Vertical Deformation Along the San Andreas Fault

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VERTICAL DEFORMATION ALONG THE SAN ANDREAS FAULT

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VERTICAL DEFORMATION ALONG THE SAN ANDREAS FAULT

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THESIS

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Chapter 1: An introduction to vertical motions along the San Andreas Fault System: 3-D deformation model, geologic, geodetic and tide gauge velocities

1.1 Introduction

There have been numerous M 6+ earthquakes along the San Andreas Fault System (SAFS) (Schwartz et al., 1984) (Figure 1.1) in the historical past. These rupture events have created millions of dollars worth of damage, and have been responsible for multiple lives lost. An improved understanding of the motions and crustal characteristics along the SAFS can lead to better hazard mitigation (Bakun et al., 2005). Horizontal crustal motions of the SAFS have been widely studied and applied to seismic hazard models (WGCEP, 2007), however vertical motions are not often utilized due to their complicated origin and sometimes large uncertainties. This study takes aim at broadening the use of vertical deformation data along the SAFS through an investigation of available datasets (geologic, geodetic, and tide gauge) and modeled motions.

The right lateral strike-slip motion along the SAFS accounts for most of the horizontal motion along the North American/Pacific Plate boundary, but there is also a prominent vertical component of deformation (~3-6 mm/yr of uplift) focused mostly around the “Big Bend” area near Los Angeles, CA (Figure 1.1). The uplift in the “Big Bend” area has formed the San Gabriel Mountains, and is due to the bend in the fault not being parallel to the direction of motion between the Pacific and North American Plates. Time-dependent vertical deformation associated with major earthquake ruptures along the SAFS has been observed (Deng et al., 1998; Pollitz et al., 2001) and is typically associated with postseismic relaxation of the crust due to a redistribution of stresses.

The vertical motion along the fault is recorded in the geologic record, by GPS devices, and in water level and sea level measurements (Turner, 1991; Deng et al., 1998; Roeloffs, 1998; Kuo et al., 2007; Niemi et al., 2008). Vertical motion measurements, combined with 3-D viscoelastic deformation models, can provide unique constraints of the rheological properties and thickness of the crust beneath faults (e.g., Pollitz et al, 2001; Johnson and Segall, 2004; Smith and Sandwell, 2006). Vertical
deformation models have also been developed to show the associations of vertical motions in the crust with sea level rise and fall (e.g. Kuo, 2007). However, no published studies to date have targeted 3-D time-dependent deformation models with geologic, geodetic, sea level, and water level measurements. This study focuses on comparing the geologic and geodetic vertical velocity data sets and examining their respective origins, applying water well log data to correct for anthropogenic effects in the GPS data, and further constraining a crustal deformation model using the geologic, geodetic, and tide gauge deformation records.

![Map of California and the San Andreas Fault System](image)

**Figure 1.1.** Map of California and the San Andreas Fault System, which represents the primary plate boundary between the Pacific and North American tectonic plates. The relative motion between the plates is shown with the arrows. Major fault strands of the San Andreas are labeled, as are major cities.
1.2 The San Andreas Fault System

The San Andreas Fault System (SAFS) represents the boundary between the Pacific and North American plates along the western edge of California (Figure 1.1). The dominant right-lateral strike slip motion along the SAFS, along with the variable geometry of the faults, leads to crustal deformation in all three directions (east/west, north/south, vertical) due to coseismic (earthquake), post-seismic (after earthquake), and interseismic (between earthquakes) motions. An improved understanding of the crustal deformation associated with these three periods of kinematic motion in the earthquake process can lead to more comprehensive earthquake hazard analyses and prevention. Seismometers stationed throughout California measure coseismic processes, however, geologic, geodetic and tide gauge data offer longer temporal observations, measuring slower deformation associated with interseismic and post-seismic motions. 3-D time-dependent crustal deformation models are capable of spanning multiple earthquake cycles, allowing for the analysis of longer time scale processes.

1.3 3-D crustal deformation model

Crustal deformation models are typically developed to estimate spatial and time-dependent rheological characteristics of the crust and mantle. Horizontal deformation recorded by GPS stations along the SAFS has been analyzed using these types of models in numerous studies (e.g., Pollitz et al, 2001; Johnson and Segall, 2004). B. Smith-Konter and D. Sandwell (Smith and Sandwell, 2003, 2004, 2006) also developed a viscoelastic model to analyze the horizontal component of motion along the SAFS, which incorporates parameters for plate thickness, mantle viscosity, fault depth, and slip rate (Figure 1.2). Using realistic earthquake data (i.e. timing, and location of major fault ruptures) over the last 500 years, this model can provide a unique understanding of the spatial and temporal aspects of vertical deformation associated with earthquake distribution (example provided in Figure 1.3). Using vertical motion data sets that sample different time spans, rheological characteristics of the crust and
mantle can be further constrained by comparing the vertical component of the model results using different combinations of elastic plate thickness, density and viscosity.

To explore vertical deformation throughout the earthquake cycle, the Smith and Sandwell model simulates the response of time-dependent fault dislocations embedded in an elastic plate overlying a viscoelastic half-space. It solves the vertical and time deformation components analytically, while the horizontal components are solved in the Fourier transform domain to expedite the computational process over a large grid. The restoring force of gravity is also included to accurately model vertical deformation. The model has been tested with 2-D analytic tests and 2-D analytic Boussinesq tests in order to assure accurate solutions (Smith and Sandwell, 2004). In this model, coseismic slip occurs on prescribed fault segments according to earthquake history. Transient deformation follows each earthquake due to viscoelastic flow in the underlying half-space. The duration of the viscoelastic response depends on the viscosity of the underlying half-space and the elastic plate thickness (Smith and Sandwell, 2004).

Figure 1.2. Sketch of the deformation model used in this study (from Smith and Sandwell, 2004) with variable thickness elastic plate (H), viscosity (η), and density (ρ).

A simulation of the vertical velocity field is provided in Figure 1.3, assuming an elastic plate thickness of 60 km, a half-space viscosity of 1.9 1e19 Pa S, and a density of 3000 kg/m³. Uplift in the regions of the San Bernardino Mountains and Mojave fault segments is due to the restraining bends
(Williams and Richardson, 1991), while subsidence in the Salton trough area is due to the releasing bend (incipient spreading). In addition, there are broad lobate regions, such as the pair seen to the east and west of the Parkfield segment, that are due to the rapid change in locking depth between the locked and creeping sections of the fault. South of Parkfield, the upper ~15 km fault has remained locked since the 1857 Fort Tejon rupture while the lower part is sliding. This introduces a vertical bending moment at the ends of the locked section that flexes the lithosphere and creates the lobate structure. The wavelength of the lobate structure is equal to the flexural wavelength of the lithosphere, which depends mainly on the thickness of the elastic plate. If another event like the 1857 rupture occurs, the moment will be removed and the vertical lobate pattern will relax. So vertical deformation occurs from two processes, the misalignment of the fault with the relative plate motion vector and changes in locking depth among the major fault segments.

Figure 1.3. Vertical velocity model of the southern SAFS (from Smith and Sandwell, 2006 and Smith-Konter et al., 2011), reflecting interseismic, coseismic, & postseismic deformation. Positive vertical velocities represent uplift and negative velocities represent subsidence. Dark solid lines represent fault segments included in this model.
One parameter that has a significant effect on the model’s vertical motion is the elastic plate thickness. The thicker the elastic plate underlying the SAFS, the less vertical motion along the fault (Figure 1.4). Over geologic time, the extent of the plate that responds elastically to stress changes (i.e., from earthquakes) will change. This is an important effect when considering data sets that sample different time periods. For example, Smith and Sandwell (2006), using a model reflecting tectonic loading limited to the past ~10 years (constrained only by modern GPS data, like that in Figure 1.5), estimated the thickness of the elastic plate to be > 60 km. Alternatively, other studies report estimates of elastic thicknesses of 23-46 km based on isostatic rebound from the draining of pluvial lakes and 30 years of leveling data of the Basin and Range (Iwasaki and Matsu’ura, 1982; Nishimura and Thatcher, 2003). Thus an estimate of a thick plate (~60 km) may be an artifact of the relatively short observation period of the data used. Although this value is in agreement with the 40-100 km estimate of Johnson and Segall (2004), the variability of these collective results highlights the need for additional vertical observations of crustal motion including those of geologic offsets and sea level changes, developed over longer time spans (> 20 years).

Figure 1.4. Example of relative vertical deformation model time series as a function of elastic plate thickness, evaluated at a point (coincident with Alameda tide gauge station) along the SAFS in northern California (see Figure 1.1 for location). The different elastic plate thicknesses are shown: 30 km (blue), 50 km (red), and 70 km (green).
1.4 Geologic and GPS vertical velocity data

Recurrence intervals of earthquakes (sampling 1 full earthquake cycle) range from 10s to 100s of years for segments along the SAFS (Working Group on California Earthquake Probabilities (WGCEP), 1995, 2003, 2007). GPS vertical data (Figure 1.5) do not cover long enough time spans (~5-20 years) to be able to measure full earthquake cycles along the SAFS; 100+ year recurrence intervals, in particular, makes GPS data insufficient for measuring interseismic and post-seismic motions along certain fault segments along the SAFS. Alternatively, geologic vertical velocities (Figure 1.6) derived from rocks between 10 Ka and 7 Ma, give vertical measurements of deformation derived from time spans longer than the recurrence intervals of earthquakes along the SAFS. In order to accurately study all three types of earthquake cycle motion, the time gap must be bridged. Thus, a combined study of geologic and geodetic vertical data, complimented by expected vertical deformation using a physical model, is particularly useful for aiding our understanding of short and long-term earthquake cycle deformation. Chapter 2 of this thesis explores this idea in further detail, emphasizing the apparent lack of correlation between geologic and geodetic data in southern California.

![Figure 1.5. Map of the 888 GPS vertical velocities from the EarthScope Plate Boundary Observatory the study region. The colors represent vertical velocities and are saturated at +/- 2 mm/yr. See Figure 1.6 for regional fault labels.](image)

![Figure 1.6. Map of the 1627 geologic vertical velocities from the SCEC Vertical Motion Database (VMD) (Niemi et al 2008) used in this analysis. The colors represent vertical velocities saturated at +/-2 mm/yr.](image)
1.5 Tide gauge data

In addition to geologic vertical rates of deformation, tide gauge data collected along the western coast of North America can provide a glimpse into long-term vertical deformation along the SAFS. Tide gauges measure relative changes in sea level with sub mm precision (Douglas, 1991). Time series of sea level data is relatively variable (Figure 1.7) when compared to model time series (Figure 1.8). This variability is derived from a number of sources including seasonal events (e.g. El Niño, La Niña), storms, flooding, and tectonic events. Tide gauges can record a vertical response of the crust from tectonic events, which is the primary emphasis of Chapter 3 of this thesis. Processing of tide gauge records (details provided in Section 3.2) yields relative crustal motions at the tide gauge stations that we can then compare to relative crustal motions generated by earthquake cycle processes of the SAFS (Figure 1.8). A full comparison of the tide gauge and model time series is illustrated in Chapter 3.

Figure 1.7. Relative sea level (mm) vs. time for the San Francisco tide gauge station (see Figure 1.1 for location).

Figure 1.8. Relative vertical displacement (mm) at the San Francisco station location for the tide gauge record (black line, sea level rise rate of 1.8 mm/yr removed) and three model results with different elastic plate thicknesses (30 km blue line, 50 km red line, 70 km green line).
1.6 Thesis organization

This thesis examines vertical crustal deformation along the SAFS. This comprehensive analysis includes a comparison of the geologic and geodetic vertical velocities in Southern California, used in concert with a 3-D vertical deformation model to constrain rheological characteristics of the crust in Chapter 2. Time-dependent vertical land motions estimated by tide gauge stations are compared to model predicted results in Chapter 3. The overarching conclusions resulting from these steps are provided in Chapter 4. Chapters 2 and 3 are being prepared for publication, so their references are listed separately.

1.7 References


Schwartz, D., Coppersmith, K., (1984), Fault Behavior and Characteristic Earthquakes’


Chapter 2: Investigating vertical motion discrepancies in Southern California using geologic, geodetic and well log data

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Abstract

Geodetic and geologic vertical velocity measurements record uplift and subsidence throughout Southern California that provide, in some regions very different perspectives of vertical deformation. In this study we compare vertical geologic velocities from the SCEC Vertical Motion Database and GPS velocities from the EarthScope Plate Boundary Observatory. Analyzing the relationship between geologic and geodetic vertical data is nontrivial, as these data sets differ in geographic coverage area, spatial resolution, signal source, and associated uncertainties. As these data sets are not spatially co-located, several different interpolation techniques are utilized for optimal analysis of the data. Our major finding is that the geologic and geodetic vertical motions in Southern California are not well correlated, regardless of the technique used to compare the data sets. In particular, we identify significant discrepancies north of San Diego and north and west of Los Angeles. Since anthropogenic effects may contaminate some of the geodetic data signal, a first-order groundwater correction was developed to isolate groundwater deformation recorded in the GPS data. Our results suggest a slight improvement in the correlation between geologic data and GPS data corrected for groundwater deformation, although future work is needed to develop a heterogeneous ground water to vertical motion ratio model. These results also suggest that geologic and GPS data show modest correlation when limiting the data to vertical motions that are dominated by large-scale tectonic motions.

2.1. Introduction

Geologic and geodetic measurements of vertical deformation record localized zones of uplift and subsidence that document critical components of both long and short-period earthquake cycle deformation. Several factors can contribute to vertical crustal motions, such as tectonic motions, glacial
rebound, groundwater pumping and hydrocarbon extraction. In particular, tectonic motions of faults within active plate boundary zones can be responsible for a significant amount of vertical deformation (Wyatt, 1989; Heki, K. 1996; Shimada and Bock, 1998; Soudarin et al., 1999). Understanding how such deformation is measured and reported in geologic and geodetic data sets has important implications as earthquake cycle models that help quantify seismic hazards rely on parameters like fault depth, lithospheric thickness, and mantle viscosity that are sensitive to vertical motions (Smith and Sandwell, 2006). Moreover, as recent studies have suggested a significant discrepancy in geologically and geodetically-determined horizontal slip rates (i.e., Sauber et al., 1994; Miller et al., 2001; Becker et al., 2004; Meade and Hager, 2005; Oskin et al., 2008), it is also important to understand how vertical rates may differ. Investigating the relationship between the geologic and geodetic vertical data however, is relevant, as these data sets often differ in geographic coverage area, spatial resolution, and associated uncertainties. In addition, tectonic signals in GPS data may be contaminated by anthropogenic activities such as groundwater pumping and hydrocarbon extraction.

Geologic and geodetic data in Southern California provide a natural laboratory for assessing possible vertical motion discrepancies. Southern California occupies an active strike-slip boundary between the Pacific and North American tectonic plates, comprising the San Andreas Fault System (SAFS), including the San Andreas, the San Jacinto and the Elsinore faults (Figure 2.1). The SAFS formed 29 Ma when the Farallon plate became completely subducted in the region of present-day California. At 5 Ma, when the triple junction at the northern end of the SAFS jumped east to its current location at Cape Mendocino, the transpressional environment that exists today began. This transpression is caused by the right-handed bend (“Big Bend”, Figure 2.1) along this right-lateral strike-slip fault (Griscom and Jachens, 1989). Currently, the relative motion between the Pacific and North American plates is ~46-50 mm/yr (DeMets et al., 1990, 1994; WGCEP, 1995, 1999), which coupled with the bends in the fault, leads to restraining and releasing environments along the SAFS.
Geologic vertical velocity data capture these tectonic motions in Southern California over long time periods (10 Ka to 7 Ma), but offer sparse spatial coverage. Vertical geologic rates are derived from analysis of rocks in marine terraces, incised river terraces and stratigraphic surfaces, and from thermochronology studies (i.e., Niemi et al., 2008). In contrast, vertical geodetic (i.e., GPS and InSAR) data in California offer ideal spatial coverage but sample only a limited time span (~5-20 years). Large uncertainties (up to 30+ mm/yr) and anthropogenic effects (Watson et al., 2002) have discouraged the use of vertical GPS velocities in the past (Bock et al., 1997). Complicating these data, vertical land motions have been linked to groundwater pumping for over half a century (Shuman and Poland, 1969; Bouwer, 1977; Chi and Reilinger, 1984; Galloway et al., 1998; Bawden et al., 2001). One of the most famous cases of groundwater derived subsidence was reported for the San Joaquin Valley in central California where between 0.3 and 9 meters of subsidence was identified over a 45 year time period (Poland et al., 1975). Hydrocarbon extraction has also been shown to affect land surface elevations (Ostanciaux et al., 2012), possibly impacting GPS uplift/subsidence measurements.

In this study we investigate the relationship between the geologic and geodetic vertical motion data in Southern California. Geologic data are derived from the Southern California Earthquake Center (SCEC) Vertical Motion Database (VMB) (Niemi et al., 2008) and the GPS data are derived from the Earthscope Plate Boundary Observatory (PBO). As these data sets are not spatially co-located, we explore several different interpolation techniques for optimal analysis of the data. We also develop a simplified groundwater correction from regional well log data to investigate the first-order groundwater deformation signal recorded in the geodetic data. Finally, we compare vertical velocities of all data sets to vertical tectonic motions estimated by a 3-D viscoelastic earthquake cycle model.
2.2. Vertical deformation data

2.2.1 Geologic vertical data

In this study, we use geologic vertical velocities from the SCEC VMD (Niemi et al., 2008), which comprises over 1700 data points in Southern California. These velocities range from -7 to 19 mm/yr (Figure 2.A1) with an average uncertainty of 0.3 mm/yr. Of these data, we use 1627 measurements (Figure 2.1) that fall within two standard deviations of the mean to eliminate statistical outliers. These resultant velocities range from -3.5 to 3.6 mm/yr with a mean velocity of -0.02 mm/yr, standard deviation of 0.85 mm/yr, and an average uncertainty of 0.13 mm/yr. The relatively low uncertainties are discussed further in the Appendix. These data are derived from analyses of marine terraces, incised river terraces, thermochronological ages, and stratigraphic surfaces, reflecting motions of rocks ranging in age from 10 Ka to 6.8 Ma. Ages of the rocks are determined, along with their current vertical position relative to their position at deposition, which together provide a vertical velocity estimate.

![Figure 2.1. Map of the 1627 geologic vertical velocities from the SCEC VMD from Niemi et al. (2008) used in this analysis. The colors represent vertical velocities saturated at +/-2 mm/yr.](image-url)
2.2.2 Geodetic vertical data

In this study we also explore a subset of GPS vertical velocities from the Earthscope PBO (T. Herring, pers.com.). These data range from -44 to 88 mm/yr (Figure 2.A2) with an average uncertainty of 1.7 mm/yr. For this analysis, we divide this data set into the approximate geographic area of the geologic data and eliminated outliers by removing data that are outside two standard deviations of the mean, resulting in 888 GPS velocities (Figure 2.2). We also removed the mean from the data (1.47 mm/yr) to eliminate the possibility of a reference frame bias, following the approach of Shen et al. (2011). The resulting velocities range from -9.5 to 8.9 mm/yr with a mean velocity of 0 mm/yr, standard deviation of 2.33 mm/yr, and an average uncertainty of 1.5 mm/yr. We discuss the physical interpretation of these values in context with the SCEC Crustal Motion Model (Shen et al., 2011) in Section 2.5. These data reflect average vertical motions over the past ~20 years associated with both tectonic events and anthropogenic effects such as groundwater pumping and hydrocarbon extraction (Bawden et al., 2001; Watson et al., 2002).

These data sets demonstrate very different vertical motion behavior along the SAFS. For example, the GPS data reflect a combination of negative (subsidence) and positive (uplift) values along the San Andreas fault near the Big Bend, while the geologic data (although limited along the fault) are predominantly positive. The highest (8.9 mm/yr) and lowest (-9.5 mm/yr) geodetic velocities are both located just northwest of the Big Bend region where uplift is expected, indicating a signal that is not solely due to tectonic effects. Similarly, the highest (3.6 mm/yr) and lowest (-3.5 mm/yr) geologic velocities are both located along the Santa Clara River (Figure 2.1) in the vicinity of the Ventura anticline, showing that local scale tectonics are represented in the geologic record. One other noticeable area of consistent uplift is along the San Gorgonio Pass, where the dip of the fault is low enough that thrusting occurs (Yule and Sieh, 2003).
2.3. Data comparison techniques

While qualitative inspection of the vertical geologic and geodetic velocities (Figures 2.1-2.2) reveal obvious variation, we also note several inherent differences that exist between the two data sets. While the GPS data have relatively high uncertainties, they provide a more evenly spaced, broader coverage area than the clustered locations of the geologic data. The geologic data are more numerous but are primarily situated along the coastline and confined to regions west of the SAFS. Thus, as these data sets are not spatially co-located, several different interpolation techniques were tested to optimally compare the distribution of vertical velocities in Southern California. We utilize Generic Mapping Tools (GMT, http://gmt.soest.hawaii.edu) for many of these tasks, as described in detail below.

2.3.1 Block median surface interpolation mapping

We first inspected a smoothed representation of the respective data sets by constructing surface interpolation maps. We employed GMT’s block median function to interpolate the two data sets (Okubo et al., 2004; Talley et al., 2005), which weights each arbitrarily located vertical velocity by its associated uncertainty and calculates a median velocity and position within a specified grid with 22 km cell
spacing. We then use GMT’s surface function to interpolate the data sets over an 11 km spaced grid and
the grdsample function to generate grids with 1 km spacing, spanning an area limited by the geographic
extent of the geologic data (Figure 2.1). We experimented with several alternate block median/surface
grid cell spacing sizes to suppress surfacing artifacts, however visual inspection indicated that this
combination provided the optimal preservation of the two data sets, because of the original spacing of
the geodetic data (~10 km, Wei et al. (2010)).

The results of the surface interpolation algorithm provide continuous (and smoothed) velocity
fields over a comparable geographic area (Figures 2.3 and 2.4). Some of the features in these surface
maps are sharpened or softened compared to the uninterpolated data (Figures 2.1 and 2.2) due to
weighting based on the data uncertainties. We also note that the uninterpolated data (particularly the
geologic data) have many data points that are located very close together and sometimes plot on top of
each other, hence masking some of underlying data. The geologic interpolated velocity field (Figure 2.3)
indicates uplift northwest of Ventura in the vicinity of the Ventura Anticline, and along the SAF in the
San Bernardino Mountains, however limited data exist here; a large zone of subsidence is observed in
the LA Basin and smaller scale subsidence is observed north and west of Riverside. Alternatively, the
GPS interpolated data (Figure 2.4) reveal a very heterogeneous vertical velocity field. Along the SAF,
pockets of local subsidence are interspersed between areas of uplift. Visual comparison with the
geologic interpolation map reveals some common zones of uplift along the San Bernardino Mountains
and northwest of Ventura, and also zones of subsidence north and west of Riverside. The LA Basin, as
observed by the GPS data, shows regions of both subsidence and uplift.

While there are several obvious uplifting and subsiding regions in the surface maps that
correspond to the raw data, we also note several troubling artifacts in regions with poor coverage. For
example, both interpolated maps suggest questionable velocities off the coast of California where
velocity measurements are scarce. Interpolation of the geologic data near Ventura and San Nicholas
Island generate large positive velocities as far south and west as the Channel Islands, however no data exist in the midsection of this region to corroborate these results. A similar behavior is noted in the GPS interpolation map near San Diego where a large uplifting feature is suggested, due a steep gradient introduced by the block median function between a large uplift velocity near Point Loma (southern San Diego), and more distant, small subsidence values. An additional zone of anomalous velocity is noted in the GPS interpolated map east of the SAF (northeast of Bakersfield), where a large negative feature dominates the region due to relative absence of GPS stations here.

To better illustrate the discrepancies between the two data sets, we also computed a residual map (geologic – GPS) (Figure 2.5) of the two interpolated velocity fields. While there is significant variation amongst the two interpolated data sets, the mean difference of these data is surprisingly low (1.4 mm/yr), with residuals ranging from -5.9 to 13.6 mm/yr. We note that features observed in this residual map agree with the interpolated results of Niemi et al. (2008), although their results are presented with a stretched colorscale, which suppresses some of the vertical variation.

From the interpolated geologic and GPS velocity fields (Figures 2.3 and 2.4), we extracted 643 velocities from each surface over an evenly spaced 11 km grid spanning the coverage area shared by both data sets (grey outline in Figure 2.4). We computed a simple linear regression of these data using a least squares approach, which yields an RMS residual of 1.755 mm/yr, a slope of 0.1043, (Figure 2.6) and a computed R correlation value of 0.204. We also used a Gaussian kernel density estimator (Rosenblatt, 1956) to generate a probability density distribution to illustrate 1σ and 2σ confidence levels of the agreement between the two data sets, taking into account average associated uncertainties. The Gaussian kernel density estimator sums Gaussian kernels for each velocity at the gridded data locations from each dataset. The area on the scatter plot where the data lie within one standard deviation (68% confidence; indicated with a white outline) and two standard deviations (95% confidence; indicated by the shaded cloud) are shown in Figure 2.6. The Gaussian kernel dimensions were defined by the
average GPS and geologic uncertainties (1.55 and 0.13 mm/yr respectively), and discretized onto a 300 x 300 point grid. We double the geologic uncertainty (from 0.13 to 0.26) in order to create more continuous density clouds (which only affects the aesthetics of the graph, not the computed statistics).

**Figure 2.3.** Surface interpolation map of the geologic vertical velocity data. The limited scope of the geologic data set is shown here with the interpolation only performed within an ~80,000 km² area bounded by the geologic measurement locations. Grey regions represent no data. Velocities are saturated at +/- 2 mm/yr.

**Figure 2.4.** Surface interpolation map of the GPS vertical velocity data. The full coverage area of the GPS data is shown while the grey outline represents the extent of the geologic velocity interpolation map (Figure 2.3). Grey regions in southwest corner of the map represent no data. Velocities are saturated at +/- 2 mm/yr.

Based on the correlation coefficient alone, it is clear that the geologic and geodetic data have a very poor correlation. The center of the probability density cloud (indicated by the warm colors) is offset in the positive y-direction (geologic data) and slightly in the negative x-direction (GPS data), indicating that the median geologic velocities are more positive than the median GPS velocities. If these two data sets reflected identical deformation signals (in space and time), then the ratio between them would be expected to be 1:1 (indicated by the dashed line), however the true ratio of these data (thick black line) suggests significant disagreement.
2.5. Vertical velocity residual map (geologic – GPS). This map reveals significant variations in the two data sets, with colors saturated at +/- 10 mm/yr.

2.6. Geologic vs. GPS vertical velocities of the 643 velocities extracted from evenly spaced grids (11 km spacing) from both geologic and GPS surface interpolation maps (Figures 2.3-2.4), within the coverage area of the geologic data. A Gaussian kernel density estimator is used to plot the probability density function of the velocities at collocated points in both data sets. The warmer colors show the highest density of velocities, with the perimeter of the shaded cloud indicating the boundary of 95% (2σ) confidence level and the white line outlining the boundary of the 68% (1σ) confidence level. The dashed line represents a 1:1 ratio, or a perfect fit, while the solid black line shows the actual ratio between these data sets.

2.3.2 Masking co-located data

While the surface interpolation approach described in the previous section provides a continuous velocity map of both data sets, the poor correlation estimated by this approach emphasizes its inadequacies. Next we apply a more rigorous masking approach by utilizing velocities confined to co-located points within a 11 km spatial domain. This technique utilizes the surface interpolated gridded
results (Section 2.3.1), however we also use the raw block median results to define a mask representing regions that contain data voids. In this step, instead of interpolating numbers from neighboring cells when values within the cells do not exist, we replace the cells without velocities with NAN (not a number). Applying this process to both data sets, we then construct and apply a cumulative mask of NANs to both surface interpolated grids. This mask effectively disqualifies (or eliminates) velocities of grid cells that are not common to both data sets, leaving only co-located points (or cells) (Figure 2.7). We tested various mask grid cell sizes and determined that 11 km spaced grid cells offer the highest number of co-located cells (16). For example, a grid spacing of 6 km results in 3 co-located blocks, while a grid spacing of 22 km results in 1 co-located block. The 11 km spacing is optimal because it preserves the original spacing of the GPS data (~10 km, Wei et al. 2010) and helps manage the limited geographic extent of the geologic data.

The results of this masking approach clearly demonstrate the limitations of the data sets’ respective geographical locations but also highlight some subtle trends. Of the 16 co-located cells, 8 cells positively correlate (5 cells reflect subsidence in both the geologic and GPS data sets, 3 cells show uplift in both). Proximity to the SAF may be responsible for the positive correlation amongst the uplifting locations (Figure 2.7, cells 3, 9, and 10). Likewise, 4 of the 5 subsiding locations (cells 7, 8, 12 and 13) are closely situated in the LA Basin, a region of well-documented subsidence (Turcotte and McAdoo 1979; Ingersoll and Rumelhart, 1999; Bawden et al., 2001). Alternatively, 7 cells show subsidence/uplift and 1 cell shows uplift/subsidence in the GPS/geologic velocities. Of these negatively correlating regions, cells 2 and 4 (Figure 2.7 and 2.8) are extreme outliers.
Figure 2.7. 16 co-located points resulting from the masking technique. GPS velocities are represented on the left side of each circle and geologic velocities are represented on the right. The numbers labeling the velocity circles correspond to the numbers plotted in Figure 2.8.

Figure 2.8. Geologic vs. GPS vertical velocities of the 16 co-located cells shown in Figure 2.7. A Gaussian kernel density estimator is used to plot probability density function of the velocities at co-located cells in both data sets, as in Figure 2.6.

We further explore a possible correlation between the co-located cells resulting from the masking approach with a simple x-y scatter plot (Figure 2.8) using the same density estimator approach applied to Figure 2.6. A linear regression analysis yields an RMS residual of 3.22 mm/yr with a slope of 0.0807, and a computed R value of 0.152 for these data, indicating that the resulting correlation using this technique is also poor. Based on the R values alone, it is clear that the geologic and geodetic data again have very poor correlation. The probability density cloud (Figure 2.8) is primarily offset in the negative x-direction (GPS data), indicating that this technique primarily sampled co-located blocks with subsiding GPS velocities. After applying this masking approach, the ratio between these two data sets is still far from 1:1 (thick black line), indicating these data sets do not have a strong linear correlation.

2.3.3 Delaunay triangulation

As an additional approach, we utilized the Delaunay triangulation technique (Lee and Schachter, 1980; Chew, 1989) to interpolate average geologic velocities located within triangular grid cells defined by GPS station locations. Using this technique, 1718 triangles cells were constructed, with apices defined by the locations of all GPS station locations (Figure 2.9). Of these cells, 147 contained at least 1
geologic velocity measurement, which we use to define our sample size for this approach (cells with no geologic data were omitted). Average geologic velocities were calculated for each triangle using the available data points confined by each triangle, while average GPS velocities were calculated for each triangle using the velocities of stations that defined the 3 apices of each triangle.

We again explore a possible correlation between the triangulated cells resulting from this approach with a simple x-y scatter plot (Figure 2.10) using the same density estimator approach as in previous figures. For this technique we calculate an RMS residual of 2.07 mm/yr. A linear regression analysis yields a slope of 0.1273, and a computed R value of 0.220 for these data. While the correlation of using this approach is low, it is in the range of the other techniques. The center of the probability density cloud again shows that the median geologic velocity is slightly positive, while the GPS median velocity is slightly negative. This technique shows a y-intercept closer to 0 than the previous two, but with a negative slope far from 1, indicating an ambiguous relationship between the sources of these two data sets.

**Figure 2.9.** Delaunay triangulation schematic. 1718 triangles (white lines) are drawn between the 888 GPS velocity measurements, with black dots showing the locations of the geologic vertical velocities. 147 of the GPS triangles contain geologic data, the comparison of which is shown in Figure 2.10.

None of these three approaches for comparing the data were able to identify even a modest correlation between the GPS and geologic vertical velocities, indicating that there may be an additional
source of deformation influencing the geologic and GPS discrepancy. We further discuss possible origins of discrepancies in Section 2.5, however an obvious source of non-tectonic deformation reflected in the GPS velocities may be the effect of groundwater pumping, which we will address in the next section.

2.4. Simple groundwater correction of GPS data

While geologic measurements should theoretically reflect deformation of rocks pre-dating modern times, the relatively short observational timescale of the GPS record (~20 years, Bock et al. (1997)) could permit anthropogenic activities to influence GPS velocities (vanDam et al., 1994; Munekane, et al., 2004; Jia et al., 2007). Thus to obtain a “clean” GPS signal, deformation arising from groundwater and hydrocarbon pumping should be removed from the data (Bawden et al., 2001; Watson et al., 2002; Burgmann et al., 2006), however this continues to be a complex problem with the full extent of pumping, for example, often unknown. While we cannot estimate the effects of hydrocarbon pumping without site-specific data from industrial partners, we can make a first-order attempt to correct for deformation due to groundwater extraction and recharge. In this study, this correction is made using water well log records and an assumed ground subsidence/groundwater ratio (VGR). Water level changes in aquifers can lead to significant subsidence and uplift of the ground (Shuman and Poland, 1969; Bouwer, 1977; Chi and Reilinger, 1984; Galloway et al., 1998; Bawden et al., 2001).

The California Department of Water Resources (http://www.water.ca.gov) provides water level data for tens of thousands of wells throughout the state. Most of these water level records contain a single measurement, so time-series data are not obtainable. However, several tens of wells have water level records spanning 1 to over 80 years. These time-series provide us with rates of change in water level over time, which is simply the amplitude of the difference in water level, divided by the number of years the record spans. After gathering these rates from 107 wells throughout Southern California, we used 103 (Figure 2.11a) that were within two standard deviations of the mean for this analysis (the four
points we excluded had rates exceeding +2.5 m/yr and −5 m/yr). Agriculturally dominant areas such as the Salton Trough show declines in groundwater levels, while areas along the coast northwest of Los Angeles, show increases in groundwater levels, possibly due to artificial recharge (Figure 2.11a) (Clark et al., 2005).

When an aquifer is dewatered, the pressure provided from the water is lost. This absence of pore water pressure can lead to compaction of sediments/grains. This compaction may lead to ground subsidence based on various hydrologic characteristics of the underlying aquifer such as compressibility, particle size/shape and geochemistry of the pore water (Sun et al., 1999). In addition to lowering water tables leading to subsidence, an increase in water level can have the opposite effect, causing uplift (Sun et al., 1999). Several groups have modeled aquifer-specific subsidence due to groundwater recharge and extraction (i.e. Galloway et al., (1998), Sun et al., (1999), Amelung et al., (1999), and Bawden et al., (2001)). These studies suggest an empirically derived ratio of vertical land motion due to groundwater (VGR) ranging from 0.002 – 0.015. In Southern California, Galloway et al., (1998) and Bawden et al., (2001) observed vertical land motion due to groundwater pumping over the full range of these values from basins in and around Los Angeles. This basin-to-basin variation creates the need for using caution when lumping all of Southern California into one large aquifer basin, as the VGR value is highly dependent upon the aquifer being studied, however, compiling a heterogeneous model of VGR values for Southern California is a tedious task. We are not aware of any such model at the scale and resolution of this analysis, thus we make a very simplifying assumption that the VGR value is homogeneous in Southern California. This assumes that all rocks in Southern California have the same response to ground water changes, which is an obvious source of error, however it is beyond the scope of this study to develop a heterogeneous VGR model.

We convert groundwater level rates (WL, m/yr) to subsidence (SUB, mm/yr) using the following equation:
\[ \text{SUB} = \text{WL} \times \text{VGR}. \] (1)

We note that this equation preserves the sign of the groundwater level rates (uplift (+), subsidence (-)), but scales the effective land deformation by the VGR. As an initial test, we use an average VGR value adopted from published studies (0.006), however we also evaluate models spanning the range of published VGR values (0.002 – 0.015) to test the sensitivity of our results to this parameter. Using Equation 1, aquifer water level rates (Figure 2.11a) are converted to effective rates of subsidence and uplift (Figure 2.11b). We then constructed an interpolated map of these rates using GMT’s block median and surface tools (Figure 2.11c), using the same technique as described in Section 2.3.1.

As our goal is to correct the GPS vertical velocity field for groundwater-induced deformation, we subtract the interpolated groundwater deformation rates (Figure 2.11c) from velocity measurements at each GPS station location (Figure 2.2) located within the area of the groundwater correction. The corrected GPS velocities are shown in Figure 2.A3 and the resulting interpolated GPS vertical velocity field with the groundwater correction is shown in Figure 2.12. While the original GPS velocities contained 424 out of 888 measurements with negative velocities, the groundwater correction modifies this number to 345 and increases the overall mean of the GPS data from 0 mm/yr to 0.53 mm/yr. This effect is also apparent in Figure 2.12 where many of the areas of uplift in the original data (Figure 2.4) have increased in magnitude and area in the groundwater corrected GPS velocities. The impact of these results on the geologic/GPS correlation analyses is discussed further in Section 2.5. We emphasize that this groundwater correction relies on numerous hydrogeologic assumptions, and requires further development to be applied in different locations and over different geographic scales.
2.5. Discussion

Using three different quantitative techniques, we generated maps and regression statistics that all suggest a consistent and weak agreement between vertical GPS and geologic data. Before discussing these results in detail, we first note prominent key similarities between the geologic and GPS data sets (Figures 2.1, 2.2, 2.5 and 2.7). The area just west of Los Angeles shows significant subsidence in the geologic data, largely due to measurements made along the Santa Clara River (Figures 2.1, 2.3). These
measurements are amongst the largest subsidence values in the geologic data set, with values as low as -3 mm/yr, with GPS velocities in the same area ranging from -2 to 4 mm/yr (Figures 2.2, 2.4). Further to the northwest, geologic velocities along the Ventura Anticline exceed 3 mm/yr of geologic uplift (Figures 2.1, 2.3), and GPS velocities also show uplifting motions ~ 2-3 mm/yr (Figure 2.2, 2.4) in the same region. The residual in this region is mostly positive (Figure 2.5), reflecting slightly higher velocities of geologic uplift, and also reflecting the short wavelength variation in the interpolated GPS velocity map. An additional area where the data appear to agree slightly (indicated by light yellows and blues in Figure 2.5) is northeast of Los Angeles, along the SAF. Blocks 3, 9, 10 (Figures 2.7 and 2.8) also show the most agreement of uplift between the geologic and GPS data. In particular, cell 10 (located right along the SAF at the San Bernardino Mountains) shows nearly 1:1 positive agreement between the two data sets (2.7 mm/yr from geology and 2.3 mm/yr from the GPS data).

A primary quantitative observation we can make from the regression analysis of the geologic and geodetic (original data, not corrected for groundwater deformation) data is that their correlation is quite weak (Figures 2.6, 2.8 and 2.10). First, this analysis shows that the two methods with larger sample sizes (block median surface interpolation and triangulation techniques) have a more concentrated percentage of data near the x-y origin. In addition, the RMS residuals for both these methods are near, or below 2 mm/yr, while the RMS residual for the masking approach is over 3 mm/yr (this not a surprise given the small sample size). The correlation coefficients for the masking and block median surface interpolation methods are very low (0.15 – 0.2 respectively) while the triangulation technique’s correlation is negative. These statistics reveal a consistent lack of correlation between geologic and GPS datasets, regardless of the processing technique. Of the three techniques, the block median surface interpolation approach provides the smallest RMS residual and highest correlation coefficient when comparing the geologic and GPS data (not accounting for groundwater deformation).
Each processing technique samples the data in a different way and over different geographic extents, which lead to variations in regression statistics. The blockmedian surface interpolation comparison method has the largest sample size (643 points), but this comparison is fairly crude, as it does not weight regions with a higher density of samples (like the coastlines) any differently than regions of sparse data. The masking technique improves upon this inadequacy by limiting the comparison of points to 11 km square blocks that contain both a geologic and a GPS observation. While this technique limits the sample size to only 16 points (common blocks), it provides a quantitative approach for treating the spatial resolution variability between the two data sets. The triangulation technique also limits the comparison to co-located points, but through triangular cells of variable sizes. For small triangular cells (representing a dense GPS array), this approach is ideal as averaging of velocities takes place over small spatial regions. For large triangular cells (representing sparsely located GPS stations, like the distances separating island and mainland station locations), however, this is less than ideal as averaging of GPS data at such far distances may yield inconsistent results.

Our major finding from this study is that geologically and geodetically derived vertical motions in Southern California are not strongly correlated. This observed discrepancy could arise from several factors beyond techniques used to analyze the data. First, the geologic dataset is incomplete, as data are primarily limited to geographic regions west of the SAFS. Second, unique crustal relaxation time scales are reflected in the geologic and geodetic data, which could play an important role in how vertical velocities are recorded. Third, the GPS data are contaminated by an unknown amount of anthropogenic effects. Fourth, uncertainties in the GPS reference frame could propagate errors into reported vertical velocities. Next we assess the roles of each of these issues.

2.5.1 Geodetic and geologic data discrepancies: Space and time

While the geologic and geodetic vertical data discrepancies may suggest relevant tectonic differences, they may also be due to spatial limitations of the data sets. The geologic data are confined
to geologic units that contain uplift markers (e.g. marine terraces, river terraces, outcrops) (Neimi et al., 2008)). In the case of the SCEV VMD, the geologic velocities are primarily confined to a region west of the SAF, and largely cluster near the coastline, with 1627 measurements spanning an ~ 80,000 km² region. Alternatively, the EarthScope PBO array is roughly evenly spaced (~ 10 km resolution, Wei et al., 2011), with 888 measurements spanning a ~ 300,000 km² region. Surface interpolation and triangulation averaging of such a data set is much easier to justify than that of the geologic vertical motion data set, as discrepancies may arise because of erroneous extrapolation over large regions of no geologic data.

Discrepancies may also be related to different timescales of crustal and topographic loading. The geologic data, in theory, account for tens of thousands to millions of years of motion; GPS data only sample the last ~20 years of deformation, which may not provide a long enough time-series to obtain well-resolved (and non-transient) tectonic rates. For example, the elastically strong portion of the crust and mantle that is responsible for supporting topographic loads typically achieves isostatic equilibrium in 1 – 10 Ma (Nishimura and Thatcher, 2003). Observations that sample this “geologically long” time period may result in lower velocities than observations of stress relaxation over much shorter times. A similar relationship is observed for thermally activated viscoelastic processes (Watts, 2001), where loading ages of, for example, 2000 years and 1 Myr, yield reduced effective plate thickness of approximately factors of 2 and 10, respectively. Thus higher GPS velocities, in some regions, may be an artifact of the relatively short observation period of the data. The impact of earthquake events on GPS velocities may also affect velocities observed by GPS stations (Oskin et al., 2007). For the EarthScope PBO dataset, it appears that observations were omitted for the period following each earthquake in which non-linear motion was observed in station time series; these periods range from 2.2 – 3.5 years for significant earthquakes over the past 15 years (Shen et al., 2011). Thus no coseismic/postseismic model parameters were used to treat earthquake deformation in this GPS dataset.
In addition, error estimates in both geologic and GPS processing techniques could add to the differences between the data sets. Acquisition and estimation of geologic velocities includes numerous steps susceptible to error, such as paleoelevations and age dating of the rocks (Niemi et al., 2008). GPS velocity measurements may contain errors derived from differing antenna, residual variances, and troposphere time delays, among others (Shen et al., 2011). GPS observations with short time series may have larger vertical rate errors than those with longer time series, due to the inability to realize potential inconsistencies or outliers with a short sample size (Shen et al., 2011). In addition, GPS velocities may include anthropogenic effects such as groundwater and hydrocarbon pumping, in addition to the tectonic signals they record.

2.5.2 Impact of groundwater

To further investigate potential discrepancies between observed geologic and GPS vertical motion, we explore the impact of groundwater deformation on GPS vertical velocities in Southern California (Bawden et al., 2001; Watson et al., 2002). Due to the limited groundwater data and models available, we have made several simplifying assumptions to provide an estimate of groundwater deformation rates, which we use to “correct” the GPS vertical velocities, as most of the groundwater rates used in this correction were derived from the past decade, which is consistent with the timescale of the GPS observations. The vertical land motion to groundwater level ratios (VGR) we explore were derived empirically from results not specific to Southern California, but from studies spanning the continental U.S. While the unjustified assumption that all rocks in California respond homogeneously to such a ratio provides an expedited process for determining a groundwater deformation correction, it does not take into account the true variability of important hydrologic parameters like compressibility, particle size and shape, and the geochemistry of pore water (Sun et al., 1999). These variable parameters effect ground subsidence ratios of the numerous rock types in aquifers across Southern
We emphasize that this correction is not designed to be used on small-scale projects, but provides a broad, first-order estimate for future studies to build upon.

![Figure 2.13. (a) Geologic vs. geodetic vertical velocities of the 643 velocities extracted from evenly spaced grids using the block median surface interpolation approach (as in Figure 2.6), but now with a groundwater deformation correction applied to GPS velocities. (b) Comparison of the 16 co-located blocks, resulting from the masking approach (as in Figure 2.8), but now with a groundwater correction applied to the GPS velocities. The numbers labeling the points correspond to in the collocated blocks shown in Figure 2.7. (c) Comparison of the 147 velocities resulting from the Delaunay triangulation technique (as in Figure 2.10, but now with a groundwater deformation correction applied to the GPS velocities.)](image)

The results of the GPS groundwater correction can be seen qualitatively (Figures 2.2, 2.4, 2.12, and 2.13), and quantitatively (Figures 2.6, 2.8, 2.10, and 2.13). The statistics for the regression analyses, both before and after the groundwater correction, are provided in Table 2.1. These results indicate that a small improvement to the correlation between the geologic and GPS data can be made if groundwater deformation is accounted for in GPS velocities (highlighted cells). When adopting a VGR of 0.002, both the masking and triangulation technique show reduced RMS residual and an increased correlation coefficient. VGR of values up to 0.006 also provide an improved correlation and decreased
RMS residual for the masking approach, suggesting that the specific regions sampled by this technique are more responsive to groundwater variations. Results for the block median surface interpolation technique, however, do not improve for any VGR (values as low as 0.008 were tested, which provide statistics equivalent to the original data results), suggesting that a homogeneous VGR applied to a large region like Southern California (best sampled by the block median analysis) is not an ideal approach. Based on these results, we are cautiously optimistic that a first-order groundwater correction can eliminate a portion of the anthropogenic effects in the GPS data, however a heterogeneous VGR model should be used to inspect smaller scale deformation due to changing groundwater levels. In the future, this correction could be applied to eliminate the groundwater affects detected by InSAR (Lu and Danskin, 2001).

### Table 2.1

<table>
<thead>
<tr>
<th>Processing Technique</th>
<th>Original data</th>
<th>Corrected data VGR=0.006</th>
<th>Corrected data VGR=0.002</th>
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<td>RMS residual</td>
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<td>Triangulation</td>
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<td>-0.13</td>
<td>-0.09</td>
<td>2.17</td>
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#### 2.5.3 Hydrocarbon effects

It is important to also acknowledge the possible impact of hydrocarbon withdrawl on a study like this. The state of California is the fourth largest producer of oil in the United States (Sheridan, 2006). Hydrocarbon extraction and associated fluid injection can lead to land subsidence/uplift similar to groundwater pumping (Bawden et al., 2001). These effects are likely to occur in geographic areas that are most highly pumped for oil in the state: the counties of Kern, San Luis Obispo, Los Angeles, and Ventura, among others (Sheridan, 2006). These counties all reside within our study area, so the possibility of additional deformation due to hydrocarbon withdrawl activity may exist in the GPS vertical data. While not quantified here, we note that the potential presence of hydrocarbon pumping
has been shown to contribute to deformation depending on the porosity of the sediment being pumped (Poland and Davis, 1969).

### 2.5.4 Comparison of PBO and CMM4 GPS data

GPS reference frame issues are common when using vertical deformation velocities. Shen et al., (2011) removed the mean vertical velocity from their SCEC Crustal Motion Model (CMM4) in order to eliminate any possible reference frame bias. The resultant velocities are then relative to local motions (like that of the SAFS), rather than to the Stable North American Reference Frame (SNARF). The Earthscope PBO GPS velocities we use in this study are also relative to the SNARF, so we used a similar approach as Shen et al. (2011) to eliminate reference frame bias in the vertical GPS velocities. Originally, PBO vertical data reflected an abundance of negative velocities, but after removing the mean, a classic bell curve distribution centered at 0 mm/yr was represented by the data. To investigate additional variations between the Earthscope PBO data and the SCEC CMM4 data, we compared vertical measurements at 379 stations common to both data sets (Figure 2.14, no groundwater correction applied). This map shows the difference between the two data sets (PBO – SCEC), with positive values indicating higher PBO velocities, and negative values indicating lower PBO velocities.

![Figure 2.14. GPS velocity difference (SCEC CMM4 – EarthScope PBO) in mm/yr. The positive values represent higher velocities in the PBO data, as negative values represent lower velocities in the PBO data.](image)
While no systematic pattern is evident, the greatest outliers are centered near Los Angeles, were residual extrema are both located. Shen et al. (2011) provided similar results as auxiliary material (demeaned), identifying the largest variations near Parkfield, Long Beach, Ventura, and east of the San Gabriel mountains. In particular, the cluster of residual velocities at Parkfield may be due to effects of the 2004 Parkfield earthquake, which would be present in the PBO velocities (Shen et al., 2011). We also note some broad similarities between the geologic residuals illustrated in Figure 2.5. Future work will be aimed at investigating how the SCEC CMM4 vertical velocity data compare to geologic data of the SCEC VMD.

2.5.5 SAFS earthquake cycle vertical velocity model

To investigate large-scale deformation in Southern California arising from earthquake cycle motions of the SAFS, ideally represented by both geologic and geodetic data sets, we compute vertical velocity estimates from a 3-D viscoelastic earthquake cycle deformation model (Smith and Sandwell, 2003; 2004; 2006). This semi-analytic model simulates the response of time-dependent dislocations embedded in an elastic plate overlying a viscoelastic half-space, where the restoring force of gravity is also included to accurately model vertical deformation. Interseismic deep slip (prescribed by geological slip rates, WGCEP (1995; 2007) below locked fault patches generates the first-order orientation and magnitude of the velocity field. Coseismic slip is prescribed along active fault segments throughout historical (the last ~200 years) and prehistorical (1000 A.D.+ ) times (e.g., Grant and Lettis, 2002). Transient deformation follows each earthquake due to viscoelastic flow in the underlying half-space; the duration of the viscoelastic response, characterized by the Maxwell time, depends on the viscosity of the underlying half-space and the elastic plate thickness (see Smith and Sandwell (2004) for additional details) The model is purely kinematic in that apparent locking depths, elastic plate thickness, and mantle viscosity were adjusted to match GPS horizontal velocity measurements to an uncertainty of less
than 2 mm/yr (Smith and Sandwell, 2006). Model parameters (i.e., slip rates and locking depths for fault segments in this current model) are provided by Smith-Konter and Sandwell (2009) and Smith-Konter et al. (2011).

An example of the present-day vertical velocity field from this model is shown in Figure 2.15. Uplift in the regions of the San Bernardino Mountains and San Gabriel Mountains (Mojave segment) is due to the restraining bends (Williams and Richardson, 1991), while subsidence in the Salton trough area is due to several small releasing bends (incipient spreading). In addition, there are long wavelength low magnitude lobate regions, such as the negative regions in the north and south quadrants, and the positive regions in the east and west quadrants, of Figure 2.15. These features are due to the rapid change in locking depth between the locked and creeping (or shallowly locked) sections of the fault system at the present-day stage of the earthquake cycle. South of Parkfield, the upper ~15 km fault has remained locked since the 1857 Fort Tejon rupture while the lower part is sliding. This introduces a vertical bending moment at the ends of the locked section that flexes the lithosphere and creates the lobate structure. If another event like the 1857 rupture occurs, the moment will be removed and the vertical lobate pattern will relax over the Maxwell timescale. These types of vertical deformations will occur whenever there is a transition in locking depth that occurs over a distance shorter than a ~1/4 flexural wavelength or ~80 km in this model. So vertical deformation occurs from two processes, the misalignment of the fault with the relative plate motion vector and changes in locking depth among the major fault segments.

In short, this model provides a first-order representation of large-scale vertical deformation resulting from only the dominant strike-slip component of motion along the SAFS. While the emphasis of this paper is not to perform a series of rigorous modeling exercises to fit the geologic or geodetic data (we reserve this work for a companion paper in preparation), we can use the model to illustrate key tectonic deformation processes and possibly discriminate against questionable vertical GPS velocities.
that might be significantly contaminated by anthropogenic effects. Keeping all other parameters constant, we systematically adjust the elastic plate thickness \( (H = 30\text{-}90 \text{ km}) \) and viscosity of the model \( (\eta = 1\times 10^{18}\text{ to } 3\times 10^{19} \text{ Pa s}) \) to explore the sensitivity of model to these parameters. We find that both geologic and geodetic data sets are best replicated using a thick elastic plate \( (> 60 \text{ km}) \) and a mantle viscosity of \( 1 \times 10^{19} \text{ Pa s} \), consistent with previous horizontal GPS modeling results (i.e., Smith and Sandwell, 2006). Based on these results, we use a model of \( H = 60 \text{ km} \) and \( \eta = 1\times 10^{19} \text{ Pa s} \) to guide the remaining points of this discussion.

![Model velocities 60 km](image)

**Figure 2.15.** Vertical velocity model of the southern SAFS (from Smith and Sandwell, 2006 and Smith-Kontner et al., 2011), reflecting interseismic, coseismic, & postseismic deformation over the past 1000 years. Positive vertical velocities represent uplift and negative velocities represent subsidence. Dark solid lines represent fault segments included in this model.

While there are several approaches we could take to compare model and data vertical motions, our first attempt is to classify the vertical data according to whether each measurement positively correlates with the model. This approach filters the data in a way that omits inversely correlated data, where model and data have reversed sign. The methodology behind this approach is that the velocity magnitude and spatial extent of key features in the model will decrease or increase depending on the primary input parameters, however the present-day model velocities (as reflected in Figure 2.15) will never reverse sign; a restraining bend will always show some magnitude of uplift (positive) and a releasing bend will always show some magnitude of subsidence (negative). Earthquake cycle effects,
however, will alter vertical long-wavelength features (as described above), however these are likely to be second-order features and will not vary for the present-day epoch model.

To compare the model with positively correlated data, we extract model velocity values at each geologic and GPS measurement location (Figures 2.1 and 2.2). We find that 60% (973 out of 1627) of geologic velocities positively correlate with the model, while 59% (526 out of 888) of GPS velocities positively correlate with the model. After applying the groundwater correction to the GPS data, the number of positively correlating GPS velocities decreases to 58% (517 out of 888) over the whole region of the GPS data. This slight decrease in correlation does not signal failure of the groundwater correction, as it is statistically insignificant, as only 9 corrected GPS velocities change sign of motion. We also note the surprisingly high number of geologic velocities that positively correlate with the model; the model does not account for deformation from subsidiary faults in the Los Angeles basin nor along the Pacific coastline, where the majority of the geologic measurements are derived.

We can extend this analysis one step further by investigating the correlation of the geologic and GPS data by using the model as a “tectonic filter”. We extract geologic and GPS velocities from the block median surface interpolation maps of Figures 2.3 and 2.4, using the same 11 km spaced grid for the extraction as used in Figure 2.6. For this filtered analysis however, we only compare geologic and GPS data that positively correlate with the modeled vertical estimates. These results are provided in Figure 2.16. For this comparison, we calculate an RMS residual of 1.5 mm/yr. A linear regression analysis yields a slope of 0.22 and an R correlation coefficient of 0.45. The correlation and RMS residual between these points is significantly better than the other analyses attempted, although this expected given the model filter that we apply. These results suggest that geologic and GPS data show modest correlation when limiting the data to vertical motions that are dominated by large-scale tectonic motions.
2.6. Conclusions

The goal of this study is to better understand the vertical motions in Southern California as observed by geologic and geodetic data. While the number of available data points for comparing the geologic and geodetic vertical velocity data largely depends on the chosen comparison technique, all methods suggest that the relationship between the data sets is not 1:1, and the correlation is poor. This lack of correlation could arise from several factors. First, the geologic dataset is incomplete, limited by the geographic coverage of measurements. Second, unique crustal relaxation time scales are reflected in the geologic and geodetic data, which could affect the recorded velocities. Third, the GPS data are contaminated by an unknown amount of anthropogenic effects. Fourth, uncertainties in the GPS reference frame bias could propagate errors into reported vertical velocities (Shen et al., 2011; T. Herring, per.comm.).

The groundwater correction developed, while rudimentary, provides a filter for some of the anthropogenic effects in the vertical GPS data in Southern California. We intend to refine the
groundwater correction so that it can be used to improve GPS and InSAR vertical velocities over multiple geographic scales with greater confidence. One intention of this simple study is to encourage other groups to pursue more elaborate groundwater corrections in the future. To further filter anthropogenic effects from the GPS data, a hydrocarbon correction could be pursued with pumping data from oil companies.

We will continue exploring the vertical motion discrepancies of Southern California using the SCEC CMM4 vertical velocity dataset and regional tide gauge records along the California coast. In tandem, the deployment of geodetic arrays in locations complimentary to existing geologic observations would further aid this analysis. A more complete geologic dataset, particularly spanning active faults of the SAFS, could significantly advance the comparison of geologic and GPS vertical motion velocities, ideally including data from more marine and river terraces, in addition to other paleoaltimeters, such as basalt paleoaltimetry (Oskin et al., 2007).

Appendix

Here we compare histograms of the geologic (SCEC VMD) and GPS (EarthScope PBO) vertical velocity data sets. Figure 2.A1 shows the range of geologic velocities. Outliers exceeding -3.5 and 3.6 m/yr were omitted from this study. 47% (817/1726) of the geologic data points have negative velocities. Figure A2 shows a large range of velocities in the original PBO GPS data (blue), with outliers exceeding -9.6 and 8.9 mm/yr omitted, leaving 78% (1115/1423) GPS data points have negative velocities. Shen et al. (2011) discuss a similar behavior in the SCEC CMM4 vertical velocity data, which they attribute to a reference frame bias (See Section 2.5.4). To eliminate this bias, we subtract the mean velocity from the PBO data to arrive at the mean removed velocities (yellow). This process leaves us with 48% (424/888) of the GPS measurements within the study area having negative values.
Figure. 2.A1. Distribution of the geologic vertical velocities (bin size 0.5 mm/yr).

Figure. A2. Distribution of the GPS vertical velocities (bin size 0.5 mm/yr). The raw GPS data is shown in purple, with the mean removed data set shown in yellow. Notice the right (positive shift) in the values of the mean removed data set.

These data sets show the original raw data for the geographic study region 32-36° N, 115-121° W. We removed data points that were outside two standard deviations of the mean in order to provide a smoother (more regional) perspective on the vertical velocities in Southern California. The geographic
and statistical trimming cut the GPS data from 1423 points to 888. The geologic data was trimmed from 1726 to 1627 points.

We also note that the geologic data have relatively low uncertainties due to the large uplift values (e.g. 100-2000+ meter offsets) (M. Oskin, pers. comm.). These rocks range between 10 Ka and 6.8 Ma, and their vertical positions have little associated uncertainty, so when converted into rates of mm/yr, the uncertainties on the rates of uplift or subsidence become relatively small (0.13 mm/yr on average).

The GPS vertical velocities, after applying a groundwater correction with an assumed VGR of 0.002 (see Section 2.4), are shown in Figure 2.A3. A table of these velocities is also available (Table 2.A1).

Figure 2.A3. GPS vertical velocities corrected for groundwater effects.
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Chapter 3: Pacific coast tide gauges: Where they are, how they record, and what they tell us about vertical deformation along the San Andreas Fault System

3.1 Introduction

Tide gauge records have been shown to measure relative crustal motions (Larsen et al., 2003; Melini et al., 2004). The continuity and accuracy (< 0.1 mm/yr error) make this data ideal for modeling vertical motions along the San Andreas Fault System (SAFS). After processing, these records offer long-term (30-100+ years) observations of the vertical deformation along the California coastline (Figure 3.1). These relative motions can be compared to 3-D crustal deformation models in order to constrain rheological characteristics of the crust.

![Figure 3.1. Locations along the Pacific coast of the 9 tide gauge stations used in this study.](image-url)
3.2 Data Processing

Fourteen tide gauge stations record sea level measurements along the Pacific coast in our study region according to the Permanent Service for Mean Sea Level (PSMSL) (Table 1). Of these stations, we use 9 records acquired from the PSMSL (http://www.psmsl.org) from Point Reyes, CA in the north, to Ensenada, Mexico in the south to be used in this study (Figure 3.1) [These data sets represent a range of time periods of observations, between 3 years (San Mateo station) and 166 years (San Francisco station) of data]. The tide gauge stations we omitted for this study were due to either a short recording time, or numerous, large gaps in measurements in either the monthly or annual sea level measurements (e.g. San Mateo, Rincon Island stations). Of the tide gauge records used in this study, there are some with small gaps in their annual sea level measurements (Table 3.1). For these short time gaps, we augmented the time series with a 2-year running average that brackets each measurement gap. This process maintains the overall trend of the sea level measurements, which is the critical piece of information that we use to compare with our model time series results.

After the data gaps were filled, the 9 tide gauge annual average sea level records (in mm) were adjusted for both the average rise in global sea level (~1.8 mm/yr) (Douglas et al., 2001) and the Global Isostatic Adjustment (GIA) estimate for each station (Table 3.1). We also computed a 5-year running mean for each data point to smooth out yearly weather disturbances (Figure 3.2a). This running mean removal smoothes storm and weather effects, while, like the gap filling process, not effecting the overall trend of the data. Averaging sea level records can eliminate oceanographic “noise”, while maintaining rapid crustal motions (Larsen et al., 2003, Melini et al., 2004).
Table 3.1. Showing the coverage years, gaps, and Global Isostatic Adjustment (GIA) for each station along the California coast. Stars indicate stations not used in this study due to short, or gapped coverage.

<table>
<thead>
<tr>
<th>Station</th>
<th>Coverage years</th>
<th>Year gaps</th>
<th>GIA (mm/yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Point Reyes</td>
<td>1975-2010</td>
<td>1992</td>
<td>0.18</td>
</tr>
<tr>
<td>San Francisco</td>
<td>1854-2010</td>
<td></td>
<td>0.26</td>
</tr>
<tr>
<td>Alameda</td>
<td>1939-2010</td>
<td>1988, 1996</td>
<td>0.26</td>
</tr>
<tr>
<td>San Mateo*</td>
<td>1985-1988</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Monterey</td>
<td>1973-2010</td>
<td></td>
<td>0.11</td>
</tr>
<tr>
<td>Santa Monica*</td>
<td>1933-2010</td>
<td>1966-73, 1983, 1999</td>
<td></td>
</tr>
<tr>
<td>Alamitos Bay*</td>
<td>1953-1965</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Los Angeles</td>
<td>1923-2010</td>
<td>1979, 1999</td>
<td>-0.01</td>
</tr>
<tr>
<td>Newport Bay</td>
<td>1995-1990</td>
<td></td>
<td>-0.01</td>
</tr>
<tr>
<td>San Diego</td>
<td>1906-2010</td>
<td>1926, 1962, 1999</td>
<td>-0.01</td>
</tr>
<tr>
<td>Ensenada</td>
<td>1956-1990</td>
<td>1961</td>
<td>0.0</td>
</tr>
</tbody>
</table>

Figure 3.2. (a) Relative sea level observed by the San Francisco station before processing (blue line), after 5-year running mean removal (red line), and after removal of annual sea level rise and GIA (black line). Notice how the processing smoothes the annual sea level measurements, but maintains the overall trend of the data. (b) Relative sea level at the San Diego station raw (blue line), after removing 5-year running mean (red line) and post GIA (black line). (c) Land motion at the San Francisco tide gauge station relative to the San Diego station.
Major ocean-climate signals, such as the interannual El Niño related signal, are considered regional features of the Pacific. To remove the oceanographic effects from the sea level record, we assume oceanographic effects are common to all California stations. Then to isolate the tectonic components, we subtract a reference sea level record (San Diego station) from all the other stations. The San Diego station is used as a reference station due to the long, reliable record of measurement (Table 3.1) in an area with relatively low earthquake activity. An example is shown in Figure 3.2c for processed sea level at San Francisco relative to sea level at San Diego. We interpret this difference as being due to vertical tectonic motions related to the earthquake cycle.

### 3.3 Model Comparison

To explore the effects of earthquake deformation on tide gauge measurements, we compare relative displacements derived from sea level measurements to a suite of geodetically constrained displacement models of present-day deformation and historical seismic events (see Section 2.5.5 for model details). Cumulative model displacement through time can be seen in Figure 3.3. As time progresses, the amount of accumulated displacement builds up. Each model-derived time-series (Figure 3.4) reflects earthquake activity (and subsequent displacements) dating back 1000 years, extracted from each representative model at the geographic positions consistent with the California tide gauge station locations. To compare the model results with the tide gauge time-series, the model time series were processed the same as the tide gauge time series, yielding model results relative to San Diego. The basic theory for this comparison is that earthquake-induced land movements should inversely offset the local sea level observations at each tide gauge station.
Previous comparison of the geologic and geodetic data to the deformation model (Section 2.5.5) provided consensus on a model viscosity of $1 \times 10^{19}$ Pa s. Likewise, the elastic plate thickness most preferred by the geologic and geodetic data was thick (60+ km). To further constrain the elastic plate thickness, we adopt variable elastic plate thicknesses of 30, 50 and 70 km as model parameters. Time-series results of the model and tide gauge displacements are illustrated qualitatively in Figure 3.4. The black lines represent relative tide gauge displacements for each station, with the model displacements as a function of elastic thickness shown as blue, red and green lines (30, 50 and 70 km, respectively). RMS residual misfits of our set of trial models vs. relative tide-gauge data are listed in Table 3.2.
From Figure 3.4, the slope of modeled vertical displacements are largely controlled by interseismic motions, however displacements from major earthquakes are evident and may be required to explain some of the unique signatures in the sea level data. For example, some small earthquake displacements can be seen in both models and data (i.e., small slope change in Los Angeles data due to 1940 Imperial earthquake, 1987 Superstition earthquake, 1992 Landers earthquake). Note that the San Francisco data appear largely flat over the extent of the 100-year time series, however these data are plotted with an expanded vertical axis and do vary ~20 mm over the past century; alternatively, the Los Angeles station has a slope of nearly 100 mm over the course of 70 years.
Figure 3.4. Relative tide gauge data and vertical displacement model profiles extracted at 8 sample stations. Tide gauge data (black) and models are plotted relative to the San Diego data/model, with the mean removed. Model results show effect of 3 different elastic plate thicknesses: 30 km (blue line), 50 km (red line), and 70 km (green line).
Table 3.2. RMS residuals for variable elastic plate thickness model results (30, 50 and 70 km) for each tide gauge station, plus the average RMS residuals, for all 8 stations. The lowest RMS for each station is highlighted in grey, and the overall lowest average RMS residual is highlighted in purple.

<table>
<thead>
<tr>
<th>Station</th>
<th>Model RMS 30 km</th>
<th>Model RMS 50 km</th>
<th>Model RMS 70 km</th>
</tr>
</thead>
<tbody>
<tr>
<td>Point Reyes</td>
<td>22.9</td>
<td>13.1</td>
<td>10.6</td>
</tr>
<tr>
<td>San Francisco</td>
<td>42.5</td>
<td>15.4</td>
<td>11.1</td>
</tr>
<tr>
<td>Alameda</td>
<td>12.7</td>
<td>19.7</td>
<td>23.5</td>
</tr>
<tr>
<td>Monterey</td>
<td>27.6</td>
<td>12.2</td>
<td>7.9</td>
</tr>
<tr>
<td>Port San Luis</td>
<td>9.5</td>
<td>8.7</td>
<td>7.4</td>
</tr>
<tr>
<td>Los Angeles</td>
<td>32.3</td>
<td>8.8</td>
<td>17.6</td>
</tr>
<tr>
<td>Newport Bay</td>
<td>23.4</td>
<td>11.3</td>
<td>8.7</td>
</tr>
<tr>
<td>Ensenada</td>
<td>17</td>
<td>16.5</td>
<td>16.7</td>
</tr>
<tr>
<td><strong>Average</strong></td>
<td><strong>23.5</strong></td>
<td><strong>13.2</strong></td>
<td><strong>12.9</strong></td>
</tr>
</tbody>
</table>

3.4 Discussion

The comparisons between the sea level data and the deformation models presented in Figure 3.4 suggest that the two are highly compatible, however major model refinement and exploration is certainly necessary. Inspection of the overall first-order trends at each station demonstrate an intriguing dependence of vertical displacements on interseismic deformation, however displacements from significant earthquakes are also required to explain deviations in the sea level data. Of the three models plotted in Figure 3.4, the model representing a thin plate (30 km) is the least preferred by our previous GPS/geologic data analyses. Similarly, this model provides a poor match to nearly all the tide gauge data, with the exception of the Alameda station, where it appears to fit quite well. In general, thicker plates demonstrate an improved fit to the tide gauge data, with a remarkably good match to the Newport, Los Angeles, Pt San Luis, Monterey, and San Francisco data.

The average RMS residuals for each model and tide gauge time series are 23.5 mm/yr (30 km plate), 13.2 mm/yr (50 km plate), and 12.9 mm/yr (70 km plate) (Table 3.2). These residuals agree with the qualitative analysis that the tide gauge measurements of vertical displacement along the Pacific coast prefer a thick (50-70 km) elastic plate. This follows the same trend of the geologic and geodetic data preferring a thick (60+ km) elastic plate. Moreover, these results suggest that tide gauge data are sensitive to earthquake cycle deformation and can provide vertical crustal velocities as reliable as the
rock history and GPS measurements. Additional continuously recording tide gauge stations along the Pacific coast could provide a more comprehensive data set of vertical crustal motions in California.

In this study, we made several approximations for eliminating signals within the tide gauge data that are considered “nontectonic” (not arising from earthquake cycle deformation). One of these, the GIA, is an inherent source of error in our tide gauge analysis due to the uncertainty in its estimation (Douglas et al., 2001). Glaciation, and the eventual melting of glaciers lead to vertical motions measureable at the Earth’s surface, affecting land on which tide gauge stations record sea level measurements (Peltier and Tushingham, 1989). Models that are developed to estimate how much the Earth responds to the loading and unloading from glaciers forming and retreating take into account the viscoelasticity of the Earth over geologic timescales. The viscosity of the Earth’s lower crust and mantle is a quantity that scientists are still trying to estimate (as shown in Chapter 2). Therefore, unknown characteristics in numerical models of GIA make it a non-exact estimation of the Earth’s response to the weight of glaciers. Furthermore, the precise extent of the last glacial maximum is unknown, and it also varies with latitude, causing GIA estimates to decrease with decreasing latitude (Table 3.1) (Clark and Mix, 2002). All together, these uncertainties provide potential error in our analysis of the vertical deformation in California.

3.5 Conclusions

This study investigated the long-term extent and implication of tectonic deformation on sea level change recorded over that past 100 years along the California coastline. Using coastal tide gauge time series data, we explored the agreement between sea level variations and estimates of vertical displacements produced by a 3-D earthquake cycle deformation model that is constrained by geologic slip rates, geodetic velocities, and historical seismic data along the San Andreas Fault System. Overall, our results suggest that a moderately thick plate provides optimal fit to coastal tide gauge data, and these parameters are also preferred by the GPS epoch-year data.
As demonstrated by this first-order comparison of model and data trends, there is evidence for compatibility, however no single model demonstrated here fits all data time-series equally well. Potential sources of misfit may also be due to unmodeled phenomena present in the sea level data, such as bay focusing, river discharge, and ground water extraction. These unmodeled phenomena will require further inspection, and may be assisted by supplementary analyses of geologic and InSAR data. An incomplete earthquake database may also provide anomalous model estimates, in that modeled earthquakes incorporated into this study were restricted to only those (M≥6) directly associated with faults of the primary SAFS. Significant events such as the Long Beach (1933), San Fernando (1971), and Northridge (1994) earthquakes were not included in this model, and inclusion of deformation from these events will likely alter vertical motions predicted by models. Cumulative effects of smaller seismic events (< M6) may also be important over a long time period. Future work will be aimed at rigorously examining these sources of misfit.

3.6 References

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Chapter 4: Conclusions and future work

While the growing archive of geodetic data in California provides precise deformation measurements with a high degree of spatial coverage, these data supply only a ~20-year record of crustal motions and therefore weakly constrain time-dependence of crustal motions. Geologic data provide a 1000+ year record of deformation, however these data are often sparsely located and provide only a single (averaged) rate, not a time series of deformation changes. Alternatively, sea level change has been continuously recorded at several tide gauge stations for over the past 30-100+ years. Geologic processes, including displacements from large earthquakes, have been shown to produce sea level variations, as vertical interseismic strain accumulation can contribute uplift and subsidence rates of 1-3 mm/yr. Therefore, long-term tectonic contributions to sea level variations must also be considered in crustal deformation studies.

All three vertical deformation data sets used in this study (geologic, geodetic and tide gauge) prefer a thick (60+ km) elastic plate, with a viscosity of 1e19 Pa S. The agreement of these data sets is very encouraging, suggesting that the deformation model used in these studies, originally tuned to horizontal GPS velocities, is truly 3 dimensional in its representation of the deformation along the SAFS.

In order to further constrain rheological characteristics of the crust along the SAFS (and eventually other fault systems), more geologic vertical velocity measurements, and more tide gauge stations would aid this type of analysis by providing more comprehensive depictions of motions along the SAFS. A well understood, consistently processed set of GPS vertical velocities would provide continuity among geophysical modelers, assuring they are comparing their model results to the cleanest, most reliable GPS data set; horizontal motions are well defined and scrutinized by the geodetic community, however vertical motions are not as well-studied. An ideal GPS vertical velocity data set
would include corrections for groundwater and hydrocarbon extraction. Ignoring anthropogenic effects in vertical GPS data can have a significant (and erroneous) impact on the way geodetic modelers interpret crustal motions and model predictions.

Finally, while current purely strike-slip models estimate crustal velocities well, it is understood that a fault system may contain more than one sense of slip. Deformation models that include thrust, normal and strike-slip faults will allow for a more local scale investigation of crustal velocities, strain accumulation, and fault mechanics. We realize that varying fault geometries will complicate models, however it is our hope that with increased computing power, more efficient processing (e.g. models developed in the Fourier domain) and data sets that can resolve these types of deformation, this will be achieved in the near future.
Curriculum Vita

Garrett was born on April 27th, 1985 in Yakima, WA. After graduating high school in May 2004 at Helena High School in Helena, MT he attended the University of Idaho. He worked as a teaching assistant his senior year before receiving a B.S. in geology from the University of Idaho in May of 2010. His studies then took him to the University of Texas at El Paso, where he worked as a teaching assistant in the fall of 2010. From January 2011 until May 2012 he has worked as a research assistant, and is planning on graduating in May 2012 with an M.S. in Geology.

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