intense tectonic activity in the transition between the western Basin and Range and the Sierra Nevada block [Wallace, 1984] (Figure 2).

These two regions are part of a larger set of deforming regions that form the wider plate boundary through which the relative displacement of the Pacific (PA) and the North America (NA) plates is accommodated.

An estimated 20 to 25 % of the overall PA/NA relative motion is accommodated by deformation across the ECSZ [Bennett et al., 1997], and about 7% by deformation across the CNSB [Thatcher et al., 1999; Wernicke et al., 2000; Hammond and Thatcher, 2004; Hammond and Thatcher, 2005].
Figure 2: GPS velocity field of Western North America east of the San Andreas system. Velocities represent motion of the stable Basin and Range (3mm/yr), Oregon block (7mm/yr) and Sierra Nevada Block (13mm/yr) with respect to stable North America. The Walker Lane (WL), Central Nevada Seismic Belt (CNSB) and Eastern California Shear Zone (ECSZ) accommodate 10 mm/yr of the Pacific North America relative plate motion.

As a consequence, the ECSZ and the CNSB are the location of intense and complex faulting.

Within the ECSZ and CSNB, 8 faults system have been identified: the Owens Valley – White Wolf – Airport Lake (OWA) fault-system, the Saline Valley – Hunter Mountain – Panamint Valley (SHP) fault-system and the Fish Lake Valley - Death Valley – Furnace Creek (FDF) in the ECSZ, and the Pleasant Valley, Dixie Valley, Fairview Peak and Rainbow Mountain-Stillwater faults in the CNSB, and the Cedar Mountain fault lying between the ECSZ and CNSB (Figure 3).
Over an 82 years time period, between 1872 and 1954, 6 of the 8 fault zones have ruptured, in a series of 11 earthquakes of M>6.5. The ruptures affect the entire length of both the ECSZ and the CNSB, 400 km of faulting, and the cumulative magnitude of the earthquakes amounts to M8.1. There are two gaps in the seismicity, the Stillwater seismic gap and the White Mountain Seismic gap.

The recurrence interval of the individual faults of the ECSZ ranges from 375 to 5000 yrs [Zhang et al., 1990; Brogan et al., 1991; Klinger and Sarna-Wojcicki, 2001; Lee et al., 2001; Oswald and Wesnousky, 2002]. It is significantly larger for the CNSB with values ranging from 3000 to 50 000 yrs [Wallace, 1984; Caskey et al., 2004; Bell et al., 2004]. The Cedar Mountain Fault, which lies in

Figure 3: Surface faulting during the 1915 Pleasant Valley (top), 1954 Dixie Valley (bottom left) and 1954 Fairview Peak (bottom right) earthquakes.
the transition zone between the ECSZ and the CNSB as a recurrence interval of 3600 yrs [Bell et al., 1999].

The contemporary surface deformation of the Basin and Range – Sierra Nevada transition region has been measured by Very Long Baseline Interferometry (VLBI), Slant Laser Ranging (SLR) studies [Argus and Gordon, 1991; Dixon et al., 1995], trilateration [Savage and Lisowski, 1995], and more recently and in more details by GPS studies [Bennett et al., 1997; Thatcher et al., 1999; Wernicke et al., 2000; Hammond and Thatcher, 2004; Hammond and Thatcher, 2005; Dixon et al., 2000; McClusky et al., 2001; Dixon et al., 2003; Hammond and Thatcher, 2007].

1.2.2 Central Nevada Seismic Belt deformation

1.2.2.1 Long-term fault motion

The averaged Holocene slip rate, integrated over the parallel strand of the CNSB, is estimated to be between 0.5 and 1.3 mm/yr from paleoseismic and exploratory trenching of and around the historical ruptures of 1915-1932-1954 [Bell et al., 2004]. The ratio of strike-slip to extension is ~2-3 and has been estimated from geodetic analysis of the 1915-1932-1954 earthquake sequence [Hodgkinson et al., 1996].
1.2.2.2 Present-day fault motion

The present day surface deformation across the CNSB is known from GPS measurements [Thatcher et al., 1999; Wernicke et al., 2000; Hammond and Thatcher, 2004; Hammond and Thatcher, 2007; Svarc et al., 2002] (Figure 4). The CNSB is the western boundary of the Basin and Range and is in close proximity with the Walker Lane (WL), the next deforming region to the West. At the CNSB location, the surface deformation is characterized by highly-localized strain, which contrast with the relatively slow deforming Basin and Range. Svarc et al., [2002] model the surface deformation with a combination of right lateral simple shear and extension normal to the strike of the range bounding faults. Hammond and Thatcher [2004] model the surface deformation with a combination of $2.3 \pm 0.3$ mm/yr right lateral and $3.0 \pm 0.4$ mm/yr extensional so a total rate of $3.8 \pm 0.5$ mm/yr. The strike-slip to extension ratio is 0.7 to 1.

Figure 4: East-west velocity profile across the Basin and Range province [Hammond and Thatcher, 2004]. Note velocity change across the Central Nevada Seismic Belt (CNSB) and velocity change between West Longitude 117 and 113.
The present day deformation across the CNSB is puzzling for two reasons: First, the deformation rate during Holocene time is lower than the 2 to 4 mm/yr obtained from the results of GPS measurements (5-7); Second, the present day ratio of strike-slip versus extension is two times higher than the ratio over one entire seismic cycle. Also, GPS measurements reveal a zone of east-west contraction east of the CNSB [Wernicke et al., 2000; Hammond and Thatcher, 2004] that is difficult to reconcile with current geodynamic models of the region, which involve east-west extension and right-lateral shear.

One possible explanation for these two discrepancies is that the GPS data do not record just the long-term deformation but that they also include transient deformation associated with viscous or visco-elastic relaxation of the lower crust or upper mantle following the last century’s earthquakes [Svarc et al., 2002; Hetland and Hager, 2003].

1.2.3 Eastern California Shear Zone deformation

1.2.3.1 Overview

GPS velocity measurements are used to study the present day deformation across the ECSZ (Figure 5). The total present day velocity change across the ECSZ is estimated at $11.1 \pm 0.3 \text{ mm/yr}$ [McClusky et al., 2001], 12 mm/yr [Meade and Hager, 2005], 12.8 mm/yr [Dixon et al., 2003] or 13 mm/yr [Miller et al., 2001].

The total ECSZ velocity change is invariably modeled by slip rate along three main fault systems across the ECSZ as followed: to the west, the Owens Valley –
White Wolf – Airport Lake (OWA) fault-system slip rate is estimated at 0-7 mm/yr [Bennett et al., 1997], 2.1 ± 0.3 mm/yr [Dixon et al., 2003], 3.5-7.3 mm/yr [Meade and Hager, 2005], 5.3 ± 0.7 mm/yr [McClusky et al., 2001], or 6-7 mm/yr [Savage and Lisowski, 1995]; in the center the Saline Valley – Hunter Mountain – Panamint Valley (SHP) fault-system slip rate is estimated at 2.5 ± 0.8 mm/yr [McClusky et al., 2001], or 4.2 ± 1.3 mm/yr [Meade and Hager, 2005]; to the east, the Fish Lake Valley - Death Valley – Furnace Creek (FDF) fault-system slip rate is estimated at 8.3 ± 1.2 mm/yr [Dixon et al., 2003], 2.8 ± 0.5 mm/yr [McClusky et al., 2001], between 3 and 5mm/yr) [Bennett et al., 1997], 5 mm/yr [Miller et al., 2001], 2.4 ± 1.2 mm/yr [Meade and Hager, 2005].
Despite a velocity field from dense GPS networks, the OWA and FDF fault slip estimations differ beyond error estimate between studies; the SHP fault slip estimations are more consistent, although less numerous and although the complete fault system is poorly covered by geodetic measurements. All fault systems are modeled as vertical strike slip fault.

1.2.3.2 The Saline Valley - Hunter Mountain – Panamint Valley Fault System

The present day slip rate of the SHP is poorly known, in part because of the little coverage of geodetic survey across the fault system and because of its possibly complex, and therefore, difficult to model, geometry.
1.2.3.2.1 Geometry

The geometry of the SHP fault system is matter of debate. Field reconstruction and geophysical exploration suggests that the SHP is a low angle normal fault system composed by the Saline Valley fault dipping east, the Panamint Valley dipping west, and the Hunter Mountain (HM) fault a strike slip transfer fault between the two low angle normal faults [Oswald and Wesnousky, 2002; Biehler, 1987] Figure 6. Modeling of GPS data suggest that the ground motion in the region of the SHP is parallel to the HM fault suggesting that the system is active [Bennett et al., 1997; Miller et al., 2001].

1.2.3.2.2 Fault Slip Rate

The average slip rate during the lifetime of the SHP fault system is obtained by extrapolation of the long-term slip rate of the HM fault from its total offset and inception time. The total offset on the HM fault is measured by the offset of the Hunter Mountain batholith offset by the HM fault, and by reconstruction of the Panamint and Saline valleys opening, the total offset is estimated at $9.3 \pm 1.4 \ km$ [Sternloff, 1988; Burchfiel et al., 1987]. The inception time of the HM fault is bracketed between 4 Ma [Burchfiel et al., 1987; Hodges et al., 1989] and 2.8 Ma [Lee et al., 2009], which corresponds to a long term average slip rate on the HM of 2.3 and 3.3 mm/yr respectively.

The Holocene slip rate of the SHP fault system has been measured along both the Panamint Valley and the HM faults. Zhang et al., [1990] measure a rate of $2.36 \pm 0.79 \ mm/yr$ from offsets of alluvial fans along the Panamint Valley fault.
Oswald and Wesnouski [2002] measure a rate of 3.3 to 4.0 mm/yr from offset of Holocene deposits along the HM fault.

1.3 Discussion

1.3.1 Slip rates
The CNSB and ECSZ fault slip present a discrepancy between the long term and the present day value. In the case of the CNSB the present day slip rate is well determined and it has been suggested that post-seismic deformation is playing a significant role on the present day deformation [Hetland and Hager, 2003]. In the case of the ECSZ, the present day slip rates are highly variable and therefore prevent any sensible comparison with long-term values.

1.3.2 Geometry of the SHP fault system
The proposed low angle geometry of the SHP fault system is contested on the basis that low angle normal faulting is contrary to the principle of rock mechanics and to observation that seismologically active faults had a dip between 30 and 60° [Jackson and White, 1989]. The suggested explanation for the widely observed low angle contacts, for example in the Basin and Range [Wernicke, 1995], is that low angle structures are derived from pre-existing “normally” dipping active fault, by rotation caused by isostatic readjustment. This view is challenged by an alternative to the Andersonian theory of rock fractures [Melosh, 1990], recent studies of fault properties [Numelin et al., 2007], and observation of seismicity along low angle faults [Floyd et al., 2001]. Despite the possibly complex geometry of the SHP fault system, the surface deformation field across
the entire ECSZ has been interpreted with a fault model in which the SHP was
a single or set of strike slip faults. This of course has in turn implications on the
fault slip rate determined.

A proper determination of the SHP geometry from the currently available
information about surface deformation is difficult because of the close proximity
of other major faults and is complicated further by the fact that the surface
defformation associated with such a shallow dipping faults is likely to produce a
broad surface displacement field.

In an active low angle normal SHP fault system, the HM fault would be a shallow
locked vertical right lateral strike slip system. The associated surface deformation
will in that case have a short wavelength that will be distinguishable from the
broader type deformation associated with the nearby low angle normal fault and
the OWA and FDF large strike slip systems.

1.3.3 How rigid is the Basin and Range?
Present day background seismicity and surface deformation suggests that the
Basin and Range is rigid and behaves as a block. However, faults are widely
present through the entire Basin and Range, and recent geodetic surveys show
small but significant strain accumulation within the area, and a recent M6
earthquake in Wells, Nevada, suggests that the Basin and Range is deforming.

1.3.4 Measuring Low Strain Rate with Synthetic Aperture Radar
Points 1.3.1-3 can be addressed by knowing better the surface velocity of the
region bounded by the ECSZ and CNSB, this demand an improvement of the
InSAR methods to measure surface deformation. The limitations of the previous methods are one or more of the following. (1) They omit long-wavelength deformation, (2) they make assumptions about the spatial variation of the deformation, (3) they make assumptions about the time-dependency of the deformation (linear deformation), and (4) they use approximate orbital models. The goal of the present work is to provide a method that will combine those approaches, specifically, using time series of GPS to take into account deformation at each SAR epoch.
CHAPTER 2  InSAR - GPS Integration for the Study of Large-Scale Processes

2.1 Background

Interferometric Synthetic Aperture Radar (InSAR) has been used successfully to measure and study surface deformation such as glacier movements [Goldstein et al., 1993], earthquakes [Massonnet et al., 1994] and volcano inflation [Amelung et al., 2000]. The measurement of subtle, long wavelength deformation (>50 km), such as inter-seismic and post seismic deformation [Massonnet, 1997; Pollitz et al., 2000], remains a challenge. The precision of the InSAR measurement is affected by surface decorrelation, unmodeled phase contributions due to atmospheric water vapor and ionospheric effects, and uncertainties in the position of the satellites. The uncertainties in the position of the satellites (orbital errors) degrade the precision of the measurements from millimeters (the instrumental precision) to centimeters or more. Accurate orbits, such as the ones provided by the DELFT Institute for Earth-oriented Space Research (DEOS), have an accuracy of the order of 15 centimeters [Scharroo, 2002].

Since the end of the 1990’s, techniques for the simultaneous analysis of large numbers of SAR acquisitions (there are more than 150 acquisitions for most of Europe) have led to time-dependent measurements [Ferretti et al., 2001; Hooper...
et al., 2004; Berardino et al., 2002; Lanari et al., 2004b; Casu et al., 2008].

These time-series techniques have significantly improved measurement precision. First, stable targets can be extracted using statistical analysis on a large dataset [Ferretti et al., 2005]. Second, the atmospheric contributions, or atmospheric phase screen (APS), can be extracted and removed by applying spatial-temporal filtering on the time series [Ferretti et al., 2001; Berardino et al., 2002].

Large-scale tectonic processes, such as inter-seismic deformation across entire fault zones and post-seismic deformation following large earthquakes, can be studied using multiple, consecutive radar frames [Peltzer et al., 2001a; Fialko, 2006; Wright et al., 2004b; Biggs et al., 2007]. In this case, the measurement precision is limited mainly by long-wavelength phase contributions related to orbital errors (we refer to them as orbital phase errors), Burgman et al., [2006] estimate a gradient of velocity error of 0.094 mm/yr/km. Subtle, long-wavelength deformation can be resolved using a priori information about the surface displacement field. For example, Peltzer et al. [2001a] calibrate InSAR data across the Eastern California Shear Zone using a model of long-term plate motion. Fialko [2006] uses more than 50 GPS velocities to remove a linear ramp from a stack of interferograms for the southern San Andreas Fault. Burgmann et al. [2006], working in the San Francisco Bay Area, use a GPS-constrained tectonic model to remove a ramp from the InSAR velocities obtained using persistent scatterer methods.
If *a priori* information about the displacement field is not available, long-wavelength deformation can be retrieved by simultaneous inversion for models for the tectonic deformation and for the orbital phase errors [Wright *et al.*, 2004b; Biggs *et al.*, 2007].

In most studies, the orbital phase errors are approximated by a first or second order two-dimensional polynomial. An exception is the ROI_PAC software package from the Jet Propulsion Laboratory, which uses a physical orbital model and topographic information to estimate the orbital phase errors. Kohlhase *et al.* [2003] remove the orbital phase errors using a semi-physical orbital model and a network of SAR data.

Kohlhase *et al.* [2003], and Biggs *et al.* [2007] use a network approach to resolve the orbital phase errors. This approach potentially retains the temporal resolution of the InSAR data although Biggs *et al.* assume a linear deformation model.

The limitations of the described methods are one or more of the following. (1) they omit long-wavelength deformation; (2) they make assumptions about the spatial variation of the deformation; (3) they make assumptions about the time-dependency of the deformation (linear deformation); and (4) they use approximate orbital models.

In this paper we present a new method for the measurement of subtle, long-wavelength deformation. The method identifies and removes orbital phase errors from the InSAR data using GPS and a physical orbital model, without making assumptions about the deformation field in space and time. The method can
retrieve the time history of ground displacement. We apply this method to measure the inter-seismic strain accumulation across the Eastern California Shear Zone, with particular emphasis on the Hunter Mountain fault zone.
2.2 The Eastern California Shear Zone

The Eastern California Shear Zone (ECSZ), parallel to the San Andreas fault system, accommodates 20 to 25% of the total Pacific-North America plate motion [Dokka and Travis, 1990b; Dokka and Travis, 1990a; Dixon et al., 2000] (Figure 7). During the past 150 years, 4 major earthquakes have hit the region: the 1872 M8 Owens Valley, the 1932 M7.1 Cedar Mountain, the 1992 M7.3 Landers, and the 1999 M7.1 Hector Mine Earthquakes. The Hunter Mountain fault is part of the Panamint Valley - Hunter Mountain – Saline Valley (PHS) system, one of the younger faults comprising the ECSZ [Lee et al., 2009].

2.3 Results from GPS Integration of InSAR time series with CGPS – Theory

2.3.1 Differential Interferometry

Differential SAR Interferometry (InSAR) is a remote sensing technique that measures ground displacement by exploiting the measured phase difference \( \phi(t_2, t_1) \) (usually referred to as interferogram) between two SAR images acquired at epochs \( t_2 \) and \( t_1 \) [Gabriel et al., 1989; Massonnet and Feigl, 1998; Rosen et al., 2000]. \( \phi(t_2, t_1) \) depends on the phase contributions from the ground deformation between \( t_2 \) and \( t_1 \), \( \phi_{\text{def}}(t_2, t_1) \), on the difference in atmospheric delays at the time of the acquisitions \( \phi_{\text{atmo}}(t_2, t_1) \), and on the phase contribution from noise sources, \( \phi_{\text{noise}}(t_2, t_1) \):

\[
\phi(t_2, t_1) = \phi_{\text{def}}(t_2, t_1) + \phi_{\text{atmo}}(t_2, t_1) + \phi_{\text{noise}}(t_2, t_1)
\] (1)
The noise sources are uncertainties on the satellite orbit (orbital errors, discussed below), thermal noise, changes of the ground dielectric properties such as due to soil moisture or snow, surface decorrelation, and processing noise related to interpolation.

### 2.3.2 InSAR time series

To obtain the temporal evolution of ground deformation, we use many SAR acquisition of the same area and the Small BAseline Subset (SBAS) method [Berardino et al., 2002; Lanari et al., 2007]. The key idea is to select interferometric pairs with small spatial and temporal separation so as to minimize decorrelation, thus maximizing the number of temporally coherent pixels [Pepe and Lanari, 2006]. The baseline thresholds (the maximum spatial baseline and the maximum time span between acquisitions) depend on the type of environment. Sparsely vegetated environments with little topography, allow for larger thresholds than heavily vegetated environments with significant topography.

In the SBAS algorithm, a set of Q unwrapped interferograms is inverted for the phase at epoch $t$, with respect to the first acquisition, $\phi(t)$,

$$\phi(t) = \phi_{\text{def}}(t) + \phi_{\text{atmo}}(t) + \phi_{\text{noise}}(t).$$

(2)

with $t = 2, \ldots, N$. The phase due to deformation, $\phi_{\text{def}}(t)$, and due to noise, $\phi_{\text{noise}}(t)$, refer to the first epoch, the atmospheric phase, $\phi_{\text{atmo}}(t)$, is the difference in atmospheric delays between epochs $t$ and 1. The objective of crustal deformation
studies is to recover \( \phi_{\text{def}}(t) \) from the measured \( \phi(t) \). This requires the estimation of \( \phi_{\text{atmo}}(t) \) and \( \phi_{\text{noise}}(t) \).

2.3.3 Temporal coherence

We proceed to select pixels using the temporal coherence factor defined as:

\[
\gamma = \frac{\sum_{j=1}^{Q} \exp[j(\phi_p(t_2,t_1) - \overline{\phi}_p(t_2,t_1))]}{Q}, \quad 0 \leq \gamma \leq 1
\]

(3)

where \( \phi_p(t_2,t_1) \) is the phase of a pixel \( p \) of the original interferogram, and \( \overline{\phi}_p(t_2,t_1) \) is the phase of a pixel \( p \) of the synthetic between epoch \( t_2 \) and \( t_1 \) (a synthetic interferogram is formed by differencing the phase from the time series according to the epochs of the interferogram), and \( Q \) is the number of interferograms used for the time series. Low temporal coherence arises from inconsistencies of the phase between interferograms sharing common SAR acquisitions. Considering \( \phi(t_2,t_1) \), \( \phi(t_3,t_1) \), and \( \phi(t_3,t_2) \), the phase of three interferograms from processing SAR data from epochs \( t_1 \), \( t_2 \), and \( t_3 \); the temporal coherence equal 1 if \( \phi(t_2,t_1) + \phi(t_3,t_2) - \phi(t_3,t_1) = 0 \). The main causes of low temporal coherence are temporal and geometric decorrelation and error during the interferograms unwrapping. For pixels where \( \gamma \rightarrow 1 \), we expect no unwrapping errors, since a nearly perfect retrieval of the original phase has been obtained, low values of \( \gamma \) will correspond to poorly unwrapped data.
2.3.4 Atmospheric filtering

The atmospheric phase, $\phi_{\text{atmo}}(t)$, is the difference between the atmospheric phase screen at time $t$, $APS(t)$, and the atmospheric phase screen at the time of the first acquisition, $APS_1$.

The atmospheric phase screen is estimated using a spatial-temporal filter [Ferretti et al., 2001; Berardino et al., 2002]. We assume that atmospheric phase contribution are spatially correlated and temporally uncorrelated and estimate them by applying a low-pass spatial filter followed by a high-pass temporal filter to $\phi(t)$. The estimated atmospheric phase screen, $\hat{APS}(t)$, relates to the true atmospheric phase screen, $APS(t)$, as

$$APS(t) = \hat{APS}(t) + APS^\varepsilon(t)$$

with, $APS^\varepsilon(t)$, the atmospheric phase screen error. We assume that the filtering reliably estimates the atmospheric phase screen, i.e. that $APS^\varepsilon(t)$ is negligible

$$APS(t) \approx \hat{APS}(t).$$

The atmospheric phase screen for the first acquisition is given by

$$\hat{APS}_1 = \frac{1}{N - 1} \sum_{i=1}^{N} \hat{APS}(t).$$

Subtraction of $\phi_{\text{atmo}}(t)$ from $\phi(t)$ leads to the filtered phase with respect to the first acquisition $\phi_{\text{filt}}(t)$:
\[ \phi_{fil}(t) = \phi(t) + \phi_{atmos}(t). \] (7)

Substituting into (2) gives:

\[ \phi_{fil}(t) = \phi_{def}(t) + \phi_{noise}(t). \] (8)

### 2.3.5 Orbital phase error (OPE)

The initial orbits used for InSAR processing of ERS-1 and ERS-2 satellites, deviate from the true orbits by about 15 centimeters on average [Scharroo, 2002]. The initial orbit, at epoch \( t, \ o^0(t) \), relates to the true orbit, \( o(t) \) as:

\[ o(t) = o^0(t) + o^\epsilon(t) \] (9)

with \( o^\epsilon(t) \) the orbital error we are seeking to estimate. In practice, the spatial separation between the satellites during image acquisition (baseline) is used. For an interferogram between SAR images at epochs \( t_1 \) and \( t_2 \), the baseline \( b(t_2, t_1) \) is given by \( b(t_2, t_1) = o(t_2) - o(t_1) \). In the time series formulation the baseline at epoch \( t \), \( b(t) \), is the difference between the orbit at epoch \( t \) and the orbit of the first acquisition \( o_1 \):

\[ b(t) = o(t) - o_1, \] (10)

with \( t=2, \ldots, N \) and \( o_1 = o^0 + o^\epsilon \). The initial baseline at epoch \( t \), \( b^0(t) \), relates to the true baseline \( b(t) \), as
\[ b(t) = b^o(t) + b^e(t) \]  \hspace{1cm} (11)

with \( b^e(t) \) the baseline error. The baseline error relates to the orbital error as

\[ b^e(t) = o^e(t) - o^i. \]  \hspace{1cm} (12)

The orbital phase \( \phi_{\text{orb}}(t) \) at epoch \( t \) is linearly related to the baseline [Hanssen, 2001], eq. 2.4.18:

\[ \phi_{\text{orb}}(t) = b_h(t) \cdot F_h - b_v(t) \cdot F_v, \]  \hspace{1cm} (13)

with \( b_h \) and \( b_v \) the horizontal and vertical baseline components, and \( F_{h,v} \) two factors as

\[ F_h = \left( 4 \cdot \pi \lambda \right) \cdot \left( \sin \theta - \cos \theta \cdot \frac{H}{R} \right) \]  \hspace{1cm} (14)

\[ F_v = \left( 4 \cdot \pi \lambda \right) \cdot \left( -\cos \theta - \frac{H}{R} \right) \]  \hspace{1cm} (15)

\( H \) is the terrain height above the ellipsoid, \( R \) the range, and \( \theta \) the radar look angle.

The initial orbital phase based on the initial orbits, \( \phi_{\text{orb}}^o(t) \), relates to the true orbital phase \( \phi_{\text{orb}}(t) \), as

\[ \phi_{\text{orb}}(t) = \phi_{\text{orb}}^o(t) + \phi_{\text{orb}}^e(t). \]  \hspace{1cm} (16)
with $\phi_{\text{orb}}^e(t)$ the orbital phase error. We thus can express the orbital phase error in terms of the horizontal and vertical baseline errors, $b_h^e(t)$ and $b_v^e(t)$, as:

$$\phi_{\text{orb}}^e(t) = b_h^e(t) \cdot F_h - b_v^e(t) \cdot F_v.$$  \hfill (17)

The task is to estimate at each epoch the baseline error components and from this the orbital phase corrections.

2.3.5.1 **OPE estimation in the presence of deformation**

We now assume that the phase noise at epoch $t$, $\phi_{\text{noise}}(t)$, is only due to orbital errors, $\phi_{\text{noise}}(t) = \phi_{\text{orb}}^e(t)$, and obtain from (8):

$$\phi_{\text{filt}}(t) = \phi_{\text{def}}(t) + \phi_{\text{orb}}^e(t).$$  \hfill (18)

At each epoch $t$ we estimate $b_h'(t)$ and $b_v'(t)$ by minimizing:

$$\min \left\{ \phi_{\text{filt}}(t) - \phi_{\text{def}}(t) - \phi_{\text{orb}}^e(t) \right\}.  \hfill (19)$$

In the absence of deformation $\phi_{\text{def}}(t) = 0$, we estimate $b_h'(t)$ and $b_v'(t)$ from $\phi_{\text{filt}}(t)$ (i.e. from InSAR only). In the presence of deformation we estimate $b_h'(t)$ and $b_v'(t)$ using a-priori information on $\phi_{\text{def}}(t)$ as described below. For comparison, in the classical SBAS approach (Beradino et al., 2002) it is assumed that $\phi_{\text{def}}(t) = 0$, and that $\phi_{\text{orb}}^e(t)$ is a simple linear function which is estimated and then removed from $\phi_{\text{filt}}(t)$.
Continuous GPS measurements provide displacement at epoch $t$, at radar range location $r$ and radar azimuth location $a$, $\phi_{\text{def}}(t,r,a)$. Because of the pointwise nature of CGPS measurements and the number of available CGPS stations in the InSAR footpath, we approximate the along track variation of $b_h^x$ and $b_v^x$ by second order polynomials:

$$
\begin{align*}
    b_h^x(t) &= c_1(t)a^2 + c_2(t)a + c_3(t) \\
    b_v^x(t) &= c_4(t)a^2 + c_5(t)a + c_6(t)
\end{align*}
$$

with $a$ the azimuth coordinate. Equation (17) takes the form

$$
\phi_{\text{orb}}^x(t) = \left(c_1(t)a^2 + c_2(t)a + c_3(t)\right) \cdot F_h - \left(c_4(t)a^2 + c_5(t)a + c_6(t)\right) \cdot F_v
$$

At each epoch $\phi_{\text{orb}}^x$ is thus described by six parameters, $c_{1,\ldots,6}$.

At each epoch $t$ we estimate $b_h^x(t)$ and $b_v^x(t)$, by minimizing:

$$
\min \left\{ \sum_{m=1}^{M} \left( \phi_{\text{filt}}^x(t,r_m,a_m) - \phi_{\text{def}}^x(t,r_m,a_m) - \phi_{\text{orb}}^x(t,r_m,a_m) \right)^2 \right\}
$$

If deformation is known at six locations ($m=6$), we solve a system of six equations with the six unknowns $c_{1,\ldots,6}$ at each epoch $t$. Note that the estimation of $\phi_{\text{orb}}^x$ and $b_h^x(t)$ is associated with errors,

$$
\phi_{\text{orb}}^x(t) = \hat{\phi}_{\text{orb}}^x(t) + \epsilon_{\text{orb}}^x(t),
$$
\[ b_{h,v}^e(t) = \hat{b}_{h,v}^e(t) + b_{h,v}^e(t) \]  

with \( \hat{\phi}_{\text{orb}}(t) \) and \( \hat{b}_{h,v}^e(t) \) the estimated orbital phase error and baseline component errors at epoch \( t \), and \( \phi_{\text{orb}}^e(t) \) and \( b_{h,v}^e(t) \) the respective estimation errors.

### 2.4 Integration of InSAR time series and CGPS – Application to the ECSZ

#### 2.4.1 Large scale SBAS processing

In the past, the SBAS technique has mostly been applied to investigate deformation of areas typically extending \(~100 \text{ by 100 km}^2\). In this study we use the SBAS technique to study a larger area \(~600 \text{ by 100 km}^2\).
We analyze a set of 44 ERS-1/2 SAR swaths (track 442, frames 2781 to 2871), spanning the 1992-2000 time interval. To reduce the amount of data to be

Figure 9: Temporal coherence from the SBAS analysis. a) Full range. b) Coherence mask using a threshold of 0.7.
processed, the image resolution is degraded to a pixel size of 160 by 160 m$^2$, compared to 80 x 80 m$^2$ for conventional SBAS processing [Casu et al., 2008].

The interferometric pairs are selected using a maximum spatial and temporal baseline of 400 m and 1,500 days, respectively (Figure 8); by applying these constraints, 148 multilook interferograms are generated. The interferograms are phase-unwrapped and subsequently inverted for the phase with respect to the first acquisition, $\phi(t)$.

2.4.2 Temporal coherence

The temporal coherence estimated using eq. (3) is shown in Figure 9a. Most of the low topography area, including the basins, exhibit high coherence (larger than 0.7). The mountain ranges are characterized by low coherence (near 0), including the Sierra Nevada in the southwest. The loss of coherence occurs because the surface characteristics change with time. In the mountains coherence is lost because of temporary snow cover. In some valleys coherence may be lost because of flooding. Another reason for coherence loss in the mountains is geometric decorrelation related to the steep slopes. In this study, we use only pixel with temporal coherence larger than 0.7. The temporal coherence mask is shown in Figure 9b.
2.4.3 Atmospheric filtering

In order to quantify the efficiency of the atmospheric filtering, we perform a 1D covariance analysis [Hanssen, 2001] on the phase before and after filtering (\( \phi(t) \) and \( \phi_{filt}(t) \), respectively). At each epoch, we conduct an auto-correlation and model the resulting amplitude by a two-parameter Bessel function following Biggs et al., [2007] (Figure 10a,b). The figure shows that the spatial-temporal filtering reduces the correlation amplitude and length in average by 30%.
The atmospheric filtering is illustrated in map view in Figure 11. The figure shows the phase at the 23 November 1992 epoch before and after filtering (Figure 11a and Figure 11d, respectively) as well as the estimated APS for the first acquisition (1 June 1992, Figure 11b) and for the 23 November 1992 acquisition (Figure 11c).
The phase at epoch 23 November 1992 corresponds to the interferogram between the 6 June 1992 and 23 November 1992 acquisitions because the first acquisition of the interferogram is also the reference epoch of the time-series. Figure 12 shows the estimated baseline error for the 11/23/1992 epoch with and without atmospheric filtering. The differences are very small, indicating that atmospheric filtering has only little effect on the estimated baseline error.

2.4.4 Baseline errors and the effect of ground deformation on their estimation

The baseline errors for five epochs, estimated using eq. (18) with the assumption of no deformation, $\phi_{def}(t) = 0$, are shown as function of azimuth in Figure 13.

Figure 12: Horizontal Baseline correction term for epoch 11/23/1992 before, red, and after, blue, APS removal.
The estimated baseline errors ($\hat{b}_h(t)$ and $\hat{b}_v(t)$) along most of the SAR swath is less than 15 m (Figure 13a,b). This is much larger than the orbital error estimates of Scharroo [2002], but similar to values obtained using the baseline re-estimation inversion strategy based on topographic information used in JPL's ROI_PAC software. At the beginning of the SAR swath (at 40-100 km along track distance, Figure 13c) and at the end of the swath (at 400-600 km along track distance) the estimated baseline error is up to 60 m and 30 m, respectively.

Figure 13: (a) Horizontal and (b) vertical baseline error, $\hat{b}_h(t)$, for 5 epochs. (c) Zoom into horizontal baseline error for the subsiding Crescent Valley area (see Figure 19) and epochs of SAR acquisitions.
(Figure 13c). This is caused by the omission of ground deformation. The effect of surface displacement on the estimated baseline error is clearly shown in Figure 14; in the Crescent Valley subsiding area there is a positive correlation between surface deformation and estimated baseline error. We conclude from this section that $\phi_{xy}(t)$ need to be taken into account for the estimation of the baseline errors.

![Figure 14: Estimated horizontal baseline error versus surface displacement for a pixel in the subsiding Crescent Valley area.](image)
2.4.5 OPE estimation using GPS

2.4.5.1 GPS data

We use continuous GPS (CGPS) as a measure of $\phi_{\text{def}}(t)$. We use 6 CGPS stations from the PBO core network (previously referred to as BARGEN network) [Wernicke et al., 2000; Davis et al., 2006; Bennett et al., 2003] located in the area imaged by the SAR (Figure 15). The stations are distributed evenly over the entire SAR track. Two stations are located on the stable Basin and Range block (MONI and TONO), two stations are located in the vicinity of the Central Nevada Seismic Belt (GABB, NEWS), and two stations are located in the vicinity of the ECSZ (ARGU, COSO and DYER). There are two additional GPS stations with data starting in the mid nineties (LEWI and COSO) but we could not use them for reasons discussed below. The CGPS positions are referred to a stable Basin and Range reference frame defined by 26 permanent GPS stations located within the stable Basin and Range block [Schmalzle, 2008]. We use the GPS records starting in 1999 when all 6 stations were operating simultaneously.

In the following we consider the GPS displacement component in radar line-of-sight (LOS) direction, obtained by multiplying the east, north, up GPS vector with the unit vector pointing from the ground to the radar ([0.3, -0.09, 0.9] in [east, north, up]).
For the comparison of the InSAR with the GPS we use the interpolated horizontal velocity field of the Basin and Range region from Kreemer et al. [2006]. This velocity field is based on campaign and continuous data with records of a few years to over fifteen years with large station density variations (the station density is higher near the active volcanic and seismic region of the Sierra Nevada - Basin and Range boundary). The GPS velocity field includes data of the high-density, semi-permanent MAGNET network operated by the University of Nevada, Reno. Kreemer et al. obtained this spatially continuous velocity field (on a $0.2^\circ$ by $0.2^\circ$ grid) from the GPS point measurements by interpolating the

2.4.5.2 **GPS velocity field**

Figure 15: GPS based (red and blue) and InSAR (black) velocity cross-section (See location Figure 19). a – InSAR velocity with no OPE removal; b – InSAR velocity with OPE removal; c – InSAR velocity with CGPS-OPE removal. LOS velocity of the 6 CGPS stations used in the inversion (blue dots). Note that at distance 0, the InSAR, GPS model, and CGPS profiles has been adjusted to 0.
GPS velocities in a least square sense using a bi-cubic Bessel spline function. For the analysis below this velocity field is transformed into LOS direction assuming zero vertical deformation.

2.4.5.3 **Comparison between InSAR and interpolated GPS velocity field**

To compare the InSAR with the continuous GPS and the interpolated GPS velocity field, we project all the data (LOS component for the GPS) along a profile perpendicular to the ECSZ (in N60W direction) (Figure 15). The InSAR data are shown without OPE removal (Figure 15a), with OPE removal assuming zero deformation (Figure 15b), and with OPE removal accounting for deformation using the continuous GPS data (Figure 15c).

Without OPE removal there is a general disagreement between the InSAR and GPS (Figure 15a). At 400 to 600 km along the profile the difference is on the order of 4-8 mm/yr. The nature of the velocity difference is similar to Burgmann et al., [2006], although we observe a lower gradient, 0.03 mm/yr/km. With OPE removal assuming zero deformation the InSAR and GPS data are roughly consistent in the northeast (0 to 400 km along profile) but there is a discrepancy in the southeast of about 2-3 mm/yr (Figure 15b). Note that scatter is less when the OPE is removed (Figure 15a compared with Figure 15b). The case of OPE removal accounting for deformation (Figure 15c) is discussed below.
Figure 16: LOS displacement time-series from InSAR SBAS analysis and daily positions for 6 collocated GPS sites in the western Basin and Range. a - Conventional SBAS analysis after the removal of linear and second-order phase contributions (before calibrations with GPS). b - SBAS analysis after removal of a linear phase ramp determined from continuous GPS measurements (after calibration with GPS). We extrapolated the 1999-2005 GPS measurements back to the beginning of SAR measurements in 1992. The InSAR errors are estimated at ± 0.6 mm/yr [Gourmelen et al., 2007]. GPS errors of ± 0.7 are estimated assuming white and flicker noise [Dixon et al., 2000].
2.4.5.4 *Comparison between InSAR and continuous GPS*

The CGPS stations are located within the mountain ranges, where InSAR data are not available because of decorrelation. We therefore average all coherent InSAR pixel within two km radius from each CGPS stations. As a result we have to eliminate the LEWI and COSO GPS stations from our analysis. Both stations are affected by local deformation and therefore the ground within two km from the GPS benchmark may deform differently than the GPS benchmark itself. LEWI is affected by local subsidence due to water withdrawal in support of mining activities [Gourmelen et al., 2007]; COSO is affected by subsidence associated with the nearby Coso geothermal plant [Bennett et al., 2003]. We note that GPS stations affected by local deformation can well be used for the estimation of orbital phase errors as long as they collocate with coherent pixel.

The InSAR time series are shown together with the CGPS time-series in Figure 16a for the 6 GPS stations. The northernmost stations are plotted at the top and the southernmost stations at the bottom of the panel. The largest differences between InSAR and GPS are found in the south within the ECSZ. For example, ARGU’s GPS velocity is $-1.5 \pm 0.7$ mm/yr, whereas the corresponding InSAR velocity is $+1.1 \pm 0.6$ mm/yr, a difference of 2.6 mm/yr. In the north within the stable Basin and Range block MONI’s GPS velocity is $0.6 \pm 0.7$ mm/yr, whereas the corresponding InSAR velocity is $0.3 \pm 0.6$ mm/yr, a difference of 0.3 mm/yr. This pattern of southwestward increasing differences between InSAR and continuous GPS is similar to differences between the InSAR and the interpolated GPS velocity field from Kreemer et al. (Figure 15).
2.4.5.5 **OPE correction using continuous GPS**

The continuous GPS data are complete since 1999 but the InSAR data start in 1992. Therefore, we extrapolate the GPS time-series for the 1992-1999 period using the averaged 1999-2005 velocities. This approach yields predicted GPS positions at each SAR epoch.

We then assume a second order azimuth model and apply eq. (23) to estimate the OPE and remove it from the InSAR time series at each epoch. The

![Figure 17 Standard: deviation of the time-series in function of the orbital model applied. Quadratic variation of the baseline errors gives the most satisfying results. High residual standard deviation corresponds to regions of deformation (subsidence, earthquakes, ...).](image)
corrected InSAR time-series show a significant reduction of the noise level (Figure 16 right). More importantly, the difference between the InSAR and continuous GPS velocities is within the range of error.

That the estimation of the OPE using GPS works well is also clearly seen by comparing the InSAR with the interpolated GPS velocity field of Kreemer et al. After OPE removal accounting for deformation (Figure 16c), there is a broad agreement between the InSAR velocities and the interpolated GPS velocities.

2.4.5.6 **Azimuth model**

To evaluate the spatial model used to approximate the OPE, we compute the standard deviation of the phase history for each pixel. Figure 17 shows the standard deviation for the original time-series (Figure 17a), after removal of the OPE assuming a linear azimuth model (obtained by substituting eqs. (20) and (21) by a linear model, (Figure 17b), and after removal of the OPE using the quadratic models of eqs. (20) and (21) (Figure 17c). We also display the phase of the 921123 epoch as a function of the model used to remove the OPE (Figure 18). The quadratic model results in the smallest standard deviation and performs well in removing long wavelength phase residuals, suggesting that a quadratic model is most appropriate. However, the standard deviation criteria has to be taken with caution as strong regional periodical signal typically recorded by GPS will tend to increase the standard deviation. This is not the case here as we consider the GPS record in a local reference frame. Local deformation due to subsidence or earthquakes is associated with a high standard deviation in the threes cases as expected.
The average velocity for each coherent pixel, obtained from the filtered time series is shown in figure 19. The figure displays subsidence related to mining activities [Gourmelen et al., 2007], subsidence related to geothermal activity at the Coso geothermal field [Fialko and Simons, 2000], and deformation...
related to the 1993, M6.3 Eureka Valley Earthquake [Peltzer and Rosen, 1995].

The interseismic strain accumulation across the ECSZ discussed below is not
visible with the colorscale used for Figure 19. Tightening the colorscale between [-4 4] mm/yr reveals the pattern of deformation across the ECSZ (Figure 20). Little deformation is seen across the Fish Lake Valley - Furnace Creek fault system. Assuming pure strike-slip motion across the fault, and taking into account the orientation of the fault with respect to the radar look angle, we can determine an upper limit of 2 mm/yr of strain accumulation across the fault system. This upper limit can be higher however depending on the deformation's wavelength. The velocity map reveals little deformation across the White Mountain - Owens Valley fault system for the same reasons stated for the Fish Lake – Furnace Creek fault system. The upper limit here is 5 mm/yr because of the northerly orientation of the fault system. Instead, most of the deformation is located across the Hunter Mountain (HM) fault.

2.5 Modeling the strain accumulation across the Hunter Mountain fault

In order to only model the observed deformation across the Hunter Mountain fault, we remove the Eureka Valley co-seismic displacement field. For this we first divide the displacement time-series into a pre- and a post-earthquake time-series. After subtracting linear trends from both time-series, we compute the respective mean values and obtain a pre-earthquake mean position, and a post-earthquake mean position. We then subtract the pre-earthquake mean position from the post-earthquake mean position to obtain the co-seismic offset and subtract it from the displacement at each post-earthquake epoch. We then re-calculate the velocity map.
A profile perpendicular to the Hunter Mountain fault (Figure 21) reveals a LOS velocity change of $1.6 \pm 0.6$ mm/yr across the fault. The LOS velocity change
occurs progressively across a zone with a width of 5-12 km centered on the fault. The lack of a discontinuity suggests that there is no or little surface creep.

![Graph showing LOS velocity perpendicular to the Hunter Mountain fault.](image)

Figure 21: LOS velocity perpendicular to the Hunter Mountain fault. The area covered is shown in Figure 20. Fault and earth models after [Savage and Burford, 1973] in inset.

We now assume that the velocity change across the Hunter Mountain fault is the result interseismic strain accumulation along a pure right-lateral strike-slip fault. We use the classical elastic dislocation model of Savage and Burford [1973] in which the fault is driven from a freely slipping fault at depth embedded in an elastic halfspace. The model has two parameters: the far field velocity and the locking depth [Savage and Burford, 1973; Weertman and Weertman, 1964].
(Figure 12). We use a non-linear Gibbs Sampling inversion scheme to retrieve the two parameters and their probability density distributions [Johnson and Segall, 2004]. This simple model fits the data very well. The best fit is obtained for a slip rate of $4.9 \pm 0.9$ mm/yr, and a locking depth of $2 \pm 1$ km (Figure 22). Note that there is little correlation between the fault slip rate and the locking depth.

![Figure 22: Locking depth versus slip rate probability from Gibbs sampling.](image)

Additional modeling of the deformation across the Hunter Mountain fault suggests that neither surface creep nor visco-elastic relaxation is occurring. A model including shallow creep [Savage and Lisowski, 1993], predicts no displacement on the shallow creeping section. We also tested models including visco-elastic relaxation, with model parameters locking depth, fault slip rate, time since the last earthquake, recurrence interval and viscosity [Savage and Lisowski, 1998]. The best-fitting models are characterized by similar values for
the time since the last earthquake and the recurrence time, implying that the visco-elastic relaxation is completed (assuming a viscosity of $10^{19}$ Pa.s) \cite{Dixon2003, Thatcher2008}. In conclusion, the InSAR data are well modeled with a simple screw dislocation in an elastic medium, and do not require surface fault creep or visco-elastic rheology.

**Figure 23**: Model of present day tectonic for the Owens Valley – Panamint Valley – Hunter Mountain – Saline Valley faults modified from \cite{Wesnousky1994}. At the location of the InSAR profile, the depth of the Panamint fault is of the order of the locking depth for the Hunter Mountain fault.
2.6 Discussion

The InSAR data across the Eastern California Shear Zone reveal a narrow zone of deformation across the Hunter Mountain fault. The observations are well explained using the Savage and Burford [1973] elastic dislocation model with a vertical strike slip fault with a 2 km locking depth and a slip rate of 4.9 mm/yr.

The slip rate of the Hunter Mountain fault estimated from the geodetic data is higher than the geologic rates bracketed between 2.4 and 4 mm/yr [Gourmelen, 2009]. The geologic rates have been inferred from the total fault offset since fault initiation. One explanation for this difference in rate is that the fault has been accelerating. Gourmelen et al., [Gourmelen et al., 2008] proposed a model of fault evolution in which the slip rate increased as the fault matured.

Our locking depth estimation for the Hunter Mountain fault is significantly lower than locking depth estimates of nearby faults which generally range between 5 and 15 km [Bennett et al., 1997; McClusky et al., 2001; Dixon et al., 2003; Meade and Hager, 2005; Peltzer et al., 2001b; Gourmelen and Amelung, 2005].

Indication on the depth of the brittle-ductile transition in the Hunter Mountain fault area is given by the seismicity at the nearby Coso geothermal area. Recording between 1991 and 1995, Feng and Lees [1998] determine that microseismicity was localized at a depth of 3 km at the geothermal field, and at a depth of 6 km in the neighboring region. Local studies by [1980], and a regional study by Sibson [1982], find a cut-off depth of the seismicity of 5 km, shallower than for most of California. This suggests that the shallow brittle-ductile transition in the Coso region is related to a high geothermal gradient. In terms of the geometry of the
Saline Valley – Hunter Mountain – Panamint Valley fault system, the shallow locking depth of the Hunter Mountain fault could be an indicator that the low angle normal fault geometry proposed by Biehler [1987] and Wesnousky and Jones [1994] (Figure 23) is active. If this is true, the Hunter Mountain fault is a much more complex system than modeled here and may tie in with the adjacent Saline Valley and Panamint Valley low angle normal faults.
CHAPTER 3  Acceleration and Evolution of Faults: An Example from the Hunter Mountain Fault, Eastern California

3.1  Background

The processes and rates of fault zone evolution are poorly understood, but have important implications for the mechanical nature of faults, seismic process and hazard, and regional scale tectonics. Fault slip rates averaged over different time spans may help define and perhaps understand this evolution. For large offset, mature fault zones such as the Carizzo segment of the San Andreas fault in central California, geodetic rates averaged over decadal time scales, and geologic rates over longer time scales, are usually very similar. However, for other fault zones, discrepancies between geodetic and geologic rates have been observed, leading to suggestions of complex behavior such as slip pulses, strain waves, and possible relations to earthquake clusters and other non-linear spatial-temporal deformation processes [Peltzer et al., 2001a; Meade and Hager, 2005; Bennett et al., 2003; Marco et al., 1996; Friedrich et al., 2004a; Dolan et al., 2007]. One explanation for discrepancies between geologic and geodetic rates that may apply to some young fault zones is that a new fault by definition must experience a finite period of acceleration, during which rate differences over different time spans are expected. However, it has been difficult to document this phenomenon, due largely to data limitations. Even definitions of mature and immature faults are lacking.
Here we present new space geodetic data describing recent motion across the Hunter Mountain fault, part of the Hunter Mountain–Panamint Valley fault zone in eastern California (Figure 24), and interpret it in the context of newly available constraints on the fault initiation age. Our data indicate that present= 

Figure 24: Map of major geological features in study area. M is Mountain, V is Valley, FZ is fault zone, SAF is San Andreas Fault. Blue box: location of figure 2. “Radar Track 442” indicates location of InSAR data in Figure 20
day motion across the Hunter Mountain fault is significantly faster than available geologic estimates. We suggest that this is a consequence of the fault’s relative youth, and slip acceleration over the last few million years. Our findings have important implications for understanding the time scales and processes by which faults straighten and simplify with time, and for the interpretation of fault slip rate data and some rate discrepancies.

3.2 Geologic Background and Previous Work

The eastern California Shear Zone (ECSZ) represents a zone of significant right lateral shear in western North America, accommodating ~20-25% of Pacific-North America motion [Dokka and Travis, 1990b; Savage et al., 1990; Sauber et al., 1994]. Most of the remaining motion is accommodated by the San Andreas fault. North of the Garlock fault and south of the Mina deflection, the ECSZ represents the eastern boundary of the rigid Sierra Nevada block, accommodating northwest translation of the block relative to the interior of the Basin and Range, presumably in response to northwest Pacific plate motion. In this region, the ECSZ includes three sub-parallel fault zones: the Owens Valley-White Mountain fault zone to the west, the Death Valley-Furnace Creek-Fish Lake Valley fault zone to the east, and the central Hunter Mountain-Panamint Valley fault zone, the focus of this study (Figure 24).

Dokka and Travis [1990b; 1990a] estimated that ECSZ activity began in the eastern Mojave Desert between about 6 Ma and 10 Ma. Wernicke and Snow [1998] document a change in Sierra Nevada motion relative to stable North
America, from westerly between 16 Ma and 10 Ma, to northwest or north-northwest beginning about 10 Ma, presumably marking the initiation of ECSZ motion. Atwater and Stock [1998] document a change in Pacific plate motion relative to North America at around 8 Ma, from WNW to NW. [Reheis and Sawyer, 1997] suggest that right lateral motion on the Fish Lake Valley fault, the northern continuation of the Death Valley-Furnace Creek fault zone, also began about 10 Ma (bracketed between 11.9 Ma and 8.2 Ma). All of these studies are thus broadly consistent with initiation of the ECSZ between about 8 Ma and 10 Ma. [Du and Aydin, 1996] suggest that the ECSZ acts as a “strain bypass” related to formation of the Transverse Ranges and “big bend” of the San Andreas fault. This feature is related to the inland jump of the plate boundary to its current position in the Gulf of California beginning about 5.5 Ma [Atwater, 1989; Oskin et al., 2001], hence ECSZ activity may have accelerated after 5.5 Ma. Jones et al. [2004] suggests that the rate of motion across the ECSZ rate may have accelerated after 3.5 Ma related to changes in regional lithospheric-asthenospheric structure.

While the kinematic boundary condition for both the ECSZ and the San Andreas fault, Pacific-North America plate motion, has remained essentially constant for the last few million years [Demets and Dixon, 1999], significant rate differences for different time intervals are common for various faults comprising the less mature ECSZ, but not for the central San Andreas fault. The central San Andreas fault, near the Carrizo plain, has remained in essentially the same location (± a few km) for the last ~5 million years or longer, has accumulated
several hundred km of offset, has a geodetic rate that is similar to both the Holocene rate, and the average rate since late Miocene time within uncertainties [Meade and Hager, 2005; Sieh and Jahns, 1984; Lisowski et al., 1991; Dickinson, 1996; Liu-Zeng et al., 2006; Schmalzle et al., 2006]. While the ECSZ has also been active for the last ~ 8-10 million years, individual fault zones within it are constantly evolving, as the locus of deformation has migrated westward [Dixon et al., 1995; Dokka and Travis, 1990b; Dokka and Travis, 1990a; Jones et al., 2004; Calzia and Ramo, 2000; Stockli et al., 2003; McQuarrie and Wernicke, 2005]. This makes the ECSZ an excellent “natural laboratory” for studying fault initiation and evolution.

**Initiation Time, Total Offset and Geologic Rate.** The Hunter Mountain fault is one of the younger faults comprising the ECSZ. This northwest striking slip fault is kinematically linked to more northerly striking normal faults to the north (Eastern Inyo fault zone, bounding the west side of Saline Valley) and south (Panamint Valley fault, bounding the east side of Panamint Valley). Estimates of fault initiation age on any one of these segments therefore likely marks the beginning of activity for the entire system. The age of fault initiation is bracketed between 2.8 Ma and 4.0 Ma [Burchfiel et al., 1987; Hodges et al., 1989; Lee et al., 2009]. The latter age represents the youngest age of Pliocene basalts offset by the fault; fault initiation is reckoned to occur sometime after basalt eruption.
Figure 25: Rate estimates for the Hunter Mountain - Panamint Valley fault zone, with estimated uncertainty at one standard error. Light gray [Zhang et al., 1990]. The difference between the measured InSAR rate and the Holocene rate of [Zhang et al., 1990] may reflect spatial complexity associated with partitioning of oblique extension on the Panamint Valley fault zone into normal and strike slip components, not all of which is captured in the geologic estimate. Symbol is obtained by projecting horizontal slip rate of the Panamint Valley segment onto the Hunter Mountain fault. Shaded rectangles represent minimum geologic rate and uncertainty (vertical axis) and assumed maximum age of faulting (horizontal axis); a) based on 9.3 +/- 1.4 km offset and post 4.0 Ma faulting [Sternlof, 1988; Hodges et al., 1989; Burchfiel et al., 1987]; b) [Sternlof, 1988] (Saline Valley reconstruction with 4.6 km of post 1.4 Ma offset). Curves show two possible Rayleigh models yielding total offset of 9.3 km: Beginning at 4 Ma [Hodges et al., 1989; Burchfiel et al., 1987] (S equals 1.9 Ma); Beginning at 2.8 Ma [Lee et al., 2009] (S equals 0.7 Ma). Note that in the 4.0 Ma model, the rate is still increasing at the present time. This model predicts a maximum rate of 5.5 mm/yr, reached several million years in the future. The Holocene rate on the simpler Hunter Mountain fault [Oswald and Wesnousky, 2002] is equivalent to the InSAR rate at 95% confidence.
The basalts are not found in the low-lying Panamint Valley, but only on the surrounding hills, and hence must have erupted prior to valley formation. The most recent work [Lee et al., 2009] suggests a 2.8 initiation age for the Hunter Mountain fault, based on zircon-apatite thermo-chronometry using U-Th/He exhumation ages on the Inyo Mountains. The Inyo Mountains block forms the west side of Saline Valley and the footwall of the Eastern Inyo fault zone. These data indicate more rapid uplift and exhumation of the Inyo Mountains after 2.8 Ma, suggesting initiation of the Eastern Inyo fault zone (and by implication the Hunter Mountain fault) at that time. We use this age in the subsequent discussion.

The total offset of the Hunter Mountain fault can be estimated from offset of the Hunter Mountain pluton, the width of Panamint and Saline Valleys in a direction parallel to the Hunter Mountain fault, and piercing points associated with Miocene-Pliocene age basaltic volcanics that ring northern Panamint Valley. A widely quoted estimate (9.3 ±1.4 km, [Sternloff, 1988; Burchfiel et al., 1987]) is based on the intersection of the steep southeastern contact of the Hunter Mountain batholith and the nearly horizontal unconformity at the base of Miocene-Pliocene volcanics; the displacement of this feature across the Hunter Mountain fault is mainly horizontal, consistent with strike slip motion.

As with the age of fault initiation, rate estimates on any individual segment of the Hunter Mountain-Panamint fault system are also generally assumed to apply more broadly. This assumption is discussed below. Published geologic slip rates for the Hunter Mountain-Panamint Valley fault are of two types. First,
detailed studies of offset alluvial fans and young drainage channels allow definition of Pleistocene or Holocene-averaged rates, ranging from 2.4 to 4 mm/yr [Zhang et al., 1990; Oswald and Wesnousky, 2002]. Second, long term average rates defined over the entire time span of fault activity are defined by dividing total offset (9.3 km) by the fault initiation age, and give 1.6 to 3.3 mm/yr depending on the assumed age (e.g., [Burchfiel et al., 1987; Lee et al., 2009]. Using the most recently published age (2.8 Ma) and the total offset (9.3 ±1.4 km) gives a rate of 3.3±0.5 mm/yr. Geologic rates are sometimes referred to as minimum rates, since the age of the offset feature may be a maximum age for the period of fault activity being considered, i.e., faulting began some unknown time after formation of the dated unit. Also, such rates by definition are less than the present day rate if the fault is accelerating. In general, we expect that geologic rates averaged over Holocene time should be similar to present day geodetic rates, unless there is complex temporal behavior, or earthquake cycle effects bias the geodetic measurement (e.g., [Dixon et al., 2003; Johnson et al., 2007]), or if the fault is spatially complex and some deformation is missed in the geologic study. South of the Hunter Mountain fault, deformation may be partitioned between the Panamint Valley fault zone, a shallow, west dipping fault that crops out on the east side of the valley and includes a component of normal fault motion (forming the valley), and more steeply dipping or vertical right lateral strike slip faults such as the nearby Ash Hill fault [Walker et al., 2005]. Geologic studies on individual faults in the Panamint Valley may therefore yield rates that are low compared to the entire deforming zone that includes the paired normal
fault – strike slip fault system, or to the Hunter Mountain fault to the north, where deformation appears to be more focused.

3.3 InSAR Results

We processed 44 Synthetic Aperture Radar (SAR) images from the European Space Agency satellites ERS1/2 acquired between 1992 and the end of 2000 to produce a ground velocity map in radar line-of-sight direction (LOS) across the ECSZ (Figure 20). Interferometric SAR (InSAR) data has been used in a number of tectonic studies, including measurement of co-seismic [Massonnet et al., 1993], interseismic [Fialko, 2006], and postseismic [Gourmelen and Amelung, 2005] deformation. We apply the Small Baseline Subset (SBAS) Interferometry technique to obtain a time series of ground displacement. The SBAS technique has been widely applied in the last few years [Berardino et al., 2002; Lanari et al., 2004b; Gourmelen et al., 2007; Lundgren et al., 2001]. SBAS relies on multiple (in this case more than a hundred) conventional interferograms with small spatial and temporal baseline, reducing long spatial wavelength noise, and allowing investigation of low slip rate faults.

The new InSAR data document a remarkable zone of high velocity gradient across the Hunter Mountain fault (Figure 21). In contrast, most other active ECSZ faults imaged in this data set exhibit low velocity gradients in the immediate vicinity of the fault, consistent with strain accumulation patterns around faults that are fully locked from the surface to a depth of 10-20 km, the typical range of locking depths in this region. For a series of sub-parallel strike slip faults that are
relatively closely spaced, the corresponding overlapping strain fields have made it challenging to estimate independent slip rates from geodetic data; while the summed slip rate across the entire shear zone is well constrained, the individual fault rates are not. Because of this high velocity gradient (and the implied shallow locking depth), the new geodetic data provide an opportunity to estimate the velocity of the Hunter Mountain fault, independent of the strain effects from the adjacent Owens Valley and Death Valley-Furnace Creek fault zones.

3.4 Geodetic Slip Rate

While areas to the south (Panamint Valley) and north (Saline Valley) of the Hunter Mountain fault undergo oblique slip with a component of extensional motion, the section of the fault studied here is believed to be a simple strike slip fault with no vertical motion. We first obtain the average LOS velocity by a least-squares fit. We then assume a simple strike slip fault model, and convert the satellite line of sight range change data into purely horizontal motion. We use a simple elastic screw dislocation model, and solve for best fitting slip rate (4.9±0.9 mm/yr) and locking depth (2 +/- 1 km). The data are well fit with this simple model, and no distributed deformation or additional faults are indicated. The shallow locking depth is unusual; it might be related to elevated geothermal gradient associated with the nearby Coso geothermal field (Figure 23). Alternately, it may reflect the kinematics of the Hunter Mountain Fault, which can be considered a short transfer fault linking oblique-extensional low angle normal faults to the north (Saline Valley) and south (Panamint Valley). The Panamint
Valley fault is inferred to have a very shallow decollement depth (less than 1 km) based on the depth of alluvial fill in the valley [Burchfiel et al., 1987; Hodges et al., 1989]. This shallow decollement may influence the mechanical behavior of the connecting Hunter Mountain fault.

Long term fault slip rate estimates based on geodetic data can be sensitive to earthquake cycle effects and assumed crustal rheology, which affects the pattern of strain accumulation near the fault [Dixon et al., 2003]. However, the high velocity gradient we observe, and implied shallow locking depth, means that the slip rate estimate is not sensitive to details of the rheological model. In addition, GPS data to the south, across the Panamint Valley-Ash Hill fault zone, indicate a similarly fast rate, 4-7 mm/yr, depending on rheological assumptions of the model [Schmalzle, 2008]. These results also indicate relatively shallow locking depths (~4-8 km). We therefore consider our present-day InSAR-based rate estimate for the Hunter Mountain fault to be robust. This rate is significantly faster than most available geologic estimates, though it overlaps the Holocene estimate for the Hunter Mountain fault [Oswald and Wesnousky, 2002] at 95% confidence (Figure 25).

3.5 Discussion

Rate Discrepancies, Fault Evolution, Fault Maturity. We suggest that the various rate estimates that characterize the Hunter Mountain fault over different time spans are best considered in the context of a model for a relatively young fault that is undergoing (or has recently undergone) acceleration and evolution.
Many authors have discussed the processes by which faults mature. As total offset increases, asperities and other frictional barriers to slip are reduced, segment length increases, en-echelon segments straighten and join, and the fault straightens and simplifies, reducing overall resistance to slip [Tchalenko, 1970; Wesnousky, 1988; Stirling et al., 1996; Ben-Zion and Sammis, 2003]. Ben-Zion and Sammis [2003] describe this as an inevitable consequence of strain weakening rheology, common in crustal materials. Presuming a constant tectonic driving force, as slip resistance decreases, the fault will accelerate to some steady state slip rate, and remain in that state until the tectonic driving force changes. A fault such as the central section of the San Andreas fault has experienced these processes over millions of years and hundreds km of offset, resulting in a single, straight, well-defined, low friction fault, accommodating most of the relative motion between the Pacific plate and the Sierra Nevada block, with limited distributed deformation, and at a rate that has remained essentially constant for the last few million years. With less than 10 km of total offset, the Hunter Mountain fault may not have achieved this state, or achieved it only recently. Presumably it experienced a period of acceleration sometime after 2.8 Ma, and underwent a similar process of geometric simplification, which may still be occurring.
Specific Sources of Rate Discrepancies. Even with a simple model of evolving, accelerating faults, there are at least three explanations for differences between geodetic and geologic rates. First, geological rate estimates tend to be minimum estimates, because the offset feature formed sometime prior to fault initiation. Second, some geologic estimates may not capture total motion across the fault zone. Distributed deformation and spatial complexity are characteristic of evolving, immature fault zones. If the fault has several active, sub-parallel strands, not all of which are well-exposed, Holocene estimates across only one of the strands will underestimate total slip rate, while space geodesy captures the full deformation field. Third, if the fault is accelerating with time, the space
geodetic rate estimate will by definition be faster than geological estimates averaged over long times. We focus on this third point, but note that the tendency of faults to simplify with time provides a simple physical mechanism that promotes fault acceleration during the early stages of the fault’s life.

**Models for Fault Slip Rate Evolution.**

Pioneering analogue experiments in clay- and sand-box models by Tchalenko [1970] first demonstrated the tendency of faults to simplify with time, with initial offset accommodated by highly distributed deformation, and overall slip resistance decreasing with increasing offset as the fault zone becomes focused on a central, narrow shear zone. On a similar spatial scale, Bodin et al. [1994] showed that creepmeter measurements of slip amplitude were generally larger than measured feature offsets, even when the creepmeter spanned just a few meters on either side of the active fault strand. On the scale of a single fault segment, Fialko [2005] note that the immature fault associated with the Bam earthquake has a significant co-seismic slip deficit in the upper crust compared to slip at depth, which they attribute to distributed deformation in the upper few km of crust. On the scale of multi-segment fault zones, aftershock distributions associated with the Landers earthquake in the Mojave Desert [Liu et al., 2003] and paleoseismic work in the Central Nevada seismic belt [Bell et al., 2004] illustrate the tendency of immature faults to exhibit multiple overlapping or en echelon active fault strands. With time, as total offset increases, these faults will presumably simply simplify their geometry [Wesnousky, 1988]. We suggest that
as this simplification progresses, the fault may accelerate, and further suggest that the Hunter Mountain fault illustrates this process.

Fault zone evolution may be described in terms of slip rate as a function of time, with the fault beginning at a condition of zero offset and zero slip rate. Offset and slip rate then increase until steady state is reached, whereafter offset increases at a uniform rate and slip rate remains constant. Geodetic data define the present day rate, and geological data can give rates averaged over various time intervals, in principle placing some constraints on the rate-time path. Few studies have considered the fact that, for faults in which initiation age is known, the three key data defining fault evolution (present rate, total offset, and initiation age) are not independent. The integral of a given rate-time path, beginning at zero rate at fault initiation age, and ending with the present-day rate, yields the total fault offset. If the latter parameter is well constrained by geologic studies, then integrating possible rate-time paths provides a useful (though non-unique) model test.

A simple model involving a linear increase in slip rate through time, from zero at 3.8 Ma to 4.9 mm/yr at present (our new InSAR result), predicts the correct total offset of 9.3 km. The inferred acceleration in slip rate is ~ 1.3 mm/yr per million years. However, this model is not consistent with the younger (2.8 Ma) age for fault initiation [Lee et al., 2009]; for a fault beginning at zero rate at this time, increasing linearly to the present rate of 4.9 mm/yr, total offset is 6.8 km, well below the mapped value, 9.3±1.4 km.
We therefore consider a simple Rayleigh cumulative distribution to model fault evolution. The Rayleigh distribution is a subset of the Weibull distribution, often used to model failure in complex systems, including fracture propagation in brittle materials \cite{Kurth and Cox, 1985}. The Rayleigh distribution is the simplest Weibull distribution, defined by a single parameter, \( S \), the Rayleigh scale parameter. We imagine that prior to fault development, a large number of randomly oriented, pre-existing fractures exist in the upper crust. As the proto-fault is stressed, the probability of these fractures slipping initially increases. As this occurs, properly oriented fractures fail repeatedly and lengthen, eventually coalescing to produce a through-going fault. At this point, overall stress is reduced, reducing the probability of additional fractures being exploited. The rate of fracture exploitation follows a Rayleigh distribution, increasing rapidly from the fault initiation time, with a peak determined by \( S \), and then decaying. Fault slip rate is related to the cumulative displacement on the individual fractures, initially increasing rapidly, and then more slowly to finally reach a constant high rate. Hence the fault slip rate as a function of time, \( R(t) \), follows a cumulative Rayleigh distribution, scaled by the stationary (final) rate, \( R_f \):

\[
R(t) = R_f \left[ 1 - e^{-\frac{(t-t_0)^2}{2S^2}} \right]
\]  

(1)

The slip rate is zero at \( t=t_0 \), increasing monotonically to \( R_g \) at \( t=0 \) (t is reckoned negative before present). For the present day geodetic rate \( R_g \) at \( t = 0 \), we obtain:
\[ R_g = R \left[ 1 - e^{-\left(\frac{t_0^2}{2S^2}\right)} \right] \]

(1) then becomes:

\[ R(t) = R_g \frac{1}{1 - e^{-\left(\frac{(t-t_0)^2}{2S^2}\right)}} \left[ 1 - e^{-\left(\frac{(t-t_0)^2}{2S^2}\right)} \right] \]  

\[ S \] is the time of peak acceleration of slip rate and can be estimated from the known total fault offset by integrating (2). At this time the rate, \( R_s \), is close to one half its final value, and accumulated offset is \( D_s \). Figure 25 shows two growth curves for the Hunter Mountain fault that bracket the range of plausible values for inception time and are consistent with the measured InSAR rate. Both give total offset of 9.3 km. For the 2.8 Ma inception time (our preferred model), \( D_s \) is ~0.5 km. A similar model may apply to the San Jacinto fault zone, which initiated between 1 and 2 Ma ago [Morton and Matti, 1993], has 25km of cumulative offset, and has \( D_s \) ~0.6 km, similar to value for the Hunter Mountain fault. Despite significant differences between these two faults, the values of \( D_s \) are surprisingly similar, suggesting that this degree of offset may have some physical meaning, marking a critical point in fault evolution. Faults with offset much less than this can be considered immature, while faults with offset much more than
this can be considered mature. By this definition, the Hunter Mountain fault can be considered mature.

Given rapid growth of the Hunter Mountain fault after 2.8 Ma, it is interesting to speculate concerning its influence on adjacent ECSZ faults. Our model, as well as previous studies [Tchalenko, 1970; Wesnousky, 1988; Stirling et al., 1996] predict increasing geometric simplification of fault zones with time. Thus, it might be expected that the ECSZ will also simplify, with one fault (perhaps the Hunter Mountain-Panamint Valley system fault) becoming dominant.

Several authors have speculated that the ECSZ has actually accelerated in the last few million years, e.g., in response to the post 5.5 Ma inland jump of the Pacific-North America plate boundary to its current position in the Gulf of California, or due to post-3.5 Ma crustal delamination that led to uplift of the Sierra Nevada batholith [Jones et al., 2004]. In this case, the HMF may have accommodated most of the acceleration, with other major faults retaining a constant slip rate. However, available data are also consistent with a steady state model, with acceleration of the HMF matched by deceleration of the Death Valley-Furnace Creek fault zone, and the overall displacement rate across the ECSZ remaining constant.

Dokka and Travis [1990b; 1990a] estimated ~65 km of right-lateral offset on northwest–trending faults in the eastern Mojave Desert after 10 Ma, and suggested that this offset connects with the Death Valley-Furnace Creek fault
zone to the north. Snow and Wernicke [1989] estimate 68±4 km offset on the Death Valley-Furnace Creek fault zone. Later work supports somewhat higher offsets. Niemi et al. [2001] document ~ 100 km of N67°W displacement since 11-12 Ma across the Death Valley region; part of this includes a component of east-west extension after 12 Ma, but the majority probably reflects the post-10Ma dextral displacement on northwest-trending faults. In contrast, total offset on the HMF is less than 10 km, suggesting that the Death Valley-Furnace Creek fault zone has been the major “player” in the ECSZ for much of its history, McQuarrie and Wernicke [2005] estimate that 100±10km of total northwest displacement east of the San Andreas fault has occurred since ~10 Ma. This defines an average rate of about 10 mm/yr, most of which probably applied to the Death Valley-Furnace Creek fault zone. However, its current rate is only about 4 mm/yr [Frankel et al., 2007]. The history of the HMF suggests a simple model whereby total ECSZ rate is approximately constant over the last few million years, with a rate of ~10 mm/yr, and acceleration of the HMF after 3 Ma is approximately matched by deceleration of the Death Valley-Furnace Creek fault zone over the same period.

In general, fault evolution may be considered a balance between local frictional conditions, which provide resisting forces, and regional tectonic conditions, which provide driving forces. Our model simplifies this complex topic, dealing only with the initial growth phase of fault evolution, and the tendency to straighten and simplify with increasing offset, lowering slip resistance; the “death” phase, associated with changing tectonics, is ignored. Also, many faults will
never reach $R_t$, either because of changing kinematic boundary conditions (e.g., changing plate configuration requires formation of new faults and abandonment of existing faults) or because finite friction always leads to some distributed, off-fault deformation, unrecorded by geologic estimates of offset. Perhaps this is why the Holocene rate on the Carrizo segment of the San Andreas fault (34 mm/yr; [Liu-Zeng et al., 2006]) is slightly lower than the geodetic rate (36 mm/yr; e.g., [Schmalzle et al., 2006]). Additional studies on this and other faults will be required to determined whether such differences are significant.

### 3.6 Conclusions

Knowledge of a fault's total offset and its age of initiation, combined with an accurate estimate of the present-day rate from geodesy, can yield useful constraints on fault growth and evolution. For the Hunter Mountain fault, available geologic data (fault inception time between 4 and 2.8 Ma, total offset of 9.3 km) and geodetic data (present day rate of 4.9 +/- 0.9 mm/yr) are satisfied with a simple model of fault acceleration and evolution. In this accelerating fault model, the geodetic rate estimate is invariably faster than long-term geologic rate estimates.
CHAPTER 4 Post-Seismic Mantle Relaxation in the Central Nevada Seismic Belt

Some of the largest earthquakes in North America during the 20th century were located in the Central Nevada Seismic Belt (CNSB), one of the known actively deforming area in the Basin and Range (Figure 27). The 1915 M7.2-7.6 Pleasant Valley earthquake, the 1932 M7.2 Cedar Mountain earthquake, and the 1954 Rainbow Mountain-Fairview Peak-Dixie valley earthquake sequence (4 M6.8-7.2 events in a 6 month period) were right lateral to normal slip events, and ruptured a ~250 km long, non-continuous stretch of north-northeast striking range front faults.

The present day deformation across the CNSB is puzzling for two reasons: First, the deformation rate during Holocene time is believed to be 0.5 to 1.3 mm/yr [Caskey et al., 2004; Bell et al., 2004; Bell et al., 1999], which is lower than the 2 to 4 mm/yr measured by GPS data [Thatcher et al., 1999; Hammond and Thatcher, 2004; Svarc et al., 2002]; Second, GPS measurements reveal a zone of east-west contraction east of the CNSB [Thatcher et al., 1999; Wernicke et al., 2000; Hammond and Thatcher, 2004; Svarc et al., 2002; Bennett et al., 2003] that is difficult to reconcile with current geodynamic models of the region, which involve east-west extension and right-lateral shear. One possible explanation for these two discrepancies is that the GPS data do not
Figure 27: 1992-2000 LOS velocity map for the area of the 1915-1954 Nevada earthquakes together with epicenters (blank circles), focal mechanisms and surface rupures, Green arrows: campaign GPS velocities [Hammond and Thatcher, 2004]. Red arrows: BARGEN permanent GPS velocities [Bennett et al., 2003]. LOS velocity considered positive for decreasing distance between ground and satellite.
record just the long-term deformation but that they also include transient
deformation associated with viscous or visco-elastic relaxation of the lower crust
or upper mantle following the last century’s earthquakes [Wernicke et al., 2000;
Hetland and Hager, 2003; Bennett et al., 2003]. We use 8 years of
interferometric synthetic aperture radar data to investigate ongoing deformation
in the CNSB.

The SAR imagery covers a nearly 700 km long swath (7 conventional SAR
frames) acquired by the ERS1,2 satellites between 1992 to 2000 to investigate
crustal deformation at the CNSB. For SAR processing we used the commercial
FOCUS processor and for InSAR processing the ROI_PAC v2.2.1 software (Jet
Propulsion Laboratory). We did not re-estimate the baseline between the satellite
orbits using topographic information because this may introduce second order
phase ramps. InSAR measures changes in the radar line-of-sight (LOS) distance
between the satellite and the surface of the Earth, and is most sensitive to
vertical movement and somewhat sensitive to east-west movements (The unit
vector from the ground towards the radar is (0.384, -0.075, 0.92) in an [east,
north, up] coordinate system for descending orbit). A ground velocity map in LOS
direction is shown in (Figure 27). The map has been obtained by averaging
(stacking) 8 independent long-term interferograms, each spanning 4-7 years.
(table S1). Most of the interferograms have perpendicular baselines smaller than
100 m. We used these pairs because larger baselines lead to decorrelation of the
interferometric phase. We obtained velocity map by dividing the cumulative LOS
displacement of the interferograms by the cumulative interferogram period of 37
years. We assumed that uncertainties associated with the satellite orbits cause linear phase ramps across the interferogram and removed them from the data.

The resulting ground velocity map shows a bulge with LOS velocity as high as \(~3 \text{ mm/yr}\) of relative motion with respect to the margin of the interferograms, centered in the epicentral area of the 1915 Pleasant Valley and the 1954 Dixie Valley earthquakes. About 1-2 mm/yr is detected in the areas of the Fairview Peak and Cedar Mountain earthquakes. The map also shows an area of subsidence in the northern part of the interferogram in the area of the Lone Tree gold mine presumably caused by ground water pumping in support of open pit mining operations.

In order to test whether the observed phase signature is real deformation or a processing artifact we generated another stack of interferograms using 8 interferograms covering shorter time periods (each < 4 months, total time span \(~2 \text{ years}\)). Since no deformation is expected from such a stack, a residual signal would reveal processing, atmospheric, or orbital artifacts. To obtain comparable LOS velocities we divided the cumulative LOS displacement of the short-term stack by the cumulative time of the long-term stack (37 years). The averaged LOS velocities based on the long-term stack (Figure 28) shows a long-wavelength signal of \(~3 \text{ mm/yr}\) of LOS velocity but the short-term stack does not. This indicates that the observed phase signature represents real ground deformation. Both profiles show a short-wavelength variation of 1-2 mm/yr LOS velocity attributable to the atmospheric variability.
To test whether the InSAR data can be explained by horizontal deformation as measured by campaign GPS measurements, we compared the LOS component of the horizontal GPS with the InSAR. The GPS data [Hammond and Thatcher, 2004] show a roughly linear increase of 3 mm/yr in velocity magnitude westward across the interferogram. This corresponds to a decrease

Figure 28: A LOS velocity relative to the margins of the interferogram along a north-south profile obtained from long-term and short-term interferograms also showing the location of the GPS profile (arrows). B LOS velocity relative to stable North America along a roughly east-west profile coinciding with the GPS campaign sites (line), together with LOS component of horizontal GPS velocities (triangles). The uncertainty on the LOS component of horizontal GPS velocities is on average 0.4 mm/yr.
in LOS velocity of 0.8 mm/yr (Figure 28B). LOS velocity is with respect to stable North America using the reference frame realization of [Hammond and Thatcher, 2004]). Transferring the InSAR data into the GPS reference frame shows that the InSAR data can not be explained by horizontal motion (Figure 28). The best explanation for the bulge in the InSAR data (2 mm/yr at this latitude) therefore is local uplift. Analysis of the vertical velocities of the permanent GPS network indicates that the station NEWS, situated in the area of maximum LOS velocity, is moving upward with respect to the surrounding stations (Giovanni Sella, personal communication), in agreement with the InSAR map.

One assumption is that errors associated with satellite orbit result in linear phase ramps in the interferograms. It is also well known that orbital errors may introduce more complex large wavelength errors. We are confident however that the observed signal represents real deformation because LOS velocity maps based on the same SAR acquisitions but different interferograms showed similar results, because the velocity map based on short-term interferograms does not show any similar signature although they are based on the same acquisitions, and because the detected deformation has a maximum velocity in the area of the largest historic earthquake, which is geologically plausible and consistent with GPS. We attempted to verify the result using data from the adjacent swath to the east but we could not produce an interferogram with a similar cumulative time.
We test whether the observed deformation may be caused by post-seismic relaxation of the Earth’s crust and mantle following the 1915-1954 earthquakes. We assume linear viscoelastic rheology and consider two and three layer Earth models using the methodology of [Pollitz, 1992]. The sources for post-seismic deformation are four earthquakes, the 1915 Pleasant Valley, 1932 Cedar Mountain and 1954 Fairview Peak and Dixie Valley earthquakes. We do not include the main 1954 Rainbow Mountain earthquake. The Rainbow Mountain earthquake was M6.8-7 [Caskey et al., 2004; Doser and Kanamori, 1986], predominantly strike-slip earthquake. Modelling this earthquake showed that the post-seismic deformation in LOS direction is more than an order of

Figure 29: Misfit between observed and modeled deformation. A Two-layer earth model consisting of an elastic plate overlying a visco-elastic halfspace. B Three-layer Earth model consisting of an elastic and a visco-elastic layer over a visco-elastic half-space.
magnitude smaller than for the Dixie Valley earthquake, and thus it can be safely neglected. We used fault parameters in the range of values published after field measurements [Wallace, 1984; Caskey et al., 1996], geodetic modeling [Hodgkinson et al., 1996] and seismologic modeling [Doser, 1986; Doser, 1988] (Table 1). For the Dixie Valley fault we use a dip of $30^\circ$ [Abbott et al., 2001] because the maximum LOS velocity $>15$ km east of the surface trace of the fault suggest a low angle dipping fault. We also invert for the magnitudes of the earthquakes allowing a deviation of 0.3 from the magnitudes [Doser, 1986; Doser, 1988], and from 7.1 to 7.7 for the Pleasant Valley earthquake.

![Figure 30: A Data and best-fitting post-seismic relaxation model. B profile.](image)

Our data set consists of 11072 equally spaced (~1.7 km spacing) LOS velocity measurements. Best-fitting models are characterized by a minimum of the difference between the data and the model predictions (We use a normalized
root mean square defined as \( \text{NRMS} = \sqrt{\frac{\sum (d_i - m_i)^2}{N}} \). We varied the grid spacing and used quadtree decompositions of the data to test whether the modeling results are sensitive to the sampling method and found that this is not the case.

We first use a two-layer Earth model consisting of an elastic plate overlying a viscoelastic half-space to obtain an estimate of the elastic thickness of the crust and of the viscosity of the underlying substrate. We conducted a grid search varying the elastic thickness and the viscosity. For each grid point we conducted a linear inversion for the slip magnitude to account for the uncertainty of the earthquake magnitude. The lowest misfits (0.3 mm/yr) are found for models with an elastic thickness larger than 20 km and a subcrustal viscosity of \( 1 \times 10^{18} \) Pa s (Figure 29).

We also used a three-layer Earth model consisting of an elastic layer overlying two viscoelastic layers, representing the elastic upper crust, the viscoelastic lower crust, and the viscoelastic upper mantle. We use an elastic layer thickness of 15 km (seismogenic thickness), and a thickness of the lower crust also of 15 km so that the crustal thickness agrees with the 30 km inferred from seismic reflection data [Catchings, 1992]. We vary the viscosity of the lower crust and of the uppermost mantle. For this model the lowest misfits are found for lower crustal viscosities larger than \( 10^{20} \) Pa s and for upper mantle viscosities of \( 1 \times 10^{18} \) Pa s. The LOS velocity predicted by the best-fitting model explains the large wavelength deformation (Figure 30). We consider models with \( \text{NRMS} < 0.35 \) mm/yr as reasonable models.
The InSAR data can not be explained with post-seismic models using the published earthquake magnitudes and therefore we inverted for the magnitudes. This is desirable because the magnitudes are not well constrained by the instrumental data. In fact, our study shows that precise post-seismic deformation data can be used to estimate the magnitude of historic earthquakes as long as

Figure 31: Bottom: Magnitude of horizontal campaign GPS velocities along an east-west profile through the Basin and Range (7). Upper line: measured velocities, lower line: measured velocities minus model-predicted post-seismic deformation. Top: simplified tectonic map of the Basin and Range. Triangles: location of the GPS stations. CNSB: Central Nevada Seismic Belt, WL: Walker Lane.
an estimate of the focal mechanism is available. We found the same magnitudes for the two- and three-layer models. (Table 1). The magnitude of the Fairview peak earthquake remains unchanged at M7.2. For the Cedar Mountain and Dixie Valley earthquakes we find magnitudes of M7.1 and M6.7 corresponding to a reduction of 10% from the published magnitudes. For the Pleasant Valley earthquake we find a moment magnitude of 7.3, significantly smaller than the seismologic estimates of 7.6 [Lienkaemper, 1984] but in agreement with the value of 7.2 derived from surface faulting [Wallace, 1984]. The cumulative moment magnitude of the four modeled earthquakes is M7.55.

It is striking that the area of deformation is significantly larger than the epicentral area of the earthquakes and that the deformation field lacks short-wavelength features. The lithospheric rheology acts as low-pass filter that translates the instantaneous short-wavelength earthquake stress into long-wavelength deformation lasting several decades. This suggests that the upper parts of the lithosphere behave elastically on the time scale of our data and that viscous relaxation occurs only at greater depth. For the two-layer model we find a lower bound for the thickness of the elastic layer of 20 km and a viscosity of the underlying substrate of $1-10 \times 10^{18}$ Pa s. Using a three-layer model we find a viscosity of the substrate in the same range and a viscosity of the intermediate layer (lower crust) larger than $10^{20}$ Pa s. These results suggest that most of the crust or the entire crust of the Basin and Range lithosphere (including the lower crust) behaves elastically for at least 80 years following these large earthquakes. Relaxation of the earthquake induced stress occurs by viscous flow in the
mantle. These rheology estimates are consistent with previous studies in the Basin and Range and in the Mojave desert, which also showed an elastic or high-viscosity lower crust and a low-viscosity upper mantle [Pollitz et al., 2000; Bills et al., 1994; Kaufmann and Amelung, 2000; Nishimura and Thatcher, 2003; Dixon et al., 2004].

The GPS data collected along an east-west profile indicate an area of low-rate contraction east of the CNSB (Figure 31). A profile of secular ground velocity, obtained by removing the model-predicted post-seismic velocities from the GPS vectors (The post-seismic model predicts at the latitude of the GPS profile (~N39.3), 2 mm extension across the CNSB in roughly east-west direction. Removing this from the GPS vectors of [Hammond and Thatcher, 2004] leaves 0-2 mm/yr secular deformation, also roughly oriented east-west. Further north at the latitude of another GPS profile at N41 latitude [Hammond and Thatcher, 2005] the predicted post-seismic deformation is less than a millimeter a year.) does not show this contraction, but shows only deformation west of the CNSB (Figure 31). This suggests that the GPS-measured contraction is a post-seismic effect and supports the simple geodynamic picture for the Basin and Range in which the central Basin and Range is an essentially undeforming block with deforming boundary zones, i.e. the CNSB and the Walker Lane to the east and the Wasatch fault zone to the west [Thatcher et al., 1999; Dixon et al., 2003; Malservisi et al., 2003]. This interpretation is consistent with the geodetic microplate model for the Central Basin and Range of [Bennett et al., 2003]. The residual velocity across the CNSB itself is 0 to 2 mm/yr, in agreement with
geologic estimates of deformation [Bell et al., 2004]. This implies that the CNSB does not have the elevated seismic potential attributed based on the GPS measurements.
CHAPTER 5  Mining-Related Ground Deformation in Crescent Valley, Nevada: Implications for Sparse GPS Networks

5.1 Background

Global Positioning System measurements over the last 15 years have provided new insights into the contemporaneous deformation of the Basin and Range Province in the western USA [Thatcher et al., 1999; Wernicke et al., 2000; Hammond and Thatcher, 2005]. The dominant deformation modes are right-lateral shear and extension along the western and eastern margins (in zones referred to as the Walker Lane and Central Nevada Seismic Belt, CNSB, and Wasatch fault) [Dixon et al., 1995; Malservisi et al., 2003], whereas the Central Basin and Range behaves as a stable block [Bennett et al., 2003; Gourmelen and Amelung, 2005]. However, the motions of some GPS stations of the Basin and Range Geodetic Network (BARGEN) deviate from this simple deformation field [Wernicke et al., 2000; Hammond and Thatcher, 2005; Friedrich et al., 2003]. The most prominent example is the GPS station LEWI on the Mt Lewis in the Shoshone mountain range west of the Crescent Valley (Figure 32). LEWI moves 2–3 mm/yr southeast with respect to the neighboring stations (Figure 32a) [Wernicke et al., 2000; Bennett et al., 2003; Friedrich et al., 2004b]. This is best illustrated by plotting its velocity in a reference frame fixed to the stations NEWS and MINE (blue arrows in Figure 32, red arrows are velocities from
Figure 32: InSAR deformation velocity maps. (a) Shaded relief of the study area and permanent GPS stations from the BARGEN network. GPS motion is with respect to stable North America (red arrows) and with respect to nearby GPS stations NEWS and MINE (blue arrow). Black boxes: location of InSAR derived rate maps. (b) Rate velocity map over Crescent Valley (subset of Figure 32d). GPS station LEWI and residual velocity (blue arrow). Location of profile on Figure 33 (white dashed line) and distance marks. Color scale wrapped for better visibility. (c) Time series of deformation from 1992 to 2002, from (top) away to (bottom) within the Crescent valley subsidence area. Black circle are water level at a well collocated with the top time series showing linear relationship between surface deformation and water level. (d, e) Rate velocity maps of the subsidence at Crescent Valley, Diamond valley, Reese River Valley and Antelope Valley. Maps obtained from processing of two adjacent radar tracks (track 442 and track 170). White lettering: location of profile of Figure 33. Black dots: location of time series of deformation (Figure 32d) arranged in the order of time series (Figure 32c). The color scale is saturated at 15 mm/yr.
Bennett et al. [2003] with respect to stable North America). LEWI’s motion corresponds to crustal shortening of the Crescent Valley area, which is not consistent with the overall extension in the region [Friedrich et al., 2004b]. Although mining is occurring in the Crescent Valley possible interaction has been consider unlikely because of the distance from LEWI to the mine and because mining-induced deformation is occurring in the sediment fill. Gourmelen and Amelung [2005] used Interferometric Synthetic Aperture Radar (InSAR) to find evidence of transient deformation following a series of 1915–54 earthquakes in Western Nevada but their model does not explain the motion of LEWI. We exploit multi-temporal InSAR interferograms and generate radar LOS displacement time series and velocity maps for two adjacent SAR swaths. We invert the data in the overlapping region to retrieve the ground velocity components in the vertical and ground range direction. We show that the mining activities can produce significant horizontal deformation likely to account for part of the movement of GPS even in bedrock environment.

5.2 InSAR Data

We used ERS1,2 SAR data, frame 2799, tracks 442 and 170. The satellite tracks (descending with the satellite traveling south) are adjacent resulting in a 30 km overlap of the imaged area at this latitude (Figure 32a). InSAR measures ground displacements in the radar line-of-sight direction (LOS). The look angle varies across a full ERS SAR image from about 19° in the near range to 27° in the far range.
Figure 33: Vertical and ground range velocity inversion (see location of profiles in Figure 32d and Figure 32e). (a) East-west profiles across track 442 derived deformation rate map (v2LOS, black) and track 170 derived deformation rate map (v1LOS, red). (5 km Profiles width). For better visualization of the difference, the data for track 170 (v1LOS) have been shifted downwards by 3 mm/yr with respect to the data for track 442 (v2LOS) on profile B-B'. (b) Inversion for ground range and vertical velocities for the three profiles in (a). The B-B' profiles shows ground range and vertical deformation whereas A-A' and C-C' do not. The scatter of the points appears higher when point density is higher but is in fact similar. (c) Sketch showing the principle of the inversion based on overlapping area and angle difference of two adjacent radar tracks. (d) Filtered profile B-B' from (b). Areas of surface bedrock and sediments is indicated.

The images overlap by more than one third of the image width so that for the Crescent Valley there is a difference in the LOS direction of 4.0° (Figure 32 and Figure 33).

We processed the SAR data into LOS velocity maps using the Small BAseline Subset (SBAS) methodology [Berardino et al., 2002; Lanari et al., 2004b; Casu et al., 2006] and in CHAPTER 2. For each SAR acquisition we use the 5 interferograms with the smallest perpendicular baseline as long as it is less than 400 m and the temporal separation is less than 4 years. A temporal and spatial filtering removes linear phase contributions that are correlated in space but not in time.
such as related to errors in the satellite orbits. For this, the phase history for each pixel is smoothed using a triangular filter with a filter length of about 2 years. For each acquisition, the difference between the smoothed and the original phase is then approximated by a linear phase histories of LOS displacement with respect to the first ramp simulating the orbit error for this acquisition. Modified interferograms are formed by removing the estimated orbital phase ramps, the displacement histories and averaged velocities are then recalculated. We finally remove linear ramps from the LOS velocity fields related to residual orbital errors after filtering [Burgmann et al., 2006; Casu et al., 2006].

The LOS velocity maps show 4 areas of LOS increase of up to several centimeters per year in the Crescent Valley (Figure 32d, e), Antelope Valley (Figure 32d), Reese River Valley (Figure 32d,e) and Diamond Valley (Figure 32e). These areas are mining and agricultural fields, suggesting that we see anthropogenic ground movements.

In the Crescent Valley, the maximum LOS increase occurs 4 km east of the open-pit Cortez mine. The time histories of LOS displacement with respect to the first acquisition along a cross-section through Crescent Valley indicate up to 250 mm of displacement between 1996 and 2002 in the center of the deforming area, corresponding to a LOS velocity of 40 mm/yr (lower time series in Figure 32c). Subtle uplift of 12 mm (2 mm/yr) is detected 20 km northeast of the mine (top time series in Figure 32), likely associated with the surface discharge of groundwater pumped from the bottom of the open pit. LOS increase started in summer 1996 (Figure 32), when the exploitation of the mine began. The rate of
LOS increase was nearly linear until the end of 1998. Then, the ground velocities diminished, in particular at the edge of the deforming system as well as within the area of uplift. For simplicity, we approximate the deformation by a linear velocity for the 1996–2002 period.

The ground displacements are roughly consistent with water level measurements. No data are available for the subsiding area but well data in the uplift area in the north show a significant water level rise between 1996 and 1998 and then a stabilization of the water level (black circles and line in Figure 32c).

A profile through the deforming area of the Crescent Valley shows a subtle difference in LOS velocity of up to 3 mm/yr between the two swaths (at 15–25 km distance along profile B-B', Figure 33a). No similar difference occurs in the non-deforming areas further north and south (profiles A-A' and C-C', Figure 33), suggesting that it is related to the difference in radar incidence angle on the ground (Figure 33).

5.3 The 2-D Ground Velocity Field From Overlapping Satellite Swaths

The horizontal component of the velocity vector in LOS direction (the ground range velocity), \( v_{gr} \), and the vertical component, \( v_v \) (Figure 33) are related to the measured LOS velocities \( v_{1,2}^{LOS} \), by:

\[
\begin{pmatrix}
  v_{1}^{LOS} \\
  v_{2}^{LOS}
\end{pmatrix} =
\begin{pmatrix}
  \sin(\theta_1)\cos(\theta_1) \\
  \sin(\theta_2)\cos(\theta_2)
\end{pmatrix}
\begin{pmatrix}
  v_{gr} \\
  v_v
\end{pmatrix}
\]  

(26)
with $\theta_{1,2}$ the radar incidence angles for the two interferograms (see Wright et al. [2004a] for a more general formulation). From this we obtain:

$$v^{gr} = \frac{v_{1}^{LOS} - v_{2}^{LOS} \cos \theta_{1}}{\cos \theta_{1} - \cos \theta_{2} \cos \theta_{1}} \cos \theta_{1}$$

$$v^{v} = \frac{v_{1}^{LOS} - v_{2}^{LOS} \sin \theta_{1}}{\cos \theta_{1} - \cos \theta_{2} \sin \theta_{1}} \sin \theta_{1}$$

The $v^{gr}$ and $v^{v}$ for the 3 profiles are shown in Figure 33b and averaged curves for the B-B' profile is shown in Figure 33d. For the deforming area, $v^{v}$ varies from 6 mm/yr at the beginning of the profile to a maximum of 20 mm/yr at 16–17 km distance and is slightly positive (+1.5 mm/yr) at 18–21 km distance (Figure 33d). The latter could be the result of local faulting, uplift similar to the uplifting bulge in the perimeter region of surface loads, or is introduced by a systematic error discussed below. The pattern of the horizontal velocity $v^{gr}$ is consistent with the expected convergent motion towards a pumping source as reflected by a positive $v^{gr}$ west of, and a negative $v^{gr}$ east of the deforming center. However, $v^{gr}$ is asymmetric. The amplitude of $v^{gr}$ is higher in the east than in the west (Figure 33) and the null horizontal velocity point is offset westward with respect to the maximum vertical velocity point. This contrast with the symmetric deformation
field expected from Biot’s theory of poro-elasticity. This difference possibly reveals the laterally varying thickness of the basin fill, thicker in the east where the basin is bounded by a west-dipping normal fault. Underlying bedrock may account for some of the deformation within the basin or faults may act as barriers for ground water flow [Amelung et al., 1999].

Along the profile to the south, \( \eta^{s\sigma} \) and \( \eta^v \) are approximately zero as expected for a non-deforming area (Figure 33, profile C-C’). Along the profile to the north \( \eta^{s\sigma} \) is zero but \( \eta^v \) is slightly positive (Figure 33, profile A-A’) due to the subtle uplift associated with the release of the pumped water noted in Figure 32.

5.4 Measurement Uncertainties of the InSAR Data

The uncertainties associated with measurements of the LOS displacements \( u^{\text{LOS}} \) depends on how well atmospheric, topographic, and orbital contributions have been corrected by the SBAS processing algorithm, and on the distance of a pixel to the reference point. We find for the vicinity of the reference point in a non-deforming area a standard deviation of \( \sigma_{u^{\text{LOS}}} = 2 \) mm. The spatial gradient of the standard deviation is 0.05 mm per kilometer from the reference point.

The velocity estimates \( \eta^{\text{LOS}} \), \( \eta^{s\sigma} \), \( \eta^v \) uncertainty is a complex function of the uncertainty of the \( u^{\text{LOS}} \) measurements and their correlation, on the temporal spacing of the SAR acquisitions and on the time period covered. For a sample of 10000 pixels in a 15 km non-deforming area, we find a standard deviations of \( \sigma_{\eta^{\text{LOS}}} = 0.6 \) mm/yr, \( \sigma_{\eta^{s\sigma}} = 9 \) mm/yr and \( \sigma_{\eta^v} = 4 \) mm/yr and a negligible
dependence on the distance to the reference point. The distributions for $v^{\text{LOS}}$ are slightly asymmetric narrower than a Gaussian distribution with 85% and 97% of the samples within the 1 s and the 2 s intervals respectively. The distributions for $v^{gr}$ and $v^{o}$ on the other hand are well approximated by a Gaussian distribution with 69% and 95% of the samples within the 1 s and 2 s intervals, indicating that the standard deviations are acceptable uncertainty measures. They are 10% smaller than the expected values from linear error propagation of equations (27) and (28).

5.5  **Ground Velocity at GPS Station LEWI**

The GPS station LEWI is located 22 km to the northwest of the center of the Crescent Valley deforming area at 18 km from the Cortez mine (Figure 33b). LEWI moves 2–3 mm/yr southeast with respect to the neighboring stations. This corresponds to 1.5–2.5 mm/yr in ground range direction (the angle between LEWI’s velocity and the ground range direction is 35°). No InSAR measurements are available from the SBAS analysis for LEWI for direct comparison with the GPS data because it is located in frequently snow-covered mountains. About 9 km from LEWI at the start of the profile B-B’ we have inferred a velocity of $8 + 9$ mm/yr in ground range direction (Figure 33b and Figure 33d) Although we do not know how rapidly the deformation decays with distance from the pumping sites it seems very plausible that LEWI’s apparently anomalous motion is caused by the Crescent Valley pumping activities. This is consistent with the observation that LEWI moves towards the center of the deforming area.
5.6 Bedrock Deformation in Response to Water Level Changes

We have shown that ground water pumping is causing horizontal and vertical
ground deformation in bedrock at a distance of several tens of km to the pumping
site. This is in contrast to previous observations, for example in the Los Angeles
basin [Bawden et al., 2001] and Las Vegas valley [Amelung et al., 1999; Bell et
al., 2008]. Deformation is attributed to sediment compaction with both vertical
and horizontal displacement [Watson et al., 2002; Argus et al., 2005]. Recently,
Burbey et al. [2006] showed that pumping can be associated with horizontal
deformation of the same magnitude as the vertical deformation.

A spectacular example for pumping-induced bedrock deformation occurs in the
Tuscarora Mountains (Figure 34) 50 km northeast of the Cortez mine. The total
LOS displacement through the 1992–2001 period reaches 30 cm. The
deformation follows the mapped faults at the eastern margin of the Tuscarora
mountains. The small valleys south of the mine have no effect on the deformation
pattern suggesting a relatively deep source for the deformation (deeper than the
valley fills).

Possible mechanisms for the large-scale pumping-induced bedrock deformation
are poro-elastic deformation and the contraction or expansion of the rock mass
by closing and opening of fractures and fault gouge in response to water level
changes [Cappa et al., 2006]. We would expect horizontal deformation
perpendicular to the strike direction. This is consistent with deformation pattern in
the eastern Tuscarora mountains where the displacement gradient is
perpendicular to the faults and consistent with LEWI’s motion perpendicular to
the Shoshone Range’s strike. Another explanation would be that the bedrock composition of both the Shoshone Range (LEWI’s bedrock) and the Tuscarora Mountain contains shale and argillite, two rock type rich in clay, a mineral who’s volume is very dependant of water content.

5.7 Discussion

We have applied the SBAS-InSAR technique in order to estimate LOS mean deformation velocity maps for two adjacent SAR swaths. The presented analysis reveals a distinct deforming area, located in the Crescent Valley, Nevada, with
LOS velocities of up to 40 mm/yr caused by ground water pumping in support of mining activities. Inversion of the InSAR data for the 2-D velocity field, in the overlapping SAR swaths region, indicates subsidence and convergent horizontal deformation. Pumping-induced ground deformation may explain the enigmatic motion of the GPS station LEWI. Pumping-induced ground deformation can occur in bedrock at a distance larger than 20 km from the pumping site and needs to be evaluated prior to the geodynamic interpretation of sparse GPS velocity field
CHAPTER 6  CONCLUSION

6.1.1  CNSB Slip Rates

The high slip rate and low strike-slip to extension ratio determined from the present day velocity field across the CNSB is attributed to a combination of post-seismic deformation and man made surface subsidence in support of mining activities.

Post-seismic deformation

InSAR detects a broad area of uplift in the Central Nevada Seismic Belt. The detected deformation can be explained by post-seismic relaxation following the 1915-1954 earthquakes using typical rheological models for the Basin and Range. The post-seismic signal amounts to 2 mm/yr of extension and 2 mm/yr of uplift. When subtracted, the present day deformation can be modeled with a combination of $2.3 \pm 0.3$ mm/yr right lateral and $1.0 \pm 0.4$ mm/yr extensional so a total rate of $2.5 \pm 0.5$ mm/yr. The strike-slip to extension ratio is 1.4 to 4. These results lead to a broad agreement between geologic and geodetic strain indicators for the CNSB.

The CNSB present therefore a case study for visco-elastic post-seismic relaxation, because of the time span of the post-seismic relaxation deformation i.e. 87 years after the 1915 Pleasant Valley earthquake, and 48 years after the 1954 Dixie Valley earthquake.
Man-made subsidence

InSAR detects numerous localized ground deformation related to mining activities in Nevada. I demonstrate that a subsidence process in a nearby mine is accounts for the anomalous deformation of the continuous GPS station LEWI, that therefore does not represent anomalous fault notion nor large scale post-seismic deformation at this location. InSAR Analysis reveals a distinct deforming area, located in the Crescent Valley, Nevada, with LOS velocities of up to 40 mm/yr. Inversion of the InSAR data for the 2-D velocity field, indicates subsidence and convergent horizontal deformation. Pumping-induced ground deformation may explain the enigmatic motion of the GPS station LEWI. Pumping-induced ground deformation can occur in bedrock at a distance larger than 20 km from the pumping site and needs to be evaluated prior to the geodynamic interpretation of sparse GPS.

6.1.2 ECSZ’s Hunter Mountain Slip Rate

InSAR detects velocity change across the HM fault. The velocity change is a sharp signal, distinctive from the broad strain accumulation signal associated with nearby faults. Modeling of the deformation assumes pure right-lateral strike slip on the HM fault. Result suggests a slip rate of \(4.9 \pm 0.9 \text{ mm/yr}\). The obtained slip rate is larger than averaged long term and Holocene slip rate for the fault. The available geologic data (fault inception time between 4 and 2.8 Ma, total offset of 9.3 km) and geodetic data (present day rate of \(4.9 \pm 0.9 \text{ mm/yr}\) are
satisfied with a simple model of fault acceleration and evolution. In this accelerating fault model, the geodetic rate estimate is invariably faster than long-term geologic rate estimates.

6.1.3 **Geometry of the SHP fault system**

Modeling of the surface deformation associated with the HM fault suggests a locking depth of $3 \pm 1$ km, significantly lower than the average 10-15 km for nearby faults. The model shows no effect from visco-elastic rheology, which suggests significant time since last earthquake and/or low viscosity. Assuming that the HM fault is part of an active low and angle SHP fault system, the locking depth corresponds to the depth of the Saline Valley and Panamint Valley low angle normal fault at their intersection with the HM fault. This suggests that the system is actively deforming. Assuming only strike slip mechanism for the SHP is contrary to field observations and to our analyses of surface deformation. Considering the SHP fault system as a purely strike slip fault system for sake of simplification would lead to error in model parameters.

6.1.4 **How rigid is the Basin and Range?**

The post-seismic surface velocity across the CNSB is determined from a combination of InSAR velocity measurement and modeling. After subtraction of the post-seismic velocity field from the Basin and Range velocity field [Hammond and Thatcher, 2004], the Basin and Range appears stable in an region bounded by the Wasatch Front Fault to the East and by the ECSZ and CNSB fault systems to the east [Gourmelen and Amelung, 2005]. This supports the simple
geodynamic picture for the Basin and Range in which the central Basin and Range is an essentially undeforming block with deforming boundary zones (i.e., the ECSZ, and the CNSB to the east and the Wasatch fault zone to the west) [Thatcher et al., 1999; Dixon et al., 1995; Malservisi et al., 2003]. This interpretation is consistent with the geodetic microplate model for the Central Basin and Range of [Bennett et al., 2003].

However, the recent occurrence of the M6 Wells earthquake (Amelung, in prep), in an area of the central Basin and Range in Nevada, suggests that the central Basin and Range undergo strain accumulation. One explanation would be that strain accumulation occurs locally, in the region of Wells. Because the geodetic network of [Hammond and Thatcher, 2004] does cover an area remote from Wells, this deformation signal would not be picked-up. It is also possible that the precision of the current geodetic surveys in the Basin and Range is below the one required to measure the strain accumulation occurring in the central Basin and Range. A similar uncertainty exists in the eastern US, where a large M8 earthquake occurs in the New Madrid region but current geodetic surveys failed to identify a strain signal around the faults [Newman et al., 1999].

6.1.5 Measuring Low Strain Rate with Synthetic Aperture Radar

1. We review the theoretical framework for orbital phase errors (OPE) in InSAR time series. Orbital phase errors are related to uncertainties in the satellite position during image acquisitions. They impact the ability of InSAR to precisely measure subtle, long-wavelength deformation (over several hundred km).
2. The orbital phase error at a given epoch can be expressed in terms of the horizontal and vertical baseline errors (difference in satellite position at a given epoch with respect to the first epoch) (eq. 17). The baseline errors at a given epoch are a function of along track position. Assuming that their along track variation can be approximated by second-order polynomials the orbital phase error is described by 6 parameters (eq. 22). In the absence of deformation the orbital phase errors can be estimated directly from the InSAR data. In the presence of deformation they can be estimated using independent information such as from models or GPS (eq. 23).

3. We apply this method to the Eastern California Shear zone using 44 ERS SAR acquisitions from 1992-2001 and data from 6 continuous GPS stations starting in 1999. Using this method InSAR recovers the long-wavelength deformation of the region known from GPS (Figure 15).

4. Other applications of the method include the study of Glacial Isostatic Adjustment (GIA), with typical strain rate of 0.05mm/km/yr (Figure 35), oceanic tides, or when the SAR data orbits are poorly known like in the case of the RADARSAT satellite.
Figure 35: Vertical velocity of the North American continent [Sella et al., 2007]. Note the large wavelength of the GIA signal.
CHAPTER 7  Annex: University of Miami InSAR Processing Chain

7.1  Background

This appendix presents the University of Miami InSAR laboratory processing chain for interferogram generation (process_rsmas.pl), stacking (geodmod) and time-series generation (TSSAR) and its ability to reproduce previous results. Two datasets are used as example, the first dataset is from Gourmelen and Amelung, [2005] where stacking is performed from interferograms produced using a combination of the Vexcel and the Repeat Orbit Interferometry PACkage, ROI_PAC, softwares, the second dataset is from Gourmelen et al., [2009] where time-series were processed by the IREA group from Napoli, Italy, using the Small BAseline Subset, SBAS, algorithm [Berardino et al., 2002]

7.2  Stacking of track 213 - Central Nevada Seismic Belt

The interferograms used in Gourmelen and Amelung [2005] were processed using the code Focus from the Vexcel software for the Single Look Complex, SLC, generation and ROI_PAC for interferogram generation. Focus does not support chirp extension, the extension of the processed area in range is therefore limited. For this appendix, the interferograms were regenerated using process_rsmas.pl, which supports chirp extension (Figure 37). Similar to what was done in Gourmelen and Amelung [2005], we did not use the baseline re-estimation feature to correct for orbital phase error (the effect of orbital phase errors are discussed in chapter II).
Figure 36: Individual interferograms used in [Gourmelen and Amelung, 2005].
Figure 37: Individual interferograms used in [Gourmelen and Amelung, 2005] after removing a linear phase ramp.
Figure 38: State vector versus along track position (140 sec. record) for SAR epoch 930918 used in the processing, and second order polynomial fit.

Figure 39: Stack of interferograms shown in Figure 37.
For the interferogram processing, we use the precise DEOS orbits from Delft University. The precise orbit state vectors are provided every 60 seconds, they are then interpolated every 10 seconds within the time of the SAR data acquisition by performing an 8th order Legendre interpolation (routine inter8.f in the getorb distribution). ROI_PAC then computes the vertical and horizontal baseline between the two satellite passes. The baseline variation in function of along track position is modeled with linear or second order polynomial. Although this model might be insufficient when SAR strip of several hundreds of kilometers are processed, this has not yet been demonstrated. A second order polynomial fit to the state vectors performed on SAR epoch 930918 shows that a second order polynomial is efficient in modeling the state vector over the length of our study area (Figure 38).

The stack produced from the interferograms processed using process_rsmas.pl is identical as the one in Gourmelen and Amelung [2005] (without chirp extension) Figure 39. The additional measurement in range direction provided by the chirp extension processing shows areas of uplift east and west of the CNSB uplift described by Gourmelen and Amelung [2005]. I suspect that the small area of uplift to the west of the CNSB uplift area is the result of unwrapping errors.

7.3 Time-series analysis of track 442 - Eastern California Shear Zone

This section presents a case study for time-series inversion from the track 442, frames 2781 to 2871 and a comparison with the processing performed by the group of Riccardo Lanari at IREA, Napoli, Italy (Figure 40).
7.3.1 **TSSAR processing summary**

TSSAR is Miami's time series analysis software for inverting interferograms into displacement time-series. It is directly inspired from the Small BASeline InSAR method developed by the group of Riccardo Lanari at the Istituo Per Il Rilevamento Elettromagnetico Dell'Ambiente, IREA, in Napoli [Berardino et al., 2002; Lanari et al., 2004]. First TSSAR determine a list of interferograms forming a connected network from Delaunay triangulation of SAR data. The selection is based on the distribution of two parameters, the perpendicular baseline, \( b_{\text{Perp}} \), (minimization of the phase related to error in the digital elevation model), and the temporal baseline, \( b_{\text{Temp}} \), (minimization of temporal decorrelation), between SAR acquisitions. For a given project, thresholds \( b_{\text{Perp} \text{thresh}} \) and \( b_{\text{Temp} \text{thresh}} \) are chosen based on the characteristic of the region (ability in conserving ground characteristic over a certain time period, topographic heights range). Following the work by Pepe and Lanari [2006], we scale the temporal baseline values by the ratio \( \frac{b_{\text{Perp} \text{thresh}}}{b_{\text{Temp} \text{thresh}}} \). It has been shown by Pepe and Lanari [2006] that small variations in the ratio \( \frac{b_{\text{Perp} \text{thresh}}}{b_{\text{Temp} \text{thresh}}} \) do not significantly modify the network. After computation of geocoded interferograms, a mask is computed based on the temporal coherence. Temporal coherence is a way to quantify inconsistency in the set of interferograms, more details and possible causes are discussed in CHAPTER 2. Temporal coherence is estimated via (3). The pixels with temporal coherence over a chosen threshold are then inverted for time-series of
deformation using singular value decomposition. Finally, a rate map is determined by linear regression of the time-series.

Figure 40: LOS velocity map for track 442, frames 2781-2817 from SBAS processing at IREA a) and University of Miami TSSAR processing b).
If present, continuous GPS can be used as a final correction step to remove orbital phase residual as described in length in CHAPTER 2. TSSAR also contains code to correct for topographic errors and atmospheric phase screen.

7.3.2 SBAS processing summary

The SBAS algorithm is a sophisticated algorithm to produce high spatial and temporal resolution time-series from a set of interferograms [Lanari et al., 2004; Lanari et al., 2007] and has a number of steps not present in TSSAR.

The SBAS algorithm uses SAR data co-registered to a single master before interferogram processing. Co-registration relies mostly on a combination of topography and orbital information with little amplitude cross-correlation. A temporal coherence is computed on the set of wrapped interferograms in order to produce a mask and select pixel with high temporal coherence (typically temporal coherence higher than 0.7). The masked differential interferograms are unwrapped using a 3-dimensional (2D spatial and 1D temporal) phase unwrapping algorithms that uses the entire set of interferograms [Pepe and Lanari, 2006]. After unwrapping, the interferogram are flattened to remove the long wavelength phase contribution resulting from uncertainty in the satellite orbital parameters. Topographic errors are resolved for each pixel by iterative inversion using the altitude of ambiguity relation between phase and perpendicular baselines of the entire set of interferograms.
Figure 41: Curl of the wrapped, unfiltered, interferograms in radar geometry: 970619-970724, 970619-980115 and 970724-980115, from track 442 and frames 2781 to 2871, processed with ROI_PAC, a), and at IREA, b). The master image used for coregistration is 970724, the curl performed on geocoded interferograms shows similar curl residuals. Curl is performed as:

\[
\text{curl} = \phi_{970619-970724}^{\text{wrp}} + \phi_{970724-980115}^{\text{wrp}} - \phi_{970619-980115}^{\text{wrp}}.
\]
After inversion of the interferograms for deformation time-series using singular value decomposition, the time-series are filtered to recover the Atmospheric Phase Screen or APS and deformation rate is computed.

### 7.3.3 Co-registration errors

The processing of the track 442, frames 2781 to 2871 using the ROI_PAC, reveal a residual signal when a curl is performed on a set of three interferogram as shown in Figure 41a. The curl is performed on a trio of wrapped, unfiltered differential interferograms (Topography and flat earth removed) in radar geometry and coregistered to a common master (970724) using $\$INT_SCR/coregist_igram.pl$. coregist_igram.pl is performing an amplitude correlation (using ampcor.pl) on all SLC’s with respect to a common master chosen for its median perpendicular baseline and central date.
Figure 43: Unwrapping errors. Curl of the unwrapped interferograms 970619-970724, 970619-980115 and 970724-980115 processed in a) Miami, and b) IREA. Curl is performed as:

\[ \text{curl} = \phi_{970619-970724}^{\text{unw}} + \phi_{970724-980115}^{\text{unw}} - \phi_{970619-980115}^{\text{unw}} \]
The transformations coefficients between each SLC’s and the master SLC are then used to resample interferogram products into the master SLC’s geometry (using resamp.pl). The interferograms are wrapped therefore contain no phase residuals from unwrapping errors. Our interferograms show significant residuals. The maximum residuals occur where the terrain slopes are largest. Similar tests using only frame 2871 (Figure 42a and Figure 42b) or with interferograms processed by IREA show no significant residuals (Figure 41b). The same results have been found when using geocoded interferograms which rules out problem related to coregist_igram.pl. Possible explanations for this are errors in the calculated transformation between geometries. The apparent correlation with topography indicates that the cause may lie in the transformation of the DEM between its initial geometry and the radar geometry.

7.3.4 Unwrapping errors

Unwrapping errors can be determined by applying a curl similar as in 7.3.3, but on unwrapped interferograms. By applying a curl of the same interferograms as in 7.3.3, we show that our processed interferograms, unwrapped using snaphu, contain significant unwrapping errors (Figure 43). In this figure, the color changes from green to yellow to red correspond to a phase jump of $2\pi$, thus an unwrapping error.

7.3.5 Discussion

The major shortcomings of our InSAR time-series processing chain are inconsistencies amongst interferograms sharing the same SAR data. This
causes a poor recovery of the surface deformation as illustrated by the temporal coherence (Figure 44). These inconsistencies are due primarily to large unwrapping errors but also, in a lesser extent to mis-coregistration that appear to be correlated with topography. This seems to be related to long swath processing as illustrated by Figure 42 and Figure 45. Manual correction, further tuning of snaphu or 3-dimensional unwrapping are options to improve the unwrapping of interferograms. Only the latter will guaranty automated processing of displacement time-series from sets of SAR data. The inconsistencies that I attribute to mis-coregistration, although problematic, have lower amplitude and are possibly randomly distributed in time that if true, would not have an impact on the final rate of displacement.

7.4 Time-Series SAR (TSSAR) processing chain

This chapter presents the University of Miami InSAR laboratory processing chain from the archiving of level 0, L0, SAR data to time-series processing. The steps described from 7.4.1 to 7.4.4 are sub-program included in the master perl script $INT_SCR/process_rsmas.pl which can be run with an input template file. Subsequent parts are included into the matlab-based TSSAR time-series processing package, which is run by the master script tssar.m and a parameter file (here:

$TEMPLATEDIRHOME/ngourmelen/BandRT442ErsD.min).
Figure 44: Temporal coherence using University of Miami processed interferograms, full range a) and above 0.7 b).
Figure 45: LOS velocity for track 442, frame 2799 only. Velocity map shows subsidence in support of mining and agricultural activities.

7.4.1 Data archiving

moveL0.pl is a code that renames SAR data into a consistent and informative file name, and archive the SAR data into the $L0DIR directory. Our naming convention follows the ESA convention with a compressed tar file containing a data file and a leader file.
The name of the tar file contains the satellite name, the string "SAR_RAW_0P",
the date (YYYYMMDD) and time (HHMMSS) of the first line, the date and time
of the last line, the string "ESR", the orbit number and the string ".CEOS.tar.gz",
all separated by underscores similar to:

ER02_SAR_RAW_0P_19990520T183320_19990520T183336_ESR_021338.CEOS.tar.gz

The script moveL0.pl is included into the script getSar.pl, which itself is the first
step of the master script process_rsmas.pl. Specifically, the following has been
tested:

1 - Amended the ERS section of moveL0.pl to accept various input format (ESA,
WINSAR, GEOEARTHSCOPE, CDS, ...).

2 - Archiving of the data from ESA tracks 213 and 442 obtained from the
following sources: Miami archive, WinSAR archive, and Geo-earthscope archive.

**Code:**

```
$INT_SCR/moveL0.pl

Input file:

$TEMPLATEDIRHOME/ngourmelen/BandRT442ErsD.template

Lines 252-342

foreach $file (@filelist) {
    print STDERR "Renaming and moving $file to $trackDir \n";
```
if ($file =~ "zip") {
    ($file_dir) = split(/\..zip/,$file) ;
    `mkdir $file_dir` ;
    `unzip -o $file -d ./$file_dir` ;
}
elsif ($file =~ "tar") {
    ($file_dir) = split(/\..tar/,$file) ;
    `mkdir $file_dir` ;
    `tar -C $file_dir -xvf $file` ;
}
elsif ($file =~ "CEOS.gz") {
    ($file_dir) = split(/\..gz/,$file) ;
    `mkdir $file_dir` ;
    `mv $file ${file_dir}.tar.gz` ;
    `tar -C $file_dir -xzf ${file_dir}.tar.gz` ;
    `mv ${file_dir}/*/" ${file_dir}/.` ;
    `rmdir ${file_dir}/*` ;
}
elsif ($file =~ "CEOS.tar.gz") {
    ($file_dir) = split(/\..tar.gz/,$file) ;
    `mkdir $file_dir` ;
    `tar -C $file_dir -xzf $file` ;
    `mv ${file_dir}/*/" ${file_dir}/.` ;
    `rmdir ${file_dir}/*` ;
}
eelsif ($file =~ "LEA") {
    @dirs = split(/\/,$file) ;
}
$file_dir = @dirs[0];

}

elsif ($file =~ "\.ldr") {
    @dirs = split(/\/, $file);
    $baseDir = @dirs[0];
    @fileBase = split(/\./, $dirs[1]);
    $fileBase = @fileBase[0];
    $frame = getFramenumber("./$file", 'LED'); chomp $frame;
    @frames = split(/\./, $fileBase);
    $frame or $frame = @frames[2];
    $sat = getSatellite("$file", 'LED'); chomp $sat;
    $orbit = sprintf("%05d", getOrbitnumber("$file", 'LED')); chomp $orbit;
    $date = getStarttime("$file", 'Ers', 'LED'); chomp $date;
    $time = substr($date, 8, 6); chomp $time;
    $yymmdd = substr($date, 2, 6); chomp $yymmdd; ## Date
    $track = sprintf("%03d", orbit2track($orbit, $sat, $yymmdd));
    chomp $track;
    $file_dir = "${sat}_${track}_${orbit}_${frame}_${yymmdd}_${time}";
    `mkdir $file_dir`;
    `mv ${baseDir}/${fileBase}.ldr ${file_dir}/LEA_01.001`;
    `mv ${baseDir}/${fileBase}.raw ${file_dir}/DAT_01.001`;
    `mv ${baseDir}/${fileBase}.vdf ${file_dir}/VDF_DAT.001`;
}

$sat = getSatellite("./${file_dir}/LEA_01.001", 'LED'); chomp $sat;
$orbit = sprintf("%05d", getOrbitnumber("./${file_dir}/LEA_01.001", 'LED'));
chomp $orbit ;
$orbitn = sprintf("%06d",$orbit) ;
$frame = getFramenumber("./${file_dir}/VDF_DAT.001","VDF") ;
chomp $frame ;
$frame or $frame = getFramenumber("./${file_dir}/LEA_01.001","LED") ;
chomp $frame ;
$date = getStarttime("./${file_dir}/LEA_01.001","Ers","LED") ;
chomp $date ;
$time = substr($date,8,6) ;
chomp $time ;
$yyyyymmdd = substr($date,0,8) ;
chomp $yyyyymmdd ;
$yymmdd = substr($date,2,6) ;
chomp $yymmdd ;
$track = sprintf("%03d",orbit2track($orbit,$sat,$yymmdd)) ;
chomp $track ;
$hr = substr($time,0,2) ;
$min = substr($time,2,2) ;
$sec = substr($time,4,2) ;
if ($sec+0>7) { $start = $time-8; } else { $start=$time-48; }
$start = sprintf("%06d",$start) ;
if ($sec<52) {$stop=$time+8;} else {$stop = $time+48;}
$stop = sprintf("%06d",$stop) ;
$trackDir = "$L0Dir/$track" ;
unless (-d $trackDir) { mkdir($trackDir) or die "Can't create $trackDir:$!"; }
$folder = "${sat}_${track}_${frame}_${orbit}_${yymmdd}_${time}" ;
Message "Directory name: $folder" ;
$dumzip = "${sat}_${track}_${frame}_${orbit}.zip" ;
$sat=~ s/S/0/ ;
SLC processing

Unpacks all L0 SAR data from the $LODIR directory, determined which are needed based on track, frame, and date inputs in the .template file, and creates appropriate directories in the $SLCDIR directory. It concatenates all frames of a given orbit and converts L0 data into Single Look Complex, SLC, data.

Code:

$INT_SCR/raw2slc.pl

Input file:
7.4.3 Network Selection

From the list of SLC, $INT_SCR/SelectPairs.pl calculates a set of interferograms from Delaunay triangulation of SAR data based on the distribution of two parameters, the perpendicular baseline, \(b_{Perp}\), (minimization of topographical related phase error), and the temporal baseline, \(b_{Temp}\), (minimization of temporal decorrelation), between SAR acquisitions.

**Code:**

$INT_SCR/raw2slc.pl

**Input file:**

$TEMPLATEDIRHOME/ngourmelen/BandRT442ErsD.template

**Keywords:**

```
selectpairsopt.selectMethod = 'Delaunay'  # Triangulation method
selectpairsopt.perpBaseMax = 600        # Maximum perpendicular baseline (m)
selectpairsopt.temporalBaseMin = 0      # Minimum temporal baseline (yrs)
selectpairsopt.temporalBaseMax = 11      # Maximum temporal baseline (yrs)
selectpairsopt.startDate = '970101'     # Earliest date for SAR data (YYMMDD)
selectpairsopt.endDate = '051231'       # Latest date for SAR data (YYMMDD)
```
7.4.4 Interferogram Processing

SLCs are then used to process Interferograms using a modified version of the ROI_PAC package. The bulk of the modifications brought to ROI_PAC by members of the University of Miami InSAR laboratory consist of perl scripts to produce a large number of interferogram in an automated fashion.

**Code:**

```
$INT_SCR/slc2igram.pl
```

**Input file:**

```
$TEMPLATEDIRHOME/ngourmelen/BandRT442ErsD.template
```

7.4.4.1 Quality check

TSSAR possesses two quality check steps intended to identify poor-quality interferograms. These steps should be deactivated for automated processing.

7.4.4.1.1 Interferogram display

This step checks for poor quality interferograms that can have a variety of causes such as processing errors, strong atmospheric signal, etc. It consists of a simple visual check of each interferogram and a user prompt to remove bad interferograms from the list.

**m-file:** *Embedded in tssar.m*

**in .min file:**

```
runningopt.check_interferograms = yes  % Controls
```
7.4.4.1.2  **Network check display**

Displays the interferogram network after 7.4.4.1.1 to ensure that proper connections are there.

*m-file: lgram2baseplot.m*

_in .min file:_

```
displayopt.baselineplot = yes  % Controls
```

7.4.4.2  **Curl on interferograms**

This steps check for consistency of the interferograms. It is performed through a curl of three connected interferograms obtained during the Delaunay triangulation in _SelectPairs.m_. Curl residuals indicate inconsistency in the interferograms processing that can be related to geometric and temporal decorrelation, and to the unwrapping, resampling, coregistration, etc steps. Chapter 2 provides further description, and example residuals are shown in Figure 41 and Figure 43.

*m-file: lgramClosure.m*

_in .min file:_

```
runningopt.check_curl = yes
```

7.4.4.3  **Surface removal**

Intended to remove long wavelength phase residual from each interferograms that may occur due to orbital phase errors. It uses the matlab function _pinv_ and supports linear or quadratic models.
**m-file: RemovePlane.m**

**in .min file:**

```matlab
runningopt.remove_plane = yes % Computes the temporal correlation mask.
inputdataopt.removeplane.surf_type = linear (or 'quadratic')
```

**output file:**

`flat_seeded_interf.mat`

### 7.4.4.4 Masking

Computes a mask by computing the temporal coherence of each pixel. Temporal coherence can be estimated using (3). Temporal coherence is a way to quantify inconsistency in the set of interferograms, the possible causes are discussed in 7.4.4.2. Example of temporal coherence for track 442 using data processed by IREA is shown in Figure 46.

**m-file: UnwrapErrors.m  # Should be renamed ...**

**in .min file:**

```matlab
runningopt.tempcormask = yes % Computes the temporal correlation mask.
inputdataopt.Tcortresh = 0.7 % Lower threshold on the temporal correlation value/pixel.
```

**output file:**

`TempCorrMask.mat` % Temporal coherence mask
TempCorrel_interfK.mat  % Masked Interferograms

7.4.4.5  **Topographic errors**

Error in the digital elevation model results in phase residuals due to improper removal of the topographic phase. Digital elevation model errors are retrieved using the relation between the topographic phase and the perpendicular baseline or altitude of ambiguity. Remove a linear regression between phase and time period covered by interferogram to remove any contribution from deformation. In a subsequent step, a linear regression is performed on the residual phase versus perpendicular baseline.

The results are not convincing and I usually do not run it, a better approach would be to use this code recursively as it is done in the SBAS approach.

*m-file:* `$TSSARHOME/tssarlib/InvertInterferograms_find_dem_err_v3.m`

*in .min file:*

```
runningopt.topo_error = yes  % Computes the temporal correlation mask.
```
Figure 46: Temporal coherence from the data processed at IREA.

```matlab
int = LoadData('/RAID4/ngourmelen/from_raid6/manzo/nevada-6framesErsD_results/... aggstack_unwrappedinterferograms_nevada442_3850x675x148.dat', struct('origin', 'irea'));
[igram_diff, sum_diff, temp_correl] = UnwrapErrors(int);
figure; plot_NaNbackground(temp_correl, struct('background_color', [0.4 0.4 0.4], 'cmap', gray));
```
7.4.4.6  **Time-series inversion**

The interferograms are inverted for rate of displacement between each SAR epochs using Singular Value Decomposition. The rates are then converted to displacement time-series at each SAR epoch with respect to the first acquisition [Berardino et al., 2002].

*m-file:* $TSSARHOME/tssarlib/interferograms2timeseries.m

*output file:* first_time_series.mat

7.4.4.7  **Convert Time-series to motion**

Computes the rate of motion from the time-series using linear regression.

*m-file:* $TSSARHOME/tssarlib/timeseries2motion.m

*in .min file:*

```
```

*output file:* Motion.mat

7.4.4.8  **Orbital Phase Error removal**

This section of tssar.m applies the correction method described in chapter II of this thesis. The code first extracts InSAR displacement time-series at the location the GPS measurement, it does so by finding the InSAR matching the location of the GPS stations, or by averaging a number of coherent InSAR pixel within a
distance of the location of the GPS stations. If the time record does between
the GPS and InSAR time-series does not match, it propagates the GPS record to
the time period of InSAR time-series by applying a linear or seasonal model to
the GPS record. Finally, an orbital phase error is inverted and used to correct the
InSAR time-series.

*m-file: $TSSARHOME/tssarlib/Fit_InSAR_2_GPS_timeseries*

*in .min file:*

```
runningopt.gpsfit = yes
```

*output file:*

```
orbitGPScorrection.mat
```

7.4.4.9  **Atmospheric Phase Screen (APS) removal**

This section applies a spatial low pass Butterworth filter to the phase of each
epoch of the time-series, the output is then processed through a high pass
temporal filter to obtain the APS. The low pass filter is parameterized by a
distance parameter (in pixel unit), the high pass filter is parameterized by a time
parameter (in days).

*m-file: $TSSARHOME/tssarlib/SpatialTemporalFilter.m*

*in .min file:*

```
runningopt.APS = yes

inputdataopt.APS Spatial = 20  % In pixels
```
inputdataopt.APStemporal = 200 \text{ in days} \\

\textit{output file:} \\

APSfilteredTimeseries.mat
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