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Investigating the Dynamics of Basaltic Volcano Magmatic Systems with Space Geodesy

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INVESTIGATING THE DYNAMICS OF BASALTIC VOLCANO MAGMATIC SYSTEMS WITH SPACE GEODESY

By

Michael Scott Baker

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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INVESTIGATING THE DYNAMICS OF BASALTIC VOLCANO MAGMATIC
SYSTEMS WITH SPACE GEODESY

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Many discoveries have been made about the dynamics of active volcanoes around the world. Studies of the physical and chemical processes have lead to a better understanding of magma generation, evolution, and collection within the crust. The purpose of this research is to use space geodetic techniques to generate time series of surface displacements to study and monitor active volcanoes. Interferometric synthetic aperture radar (InSAR) and global positioning system (GPS) data provide a way to measure the displacement associated with the migration and storage of magma within a volcano at unprecedented spatial and temporal resolutions. With these data, we can determine changes in the rate of surface displacement at volcanoes, which can be an indicator for eruptive activity. Using Kilauea Volcano, Hawaii and the Galapagos Islands volcanoes as natural laboratories, this study uses InSAR time series along with GPS and seismic data to elucidate the details about the shallow magmatic systems and track the movement and storage of magma at these basaltic volcanoes.
Dedicated to my family. Thanks for all your support. A special thanks goes out to Caryl for her patients, understanding, and encouragement throughout the entire process.
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Chapter 1

Introduction

Studies of the physical and chemical dynamics of active volcanoes have lead to a better understanding of magma generation, evolution, and collection within the crust. It is believed that the source of all volcanic activity is “fundamentally basaltic” [Hildreth, 1981]. This idea stems from the notion that partial melting in the upper mantle produces basaltic magmas, providing the source necessary to drive volcanic activity. It is known that not all erupted lavas are basalts; therefore processes must alter this basaltic magma as it is transported from the source region. Studying the geophysical and geochemical phenomena associated with current and historic volcanic activity helps us to better understand these processes. The following is a synthesis of our current understanding of volcanic plumbing systems, based on extensive studies at Kilauea volcano and the Galapagos Islands volcanoes, and the role that InSAR plays in furthering our understanding.

Magma ascent and volcanic plumbing systems

Buoyancy is the main force driving magma ascent [Ryan, 1987; Tilling and Dvorak, 1993]. Tectonic stress can influence magma ascent, but to a lesser extent than buoyancy [Walker, 1993; Watanabe et al., 1999]. Partial melting in the upper mantle
produces basaltic magma, which rises through the lithosphere to shallow depths through a series of cracks and conduits [Eaton and Murata, 1960; Decker, 1987; Tilling and Dvorak, 1993; Marsh 2007]. The cracks and conduits are created from hydrofracturing and magma migrates along a series of dikes and faults [Tillings and Dvorak, 1993]. Clusters of earthquakes within the lithosphere define the magma pathways or conduits [Wright and Klein, 2007]. The density difference between magma and the surrounding lithosphere make volcanism possible, and the densities of magma and crustal rocks determine the position of magma accumulation. Physical and chemical processes, including crystallization and vesiculation, alter the magma as it ascends and ponds within the crust, resulting in different lava compositions, styles of eruption, and types of shallow volcanic plumbing systems [Wright and Fiske, 1971; Hildreth, 19881; Marsh, 2007]. Tectonic setting and the nature of the crust through which the magma migrates largely control these processes.

The classical model of a volcano consists of a magma chamber with a conduit leading to the summit of the volcano, where magma is erupted on the surface. While this model is a good generalization of a volcano, in reality, the plumbing systems beneath active volcanoes are vastly more complex. Marsh [2007] proposed a magmatic system composed of a series of vertically stacked sills and connecting conduits that link the source region to the near surface volcanic center. This model is applicable to deeper transport of magma through the crust as well, and seismic studies have confirmed this model by imaging interconnected sill complexes within the crust [Cartwright and Hansen, 2006]. It is likely that all volcanoes have, at one point in time, contained a magma reservoir that occupied a central position within the volcano and acted as storage
for magma [Walker, 1993]. The notion of a magma reservoir or chamber and the structure or geometry of the chamber can vary depending on tectonic setting.

Relatively high magmatic flux rates occur at mid-ocean ridges, especially at fast- and medium-rate spreading centers. Basaltic melts accumulate and cool in the crust at

![Figure 1.1: Schematic for an ocean island volcano (Tilling and Dvorak, 1993). a) Model of Hawaiian volcano modified from Eaton and Murata [1960]. b) Updated model by Tilling and Dvorak [1993] with more recent estimates for density and seismic velocities and details of magma transport from melt production region to shallow volcanic magmatic system.](image)
depths of ~2 to 6 km beneath the seafloor. The magma chamber beneath mid-ocean ridges has been modeled as a small area of melt (1-2 km wide by 10-100s m thick) overlying a crystal mush surrounded by solidified crust and pockets of magma [Sinton and Detrick, 1992; Hooft and Detrick, 1993]

A schematic cross-section through an ocean island volcano, such as Hawaii and the Galapagos Islands, is shown in Figure 1.1. This model was originally proposed by Eaton and Murata [1960] (Fig. 1.1a), and has remained the generally accepted model. Tilling and Dvorak [1993] updated this model with more detail about the subsurface structure and transport mechanisms from the source region to the shallow plumbing system (Fig. 1.1b). Beneath Hawaii, it is believed magma is formed at some depth interval between 60 and 170 km [Decker, 1987]. These magmas migrate to shallow depths (3-4 km) where undergo variable amounts of differentiation before eruption. Magma accumulates in the zone of neutral buoyancy and this comprises the shallow magma reservoir. The model for the reservoir is a plexus of interconnected pockets of magma represented as a network of sills and dikes [Fiske and Kinoshita, 1969]. The lavas of Hawaii and the Galapagos volcanoes range from tholeiitic to alkalic basalts with erupted compositions varying as a result of differentiation and magma mixing before eruption [Wright and Fiske, 1971; Geist et al., 1998, Naumann et al., 2002; Wright and Helz, 1996; Garcia et al., 2003; Thornber et al., 2003; Rhodes, 1984]. Geist et al. [1998] showed that differences in the Galapagos Islands volcanoes are due to fractional crystallization at different depths in the crust and mantle, resulting in the striking differences in volcano morphology between the eastern and western Galapagos
volcanoes. Even within the western Galapagos volcanoes, they find a relationship between fractionation depth and caldera morphology.

Figures 1.2 is a generalized schematic of an ocean island volcano showing typical features and the geothermal gradient beneath the surface of the volcano. At ocean islands, repeated and nearly continuous supply of hot mantle-derived magmas brings high temperature material closer to the surface, and the high supply rate maintains these temperatures. A 1.2 km deep experimental drill hole at the summit of Kilauea revealed variability in temperatures at shallow depths (<1 km) likely due to interaction with the
water table, but toward the bottom, a linearly increasing gradient was observed [Keller, 1979]. It was suggested that if this gradient persisted or increased with depth, magmatic temperatures would be expected at approximately 4 km from the surface [Zablocki, 1974]. Magmatic temperatures at these depths are also supported by geochemistry (see discussion below) [Wright and Helz, 1996]. At ocean island volcanoes, the temperatures are much higher, resulting in nearly all erupted lavas having basaltic composition.

Kilauea Volcano: a natural laboratory for volcanic processes

Kilauea is one of the most studied volcanoes and has maintained continuous eruptive activity since 1983, providing scientists the opportunity to engage in multidisciplinary research of eruptive processes. The frequent and accessible nature of eruptive activity makes Kilauea a good environment to conduct studies of the physical and chemical processes associated with volcanic activity. Many research techniques have been applied at Kilauea, ranging from seismic and geodetic studies to geochemical analysis of erupted lavas and glasses, leading to a better understanding of the underlying processes. The following is an overview of previous studies at Kilauea.

Movement of magma within a volcano generates seismicity related to pressurization and fracturing of the area surrounding magma reservoirs and pathways [Chouet, 2003; McNutt, 2000]. Numerous seismic studies have been conducted which revealed many aspects of the subsurface at Kilauea. Tomography studies located velocity anomalies and the three dimensional structure of the volcano [Rowan and Clayton, 1993; Okubo et al., 1997; Chouet 2003] and analysis of long period seismicity using dense seismic arrays mapped a shallow hydrothermal system within the caldera [Ohiminato et
Other seismic studies investigated deep magma transport within Kilauea and revealed a “tilted” conduit leading from the melt generation region up to the summit [Klein et al., 1987; Wolfe et al., 2004; Wright and Klein, 2006]. The existing broadband and short period seismic network at Kilauea provides measurements of tremor and detection of earthquake activity. This seismic network has served as the principal means for monitoring the volcano for decades and has been successful at tracking the intrusion and transport of magma within the volcano [Klein et al., 1989; Gillard et al., 1996]. Elevated seismic activity in the summit area and rift zones manifested as accelerated uplift in early 2006, leading up to the June 2007 Father's Day Intrusions in the East Rift Zone (see Chapter 3 for details). More recently, work by Dawson et al. [2010] and Chouet et al. [2010] have made use of a dense broadband seismic network in and around the summit caldera and the radial semblance method to locate sources of very-long-period (VLP) seismic energy in the shallow subsurface. Beginning in April 2008, this technique has provided near real-time VLP source locations at the Hawaiian Volcano Observatory (HVO) [Dawson et al., 2010]. These studies found an area of persistent VLP signals located 1 km beneath the caldera floor and attributed the observed activity to release of gases from the shallow magma.

Studies of volcanic gas have been widely used at Kilauea. The volatiles contained in parental (i.e. mantle-derived) basaltic magmas in order of decreasing weight percent are H2O, CO2, S, Cl, and F [Greenland et al., 1984; Gerlach and Graeber, 1985]. The principle gases released through fumaroles at Kilauea are H2O, CO2, and, SO2 [Gerlach and Graeber, 1985]. Gerlach and Graeber [1985] showed that degassing beneath Kilauea occurs as a two-stage process. The first stage is when parental magma enters the
summit reservoir where most CO$_2$ and half of the S are released. The remaining CO$_2$ and S along with nearly all H$_2$O, Cl, and F are stored in the magma as it reaches equilibrium. Second-stage degassing happens when this reservoir-equilibrated magma is nearly at the surface (40-50 m) during summit or rift zone eruptions. It has been noted that one of the challenges with gas measurements at Kilauea are the various sources of uncertainties [Casadevall et al., 1987].

Several studies have been done on the amounts of dissolved volatiles in erupted glasses [Dixon et al., 1991; Clague et al., 1995; Wallace and Anderson, 1998]. These studies concluded that magmas degassed at shallow depths before being recycled by drainback and mixed with undegassed magmas before eruption. Wallace and Anderson [1998] showed that substantial amounts of surface degassed lavas had erupted from Puu Oo with H$_2$O contents of $\geq$0.24 wt %, compared to magmas with 0.7±0.2 wt % before any degassing or drainback had occurred. Dixon et al. [1991] investigated tholeiitic glasses from the submarine portion of the Puna Ridge and found a wide range for H$_2$O (0.11 to 0.85 wt %) and S (220 to 1440 ppm by wt) contents of samples. Their preferred explanation suggests that the lavas were from mixed magmas, with one degassed component (from either subaerial eruption or a shallow magma chamber with total pressure less than lithostatic) and one that did not degas.

Geochemical analysis of erupted lavas and glasses provide quantitative measurements of magmatic processes within the shallow plumbing system at Kilauea. Wright and Fiske [1971] proposed differentiation and magma mixing models for rift zone eruptions at Kilauea. Multiple analyses of these eruptions concluded that magma in the summit reservoir is not a source of differentiated magma [Helz and Wright, 1992; Garcia
et al., 2003; Pietruszka and Garcia, 1999; Thornber et al., 2003]. Hybrid lavas are generated by varying amounts of mixing between summit-derived magmas and pockets of differentiated magma stored in the rift zones. Changes in compositions between the beginning and end of these eruptions can be produced by initially erupting the differentiated rift-stored magma, and later changes are produced with varying degrees of mixing from undifferentiated summit magmas [Helz and Wright, 1992; Ho and Garcia, 1998; Wright and Helz, 1996; Vinet and Higgins, 2010]. The beginning of the 1955 eruption contained some of the most differentiated lava ever erupted at Kilauea. Helz and Wright [1992] concluded that mixing did not occur in the conduit but rather in the reservoir within the rift. They stated that their models were only capable of determining the amounts of mixing and that it remains impossible to know when new magma first entered the reservoir.

At the summit of Kilauea, shallow reservoirs exist at 3-4 km depths where magma resides for 1-2 years before eruption (details below). The high supply rate of mantle-derived material to the shallow summit provides increased temperatures (~1346°C) at relatively shallow depths [Swanson, 1972; Clague et al., 1995; Wright and Helz, 1996]. Areas within the east rift zone are where we would expect to see greater extents of differentiation because magma resides in these bodies for longer periods of time (sometimes more than 10 years), and imaging these bodies with InSAR will provide good constraints on the volume and residence times for the magma intruded into these areas.

Sills intruded beneath the summit would presumably take longer to solidify than would normally be expected given the high thermal gradient (Fig. 1.2). In nature, sheet-like magma bodies exhibit a low aspect ratio (thickness/diameter) and typical values are
around 0.01 (laccoliths) to 0.001 (sills) [Rubin, 1993; Rubin, 1995]. A sheet-like body with a thickness of 100 m would require a diameter of 1000 km or 100 km for a sill or laccolith, respectively, which is unlikely to occur beneath active ocean island volcanoes like Hawaii and the Galapagos given that the widths of the volcanoes are on the order of 10s km. The largest of our modeled sources beneath Kilauea’s summit between 2004 and 2008 has a diameter of 7 km. Given the typical aspect ratios above, a sill with this diameter would have a maximum thickness of 7 m. The supply rate of new batches of magma to the summit is generally thought to be continuous, so these reservoirs between 3 and 4 km would be persistent bodies that would take longer to solidify than if they were emplaced elsewhere (i.e. closer to the surface or within the rift zones). Therefore it is not possible to constrain sill thickness based on solidification time. The thickness and size of these sills is best determined with modeling and fitting geodetic observations.

**Implications for InSAR studies at basaltic volcanoes**

Many of the previous studies have used various techniques to make measurements and inferences of dynamic processes at volcanic centers. Geochemical analyses provide insights into processes that modify magma compositions and are used to infer magma transport and storage, mixing and differentiation, and source generation. Seismic (earthquake location, tomography, VLP source relocation) and geodetic studies (leveling, tilt, and GPS) reveal dynamic transport processes and track the movement of magma beneath the subsurface. Kilauea provides the ideal setting to perform in situ measurements given the easy accessibility and frequency of low explosivity of eruptions, but other volcanoes prove more complicated. Other locations, such as the Galapagos, are
remote and maintaining networks similar to those in Hawaii have proven to be difficult. In order to capture the dynamics of volcanic activity, high frequency measurements over large areas are needed. InSAR provides this capability, and when combined with other geophysical and geochemical measurements, a better picture of the dynamic system is possible.

One limitation of magma mixing and differentiation models is related to timing. It has been said there is no way of knowing when new batches of magma arrive in the east rift zone reservoirs [Wright and Helz, 1996; Vinet and Higgins, 2010]. By using modern geodetic techniques (InSAR and continuous GPS) we can see when new magma is intruded as accumulation in shallow reservoirs is manifested as surface uplift. Volumes of injected magma can be estimated from geodetic measurements and provide better constraints on the mixing and differentiation models. The new seismic methods, such as the radial semblance technique [Dawson et al., 2010; Chouet et al., 2010], provide good constraints on the timing and location of VLP signals; but these are largely limited to within the summit caldera at Kilauea and target a shallow deformation source. This shallow deformation source is important because it responds quickly to changes and is even seen subsiding shortly before or concurrently with rift zone eruptions. Close coupling of seismic and InSAR measurements can provide more details about dynamic processes at the start of eruptions. Furthermore, there is clear evidence in the timing of earthquake swarms in the upper east rift zone at Kilauea and concurrent surface displacements that reveals transport pathways in the shallow subsurface. These swarms are correlated with deformation due to magma bodies beneath the summit (detailed in Chapter 3), and InSAR measurements can determine whether the swarms are a result of
injection or withdrawal of magma from the summit reservoir. The use of multidisciplinary studies at Kilauea reveals more information about volcanic processes, which can be applied to other volcanic areas for monitoring and research purposes.

While InSAR provides an unprecedented spatial resolution, one limitation of conventional InSAR is the low temporal resolution due to the repeat pass times of the satellites. More recently, InSAR time series methods have helped to overcome this limitation, and studies have shown the small baseline subset (SBAS) method [Berardino et al., 2002; Lanari et al., 2004] to be capable of measuring deformation associated with volcanic activity [Lee et al., 2010; Lundgren et al., 2004; Solaro et al., 2010; Sansosti et al., 2010]. Presented here are the results and findings from InSAR time series analyses at Kilauea Volcano, Hawaii and the volcanoes on Isabella and Fernandina islands in the Galapagos Islands, Ecuador. The InSAR time series are from multiple satellites and tracks. By using a combination of time series from multiple satellites and look geometries, the analyses provide a more frequent measurement of the surface displacement at any one volcano. This approach also emphasizes the need for better acquisition strategies from individual satellite missions to facilitate better temporal coverage.
Chapter 2

Data and methods

Synthetic Aperture Radar

The datasets for the following studies consist of SAR data from multiple beam modes (i.e. different tracks and look angles) acquired by the European Space Agency’s ERS-1, ERS-2, and Envisat satellites, the Canadian Space Agency’s Radarsat-1 satellite, the Japan Aerospace Exploration Agency’s ALOS satellite, and the German Space Agency’s TerraSAR-X and TanDEM-X satellites. In total, I processed and analyzed 1520 acquisitions on 46 separate tracks (Table 2.1) that were used to produce over 5000 interferograms. The Kilauea dataset has 30 separate tracks (994 acquisitions) acquired 1998 to 2011, and the Galapagos dataset has 16 tracks (526 acquisitions) acquired from 1992 to 2011. The number of scenes listed in the table is not the total number of acquisitions that are available for each track. Some scenes have been removed because of signal-to-noise problems related to atmospheric or other data related issues, Doppler overlap problems, or perpendicular baseline problems during the interferogram and time series processing described below. Details are given in each chapter about the specific satellites and tracks used for analysis.
Table 2.1: SAR data covering the Galapagos Islands and Kilauea Volcano

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InSAR Time Series Analysis

To generate the interferograms, we used the ROI_PAC SAR processing software developed at NASA’s Jet Propulsion Laboratory [Rosen et al., 2004] and GMTSAR developed at the University of California San Diego, Scripps Institute of Oceanography [Sandwell et al., 2011]. The interferograms were unwrapped with the Statistical-cost Network-flow Algorithm for Phase Unwrapping (SNAPHU) software developed at Stanford University [Chen, 2001]. The digital elevation models (DEMs) used for interferogram processing are from the shuttle radar topography mission (SRTM) dataset [Farr et al., 2007]. We used SRTM1 (30m resolution) data for Kilauea and SRTM3 (90m resolution) for the Galapagos. Processing scripts developed at the University of Miami allowed for automated generation of interferograms using these two software packages. The use of specialized Linux clusters purchased and configured by the University of Miami (UM) Geodesy Lab along with the High Performance Computing (HPC) core at the UM Center for Computational Science (CCS) facilitated rapid data processing.

A typical SAR track has 40-50 scenes that are compatible for interferometry. These scenes are used to generate the approximately 200 unwrapped, geocoded interferograms needed for the time series analysis. To generate one unwrapped interferogram on modern computer systems (2-3 GHz CPU, 4 GB RAM) takes about 2 hours, with the unwrapping process occasionally increasing this time to 3-6 hours. Multicore systems allow for more than one interferogram to be processed at a time, and a typical server with 8 cores can process the set of 200 interferograms need for the InSAR time series in about 2-3 days. The advantage of using clusters or HPC resources is that
the entire set of 200 interferograms can be generated at the same time, reducing the total processing time to less than one day. In theory, the set of interferograms could be generated in 2 hours if done in parallel on the HPC system, but SAR data processing is extremely input/output (I/O) intensive, which tends to be the limiting factor. To make a single interferogram requires ~300 MB of raw data, and the size of the data gets larger as higher level data products are produced (single look complex (SLC) images, full resolution interferograms, multi-looked interferograms, geocoded interferogram, etc.). In total, to make one geocoded, unwrapped interferogram requires ~20GB of disk space, therefore 200 interferograms requires ~4TB of disk space. If all processing I/O is going to a single location with a gigabit Ethernet interface (theoretical maximum speed of 125 MB/s), then to write 4 TB of data would take ~10 hours. InfiniBand connections typically have 4x to 16x higher throughput than gigabit Ethernet, which would reduce the total time for I/O operations. Assuming that the output disk can write data at those speeds, the entire time for processing the set of 200 interferograms would be close to the theoretical 2 hours. By using clusters and HPC resources for data processing, the amount of time required for generating the 46 separate time series used in this study is on the order of weeks, rather than months if conventional systems and methods were used.

Interferometric pairs are selected to minimize both perpendicular baseline and time span. Similar to the work by Pepe and Lanari [2006], we use Delaunay triangulation for selecting interferometric pairs in the perpendicular baseline-time space to create an interconnected network of interferograms. Pairs with a perpendicular baseline that are too large (600 m for ERS, Radarsat and Envisat, 2500 m for ALOS) or with insufficient Doppler overlap are removed from the network. Because pairs are removed
that do not meet these criteria and others are removed after processing due to low quality interferograms, all the additional small baseline pairs (less than 140 days and 200 m perpendicular baseline) are added to the network. This results in a well-connected network of interferograms with overlapping pairs (Fig. 2.1). Ionospheric effects are noticeable in many of the SAR acquisitions covering the Galapagos; so these acquisitions are removed and no interferograms are produced. The ionospheric noise appears to have a close correlation with time of year, most noticeable in scenes acquired between January to May and some in September.

The time series inversion requires interferograms with high coherence. In order to facilitate automated processing (going from raw data to finished time series with little to no human intervention), I developed a technique to assess the quality of the interferograms and automatically remove those with low coherence. The technique involves identifying an area in the interferograms that maintain high coherence and defining a bounding box around this area-of-interest (AOI). Setting a threshold for the percentage of pixels above a given correlation within the AOI bounding box results in

Figure 2.1: Network of interferometric pairs. a.) descending and b.) ascending paths covering the Galapagos Islands. Red dots mark the individual SAR acquisitions, blue lines connect interferometric pairs used in the time series analysis, and dashed lines are interferometric pairs that were excluded from the analysis.
low quality interferograms being removed from the network (dashed lines, Fig. 2.1). The typical threshold is 50 percent of pixels with a correlation of 0.4 or higher, but these thresholds depend on the resolution of the interferogram as well as the size and location of the AOI bounding box. One needs some *a priori* knowledge about the quality of interferograms and the pattern of correlation for the area to select appropriate parameters. Typically, defining this for one stack of interferograms will yield good thresholds, and then these can be used for all other stacks.

We use the small baseline subset (SBAS) method [Berardino et al., 2002; Lanari et al., 2004] to generate the InSAR time series. The time series inversion is done separately for each track. These time series are later used for calculating vertical displacement. To account for orbital phase error (OPE), we correct each interferogram before time series inversion by estimating and removing a plane from each interferogram. This assumption yields good results in a volcanic setting where the deformation is localized. This assumption is not well suited for studies of strain accumulation or fault motions over a broad area where a plane estimation of the OPE will likely remove deformation signal as well.

During interferogram formation, the topographic phase component is removed by using an existing DEM, in our case SRTM. Even with good quality DEMs (height accuracies of 10-20 m), significant topographic artifacts still exist in the final interferograms. In order to account for this, another topographic phase removal step is done for the time series. One approach by Berardino et al. [2002] for DEM error removal is to estimate a low-pass temporal component of deformation and solve for the topographic artifacts. The topographic error component is subtracted from each input
interferogram before the time series computation. A new approach by Fattahi and Amelung [2011] solves for the topographic artifacts in the time series using a time domain deformation model for the DEM error effects. This is our preferred approach for DEM error removal because it is computationally efficient (only one error estimate is necessary rather than 100s for each interferogram) and works better for time variable patterns of deformation due to an improved model assumption.

**Vertical InSAR Time Series**

In order to combine individual time series into one continuous time series, we compute the vertical component of motion using ascending-descending pairs. The displacement due to inflation and deflation of magma bodies will produce predominately vertical motion, so for volcanic regions, the vertical component is important. Previous studies have shown the possibility of computing three components of motion (east, north, and vertical) from InSAR, but given the look geometry of current SAR satellites, the north component is not well constrained [Hanssen, 2001; Wright et al., 2004]. The ascending-descending pairs are selected to minimize the time differences between acquisitions, with some pairs separated by only 13 hours and others by several days.

The vertical and horizontal components of displacement are computed following the equations of Wright et al. [2004]. The range look vector $p = (p_x, p_y, p_z)$ pointing from the ground to the satellite in a local east-north-up reference frame is used to compute the three components. The precise range look vector is used for each pixel, which is calculated based on the incidence angle and satellite azimuth direction. The observed change in the range direction is given by $r = -pu$ where $r$ is the line-of-sight range change
and $u$ is the vector $(u_x, u_y, u_z)^T$ with the vector component displacements (east, north, and up) in the local reference frame. To combine ascending-descending pairs, we define $R=(r_{asc}, r_{desc})^T$ with the line-of-sight (LOS) displacements for ascending and descending passes, and then $R=-Pu$ where $u$ is the vector component displacements and $P=(p_{asc}, p_{desc})$. 

Figure 2.2: Alignment and vertical computation of time series at Wolf Volcano, Galapagos Islands. The ascending data is Envisat track 61 (55 acquisitions, 12 July 2003 to 18 September 2010) and the descending data is Envisat track 140 (49 acquisitions, 23 January 2003 to 10 June 2010).
Another option for obtaining vertical InSAR time series is to assume that all the motion is vertical and directly convert LOS displacement by dividing by the vertical component of the range look vector. However, such approach can introduce an error if there is horizontal displacement that is not being accounted for.

Figure 2.3: Time series at Wolf Volcano. a.) aligned LOS time series (triangles = ascending, circles=descending), b.) converted LOS time series assuming all vertical and dividing by vertical component of range look vector, c.) vertical time series computed using ascending/descending pairs with interpolation.
Because the InSAR time series are relative measurements (unlike GPS which gives an absolute position measurement), and the time series start at different times, the individual time series need to be aligned and plotted together (Figs 2.2, 2.3). Figure 2.2a shows raw time series before alignment and the linear fit for a given time period (red and blue lines, 1 Jan 2006 to 1 Jan 2009). To align the time series, one time series is shifted by a constant to match with another time series. The alignment is done by defining an overlapping time period for each time series and fitting to each a first-order polynomial (linear) \( y = mx+b \). This yields two estimates (one for each time series), and a simple subtraction then gives the constant offset to apply to the second time series that best aligns the data (Fig 2.2b). This same procedure was done following the vertical computation to properly align all the time series. The first way to compute the vertical is to directly convert each LOS time series individually by assuming all vertical motion and dividing by the vertical component of the range-look-vector (Fig. 2.2d). The second way to get the vertical is by computing the vertical component with the ascending-descending pair described above (Fig. 2.2e). The two time series are aligned with one another at the beginning of overlap and each is linearly interpolated to provide one real data point and one interpolated point for computing the vertical (Fig. 2.2c). The aligned, interpolated time series are the input for the vertical computation using ascending-descending pairs.

Following both of these methods for computing the vertical, all the resulting vertical time series are aligned as described above to create one continuous vertical time series. Where there is GPS data available, the time series are aligned with the GPS. For the majority of cases, no GPS data is available, so one time series is selected as the “master” for aligning the other time series. A good “master” time series will have a long
time span and overlap most of the other time series. Figure 2.3 shows the alignment and vertical computation using six time series (2 ascending-descending pairs from Envisat and one from ALOS) for a point located in the center of the caldera at Wolf Volcano, Galapagos Islands. The alignment was done for all in the same systematic way by using
the same time period (1 Jan 2006 to 1 Jan 2009) to estimate the offsets. The aligned LOS time series are shown in Figure 2.3a. Figure 2.3b and 2.3a show the vertical from directly converting each LOS time series and by computing the vertical using ascending-descending pairs, respectively. The similarity between the two time series is consistent with the assumption that the vertical component is the major deformation component.

In order to make some quantitative assessment of the combined time series, we use the exponentially weighted (EW) moment functions from the Python data analysis library pandas (http://pandas.pydata.org). The functions implemented are an EW moving average (EWMA), EW moving variance (EWMV), and EW moving standard deviation (EWMSTD) which allows us to assess the quality of the time series. These functions implement rolling or moving statistics on the time series and use a spanning parameter to determine the width of the moving window. Figure 2.5 shows the aligned time series from Figure 2.3 along with the EWMA (red line), EWMV (gray line) and EWMSTD (blue line) for each. A span of 20 was used for the moving statistics corresponding to a 20-day EW moving average, variance and standard deviation.

The average variance and standard deviation is shown in the box at the bottom right of each time series (Fig. 2.5). Overall, there is an improvement in the variance in the computed vertical time series (Fig. 2.5c) compared to the others (Fig. 2.5a,b) and the mean variance and standard deviation is reduced to less than a centimeter. The LOS converted time series (Fig. 2.5b) has the highest variance and standard deviation of all three.

The higher variance for the LOS converted time series is likely related to an incorrect assumption that all the motion is vertical leading to a slight reduction of the
accuracy (Fig. 2.5b). Outliers in the time series results in spikes in the EWMV and EWMSTD and is best seen in the LOS and vertical LOS converted time series (Fig. 2.5a,b). Methods to identify and remove these outliers would improve the overall accuracy of these time series. Smaller spikes can be seen in the computed vertical time series (Fig. 2.5c), but using two measurements to compute the vertical reduces the effect of outliers, thus providing a first-order filter for noisy data.

Continuous GPS Data and Processing

The Hawaiian Volcano Observatory (HVO) in collaboration with Stanford University and the University of Hawaii has established and maintains a dense network of continuous GPS stations on Kilauea. There are a total of 36 continuous, dual frequency GPS stations covering the volcano, with 17 installed prior to 2000, and another 13 installed between 2007 and 2008. Of the 36 stations, 32 are still operational. In the Galapagos Islands, Sierra Negra is the only volcano monitored with a continuous GPS network. The network consists of 6 continuous, dual frequency stations, with 4 installed in 2002 and another 2 installed in 2009.

The GPS data are available for download from the University NAVSTAR Consortium (UNAVCO). GPS data were processed at the UM Geodesy Lab using GIPSY-OASIS software from JPL [Stephen et al., 1996]. The results provide daily point position measurements for the north-south, east-west, and vertical components in the international terrestrial reference frame (ITRF2000 for Chapter 3, ITRF2005 for all others). Ocean loading is taken into account during the processing, as well as solutions for the ionosphere and troposphere. To compare GPS and InSAR results, we removed the
plate motion from the GPS time series to account for the difference between an absolute (GPS) and a relative (InSAR) reference frame. The Pacific plate motion and Nazca plate motion are removed for Hawaii and the Galapagos time series, respectively. The result is stable plate reference frame. The plate motions are calculated with the UNAVCO plate motion calculator found at http://www.unavco.org/community_science/science-support/science-support.html. The north and east component velocities from the plate motion calculator are then used to compute the displacement due to plate motion which is removed from the respective components of the GPS time series. Once this is done, then the three components of GPS can be projected into the InSAR LOS.

For comparison of the GPS and InSAR time series, two approaches are taken that provide a way to assess the quality of the InSAR time series. The first is to project the three components (vertical, east, and north) of the GPS data into the radar LOS geometry using the range-look-vector, resulting in one time series for comparison. The second way is to compare the three-component InSAR time series, as computed above, with the three components of the GPS time series. Figure 2.5 shows the vertical, east, and north InSAR-GPS time series at station AHUP on Kilauea Volcano, Hawaii. The component with the most displacement is the vertical, with ~35 cm, followed by over 10 cm in the north component, and less than 10 cm for the east. The striking difference between the north components of GPS and InSAR confirms that InSAR is unable to adequately resolve north-south motion given the current looking geometries of SAR satellites. This issue is not a limiting factor when working in a volcanic setting because the input and withdrawal of magma will produce a predominately vertical displacement, which the InSAR will be capable of resolving. Therefore, we use the combined vertical time series
to track the deformation on the surface of volcanoes and increase the temporal frequency of observations.

![Graph](image)

Figure 2.5: Three component InSAR-GPS time series. The InSAR time series (stars) are computed using ascending-descending pairs to determine the vertical, east-west, and north-south components of motion co-located at the continuous GPS station AHUP located to the south of the summit caldera at Kilauea Volcano, Hawaii.

**Inversion Methods for Magmatic Source Parameters**

The surface displacement field from the SBAS time series is inverted to determine the parameters for shallow, magmatically induced deformation sources. We use the `geodmod` package for modeling the data. `geodmod` is part of the Miami InSAR Modeling and Interpretation Code (MIMIC, http://insar.rsmas.miami.edu/mimicwiki), Matlab-based software for analysis and modeling of InSAR and GPS data developed at UM’s InSAR lab.
Surface displacements that are indicative of magma bodies are modeled with sources simulating spherical or sill geometries. The Mogi point source [Mogi, 1958] is used to simulate a spherical magma body. A penny-shaped crack model [Fialko et al., 2001] or horizontal uniform rectangular dislocation model [Okada 1985] is used for sill-type sources. These sources are used to simulate horizontal, radially symmetric or elliptical sills. For dike events, a uniform rectangular dislocation model [Okada, 1985] is used, but the source is not constrained to be horizontal as is done for modeling sills. In all models, a homogeneous, elastic half-space is assumed with a Poisson's ratio of 0.25 and shear modulus of 3x10^{10} Pa.

The modeling is done using LOS SBAS displacements for time periods of interest. Dieterich and Decker [1975] demonstrated the difficulties in determining source geometry and depth using only vertical displacement data. Therefore, when possible, both ascending and descending LOS geometries were used. Quadtree decomposition sampling of the data is used as modeling input to reduce redundant data [Jonsson et al., 2002]. A minimum of two and a maximum of seven partitioned levels and variance thresholds between 5 and 10 mm between data points are allowed, resulting in a minimum of 16 data points and a theoretical maximum of 16,384. This sampling technique typically yields approximately 100 data points for modeling.

To determine the best fitting model, unit variance is assumed for all data points and use the normalized root-mean-square (RMS) between the data and the model defined as sqrt((d-m)^2/N) where d is the data, m is the model, and N is the number of data points. This non-linear inversion problem is solved using a Gibbs sampling algorithm. The Gibbs sampling uses a similar algorithm as for simulated annealing [Cervelli et al.,
2001], except the inversion is done for a large number of models. Gibbs sampling generates a distribution for each parameter and provides the joint distribution. We obtain posterior probability density distributions of the parameters for the preferred model from the sampling. This set of models is a proxy for the posterior probability density that allows us to evaluate the desired quantities such as mean and confidence intervals for the source parameters.
Chapter 3

Top-down inflation and deflation at the summit of Kilauea Volcano, Hawaii observed with InSAR

Summary

We use interferometric synthetic aperture radar (InSAR) to study deformation of the summit caldera at Kilauea Volcano during 2000-2008, which spanned both an east rift zone eruptive event in 2007 and the start of ongoing summit eruption in 2008. The data set consists of small baseline subset (SBAS) time series generated from 270 acquisitions on 3 separate beam modes from the Radarsat-1 satellite. We identify 12 time periods with distinct patterns of displacement that we attribute until late 2003 to secular tectonic-driven deformation and from 2004-2008 to four different sources in the summit area. We model the shallow magmatic system as a spherical reservoir at 1.9 ± 0.1 km depth below the surface to the northeast of Halemaumau (source 1) and 3 vertically stacked sills at greater depths in the southern caldera area (source 2 at the southern edge of the caldera at 2.9 ± 0.2 km depth, source 3 to the south-southeast of the caldera at 3.4 ± 0.5 km depth, and source 4 south of the caldera at 3.6 ± 0.4 km depth). The sequence for filling of and withdrawal from these reservoirs reveal a top-down process, with sequences of both inflation and deflation initiating in the shallowest source. Inflation of source 3 is
coincident with elevated seismic activity in the upper east rift zone in February 2006 and May 2007. Source 4 is elongated toward the southwest rift zone and also shows elevated seismicity that extends toward the southwest rift zone.

**Background**

Kilauea Volcano on the Island of Hawaii is one of the most active volcanoes in the world, providing a natural laboratory to study processes of basaltic magmatism. Since 1983, Kilauea has been continuously erupting predominately from the Puu Oo vent in the east rift zone (Fig. 3.1) [Heliker et al., 2003]. Recent explosive activity began at the summit in March 2008 with the opening of a vent in Halemaumau Crater [Wilson et al.,
The magmatic system at the summit of Kilauea is not a simple magma chamber geometry, but rather it consists of a complex series of dikes and sills [Fiske and Kinoshita, 1969; Dawson et al., 2004]. Within the last 15 years, the establishment of a dense global positioning system (GPS) network and acquisitions of satellite synthetic aperture radar (SAR) data sets provide a way to measure surface displacement on the volcano.

Previous geodetic studies investigated centers of inflation and deflation in and around the summit caldera using leveling, tilt, trilateration, GPS, and interferometric synthetic aperture radar (InSAR) [Cervelli and Miklius, 2003; Dvorak et al., 1983; Fiske and Kinoshita, 1969; Johnson, 1992; Lockwood et al., 1999; Poland et al., 2009]. Fiske and Kinoshita [1969] showed migrating centers of inflation inside and to the south of the summit caldera. Nearly 2.3 m of subsidence has occurred near Halemaumau since 1975 [Johnson et al., 2010], and at the same time the summit has been affected by volcanic spreading [Delaney et al., 1998]. Repeated dike intrusions occur within the rift zones and affect activity of the shallow system (e.g. subsidence near Halemaumau during the 1997 Napau intrusion [Owen et al., 2000] and 2007 Mauna Ulu intrusion [Poland et al., 2009; Montgomery-Brown et al., 2010]). Summit deflation and inflation tilt events [Cervelli and Miklius, 2003] are detected at a rate of about 5-50 per year [Poland et al., 2009] and frequently correlate with changes at the Puu Oo eruption site but their significance for the state of Kilauea’s magmatic system remains enigmatic.

Understanding the shallow system is important because measurements of the ground deformation (e.g. inflation at a particular location) could be used as an indicator for impending eruptive activity, either at the summit or along the rift zones. The summit
of Kilauea, which is visited daily by thousands of tourists, experienced explosive eruptions in the past (in 1924, and during a 300-year period since the formation of the caldera in ~1500 [Swanson et al., 2012]), so characterizing the observed deformation and resulting activity is vital for understanding the behavior of the volcano. The knowledge that is gained can be applied at other basaltic volcanic systems such as the neighboring Mauna Loa which had an intrusive period during 2002-2009 [Amelung et al., 2007] or the Galapagos volcanoes which have equally or even more dynamic shallow volcanic systems [Amelung et al., 2000; Chadwick, 2006].

A new episode of inflation at Kilauea’s summit started in 2003 [Poland et al., 2008], and, as shown below, reversed a long-term trend of secular subsidence in place since at least 1975 [Johnson et al., 2010]. The Canadian Radarsat-1 satellite acquired an excellent set of SAR imagery from 1998-2008. The frequently repeated acquisitions provide an opportunity to study the dynamics of magma accumulation and migration at unprecedented spatial and temporal resolution. We characterize the surface displacements from 2000-2008 using InSAR time series analysis and GPS measurements. Second, we infer sources of deformation in the shallow system around the summit providing parameters for location and source geometry. Third, we discuss implications for top-down inflation and deflation of magma bodies. Finally, we compare our results with historic periods of activity and previous geodetic studies.

**Data and Methods**

We use GPS and InSAR data to analyze the deformation at the summit of Kilauea Volcano from January 2000 to March 2008. The two data sets are complimentary
because they measure the same surface displacement, and combining them is advantageous because the weaknesses associated with each are lessened (i.e. poor spatial resolution of GPS and poor temporal resolution of InSAR).

**InSAR Data and Processing**

Interferograms are made using data acquired by the Canadian Space Agency's Radarsat-1 SAR satellite. At Kilauea, Radarsat-1 data have the longest continuous time span (over 10 years), and with the 24-day repeat pass of the satellite, it is the most frequently acquired SAR data. The primary data set consists of two beams, one ascending pass (standard beam S3, incidence angle 34-40 degrees, 86 scenes, 22 January 2000 to 16 March 16 2008) and one descending pass (standard beam S1, incidence angle

![Figure 3.2: Network of interferometric pairs for Radarsat-1 descending standard beam 1](image)
24–31 degrees, 96 scenes, 21 February 1999 to 17 March 2008), referred to as ascending S3 and descending S1 respectively. Due to some missed acquisitions of the ascending pass during critical time periods, a third beam completes the coverage (standard beam S6, incidence angle 45–49 degrees, 88 scenes, 31 March 1998 to 2 March 2008, used for the ascending S3 missed acquisition on 2 June 2007), referred to as ascending S6. We use the ROI_PAC SAR processing software developed at NASA’s Jet Propulsion Laboratory (JPL) [Rosen et al., 2004] to produce the interferograms.

We use the small baseline subset (SBAS) method for generating the InSAR time series [Berardino et al., 2002; Lanari et al., 2004; Gourmelen et al., 2010]. The SBAS method inverts a large number (in this case hundreds) of InSAR images, relying on the redundancy of multiple InSAR images to determine the surface displacement for each pixel through time. Interferometric pair selection is performed using a Delaunay triangulation in the perpendicular baseline-temporal baseline space to create an interconnected network of interferograms. We present here the network for the descending S1 data set (Fig. 3.2). The temporal baseline values are scaled by a ratio of the perpendicular baseline threshold (600 m) to the temporal baseline threshold (10 years) following Pepe and Lanari [2006]. Those pairs that exceed the spatial and temporal baseline thresholds and lack sufficient Doppler overlap are removed from the network (Fig. 3.2, dashed black lines). There is an unconnected subset due to large baselines, but this subset overlaps the larger network in time. SBAS employs the singular value decomposition (SVD) method, which allows these two separate networks to be linked. The interconnected network of interferograms allows calculation of displacement between any two acquisitions, regardless of the possibility of generating a particular
The same network procedure is applied to the other acquisition modes. In total, we calculate a network of 262 interferograms for descending S1, 262 interferograms for ascending S3, and 255 interferograms for ascending S6 for the time series processing. Pixels with high correlation will have phase values with lower errors, which helps to lessen the effects of phase noise and errors incurred during interferogram processing. We use only pixels with 70 percent of the interferograms having a correlation of 0.3 or higher were used. The time series are referenced to a single pixel that exhibits high coherence. This pixel is located far enough from the summit and rift zones to not be influenced by the deformation taking place there (Fig. 3.1, red square). If the reference

Figure 3.3: Co-located InSAR (green circles, Radarsat-1 descending S1) and GPS (black dots, displacements projected into descending S1 LOS) time series displacements at station AHUP. At the bottom is a histogram of shallow earthquakes (< 5 km) with magnitude greater than 0.6 in the upper rift zones and summit area of Kilauea every 2 weeks. Increases in seismic activity occurred in January 2000 (150 earthquakes in 2 weeks), February to March 2006 (550 earthquakes in 6 weeks), May 2007 (150 earthquakes in 3 weeks), and June to July 2007 (600 earthquakes in 8 weeks) with smaller increases in January 2005 and April to October 2006.
point has significant displacement, then we would not expect good agreement between
the InSAR and GPS time series around the summit (Figs. 3.3 and 3.4, described below).
We do not apply an atmospheric correction to the time series, and given the spatial and
temporal variability of this signal, it is possible that atmosphere artifacts are present at the
reference point and could influence the time series at other locations. Since there is
agreement between the InSAR and GPS time series, we are confident that atmospheric
artifacts are not an issue.

**GPS Data and Processing**

The Hawaiian Volcano Observatory (HVO) in collaboration with Stanford
University and the University of Hawaii has established and maintains a dense network
of continuous GPS stations on Kilauea (Fig. 3.1). A total of 18 stations exist with data
available during the time period covered by the InSAR data. The GPS data are available
for download from the University NAVSTAR Consortium (UNAVCO). GPS data were
processed at the University of Miami Geodesy Lab using GIPSY-OASIS software from
JPL [Stephen et al., 1996]. The results provide daily point position measurements for the
north-south, east-west, and vertical components in the ITRF2000 reference frame. Ocean
loading is taken into account during the processing, as well as solutions for the
ionosphere and troposphere. To compare GPS and InSAR results, we removed the
NUVEL 1-A plate motion [DeMets et al., 1994] from the GPS time series to account for
the difference between an absolute (GPS) and a relative (InSAR) reference frame. The
result is the Pacific plate reference frame, which can also be viewed as a reference frame
local to Hawaii. For further comparison of the GPS and InSAR time series, we project
the 3 components of the GPS data into the radar line-of-sight (LOS) for Radarsat-1
descending beam S1, showing that the measured motion is consistent for both data sets.

The GPS station showing the largest amount of displacement in the summit area is AHUP, and Figure 3.3 shows the LOS projected GPS time series and the descending S1 InSAR time series at that location. The continuous GPS stations around the summit that are operational during the time span of the InSAR time series are shown in Figure 3.4 with co-located InSAR time series. In both figures, the GPS and InSAR time series show similar results between these independent data sets.

Figure 3.4: Co-located InSAR (colored circles, Radarsat-1 descending S1) and GPS (black dots, displacements projected into descending S1 LOS) time series around the summit of Kilauea. Colored areas correspond to the four deformation episodes and vertical black lines separate the individual time periods identified in Table 1.
Based on the time-variable pattern of seismicity and surface displacement (discussed below), we categorize the activity at Kilauea's summit between 2000 and 2008.
into four episodes of deformation. We define these episodes as secular deformation from January 2000 to October 2003, summit inflation from October 2003 to June 2007, east rift zone intrusion from June 2007 to July 2007, and summit deflation from July 2007 to March 2008. Figure 3.5 shows the descending and ascending LOS displacements (left and middle columns, respectively) for these periods as well as the computed vertical motion (right column). Red colors indicate up (or toward the satellite for LOS displacements) and blue indicates down (or away from the satellite). Temporal changes in the surface displacement pattern are clearly represented in the vertical component of the displacement field which we calculate using the ascending and descending LOS displacements [Wright et al., 2004]. Using the SBAS time series displacements from different look angles on ascending and descending passes of the satellite, we can compute the three components of motion (north, east, and vertical). The northern component contains the most error and is poorly resolved because the satellite has approximately a 12-degree angle from north, so the radar is most sensitive to the vertical and east-west directions. Given that inflation and deflation of magmatic sources will produce a predominately vertical component of motion located directly above the source, we will now focus on the vertical measurements calculated from the InSAR time series. This provides a better location for the source of deformation compared to the other two components.

The secular deformation episode shows a broad area of subsidence around the summit and within the rift zones, with the maximum subsidence of 20 cm located just to the south of the summit caldera (Fig. 3.5c). The summit inflation episode shows a broad area of inflation around the summit with a maximum vertical displacement of 34 cm and
continued subsidence within the east rift zone (Fig. 3.5f). The east rift zone intrusion episode initiates deflation at the summit with a maximum subsidence of 16 cm within the caldera and 30 cm of uplift in the east rift zone at the site of the intrusion (Fig. 3.5i). The summit deflation episode shows a broad area of subsidence around the summit with a maximum of 27 cm, subsidence around Puu Oo with a maximum of 14 cm, and uplift on the south flank along the coast (Fig. 3.5l).

Changes in the GPS and InSAR time series and changes in earthquake activity (Figs. 3.3 and 3.4) revealed more details about the activity occurring during the four episodes described above. We defined 12 separate time periods (Table 3.1) that

<table>
<thead>
<tr>
<th>Period</th>
<th>Dates</th>
<th>Time (days)</th>
<th>Active Source</th>
<th>Vert. Disp. (cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Secular Deformation</td>
<td>1  22 Jan 2000 to 28 Sep 2003</td>
<td>1345</td>
<td></td>
<td>-23</td>
</tr>
<tr>
<td>Summit Inflation</td>
<td>2  28 Sep 2003 to 31 Jul 2005</td>
<td>672</td>
<td>1</td>
<td>6</td>
</tr>
<tr>
<td></td>
<td>3  31 Jul 2005 to 15 Jan 2006</td>
<td>168</td>
<td>2</td>
<td>3</td>
</tr>
<tr>
<td></td>
<td>4  15 Jan 2006 to 8 Feb 2006</td>
<td>24</td>
<td>1</td>
<td>4</td>
</tr>
<tr>
<td></td>
<td>5  8 Feb 2006 to 28 Mar 2006</td>
<td>48</td>
<td>3</td>
<td>11</td>
</tr>
<tr>
<td></td>
<td>6  28 Mar 2006 to 6 Oct 2006</td>
<td>192</td>
<td>4</td>
<td>14</td>
</tr>
<tr>
<td></td>
<td>7  27 Feb 2007 to 3 Jun 2007</td>
<td>96</td>
<td>3</td>
<td>5</td>
</tr>
<tr>
<td>East Rift Zone Intrusion</td>
<td>8  3 Jun 2007 to 27 Jun 2007</td>
<td>24</td>
<td>1</td>
<td>-16</td>
</tr>
<tr>
<td></td>
<td>9  27 Jun 2007 to 21 Jul 2007</td>
<td>24</td>
<td>1</td>
<td>6</td>
</tr>
<tr>
<td></td>
<td>11  24 Oct 2007 to 5 Jan 2008</td>
<td>72</td>
<td>3</td>
<td>-6</td>
</tr>
<tr>
<td></td>
<td>12  5 Jan 2008 to 17 Mar 2008</td>
<td>72</td>
<td></td>
<td>-4</td>
</tr>
</tbody>
</table>
categorize the displacement around Kilauea's summit from January 2000 to March 2008.

The secular deformation episode is detailed by a single period (referred to as period 1) and spans the longest amount of time at four years. Following period 1, the summit inflation episode is divided into 6 periods (period 2 through period 7) and spans the next

Figure 3.6: Vertical displacement maps for the 12 defined time periods between January 2000 and March 2008 (Table 1) showing the shifts in the center of deformation during this time. Arrows represent the horizontal GPS velocity vectors for the stations labeled in Figure 1. The black dots shown in e. through h. are the earthquake locations during swarm activity described in the text. The color scale for all periods is plotted with the same range (-10 to +10 cm). The scale for the arrows showing horizontal GPS velocities varies (either 20 cm/yr or 5 cm/yr).
3 years. The east rift zone intrusion episode is detailed by two periods, periods 8 and 9, with each covering a 24-day period in June and July 2007. The summit deflation episode is divided into 3 periods (period 10 through period 12), and covers up to March 2008, when Radarsat-1 acquisitions cease. In the following, we zoom into the summit area and describe details of the displacement pattern and how it changes through time.

Secular deformation: period 1

The first period, from 22 January 2000 to 28 September 2003, is characterized by a broad area of subsidence around the summit and upper rift zones (Fig. 3.6a). As discussed above, a maximum subsidence of 23 cm is located just to the south of the caldera. The rate and amount of subsidence decreases further down the rifts and away from the summit. The GPS stations around the summit and rift zones (AHUP, KOSM, MANE, and UWEV) show southeast motion increasing towards the south (Fig. 3.6a).

Summit inflation: periods 2 through 7

The summit inflation episode leading up to the 2007 intrusion is covered by six time periods (periods 2-6, Figs. 3.6b-g). The location of the inflation shifts between the inner caldera and an area on the southeastern edge of the caldera during three time periods, period 2 (Fig. 3.6b), period 3 (Fig. 3.6c), and period 4 (Fig. 3.6d). Inflation during period 2 centered inside the caldera to the northeast of Halemaumau Crater with a maximum displacement of 6 cm (Fig. 3.6b). During period 2, the center of inflation shifted to the southeastern edge of the caldera and produced a total of 3 cm of inflation (Fig. 3.6c). Given the long time span of periods 2 and 3, 672 and 168 days respectively, the subsidence within the rift zones observed during period 1 is still present (Figs. 3.6b, 3.6c). During period 4, the center of inflation shifted back to the inner caldera source and
inflated a total of 4 cm in 24 days (Fig. 3.6d). For all three periods, the horizontal GPS displacement around the summit (stations UWEV and AHUP) were radially outward, while stations on the flank and rift zones (stations KOSM and MANE) continued to exhibit the same seaward motion seen during period 1.

Period 5 (Fig. 3.6e) marks a shift in the center of inflation from the inner caldera to an area outside and to the south-southeast of the caldera. Coincident with a seismic swarm in the upper east rift zone, a maximum displacement of 11 cm occurred in an area centered between the upper portions of the east and southwest rift zones during this 48-day period. The high rate of inflation was sustained until 28 March 2006 at which point the inflation shifted to an area outside and to the south-southwest of the caldera marking the beginning of period 6 (Fig. 3.6f). Inflation during period 6 lasted for approximately 6 months with a maximum of 14 cm, and occurred as a broad area with an elongated pattern extending toward the southwest rift zone. During periods 5 and 6, station KOSM switched from southeast to southwest motion, and station UWEV shifted more to the north for period 5 and returned to northwest motion during period 6. Following period 6, there is no clear evidence of inflation around the summit for the latter part of 2006 and into the beginning of 2007, and seismic activity returned to normal background levels.

Another shift in the center of inflation occurs during period 7 (Fig. 3.6g) as inflation returned to the area previously seen inflating during period 5, accompanied by an increase in seismic activity in the upper east rift zone. On 24 May 2007, an M 4.7 earthquake occurred in the upper east rift zone near Puhimau Crater at 2 km depth followed by M 4.1 and M 3.9 aftershocks further down–rift at 3.3 km and 1.0 km depth respectively (white stars, Fig. 3.6g). The maximum amount of uplift was 5 cm (Fig.
3.6g), less than half the amount during period 5. The GPS stations around the summit show a radially outward pattern and the stations on the south flank continue to show southeast motion (Fig. 3.6g).

**East rift zone intrusion: periods 8 and 9**

The 17 June 2007 intrusion in the upper east rift zone created large amounts of ground displacement between the summit and Puu Oo (Fig. 3.5i), accompanied by fissure openings and lava eruptions [Sandwell et al., 2008; Montgomery-Brown et al., 2010]. Period 8 is characterized by 16 cm of subsidence concentrated inside the summit caldera to the northeast of Halemaumau (Fig. 3.6h). The intrusion produced over 30 cm of uplift in the vicinity of the intrusion (see Fig. 3.5i) accompanied by swarms of earthquakes over the first several days of the event. The GPS stations around the summit show radially inward motion in response to the rapid subsidence at the summit (Fig. 3.6h). Period 9 is characterized by an inflation of 6 cm in a small concentrated area to the northeast of Halemaumau, and the GPS stations around the summit show a radially outward pattern (Fig 3.6i). Overall, the area appears red, and is most likely related to the atmospheric noise discussed above.

**Summit deflation: periods 10 through 12**

Following the intrusion in the east rift zone, the summit area entered an episode of net subsidence comprising three time periods (Figs. 3.6i-1) and seismic activity returned to background levels. The location of displacement shifted position during this time from the inner caldera to the southeastern edge of the caldera. Period 10 is characterized by a broad area of subsidence spanning the inner caldera and southeastern portion of the
summit with a maximum of 15 cm centered on the southeastern edge of the caldera (Fig. 3.6j). A change in the rate of subsidence occurred during Period 11 evident from the change in slope in the GPS and InSAR time series (see Fig. 3.4, note stations AHUP and UWEV). The period is characterized by a subsidence of 6 cm located just outside the southeastern edge of the caldera. Unlike periods 10 and 11, period 12 (Fig. 3.6l) lacks a well-defined location of maximum displacement and is characterized by a broad area of subsidence both inside and outside the caldera with a maximum displacement of 4 cm.

**Sources of deformation**

We infer the areas of displacement active during periods 2 through 11 to be associated with four distinct sources. The time variation of the surface displacement patterns and the locations of these different sources are best visualized with maps showing the contoured vertical displacements (Fig. 3.7). Source 1 is located in the inner caldera to the northeast of Halemaumau Crater and is active during periods 2, 3, 8, and 9 (Figs. 3.7b, 3.7d, 3.7h, and 3.7i). This source is best described as a concentrated area of displacement confined to the inner caldera with periods of both uplift and subsidence.

Source 2 is active during periods 4 and 10 (Figs. 3.8c and 3.7j), and is located on the southern edge of the caldera. The area of displacement for this source is not as concentrated as source 1, and the displacement occurs over an area that spans both inside and outside the summit caldera. Source 2 shows uplift during period 4 and subsidence during period 10. Source 3 is located to the south-southeast of the summit caldera and was active during periods 5, 7, and 11 (Figs. 3.7e, 3.7g, and 3.7k, respectively) and shows both uplift and subsidence during these periods. This source is characterized by a broad area of displacement located completely outside the summit caldera and contained within
Figure 3.7: Contours of the vertical displacement for the 12 defined time periods (same as in Fig. 6). The contour interval varies in order to emphasize the location of maximum displacement and the active magma source. The four areas of deformation corresponding to magma bodies include an inner caldera source (source 1, periods 2, 4, 8, and 9), a source on the southern edge of the caldera (source 2, periods 3 and 10), a source to the south-southeast of the caldera (source 3, periods 5, 7, and 11), and a source to the south-southwest of the caldera (source 4, period 6). The black dots shown in e. through h. are the earthquake locations during swarm activity described in the text.

an area between the upper rift zones and to the northeast of the Koae fault system.

Source 4 is located to the south-southwest of the summit caldera and is only active during period 6 (Fig. 3.7f). The source shows a broad area of uplift outside the summit caldera that was elongated down the southwest rift zone.
Source Modeling

Modeling Methods

The surface displacement field shows short wavelengths and localized patterns of displacement, consistent with shallow, magmatically induced deformation. To determine the sources of the observed displacements, we employed the following modeling strategy. We select, for each source, the period with the highest signal-to-noise ratio of the data. We acknowledge that modeling of only one time rather than modeling all time periods may be selective, but this provides a first order approach to reducing errors in the modeling by only using the best data available for each source. We test models with increasing complexity and use geophysical inversion methods to select the model parameters that best fit the observations. We start with a Mogi point source [Mogi, 1958] to simulate a spherical magma chamber. For sources that do not provide a good fit with the Mogi model, we use penny-shaped crack [Fialko et al., 2001] or horizontal uniform rectangular dislocation [Okada 1985]. These sources simulate horizontal, radially symmetric and elliptical sills respectively. In all models, we assumed homogeneous, elastic half-space with a Poisson's ratio of 0.25 and shear modulus of 3x10^{10} Pa.

We run models for all the periods and the comparison of the results shows good agreement for four distinct sources. The individual time periods we use for modeling each of the four sources in more detail are selected to minimize adverse effects of noisy data and possible contributions from secular deformation. This requires the periods to have short time spans and clear deformation from only one source. Out of the four possible periods for source 1 (periods 2,4,8,9), period 8 is selected because it had a 24-day time span and large displacement signal with over 15 cm of subsidence. Out of the
two possible periods for source 2 (periods 3, 10; before and after the east rift zone intrusion and eruption), period 3 is selected because period 10 likely contains contributions from other sources. Out of the three possible periods for source 3 (periods 5, 7, 11), period 5 is selected for modeling because it provides the largest signal with the shortest time span. Source 4 is only active during period 6.

The modeling is done using the ascending and descending SBAS displacements for each period. We use quadtree decomposition sampling of the data as modeling input to reduce redundant data [Jonsson et al., 2002]. We allow a minimum of two and a maximum of seven partitioned levels and variance thresholds between 5 and 10 mm between data points that results in a minimum of 16 data points and a maximum of 16,384. The quadtree sampling results in 86 and 62 data points for the ascending and descending displacements for modeling source 1, 177 data points for the descending displacement for source 2, 82 and 69 points for the ascending and descending displacements for source 3, and 108 and 83 points for the ascending and descending displacements for source 4. The GPS data are excluded from modeling due to sparse station spacing during this time period, which results in poor coverage of the deformation sources (Fig. 3.6). Most areas containing the highest amounts of displacement are not well covered by any of the GPS stations with the exception of source 3 (station AHUP, Fig. 3.6). Since 2008, new stations have been installed in and around the summit caldera that improved the coverage provided by continuous GPS.
To determine the best fitting model, we assume unit variance for all data points and use the normalized root-mean-square (RMS) between the data and the model defined as
$$\sqrt{\frac{(d-m)^2}{N}}$$
where \(d\) is the data, \(m\) is the model, and \(N\) is the number of data points. We solve this non-linear inversion problem using a Gibbs sampling algorithm. The Gibbs sampling uses a similar algorithm as for simulated annealing [Cervelli et al., 2001], except we invert for a large number of models. Gibbs sampling generates a distribution for each parameter and provides the joint distribution. We obtain posterior probability density distributions of the parameters for the preferred model from the sampling. This set of models is a proxy for the posterior probability.

Figure 3.8: Best fitting model results for the four identified magma bodies beneath Kilauea. The four time periods used for modeling were selected to minimize time span and signal to noise levels. Profiles show the fit of the model (black line) to the data (green line). Profile locations are marked by dashed lines.
density that allows us to evaluate the desired quantities such as mean and confidence intervals.

**Modeling Results**

Table 3.2 contains the modeled parameters for each of the sources, and shows the preferred depths and 95% confidence interval obtained from the Gibbs sampling. The depths for each of the sources refer to the distance below the surface.

The best fitting model for source 1 is a point source (Fig. 3.8: b and e), rather than a sill geometry, and is located to the northeast of Halemaumau Crater inside the summit caldera at a depth of 1.9 ± 0.1 km. The estimated volume loss for this period is 9.1x10⁶ m³. The best fitting model for source 2 (Fig. 3.8h) is penny-shaped crack located close to the southeast edge of the caldera at a depth of 2.9 ± 0.2 km with a radius of 2.3 ± 0.2 km and a volume increase of 8.9x10⁶ m³. The best fitting model for source 3 (Fig. 3.8: k and n) is a penny-shaped crack located to the south-southeast of the caldera between the east and southwest rift zones at a depth of 3.4 ± 0.5 km with a radius of 3.4 ± 0.6 km and a volume increase of 11.8x10⁶ m³. The location and size of the source places the outer edges of the sill adjacent to the seismicity within the rift zones with little to no overlap between the two. The best fitting model for source 4 (Fig. 3.8: q and t) is a uniform dislocation with an overall area of 14.8 km² at a depth of 3.6 ± 0.4 km and an estimated volume increase of 10.9x10⁶ m³. The depth is in the same range as source 3, but the

<table>
<thead>
<tr>
<th>Source</th>
<th>Modeled Period</th>
<th>Model Type</th>
<th>Depth / 95% CI (km)</th>
<th>Radius / 95% CI (km)</th>
<th>Area (km²)</th>
<th>Latitude, Longitude</th>
<th>ΔV (10⁶ m³)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>8</td>
<td>Mogi</td>
<td>1.9 / 1.87-2.07</td>
<td>1.87 / 1.84-2.10</td>
<td>1.94</td>
<td>19.4069, -155.2752</td>
<td>-9.1</td>
</tr>
<tr>
<td>2</td>
<td>3</td>
<td>Penny</td>
<td>2.9 / 2.5-3.3</td>
<td>2.3 / 2.0-2.7</td>
<td>2.96</td>
<td>19.3961, -155.2700</td>
<td>8.9</td>
</tr>
<tr>
<td>3</td>
<td>5</td>
<td>Penny</td>
<td>3.6 / 2.7-4.7</td>
<td>3.0 / 1.3-4.0</td>
<td>3.64</td>
<td>19.3838, -155.2710</td>
<td>11.8</td>
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<tr>
<td>4</td>
<td>6</td>
<td>Okada</td>
<td>3.6 / 2.9-4.2</td>
<td>14.8</td>
<td>3.6</td>
<td>19.3819, -155.2891</td>
<td>10.9</td>
</tr>
</tbody>
</table>

§ Change in source volume, increase is positive
location is further to the west and the pattern of displacement is different as well, exhibiting an elongated shape extending toward the southwest rift zone. Profiles comparing the modeled source with the data show good agreement (Fig. 3.8, right side column).

Probability density distributions of latitude and longitude (point clouds, Fig. 3.9a) and depth (histograms, Fig. 3.9b) from the modeling clearly show four distinct source locations. From source 1 to source 4, the source location generally progresses from north to south and from shallow to deep. The centers of sources 1, 2, and 3 overlap, or align, in east-west extent while source 4 is farther to the west. Sources 1 and 3 have very localized point clouds showing well-constrained horizontal locations. Source 2 has a broad range of depths and a larger spread in locations, which is reasonable given the lower signal-to-noise ratio.
to-noise ratio of the data. Source 4 also has a larger spread in locations, but the depth is well constrained.

Figure 3.10 shows the correlation between the estimated source size and depth (Fig. 3.10a) and the differences in horizontal overlap progressing from the lower 95% confidence bound (L95) to the upper bound (U95) for sources 2, 3, and 4 (Fig. 3.10b). Source 1 has the least variation in depth. The size of source 1 is not inferred from the data or modeling, so for Figure 3.11b, we assume a constant radius of 500 m for source 1. In general, the relation between depth and source size shows that sources are either deep and small or large and shallow. This correlation is strongest for source 3 (Fig. 3.10a, middle), with no obvious trend for sources 2 and 4. The mean of the posterior density distribution (for both depth and radius) is the preferred model and is shown as stars in Figure 3.10a. The source depth as a function of time is shown in Figure

---

**Figure 3.10:** a) 2D scatterplot from the Gibbs sampling representing the posterior probability distribution as a function of source depth and size (radius for sources 2 and 3, area for source 4). Circles, stars, and triangles represent the lower confidence bound (L95), mean of the Gibbs sampling, and upper confidence bound (U95) respectively. b) Spatial distribution of sources for the points in a). Source 1 (blue circle) is plotted with a constant radius of 500 m, and the colors and sources are the same as in Figure 9.
3.11. From 2004 leading up to the intrusion (Fig. 3.11, vertical red line), the summit area inflated and showed a general progression from shallow to deeper sources through time (periods 2-7). Concurrent with the intrusion and eruption, the shallow summit source deflated (period 8) followed by a short period of inflation (period 9) and then continued deflation (periods 10-12), again showing a progression from shallow to deeper sources through time. This sequence represents a top-down inflation and deflation of the shallow summit reservoirs.

**Discussion**

The deformation at the summit of Kilauea Volcano shows a time variable pattern of displacement between 2000 and 2008. The locations and timing of displacements provide details about the characteristics of the shallow magmatic system beneath the summit.

**Model for the shallow magmatic system**

We suggest that the shallow magmatic system at the summit of Kilauea is best described as a series of interconnected reservoirs that are active at different points in...
time. We find four distinct deformation sources active between October 2003 and March 2008, with varying locations and times of activity (Table 3.1). Figure 3.13 depicts a schematic cross section across the summit and upper east rift zone of Kilauea. The sources get progressively deeper from north to south. Sources 2, 3, and 4 are best fit with sill-type models, whereas source 1 is best fit with a spherical model (Mogi). The inferred sources are consistent with other studies of the same period. Poland et al. [2009] and Montgomery-Brown et al. [2010] found a point source at 1.5 and 2.5 km, respectively, to the northeast of Halemaumau, which matches well with our source 1. Myer et al. [2008] modeled the same time period we have for sources 3 and 4 (periods 5 and 6) as a single source and found a similar oblong-shaped source. Their source location matches well with our source 4, except they found a shallower depth. We interpret the observed deformation to be related to magmatic processes and not hydrothermal activity due to our
source depths being much deeper than the known hydrothermal system, which is thought to be shallower than 1 km [Almendros et al., 2001; Hurwitz and Johnston, 2003], and also the fact that the deformation correlates with ongoing magmatic and volcanic activity at Kilauea.

The shallow magmatic system consists of a shallow reservoir (1.9 ± 0.1 km depth) to the northeast of Halemaumau (source 1) and a series of vertically stacked sills at greater depths (2.5-4 km) (source 2, 3, and 4). The level of neutral buoyancy (LNB) at Kilauea occupies a zone from 2-4 km [Ryan, 1987], and this corresponds well with the depths we find for our four sources. Mantle-derived magma enters the system from below and rises up to shallow depths. As the supply of magma continues and reaches the LNB, magma begins to migrate laterally creating individual sills in this depth range. These reservoirs appear to be well connected at times, yet show individual periods of activity. This suggests the connections or pathways between the magma bodies are either small and rapid changes in pressure only affect one reservoir at a time, or the connections are not continuously open and changes in pressure and stress of the system effectively open and close the pathways. The ongoing eruptive activity at Puu Oo provides evidence that magma continuously migrated away from the summit during this time.

**Top-down inflation and deflation**

The time dependence of magma storage is apparent when looking at the sequence of source initiation (Fig. 3.13, see also Fig 3.11). After sources 1 and 2 inflated (Fig. 3.13a, periods 2 to 4), the 2006 upper east rift seismic swarms (periods 5) mark when magma began accumulating at the summit in the deeper sill bodies (Fig. 3.13b, sources 3 and 4). These magma bodies continued to inflate into the beginning of 2007 (period 6),
when inflation around the summit ceased. Summit activity resumes with the inflation of source 3 (period 7), coincident with an increase of the seismicity in the upper east rift zone. The associated seismicity at ~3 km depth below the surface suggest that the conduit into the upper east rift zone originates at the depth of or slightly below source 2. The dike is fed by magma from source 1 evidenced by rapid deflation of this source (Fig. 3.13c). After a brief pulse of inflation of source 1, which may represent a rebound effect following the rapid pressure changes, source 2 deflates followed by the deflation of source 3 (Fig. 3.13d).

The summit sources clearly show a sequence of filling and emptying from the top down (shallow to deep) as magma enters and leaves the system, respectively. The shallow sources are the first to inflate beginning in late 2003, and are the first to deflate following the 2007 intrusion and eruption. In a hydraulically connected system at neutral buoyancy we would expect simultaneous pressurization and depressurization of all sources with the volume change depending on the source

![Figure 3.13: Schematic sequence of the top-down inflation-deflation beneath the summit of Kilauea (not to scale). Arrows represent the magma migration paths. Red sources indicate inflation and blue sources indicate deflation. Tables 3.1 and 3.2 list details about the time periods and source locations, respectively.](image-url)
compressibility [Rivalta and Segall, 2008]. This is in contrast to our observations, which suggests that the system is not fully hydraulically connected. The deeper sources may inflate only if the excess pressure exceeds a given threshold to open the conduit to these sources. In the same way, these sources deflate only once the excess pressure in the main plumbing system has dropped below a given threshold. This limited hydraulic connectivity could also be an expression of the conduit size. The conduits from the deeper sources to the east rift zone could be much smaller than those from the shallower sources, impeding the rapid equilibration of pressure changes for the deeper sources. The importance of conduit size for Kīlauea’s intrusions was first noted by Segall et al. [2001].

Previous periods of activity at Kīlauea are consistent with the top-down inflation that we observed in this study. Fiske and Kinoshita [1969] found the shallow source next to Halemaumau was the first to inflate prior to the 1967-1968 eruption, followed by deeper sources to the south of the caldera. This same sequence of inflating sources occurred from 2003 to 2007 leading up to the 2007 east rift zone intrusion and eruption. Periods of inflation at Kīlauea show a general trend of source migration from northeast of Halemaumau toward the south [Tilling and Dvorak, 1992]. The sources inside the caldera are shallower than those to the south, so we infer that top-down inflation is a general pattern for Kīlauea given that the inner caldera sources inflate first, followed by the south caldera sources.

Secular deformation

The summit and upper rift zones subsided at a constant rate during period 1, with the maximum rate of 4-5 cm/yr just south of the summit caldera (period 1, Fig. 3.5 a,b,and c). It is evident that subsidence continued within the upper rift zones following
this period as seen in the GPS (station KOSM, Fig. 4), but subsidence close to the summit is overprinted by the inflation of the shallow magmatic system (periods 2 through 7). At the location of the 2007 intrusion in the east rift zone, subsidence related to the secular deformation is observed before and after the intrusion (Fig. 3.5f and 3.5l). Delaney et al. [1998] found similar patterns of subsidence from 1976 to 1996 using repeated leveling surveys. We suggest that the secular deformation is predominately the result of tectonic processes with a lesser contribution from magmatic processes, consistent with the changes of gravity in that area [Johnson et al., 2010]. During period 1, the GPS stations on Kilauea’s flank show a steady seaward movement [Milklius et al., 2005], providing further evidence of tectonic processes resulting in rifting and secular deformation.

Comparison with previous studies of summit magma bodies

Previous geodetic studies of the shallow magmatic system at Kilauea using leveling, tilt, trilateration, and GPS for earlier periods of activity also detected multiple reservoirs between 1 and 4 km depths (Table 3.3). Fiske and Kinoshita [1969] found multiple centers of inflation in and around the summit caldera from 1966-1967. They concluded that a complex reservoir system existed beneath Kilauea’s summit consisting of sources located to the northeast of Halemaumau and along the southern edge of the caldera between 2-3 km depth. Dvorak et al. [1983] located 25 centers of activity at the summit from 1966 to 1970. Their models placed sources inside the caldera to the northeast of Halemaumau at 2.2 and 2.6 km depth and just to the south of the caldera at 2.4 and 3.4 km depth. Yang et al. [1992] re-examined data from 1975-1985 and found 10 centers of deflation located under the southern caldera rim in the same area as our sources 2 and 3. They identified inflation centers located in the same areas as our sources 1, 2,
and 3, but modeled these as a single inflation source and incorporated dikes into their models as well. They note that neglecting to account for dike dislocations could affect the estimated centers of inflation. From 1981 to 1985, Johnson [1992] inferred deflation sources at 2.5 km depth to the northeast of Halemaumau and deeper deflation sources at 3.5 and 4 km depth further to the south, closer to the southwest rift zone. He places the deflation source associated with the 1983 east rift intrusion, which marked the beginning of the eruptive activity that continues today, just north of the southern caldera rim. Wallace and Delaney [1995] placed a deflation source associated with the same intrusion just south of the caldera rim (near our source 2) at a depth of 3.8±1 km. Owen et al. [2000] used GPS displacements to model the deflation of a shallow caldera source at a depth of 1.87 km coincident with the 1997 Napau Crater dike intrusion and eruption. In 2000-2002, Cervelli and Miklius [2003] used electronic borehole tiltmeter data of four ~2-day interval tilt events to infer a source near Halemaumau (same horizontal position as our source 1) at a depth of 500-700 m below the surface. Using leveling and GPS data from 1996-2002, they inferred a long-term deflation source at 3.5 km depth.

Although many of these studies were limited in spatial and temporal coverage, collectively they support the idea that the magmatic sources, which we observed from late 2003-2008, are persistent features with different instances of activity (Table 3.3). Source 1 agrees well with the caldera source observed by Fiske and Kinoshita [1969]. This source was active throughout the 1960s and 1970s, and before and after the 1983 east rift intrusion [Dvorak et al. 1983; Johnson, 1992]. It was also active during the 1997 intrusion [Owen et al., 2000]. Source 2 was likely active during the pre-1983 period and was the main deflation source associated with the 1983 intrusion [Johnson, 1992;
Wallace and Delaney, 1995]. Sources 3 and 4 have less frequent periods of activity. These are part of the sources during 1967-1983 [Fiske and Kinoshita, 1969; Dvorak et al., 1983], but there is little evidence for any activity for the period from 1983-2000. The location of sources 3 and 4 is similar to that of the long-term source of Cervelli and Miklius [2003]. Their 1996-2002 leveling-measured subsidence pattern (Fig. 5 in Cervelli and Miklius [2003]) is nearly identical to our 2000-2004 InSAR-observed secular deformation pattern (Fig. 3.8a). We interpret that the source of subsidence is primarily related to the tectonic processes discussed above and with a lesser component related to the shallow magmatic sources.

**Conclusions**

1. We use InSAR time series analysis of Radarsat-1 data from 2000-2008 to describe the activity related to the migration of magma beneath the summit of Kilauea. The shallow magmatic system consists of an interconnected system of four magma bodies progressing in depth from north to south. It includes a spherical reservoir to the northeast of Halemaumau at 1.9 ± 0.1 km depth and a series of vertically stacked sills to the south of the caldera at greater depths (2.9 ± 0.2 to 3.6 ± 0.4 km). The sills overlap laterally but have distinct periods of activity, leading to the conclusion that these are independent magma bodies with varying times of activity.

2. Deformation just south of the summit caldera (area of sources 2, 3, and 4) and the seismicity in the upper rift zones provide key insights into the migration of magma at the summit. The location of earthquake swarms outlines conduits or pathways for magma migration away from the summit.
3. The timing of source activity reveals a top-down (shallow to deep) inflation before and top-down deflation following the 2007 intrusion and eruption. This general pattern of top-down inflation has been seen in previous studies, and the same sequence for deflation is clearly seen in the InSAR time series analysis as well.
<table>
<thead>
<tr>
<th>Activity</th>
<th>Time period (month/year)</th>
<th>Source location</th>
<th>Depth (km)</th>
<th>Reference study</th>
<th>Inferred source (this study)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Summit inflation, rapid extension, and swarms</td>
<td>12/1973 - 4/1974</td>
<td>Halemaumau (+)</td>
<td>2.15</td>
<td>Dvorak et al. [1983]</td>
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<td>1/1967 - 10/1967</td>
<td>South caldera (+)</td>
<td>3</td>
<td>Fiske and Kinoshita [1969]</td>
<td>2, 3, and 4</td>
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<td>1/1967 - 2/1967</td>
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<td>2.59</td>
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<td>7.2±4.4</td>
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<tr>
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<td>Johnson [1992]</td>
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<td></td>
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<td>2</td>
<td>Myer et al. [2008]</td>
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<td>6/2007</td>
<td>Halemaumau (-)</td>
<td>1.5</td>
<td>Poland et al. [2009]</td>
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<td></td>
<td></td>
<td>2.5</td>
<td>Montgomery-Brown et al. [2010]</td>
<td></td>
</tr>
</tbody>
</table>

(+): Inflating source  
(-): Deflating source
Chapter 4

Geodetic evidence for shallow magma storage within the east rift zone of Kilauea Volcano, Hawaii

Summary

Located on the Big Island of Hawaii, Kilauea Volcano is one of the most active volcanoes on Earth with continuous eruptive activity since 1983. The ongoing eruptive activity occurs predominately from the Puu Oo vent within the east rift zone, but periodic intrusions occurred in the upper east rift zone between the summit and Puu Oo. These intrusions occurred as dikes typically accompanied by fissure openings and eruptions of small volumes of lava. Interferometric synthetic aperture radar (InSAR) provided surface displacement measurements showing how the ground moves before, during, and after these intrusions. Using data from Radarsat-1, Envisat, ALOS, and TerraSAR-X satellites, we generated line-of-sight InSAR time series using the small baseline subset (SBAS) method providing dense spatial and temporal coverage at Kilauea covering the 17 June 2007 and 5 March 2011 intrusions. For these two events, the summit caldera area switched from deflation to inflation months to years before both intrusions, and just prior to the intrusions we observed increased rates of inflation accompanied by elevated seismic activity in the upper east rift zone. The area to the NE of Halemaumau Crater
subsided rapidly during both, suggesting that magma from the summit reservoir is migrating down rift to feed the intrusions. Small volumes of lava erupted through fissures along the east rift zone during both as well. The pattern of displacements for two intrusions are similar to previous dike intrusion at Kilauea, but following the 2011 intrusion, the area around Napau Crater showed rapid and sustained inflation. Localized deflation has been observed during previous intrusions in the upper east rift, which suggests sources of magma, but there have been no clear inflation events like the one following the 2011 intrusion. This inflation provides clear evidence for magma storage within the rift zone, and can provide a direct measurement of magma input into one of these reservoirs. Modeling results using either point source or sill-type geometries reveal similar depths of 4-6 km and place the horizontal location of the source just to the east of the Napau Crater. Our findings reveal some import implications about the shallow magmatic system at Kilauea: 1) a shallow caldera source is frequently activated and is a source of magma during intrusions, 2) accumulation of magma in the deeper sills south of the caldera happens less frequently, and 3) a magma body beneath Napau Crater is a source of magma storage where inflation is activated as a result of stress changes during dike intrusions.

**Background**

The east rift zone of Kilauea Volcano, Hawaii is host to periodic eruptions and dike intrusions. The latest eruption, known as the Puu Oo-Kupaianaha eruption, began on 3 January 1983 and was initiated by a dike intrusion near the Napau crater, just uprift from Puu Oo. This eruption marks one of the most voluminous outpourings of lava on
the east rift zone in the last 6 centuries. Prior to the eruption, several short-lived eruptions occurred in the east rift zone near Puu Oo and Kupaianaha from 1963 to 1969 and a longer eruption occurred on the upper east rift zone from 1969-1974. A total of 13 intrusions and a single eruption occurred between 1977 and 1983. Since the beginning, the 1983 eruption has progressed through three main epochs: 3.5 years of episodic fountaining from the central vent at Puu Oo, 5.5 years of continuous effusion from the Kupaianaha vent, and continued effusion from the flank vents at Puu Oo.

The upper east rift zone at Kilauea Volcano experienced periodic dike intrusions and eruptions since late 1990 (red lines, Fig. 4.2). In March 1991, an intrusion in the upper east rift zone occurred with no change in effusion rate at Puu Oo. In March 1992, another intrusion in upper east rift zone triggered a pause in the eruption. In late February 1993, an intrusion generated a pause in the eruption and the collapse of the Puu Oo crater floor. The 1997 Napau crater intrusion [Owen et al, 2000] diverted magma
from the conduit leading from the summit to Puu Oo causing the crater floor to collapse over 150 m and collapse on the west flank of the cone. The September 1999 intrusion [Cervelli et al., 2002] triggered another pause in eruption as well as crater floor collapse. Details of the 2007 and 2011 intrusions are discussed below.

On 17 June 2007, rapid deflation at Puu Oo and the summit caldera accompanied dike intrusion in the east rift zone about 10 km uprift (near Pauahi Crater and Mauna Ulu, Fig. 4.2) from the ongoing activity at Puu Oo-Kupaiahaha eruption [Poland et al., 2008; Montgomery-Brown et al., 2010]. The intrusion produced a small eruption in the east rift zone, increased seismicity in the upper east rift zone between the summit and Puu Oo, and elevated SO$_2$ emissions. The summit SO$_2$ emission rates rose by a factor of four from 17 June 2007 to 19 July 2007 along with rapid deflation in the summit caldera to the northeast of Halemaumau [Poland et al., 2009]. Seismic activity returned to normal levels by 23 June.

On March 19, 2008, a series of explosive eruptions began in Halemaumau Crater inside the summit caldera, the first recorded since 1924. The eruption produced ash and rock fragments that covered the area close to the summit and sustained an ash-laden plume and elevated SO$_2$ emission rates. In September 2008, a lava lake within Halemaumau became visible from the surface. Since then, the lava lake has been sustained and numerous collapses and rock falls occurred within the summit vent. There have been times of low activity (decreased gas emissions and tremor levels) that lead to speculation that the eruption ended, but shortly afterward the eruptive activity returned. A seismic swarm occurred in early 2009 and the earthquakes were concentrated beneath
the summit caldera. This swarm was different than the ones in 2006 and 2007 where the seismic activity was predominately in the upper rift zones.

On 5 March 2011, the floor of the Puu Oo began to collapse following the onset of rapid deflation at Puu Oo and increased tremor along the middle east rift zone. Shortly after, the summit began to deflate. An earthquake swarm in the middle east rift zone near Makaopuhi and Napau Craters accompanied the collapse of the Puu Oo crater floor (the

Figure 4.2: Approximate location of upper east rift zone intrusion over the last 20 years. Black dots are catalog earthquake locations (less than 6 km depth). Histogram at bottom shows the number of biweekly earthquakes and red vertical lines mark the times of the mapped intrusions.
level dropped at least 110 m) and the lava lake with Halemaumau continued to drain. Fissure eruptions in and around the Napau Crater produced lava fountains and vigorous spattering which fed channelized a’a flows (80-290 m wide) that advances at least 2.6 km (Fig. 4.2). Sulfur dioxide emission rates averaged 10,000 tonnes/day, the highest measured since July 2008 as a result of the 2007 intrusion-eruption. A pause in eruptive activity occurred on March 10 accompanied by decreased SO$_2$ emission rate and seismic activity.

**Geodetic data for the 2011 intrusion at Napau**

Using InSAR and GPS for the 2011 dike intrusion and eruption, *Lundgren et al.* [2012] find a dike dipping at 72 degrees to the southeast that plunges uplift toward Puu Oo. Distributed opening models showed that the dike was confined to the upper 3 km with a maximum opening of 2.1 m concentrated around the deeper part beneath Napau Crater on March 6 (volume of $12 \times 10^6$ m$^3$), and maximum opening of 2.8 m on March 10-11 (volume of $16 \times 10^6$ m$^3$). For the post-diking deformation around Napau Crater, their model consisted of a shallow line dike (6 km long, 0.4 km wide at 1.5 km depth), a deep sill gently dipping to the north at 7 km depth beneath Napau, and slip of approximately 0.2 m along the decollement detachment fault. Their findings also show that co-eruptive stress change from the dike would promote horizontal sill opening in the area.

**Dataset**

The InSAR time series present in this chapter are from Radarsat-1 (see Chapter 3 for processing details), Envisat-1, ALOS, and TerraSAR-X/TanDEM-X. Details for the
interferogram and time series processing procedures are given in Chapter 2 except where noted. In order to increase the temporal frequency, the individual LOS time series were combined by using ascending-descending pairs to calculate the vertical component of motion at the summit of each volcano as detailed in Chapter 2.

**Seismic Activity**

The movement of magma within a volcano generates seismic activity and earthquakes related to pressurization and fracturing of material surrounding magma reservoirs and pathways [McNutt, 2000; Chouet, 2003]. Therefore, we use the occurrence and locations of earthquakes as an indicator for subsurface magma transport and correlate the activity with the observed surface displacements. The histogram at the bottom of Figure 4.2 represents the biweekly number of shallow earthquakes (<6 km) beneath the summit and upper rift zones from 1990 to 2012. Spikes in seismic activity occur during intrusion in the upper east rift zone (marked by vertical red lines in Fig 4.2) as well as other time. The normal background level of seismicity can be estimated from the times between these spikes, such as in 1994 or around the beginning of 1999, and is less than 10 earthquakes per week.

Time periods with elevated seismicity that are covered by the InSAR time series presented in this chapter are in January 2000 (150 earthquakes in 2 weeks), February to March 2006 (550 earthquakes in 6 weeks), May 2007 (150 earthquakes in 3 weeks), June to July 2007 (600 earthquakes in 8 weeks), December 2008 (350 quakes in 4 weeks). With the exception of the January 2000 and December 2008 swarms, the seismic swarms are accompanied by clear surface displacement present in both the InSAR and
GPS data. Beginning in 2004, the seismic activity becomes elevated and seismic swarms occurred in the upper east rift zone between the summit and Pauahi crater in 2006 and May 2007. We showed in Chapter 3 that the 2006 and May 2007 swarms are associated with subsurface magma close to the summit, and the 2007 swarm is associated with an intrusion and subsequent eruption in the east rift zone near Mauna Ulu [Sandwell et al., 2008; Poland et al., 2009; Montgomery-Brown et al., 2010]. The December 2008 seismic swarm occurred beneath the summit caldera. The earthquakes were concentrated to just around the summit, in contrast to the swarms in 2006 and 2007, which were primarily in the upper rift zones. Slightly elevated seismicity occurred before the 2011 intrusion, and accelerated uplift was seen inside the summit caldera. Again, these earthquakes were in the upper east rift zone between the summit and Pauahi crater, the same as in 2006 but with fewer earthquakes. Other periods prior to 2000 with elevated seismicity are related to intrusions in 1991, 1992, 1993, and 1997 (marked by vertical red lines, Fig. 4.2).

**Surface Displacement Activity: 2007 to 2011**

Figure 4.3 shows the LOS displacement for descending SAR data for periods before, during, and after the 2007 and 2011 intrusion-eruptions (notice difference in scales between the two sides). Before the 2007 intrusion, inflation of ~35 cm at the summit began in late 2003 (~4 years before the intrusion) with the majority of the deformation located outside the summit caldera (Fig. 4.3a). Before the 2011 intrusion, inflation of ~5 cm at the summit began in late 2009 (~1.5 years before the intrusion) and the deformation was contained within the summit caldera (Fig. 4.3d). The deformation
pattern associated with each of the intrusions is typical for dike intrusions (red areas in the east rift zone, Fig. 4.3b, 4.3e). During both of the intrusions, the area to the northeast of Halemaumau Crater subsided rapidly (blue area inside the caldera, Fig. 4.3b, 4.3e).
The amount of subsidence was approximately the same for both, 16 cm in 2007 and 19 cm in 2011.

There is a very striking difference in the pattern of deformation following these two intrusions. Following the 2007 intrusion, a short period (~30 days) of inflation occurred inside the summit caldera (see Fig. 3.6i), then a sustained period of deflation, with a maximum of ~45 cm, centered outside the caldera began (Fig 4.3c). During this period, the area surrounding Puu Oo also experience continued subsidence as well (Fig. 4.3c). There is a considerable amount of south flank motion following the 2007 intrusion as well (red area in Fig. 4.3c). The deformation following the 2011 intrusion consisted of sustained inflation at the summit caldera (~15 cm), and inflation within the east rift zone (~10 cm) in the same location as the dike intrusion (Fig. 4.3f). The subsidence around Puu Oo and the flank motion seen after the 2007 intrusion is not present following the 2011 intrusion.

Figure 4.4 shows the vertical InSAR time series for a point inside the summit caldera, NE of Halemaumau Crater (Fig. 4.4a), and a point inside Napau Crater (Fig 4.4b). The time series show a highly variable pattern of uplift and subsidence through time at both of these locations. The summit caldera time series has the larger amplitude of inflation-deflation, covering a range of over 40 cm (Fig 4.4a). Chapter 3 provides details of summit deformation related to the 2007 intrusion and shows multiple magma bodies active before and after the event, revealing a top-down filling and emptying sequence of the summit reservoir. The time series from Chapter 3 were extended with Envisat, ALOS, and TerraSAR-X data (Fig. 4.4), and shows a continued subsidence at the summit, and as time progressed, the slope of the displacement time series gradually
decreased, corresponding to a decrease in subsidence velocity. In mid 2009, the summit begins to gradually inflate, followed by an increase in inflation velocity beginning in late 2010 just prior to the 2011 intrusion (Fig. 4.4a). There is rapid subsidence during the 2011 intrusion, same as in 2007, but immediately following the intrusion there is a sustained period of rapid inflation that lasted approximately 4 months followed by a decrease in the rate of uplift (Fig. 4.4a).

Figure 4.4: Vertical InSAR time series above a.) summit source and b.) Napau source. Vertical red lines mark the times for the 2007 and 2011 intrusions. The time series are plotted at the same vertical scale.
The Napau Crater time series covers a range of approximately 15 cm (Fig. 4.4b), with the Radarsat-1 data having nearly 10 cm of scatter (Fig. A4). Prior to the 2007 intrusion, the high variance of the time series makes it difficult to interpret any subtle deformation that may be occurring. Following the 2007 intrusion, the time series have a much lower variance making it possible to see the deformation that occurred. There is a

Figure 4.5: Post 2011 intrusion deformation. a.) Vertical displacement field computed from TerraSAR-X interferograms (descending track 24, strip_007, 11 Mar 2011 to 6 Oct 2011 and ascending track 32, strip_008, 25 Apr 2011 to 7 Oct 2011). b.) Time series for a point to the NE of Halemaumau. c.) Time series for a point in Napau Crater. Green stars mark locations of the time series in b. and c. White triangles mark the locations of continuous GPS stations.
gradual subsidence until early 2010 with little to no deformation leading up to the 2011 intrusion. This same pattern is seen in the time series around Puu Oo (Fig. A3). During the intrusion, there was rapid subsidence at Napau followed by a sustained period of inflation until the end of the time series (Fig 4.4b).

The 2011 post-intrusion deformation is best seen in the vertical TerraSAR-X displacement and time series (Fig. 4.5). The displacement map shows inflation within the summit caldera and around Napau Crater (red areas in Fig. 4.5a). The vertical time series inside the caldera clearly show the increased rate of inflation several months before the 2011 intrusion and rapid re-inflation following the intrusion (Fig 4.5b). The time series in Napau Crater show no real sign of deformation before the intrusion, less than 5 cm of deflation during the intrusion, and ~13 cm of inflation following the intrusion (Fig. 4.5c). The rate of inflation was higher immediately following the intrusion and gradually decreased and seen by the changing slope in the time series. This inflation is likely due to an area of secondary magma storage within the rift zone, which we refer to as the Napau source, and elastic modeling of the displacements provides further details (below).

**Source Models**

Elastic half-space models for the summit source and the east rift zone source were run to determine the depth and location for each. We refer to the summit source as source 1 as we did in Chapter 3; the east rift zone sources are referred to as the Napau source and Puu Oo source. Details of the modeling methods and strategies are given in Chapter 2. Individual interferograms that exhibited good signal-to-noise were selected for modeling. For source 1 during the co-eruptive period, the ascending TerraSAR-X
track was selected (track 32, beam strip_008, 1 March 2011 to 3 April 2011). The quadtree decomposition resulted in 164 data points for modeling. For the post eruptive period, the descending track was selected because of good time coverage, with the interferogram covering a 6-month period starting on 11 March, about the same time as

Figure 4.6: Elastic half-space models for source 1 located to the NE of Halemaumau Crater. a.) deflation during the 2011 intrusion-eruption, b.) period of inflation following the intrusion-eruption. The bottom plots show the depth distribution of each model from the Gibbs sampling and the lower ($L_{0.95}$) and upper ($U_{0.95}$) 95% confidence limits.
the pause in activity marking the end of the eruption. This interferogram (track 24, beam strip_007, 11 March 2011 to 6 October 2011) was used to model both the inflation of the summit caldera source and the inflation of the Napau source. For the Puu Oo source, one ascending (track 322, beam S1, 26 Dec 2007 to 3 Feb 2010) and one descending (track 429, beam S2, 28 Nov 2007 to 10 Feb 2010) Envisat interferograms were used. The quadtree decomposition resulted in 155 data points for modeling the summit source, 110 data points for modeling the Napau source. Details of the modeling and quadtree methods are discussed in Chapter 2.

Table 4.1 lists the depths and locations for source 1 during and after the intrusion, Puu Oo following the 2007 intrusion, and the Napau source after the intrusion. The models for source 1 (Fig. 4.6) show good agreement with source 1 from Chapter 3 and with previous studies (see Table 3.3 for details). The depth for the co-eruptive period is 2.37±0.5 km (Fig 4.6a) and the depth for the post-intrusion period is 1.73±0.45 km (Fig.

Figure 4.7: Elastic half-space models for the Puu Oo source. The source was modeled with a Mogi source to simulate a spherical source geometry. Left: Descending and ascending data, middle: model, right: profile. Red star marks the location of the source and white dashed line shows the location of the profiles.
4.6b). For the co-eruptive period, we find a source slightly deeper than that of Baker and Amelung [2012] (1.9±0.1 km), and for the post-intrusion period a shallower source. We believe these differences are not significant (all source depths are within the 95% confidence intervals), and the observed deformation is associated with the same source. For Puu Oo, a point source was used to model the deflation. The depth of the source is 3.8±0.3 km and the location is just to the southeast of the Puu Oo cone (Fig. 4.7).

For the Napau source, two different types of source models were used to investigate the location and geometry of the source. The first source model used was a

![Elastic half-space models for the Napau source in the east rift zone, uprift from Puu Oo. The source was modeled as planar source (okada) to simulate a sill source geometry (center) and with a Mogi source to simulate a spherical source geometry (right). The bottom plots show profiles (marked by dashed line) for the data and models (left), and the depth distribution of each source from the Gibbs sampling and the lower (L95) and upper (U95) 95% confidence limits.](image)

Figure 4.8: Elastic half-space models for the Napau source in the east rift zone, uprift from Puu Oo. The source was modeled as planar source (okada) to simulate a sill source geometry (center) and with a Mogi source to simulate a spherical source geometry (right). The bottom plots show profiles (marked by dashed line) for the data and models (left), and the depth distribution of each source from the Gibbs sampling and the lower (L95) and upper (U95) 95% confidence limits.
Mogi source, which simulates a spherical geometry. The location of the source is just to the east of Napau Crater with a depth of $4.9 \pm 1.3$ km (Fig 4.8a). The second source model is a planar source, which simulates sill-type geometries. Again, the location is just to the east of Napau Crater with a size of 2.6 km by 2.2 km at a depth of $6.3 \pm 1.4$ km (Fig. 4.8b). The two source models overlap in both horizontal position and depth, with the model using the point source being slightly shallower. The InSAR coherence is low around Napau Crater, which is connect by a small bridge of coherent data. The profiles show some steps is the data (black line) that could be related to unwrapping issues in the area, but the overall shape of the profiles for data and models agree, with the Okada model providing a slightly better fit to the data.

**Discussion**

Figure 4.9 shows a schematic cross section (not to scale) across the summit and upper east rift zone of Kilauea. The locations for the 2007 and 2011 intrusions are shown as well as the summit magma bodies discussed in Chapter 3 and the Napau source discussed below. The area of swarms and dike intrusions is a zone in the top part of the rift (from the surface down to $\sim 3$ km) that is rigid and can accommodate earthquake swarms and fracturing. The shallow east rift conduit lies below this zone and provides a pathway for magma from the summit to areas downrift. The dunitic cumulate body at the decollement represents a zone of olive crystal accumulation in the deep rift [Vinet and Higgins, 2010].
With the Napau source, we have for the first time clear geodetic evidence for input of magma into a storage reservoir in the upper east rift zone. It is well accepted that magma bodies exist in the east rift zone. The location of a shallow magma body was revealed in the lower east rift zone when a drilling operation at the Puna Geothermal Venture wellfield encountered a dacite melt at ~2.5 km depth [Teplow et al., 2008]. Leveling surveys in the area beginning in 1992 measured subsidence rates of 0.71 cm/yr that was interrupted by a small inflation anomaly of 0.5 cm in 2005 [Anderson, 2008]. The use of precise leveling techniques provided the necessary resolution to measure this anomaly. The small amount of inflation suggests that there was very little magma being supplied to the area. Cervelli et al. [2002] suggested the existence of a shallow magma reservoir in the upper east rift zone based on analysis of the 1999 dike intrusion. They concluded that the volume of magma required to fill the dike could not have come from

![Figure 4.9: Schematic cross-section of Kilauea's shallow magmatic system beneath the summit and upper east rift zone showing the location of magma bodies and dike intrusions (represented as red bodies) and pathways for magma migration (represented as solid and dashed arrows signifying very likely and probable pathways respectively).](image-url)
the shallow summit and Puu Oo reservoirs alone and required input from additional reservoirs. Using b-value anomalies within the east rift zone, Wyss et. al. [2001] find evidence for active magma chambers between 4 and 8 km depth, and Cayol et al [2000] interpreted EDM and leveling data as deep rift dilation from the decollement depth to 3 km, which fits well with our location for the Napau source.

Dieterich et al. [2003] concluded that the 1983 eruption initiated as a deep-rift dike propagating to the surface based on horizontal deformation data on the south flank. Vertical propagation of dikes is inconsistent with seismically determined dike propagation directions in the upper east rift zone. Klein et al [1987] and Cervelli et al [2002] evaluated dike propagation directions, which reveal important information about the magmatic source. Of the 23 intrusions from 1963 to 1983 examined by Klein et al [1987], only three show uprift (east to west, from Puu Oo) propagation, seven show both uprift and downrift (west to east, from the summit), and the remaining 13 only show downrift propagation. The analysis of Cervelli et al [2002] favored west to east propagation for the 1999 intrusion as well. Reanalysis of the seismic swarm associated with the 1983 intrusion reveal that the summit began deflation and the swarm initially migrated downrift, but abruptly stopped near Napau crater [Rubin et al, 1998]. For 1983 intrusion, Okamura et al [1988] used evidence from tiltmeter data to conclude that the dike tip intersected a magma body slightly beyond Napau Crater, causing a pressure pulse to travel through the body that was seen in the tiltmeter data. This evidence suggests that dikes propagate horizontally rather than vertically, and is consistent with a magma body near Napau crater.
The Napau source showed little to no deformation before the 2011 intrusion, and we observed subsidence during the intrusion. The nearly 15 cm of uplift following the intrusion provides evidence for an active magma storage area beneath Napau Crater. This is further supported by the absence of seismicity down-rift from Napau Crater (Fig 2.2), suggesting an area more conducive to ductile deformation and elevated temperatures related to the shallow magma bodies. InSAR provides the only measurement we have for this area because it is an area with no continuous or campaign GPS stations. The closest GPS stations are around Puu Oo on the outer edge of the fringe pattern produced by the deformation (Fig 4.5). Uplift is seen at some of the stations around Puu Oo (Fig. A3) and those stations are ones located on the outer edge of the deformation. For the other stations, the deformation at Puu Oo would likely dominate any signature of this inflation source in those GPS time series. The aseismic area starts east of Napau Crater, between Napau and Puu Oo (Fig. 4.2), and is exactly where our modeled source for the post-intrusion inflation is located. There are no shallow or deep rift earthquakes in this area, and it is reasonable to believe that the area is the location of a persistent magma reservoir.

Geochemical studies of erupted lavas and glasses have shown various models for differential and magma mixing for rift zone eruptions at Kilauea [Wright and Fiske, 1971; Garcia and Wolf, 1988; Helz and Wright, 1992; Garcia et al., 2003; Pietruszka and Garcia, 1999; Thornber et al., 2003]. These studies concluded that the magma from the summit reservoir cannot be a source of differentiated magma and have proposed pockets of magma stored in the rift zones as sources for the differentiation. These rift zone stored magmas contribute to the changes in compositions between the beginning and end of eruptions by initially feeding these eruptions and then later mixing with summit derived
magma to produce hybrid lava towards the end of the eruptions. Garcia and Wolf [1988] found that erupted lavas from the beginning of the 1983 eruption were produced by crystal fractionation of magma stored within the east rift zone. The beginning of the 1955 eruption contained some of the most differentiated lava ever erupted at Kilauea, and Helz and Wright [1992] concluded that mixing did not occur in the conduit but rather in the reservoir within the rift. They stated that their models were only capable of determining the amounts of mixing and that it remains impossible to know when new magma first entered the reservoir. We have for the first time evidence for when these reservoirs received input of new magma.

During the eruptions, the surface displacements are predominately influenced by the shallow dike opening, making it difficult to discern whether or not the Napau source was active during this time. Figure 4.3e shows an area of subsidence at Napau Crater, but it is unclear whether this is due to deflation of the source or just displacement related to the dike opening. Localized areas of subsidence detected by InSAR in 2007 [Sandwell et al., 2008], together with the Makaopuhi source of Owen et al. [2000] in 1997 suggests that deflation of shallow magma sources during intrusions is common, but the inflation of these sources has not been widely observed. The model of the dike by Lundgren et. al. [2012] shows that the dike is confined to the upper 3 km and deepens to the SW beneath Napau crater (represented as east-west slanted dike in Figure 4.9) and has a maximum opening of 2.1 m. Their calculations of stress change due to the 2011 dike show a positive Szz stress change that promotes horizontal sill intrusion, which would explain why we see rapid inflation of the Napau source following the eruption. Given this, our preferred model for this source is a sill (Okada), rather than the slight shallower Mogi
source, at a depth of 6.3±1.4 km with a volume of 7.3x10^6 m^3 (Fig. 4.8). There is no indication of this source inflating following the 1997 intrusion-eruption near Napau Crater, and Owen et. al. [2000] modeled a shallow deflating source near Makaopuhi Crater (uprift from Napau) indicating that the Napau source was not active during the eruption either. While both the 1997 and 2011 dike intrusions were near Napau Crater, the size and locations are slightly different [Owen et. al., 2000; Lundgren et. al., 2012]. These differences might result in different patterns of stress change in the area, which would explain why the opening of sills beneath Napau Crater is more likely in 2011 and not in 1997.

The Puu Oo source showed a sustained period of deflation following the 2007 intrusion, similar to what was seen at the summit (Fig. A3). There is no clear evidence for inflation or deflation prior to the intrusion. The deflation following the intrusion is likely a magmatic source and not related to secular deformation because the slope of the time series changes. These changes signify variations in the rate of deflation, where as secular deformation would be a constant rate without any changes. Furthermore, the lack of deflation before the 2007 intrusion suggests that the secular deformation observed at the summit is not occurring downrift at Puu Oo. The depth of the Puu Oo source is 3.8±0.3 km, which is slightly deeper than the sill-type magma bodies beneath the summit (deepest is 3.6±0.4 km) but shallower than the Napau source (6.3±1.4 km). The depth of the source places it close to, but slightly deeper than, the conduit in the east rift zone leading from the summit and within the level of neutral buoyancy. We interpret this source as an area of persistent magma storage beneath Puu Oo (Fig. 4.9). This magma body is fed from the deep rift with possible contributions from the east rift conduit and
contributes to the continuous effusion of lava from the flank vents. The period of
deflation correlates with a time of elevated lava effusion from Puu Oo, evidence that this
magma body provides a significant contribution to the erupted lavas.

Source 1 to the northeast of Halemaumau has been well imaged with geodetic
methods. Previous geodetic studies have shown a range of depths for this source (1.5-2.5
km), which is in good agreement with our findings. There is a slight difference in source
depths between the intrusion period (deflating source) and post-intrusion period (inflating
source), but we associate the activity during the two periods with the same magmatic
source. During the 2007 and 2011 intrusions, source 1 responded with rapid deflation
localized to the NE of Halemaumau Crater with approximately the same amplitude (15-
20 cm). This same source deflated during the 1997 intrusion as well, which suggests that
this type of deformation is normal even though the locations of the dike intrusions are
different. This implies that regardless of where the dikes intrude in the upper east rift
zone, there will be some contribution of magma coming from the summit reservoir.

The location of source 1 is also consistent with the location of the conduit in the
shallow system that was inferred from broadband seismic data by Dawson et al. [2010]
and Chouet et al. [2010]. These studies used the radial semblance method and waveform
inversions to locate the source of very-long-period (VLP) seismic energy at a depth of 1
km beneath the caldera floor to the northeast of Halema’uma’u crater. The source of the
VLP signal is attributed to degassing and fluid migration within the magma transport
conduit. The location we find for source 1 with modeling of the InSAR data is directly
beneath this area. This further suggests that the VLP seismicity represents the migration
of magmatic fluids and gases from the magma body into the shallow hydrothermal system within the summit caldera.

Prior to both intrusions, there was inflation at the summit of Kilauea. The amount of inflation was quite different for both, with over 35 cm before 2007 and only 5-10 cm before 2011. The inflation prior to 2007 revealed a top-down inflation of multiple magma bodies in and around the summit, which happened over the course of ~3.5 years. The shallowest magma body, source 1, was the first to inflate, followed by deeper sills to the south. Prior to 2011, only source 1 inflated, and the inflation occurred over ~1.5 years. In both cases, there was an increase in inflation rate just prior to the intrusions seen as a sharp increase in the slope of the time series. The difference in the amount of inflation reveals that a much larger volume of magma accumulated at the summit prior 2007, which might explain the striking difference between the post-intrusion deformations. Prior to the 1997 intrusion-eruption, Owen et. al. [2000] find contraction across the summit caldera (indicating deflation), which is in contrast to the accelerated inflation that we observed prior to 2007 and 2011 eruptions.

Following the 2007 intrusion, source 1 inflated for a very short period, then the entire summit area entered a prolonged period of subsidence and revealed a top-down deflation of the summit. This deflation period lasted until late 2009, when the summit area began to inflate prior to the 2011 intrusion. Following the 2011 intrusion, source 1 sustained inflation for a period of several months before the inflation stopped, seen as a flattening the in the InSAR time series. This same pattern was observed by Owen et. al. [2000] following the 1997 intrusion as the summit sustained inflation for several months until it returned to near pre-intrusion levels. This suggests that the 2007 post-intrusion
deformation is abnormal, possibly related to the higher volume of magma that accumulated prior to the intrusion. Montgomery-Brown et al. [2010] classify 2007 as an “active” intrusion whereas 1997 and 1999 were passive ones, which might explain the differences in post-intrusion deformation.

**Conclusions**

The 2007 and 2011 intrusions show similar patterns of deformation with both showing increased rates of inflation at the summit just prior to the intrusion accompanied by elevated seismicity in the upper east rift zone. There is a striking difference in the deformation at the summit following the intrusions, with a short period of inflation then a period of sustained deflation in and around the summit caldera following 2007, and a period of rapid inflation following 2011. After the 2011 intrusion the area around Napau Crater had a period of sustained inflation. This provides some of the first geodetic evidence for input of magma into an area of secondary magma storage in the upper east rift zone. The existence of this source helps explain the observations of highly differentiated lava that have been analyzed by numerous geochemical studies. Given the depth of the source, it is unclear whether magma migrated from the summit area to fill this source or if the magma came from the deep rift, and gas measurements could help with constraints on the source of the magma input. These results suggest that the shallow plumbing system at Kilauea is an intricate plexus of magma pathways and magma bodies with time variable patterns of activity.
Table 4.1: Time periods covered in Figure 4.3.

<table>
<thead>
<tr>
<th>Period</th>
<th>Dates</th>
<th>Time (Days)</th>
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<tbody>
<tr>
<td>Pre-Intrusion 2007</td>
<td>2005-04-08 to 2006-11-24</td>
<td>595</td>
</tr>
<tr>
<td>Intrusion 2007</td>
<td>2006-11-24 to 2007-06-22</td>
<td>210</td>
</tr>
<tr>
<td>Post-Intrusion 2007</td>
<td>2007-06-22 to 2010-09-24</td>
<td>1190</td>
</tr>
<tr>
<td>Pre-Intrusion 2011</td>
<td>2010-07-12 to 2011-02-17</td>
<td>220</td>
</tr>
<tr>
<td>Intrusion 2011</td>
<td>2011-02-17 to 2011-03-11</td>
<td>22</td>
</tr>
<tr>
<td>Post-Intrusion 2011</td>
<td>2011-03-11 to 2011-10-06</td>
<td>209</td>
</tr>
</tbody>
</table>

Table 4.2: Model parameters for the summit and Napau sources in 2011 and Puu Oo.

<table>
<thead>
<tr>
<th>Source</th>
<th>Modeled Period</th>
<th>Model Type</th>
<th>Depth (km)</th>
<th>Latitude, Longitude</th>
</tr>
</thead>
<tbody>
<tr>
<td>Source 1</td>
<td>Intrusion</td>
<td>Mogi</td>
<td>2.37±0.5</td>
<td>19.4088, -155.2757</td>
</tr>
<tr>
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<td>Post-intrusion</td>
<td>Mogi</td>
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<tr>
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<td>Post-intrusion</td>
<td>Mogi</td>
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<td>19.3718, -155.1353</td>
</tr>
<tr>
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<td>Post-intrusion</td>
<td>Sill*</td>
<td>6.3±1.4</td>
<td>19.3728, -155.1362</td>
</tr>
<tr>
<td>Puu Oo</td>
<td>2007 post-intrusion</td>
<td>Mogi</td>
<td>3.8±0.3</td>
<td>19.3802, -155.0904</td>
</tr>
</tbody>
</table>

*Preferred model
Chapter 5

Measuring deformation associated with magmatic processes in the Galapagos Islands with multi-satellite InSAR time series

Summary

The Galapagos Islands are an active volcanic island chain in the eastern Pacific Ocean off the west coast of Ecuador. The Galapagos Islands are home to some of the most active volcanoes in the world, but given their geographic location and difficult working conditions, the volcanoes have gone largely unmonitored with geophysical methods until the last 19 years. Since 1990, Isabela and Fernandina islands have experienced seven eruptions from four different volcanoes, and widespread deformation at many of the calderas has been detected using interferometric synthetic aperture radar (InSAR). Using radar data from ERS-1/2, Radarsat-1, Envisat, and ALOS, we show the time history of vertical deformation at the summit of each volcano from 1998 to 2011 by combining ascending-descending InSAR time series pairs. The line-of-sight InSAR time series were generated with a total of 1782 interferograms from 542 acquisitions over a span of 6799 days from 15 June 1992 to 12 February 2011.
Changes in activity at individual volcanoes are easily seen in the time series, as well as variable patterns of deformation from the displacement maps. By combining multiple satellites in this fashion, the temporal frequency of observations is increased which can lead to quicker detection of changes at any of the volcanoes. For the 2008 eruption at Cerro Azul and the 2009 eruption at Fernandina, acquisition just prior to the eruption and co-eruptive acquisitions provided a means to track the progress of the eruption.

Identifying the normal behavior of a volcano is easily realized with the combined InSAR time series. There are differences in behavior between these seemingly similar volcanoes where Sierra Negra, Fernandina, and Alcedo showed quite a lot of variation through time, Wolf had an overall increase or inflation; yet only Fernandina, Sierra Negra, and Cerro Azul experienced eruptions during this time. Because the activity at each volcano is so different, it is not possible to use the activity at one as a model for another. Identifying these subtleties leads to a better understanding of the underlying magmatic process that influences the deformation.

**Background: Recent Hotspot Volcanism in the Galapagos**

Currently, only the volcanoes in the western Galapagos are thought to be active. These include the volcanoes found on Isabela Island (Wolf, Darwin, Alcedo, Sierra Negra, and Cerro Azul) and Fernandina Volcano on Fernandina Island (Fig. 5.1). Four of these volcanoes (Alcedo, Cerro Azul, Fernandina, and Sierra Negra) erupted one or more times within the last 20 years. The use of interferometric synthetic aperture radar (InSAR) has made it possible to study these volcanoes in greater detail during this time to
get a better understanding of the causes of deformation and ultimately the geometry and spatio-temporal distribution of the underlying magmatic system.

The volcanoes exhibit unique circumferential and radial dike patterns that in not widely seen at other volcanoes. All of the active volcanoes on Isabela and Fernandina exhibit this characteristic and the eruptive fissure pattern seen on these volcanoes is persistent through time [Chadwick and Dieterich, 1995]. Beneath Galapagos volcano calderas, shallow sills are thought to hold the majority of magma before eruption [Amelung et al., 2000; Geist et al., 2006a; Yun et al., 2006], and research has focused on the origin of these magmas. Some lava flows erupted from flank vents facing adjacent volcanoes are determined to be anomalous because these lavas are compositionally

Figure 5.1: Basemap of the Galapagos Islands, Ecuador. Colors represent the LOS displacement from Envisat track 61, beam S2 from 2003 to 2010. Yellow dots show the catalog earthquake locations from 1970 to 2011. White stars mark the location of time series shown in Fig. 5.2 (see Table 5.1). Boxes outline the areas covered in Fig. 5.3.
indistinguishable from lavas of the adjacent volcanoes, therefore the lavas are thought to be the result of laterally intruded magmas from the adjacent volcano [Geist et al., 1999].

Over the last 20 years, Fernandina has had 5 subaerial eruptions (1988, 1991, 1995, 2005, and 2009), the most of all the volcanoes. The submarine portion of Fernandina plays host to an active rift zone consisting of craters and small, cratered cones [Geist et al., 2006]. Movement of magma within the subaerial portion of Fernandina volcano is accomplished with both radial and circumferential dikes that contribute to the growth of the volcano. The circumferential diking is concentrated close to the caldera and the radial diking is displayed on both the submarine and subaerial flanks [Geist et al., 2006]. The 1995 eruption was characterized by a radial dike intrusion on the southern flank of the volcano and the recent 2005 eruption showed evidence for circumferential dikes close to the caldera rim.

The latest eruption of Sierra Negra was in September of 2005. The source of magma for the eruptions and deformation of the volcano is located in a shallow chamber beneath the caldera [Yun et al., 2005; Amelung et al., 2000]. Yun et. al. [2005] find the betting fitting model for the shallow chamber to be a sill at 2 km depth. Modeling of observed displacement in the caldera lead to the conclusion that high excess magma pressures from the shallow sill provided a mechanism for trapdoor faulting and a method of magma transport to the surface which is observed using InSAR [Amelung et al., 2000; Jonsson et al., 2005].

Cerro Azul is the southernmost volcano on Isabela Island and is thought to be the youngest. The most recent eruptions occurred on September 13, 1998 and May 29, 2008. The eruptions consisted of inter-caldera flows and dike intrusions and fissure eruptions
on the flank [Teasdale et al., 2002]. The style of eruption (flank flows and inter-caldera flows) has not changed in recent history and is responsible for the building of this volcano [Naumann and Geist, 2000].

**Data and Methods**

The data set for this study consists of SAR data from multiple beam modes (i.e. different tracks and look angles) from the European Space Agency’s ERS-1, ERS-2, and Envisat satellites, the Canadian Space Agency’s Radarsat-1 satellite, and the Japan Aerospace Exploration Agency’s ALOS satellite. In total, there are 17 different tracks from the 5 satellites consisting of 542 acquisitions over a span of 6799 days from 15 June 1992 to 12 February 2011 (see Table 2.1). The Envisat data were acquired through background tasking that resulted in high acquisition density from 2005 to 2009 with approximately 7 acquisitions every 35 days. To generate the interferograms, we used a modified version of the ROI_PAC software [Rosen et al., 2004] for Radarsat-1 (starting from SLC produced by Vexcel software) and modified version of GMTSAR [Sandwell et al., 2011] for the other satellites (starting from raw data). A total of 1782 interferogram were generated for use in the time series inversions.

The final vertical time series were aligned and plotted together (Fig. 5.2). White stars in Figure 5.1 show the location for the time series at each volcano. The locations are close to the typical maximum displacement within the caldera. For Wolf, Darwin, Fernandina, and Sierra Negra this location is close to center of the caldera. For Cerro Azul, the point is on the edge for the caldera due to low coherence in the center. At Alcedo, the maximum displacement is offset from the center of the caldera and is likely a
result of structure and caldera geometry influencing the deformation within the caldera. Due to poor quality ascending-descending ERS pairs for Wolf, Darwin, Alcedo, and Fernandina (Fig. 5.2a-d), only the Radarsat-1 LOS converted vertical time series are shown (like Fig. 2.3b) instead of the computed verticals since the descending Radarsat-1 starts in 2000. For Cerro Azul and Sierra Negra (Fig. 5.2e,f), the ERS data quality was better and the computed vertical is shown.

Overview of Ground Deformation 1998-2011

Figure 5.2 shows the multi-satellite vertical InSAR time series at each volcano. Given that input and withdrawal of magma will produce a predominately vertical signal, the combined vertical time series provide increased temporal frequency of observations for tracking deformation related to magmatic processes. The pattern of deformation at the volcanoes is highly non-linear in nature with perhaps the exception of Wolf. Each volcano has periods of inflation and deflation on irregular intervals, and with the use of the high temporal frequency InSAR time series, we are able to identify changes in displacement rates, locations and patterns of deformation, and locate other areas of possible deformation. The east-west component of the time series (Fig. A4) also reveals information about the deformation. Although the signal-to-noise is higher in the east-west component, significant horizontal displacement is observed at Cerro Azul (Fig. A4e), Fernandina (Fig. A4d) and Sierra Negra (Fig. A4f), with less east-west displacements at the others.
The last eruption of Wolf occurred in 1982 from the central vent and contained a radial fissure. Since the beginning of our InSAR time series, Wolf has maintained a
steady rate of inflation of approximately 4 cm/yr nearing a total of 45 cm until an abrupt stop coincident with the 2009 eruption at Fernandina (Fig. 5.2a). The inflation is contained within the caldera with little evidence for displacement on the flanks (Fig. 5.3a).

**Darwin**

There is uncertainty about the last known eruption of Darwin but records show a relatively short eruption in June 1813. Darwin shows the least amount of deformation of all the volcanoes with about 5 cm of inflation until 2005 when it began a period of gentle deflation of about 5 cm (Fig. 5.2b). The displacement is evenly distributed within the caldera and the pattern of inflation (Fig. 5.3b) and deflation (Fig. 5.3c) are the same, suggesting that the same source is responsible for the observed deformation.

**Alcedo**

Deformation at Alcedo is characterized by localized displacement confined to the inner caldera (Fig. 5.3 d-h). The last eruption occurred in 1993, but due to the lack of SAR acquisitions, the deformation associated with this eruption is not evident in the time series. Amelung et al. [2000] show more than 80 cm of uplift within the summit caldera from 1992 to 1997, but beginning in 1998 when there is more continuous InSAR coverage, we show Alcedo experienced a period of steady deflation from 1998 until early 2007 (Fig. 5.2c). The displacement is concentrated in the southern portion of the caldera and is likely influenced by faults and fractures in this area (Fig. 5.3d). In 2007, the caldera began a period of rapid inflation marked by an earthquake event in February 2007 (Fig. 5.3e). The inflation appears to slow or even stop altogether coincident with the beginning of the 2009 Fernandina eruption (Fig 5.2c). In June 2010, a discrete
A subsidence event occurred and caused approximately 10 cm of subsidence localized in the southern portion of the caldera (Fig. 5.3g) that reinitiated the inflation seen before the beginning of the 2009 eruption at Fernandina (Fig. 5.3h).

**Fernandina**

Over the entire period covered with InSAR, Fernandina shows a highly non-linear time series, with steady uplift interrupted by brief periods of subsidence correlated with earthquakes and eruptive activity. Fernandina experienced eruptions in 1995, 2005 and 2009. These eruptions were accompanied by rapid subsidence inside the caldera followed by post-eruption inflation (Fig. 5.2d). A magnitude 5.6 earthquake on 29
August 2007 caused rapid subsidence of over 20 cm within the caldera immediately followed by rapid re-inflation (36 cm/yr) which recovered in 5 months. The time series displacements also show deflation on the flanks during the 2005 eruption and 2007 earthquake events, and this broader pattern of subsidence is evidence for a deep source beneath the summit.

**Sierra Negra**

Sierra Negra has the largest amount of displacement of all the volcanoes in the Galapagos. At the end of the InSAR time series, a net gain of nearly 80 cm occurred, but trap-door faulting events in the southwest part of the caldera [Amelung et al., 2000; Chadwick et al., 2006; Jonsson, 2009] and the deflation of a shallow magma body beneath the caldera during the 2005 eruption [Yun et al., 2006; Yun et al., 2007] created fluctuations of over 4 m inside the caldera. Since the 2005 eruption, the inner caldera has re-inflated nearly 3 m with changes in the rate through time (Fig. 5.2f.). The GPS station (GV02) was not operating immediately following the eruption, but good InSAR coverage shows a rapid inflation until 2007, and continued inflation through 2011 along with the GPS.

**Cerro Azul**

The deformation from the last two eruptions of Cerro Azul was measured with InSAR (see Chapter 6 for details). During both, the summit area showed a broad area of deflation. Both eruptions produced dike intrusions on the eastern flank, with the 2008 dike being more complex and occurred further down the flank. Rapid post-eruption inflation occurred in 1998, but is almost absent following the 2008 eruption. Following the 2008 eruption, the area directly above the dike shows rapid subsidence likely related
to cooling and contraction of the intruded magma or a viscoelastic effect of the surrounding rock (Fig. 5.3o). Continuous InSAR coverage between the two eruptions shows rapid re-inflation following the 1998 eruption with a decrease in inflation rates over time. There is no clear pre-eruptive signal before either of the eruptions.

**Conclusion**

Changes in activity at individual volcanoes are easily seen in the time series, as well as variable patterns of deformation from the displacement maps. By combining multiple satellites in this fashion, the temporal frequency of observations is increased which can lead to quicker detection of changes at any of the volcanoes. For the 2008 eruption at Cerro Azul and the 2009 eruption at Fernandina, acquisition just prior to the eruption and co-eruptive acquisitions provided a means to track the progress of the eruption.

Identifying the normal behavior of a volcano is easily realized with the combined InSAR time series. There are differences in behavior between these seemingly similar volcanoes where Sierra Negra and Alcedo (Fig. 2f and 2c, respectively) show quite a lot of variation through time, Wolf and Fernandina have an overall increase or inflation (Fig. 2a and 2d); yet only Fernandina and Sierra Negra experienced eruptions. Because the activity at each volcano is so different, it is not possible to use the activity at one as a model for another. Identifying these subtleties leads to a better awareness of unusual activity and assessment of associated hazards.

Interactions between volcanoes are also realized with InSAR time series. There is evidence from the time series that inflation at Sierra Negra wanes during the 2008
eruption at Cerro Azul and continues shortly after (Fig. 2f). Likewise, the 2009 eruption at Fernandina marks an abrupt end to the rapid inflation at Alcedo since 2007 (Fig. 2c.), and the steady inflation at Wolf flattens out to nearly nothing as well (Fig. 2a.). These clear changes in the behavior at neighboring volcanoes means that small stress changes at one volcano will have an effect at others that are in close proximity. From December 2006 to August 2007, a series of earthquakes located between Fernandina and Alcedo coincide with changes and increases in the deformation at each volcano. The earthquakes mark the beginning of rapid inflation at Alcedo in February 2007, and rapid subsidence and re-inflation occur at Fernandina in December 2006 and again in August 2007. Due to the sparse seismic network in the Galapagos, the nature of the earthquakes in this area is largely unknown, but surface displacements show a clear correlation with these events.

The use of InSAR as a method for monitoring volcanic hazards has not been widely used given the low temporal frequency. With the increase of current and planned SAR missions, combining multi-satellite time series into a single continuous time series brings InSAR in to the realm of volcanic hazard monitoring tools. With the combined time series, the next repeat-pass of the satellite is no longer the limiting factor, and each acquisition (regardless of satellite, track, or beam mode) will extend the time series. In volcanic island settings such as the Galapagos and Hawaii, each pass of the satellite can acquire data, providing a range for the period between time series measurements from 13 hours (difference between an ascending and descending pass) to at most 5 days.
Table 5.1: Location of time series in Figure 5.2.

<table>
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<th>Latitude (deg)</th>
<th>Longitude (deg)</th>
<th>Elevation (m)</th>
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<td>1111</td>
</tr>
<tr>
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<tr>
<td>Fernandina</td>
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<td>663</td>
</tr>
<tr>
<td>Sierra Negra</td>
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<td>955</td>
</tr>
<tr>
<td>Cerro Azul</td>
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<td>-91.387</td>
<td>1185</td>
</tr>
</tbody>
</table>
Chapter 6

Details of eruptive activity at Cerro Azul Volcano, Galapagos Islands

Summary

Cerro Azul Volcano is located on the southernmost tip of Isabela Island, Galapagos Islands, Ecuador. Cerro Azul is one of 6 active shield volcanoes that make up Isabela Island, and the two most recent eruptions in 1998 and 2008 are covered by InSAR measurements. Using SAR data from ERS-1, ERS-2, Radarsat-1, Envisat, and ALOS, the deformation before, during, and after the eruptions are measured and analyzed. The eruptions are characterized by deflation of a deep (~6 km depth) source beneath the summit and dike intrusions producing fissures and lava flows on the eastern flank.

Background

Cerro Azul Volcano (Fig. 6.1) has erupted 11 times from 1932 to 2008 for an average of one eruption every 6.9 years. Cerro Azul is unique compared to the morphology and composition of the other active Galapagos volcanoes. This is thought to be a result of the age of the volcano itself, representing an early evolution or building stage that the other volcanoes have already completed. Assuming this to be true, Cerro
Cerro Azul coupled with the other active volcanoes of the Galapagos provides a unique opportunity to study their evolution first hand.

Since the start of SAR acquisitions, there have been two eruptions within a 10 year period. The eruptions occurred on September 15, 1998 and May 29, 2008 and lasted 51 days and 13 days respectively. Radarsat-1 and ERS data covered the 1998 eruption, and the 2008 eruption is covered by Envisat and ALOS data. The coverage for the 1998 eruption is sparser than the coverage for the 2008 eruptions due to lower frequency acquisitions, but still allowed for measuring the deformation during the eruption. Nearly continuous radar data coverage between the eruptions allowed us to see the state of the volcano following the 1998 eruption and leading up to the 2008 eruption. Using data from 8 different Envisat tracks in 2008, we measured the deformation associated with
two separate eruptive phases at the volcano and tracked the progression of dike intrusions on the flank.

Detailed analysis of the 1998 eruption from fieldwork and satellite remote sensing documented the entire event. The 1998 eruption at Cerro Azul lasted 36 days and produced intra-caldera eruptive vents and fissure eruptions on the eastern flank (~5.5 km from the eastern rim) \cite{NaumannGeist2000, MouginisMark2000, Rowland2003, Teasdale2005}. The erupted lavas covered a wide range of compositions with evidence of mixing between the 1998 and 1979 magmas. The total estimate for erupted lavas from both intra-caldera and flank vents is $1 \times 10^8$ m$^3$. By the end of the eruption, the flank complex had grown to approximately 80 m high and 1 km in length \cite{Teasdale2005}. The flank lava flows were estimated to be 16 km long, covering 16 km$^2$, and having a dense rock equivalent volume of $54 \times 10^6$ m$^3$ \cite{Rowland2003}.

The 2008 eruptions began on 29 May and lasted until 11 June. Lava primarily erupted from the SE flank with some lava from a vent on the NE wall of the caldera. The flows were 2-3 km wide and up to 10 km in length. The eruption is separated into two phases, the first is from 29 May-1 June, and the second from 3 June to 11 June. Phase 1 consisted of rapid lava emissions from circumferential and radial fissures outside the caldera. The second phase emitted lavas over the remainder of the eruptions from a separate fissure lower on the SE flank. The MODVOLC thermal anomalies data measured a peak 90-pixel anomaly on 5 June at 0420 that was reduced to a 5-pixel anomaly later that day.
Methods

The data used to study the 1998 eruption is from ascending Radarsat-1 beam S5 and ERS tracks 140 (descending) and 61 (ascending). For the 2008 eruption, 7 tracks from Envisat and 2 tracks from ALOS were used. The Envisat data were acquired through background tasking that resulted in high acquisition density from 2005 to 2009 with approximately 7 acquisitions every 35 days. For the 2008 eruptions, ascending-descending pairs acquired on days 1, 2, 4, 5, 7 and 8 allowed for computing vertical time series (see Ch. 2 for details) providing detailed observations as the eruption progressed. The ALOS had more sparse acquisitions, but contributed to pre- and post-eruption deformation measurements. The ALOS data was acquired from ASF through the US Government Research Consortium (USGRC) Palsar Datapool. All Envisat and ALOS data were processed with GMTSAR and ERS and Radarsat-1 were processed with ROI_PAC. For all interferograms, SRTM3 was used for topographic phase correction.

Figure 6.2: Eruptive deformation at Cerro Azul. a.) LOS displacements from 13 Sept 1998 to 4 Jun 1999 from Radarsat-1 ascending beam S5. b.) LOS displacements from 26 Apr 2008 to 5 Jul 2008 from Envisat ascending track 61 beam S2. 1 fringe is equal to 3 cm of LOS displacement.
The SBAS time series analysis and computation of vertical InSAR time series follow the methods described in Chapter 2.

**Surface displacement activity: 1998 and 2008 eruptions**

The 1998 and 2008 eruptions at Cerro Azul show differing patterns of surface deformation (Figs. 6.2). Both eruptions have a deep (5-7 km) magma body beneath the summit that deflated during the eruption. A few key differences existed between the two eruptions. One is the locations and size of the dike intrusions on the flank. The 1998 eruption consisted of a single dike on the upper southeast flank close to the summit (Fig 6.2a). The 2008 eruption shows first circumferential diking near the summit followed by

![Figure 6.3: Vertical InSAR time series covering 2008 eruption. The locations for the time series are shown as black stars in Fig. 6.1. a.) time series at the summit caldera b.) time series on eastern flank at the location of the dike intrusion during phase 2 of the eruption. Vertical blue, dashed lines mark the beginning and end of the 2008 eruption. Time period used for aligning the time series was 1 Dec 2007 to 1 May 2008.](image-url)
radial dike further down the flank (Fig. 6.2b). Using the frequent InSAR coverage during 2008, we see the progression of the flank dike intrusions and capture two separate segments (Figs. 6.3 and 6.4).

Another key difference is the apparent lack of recharge following the 2008 eruption. The three months immediately following the 1998 eruption show a rapid inflation (5-10 cm) for a deep source at the summit (Fig 6.2d). Inflation continues near the summit and a total of 30 cm accumulates leading up to 2008. Following the 2008 eruption, there is only a slight amount of inflation at a rate much less than that in 1998. The frequency of SAR acquisitions is much less in 1998, but we can see that the deformation due to the dike intrusion is done by day 13, but the deep source beneath the summit continues to throughout the eruption (Fig. 6.2b). Given the high frequency of
Envisat observations for the 2008 eruption, we see that all deformation due to dike intrusion is finished by day 5 of the eruption. The deep summit source appears to stop deflating at that time as well. This suggests that the lavas that continued to erupt for the remainder of the eruptions are being fed from the injected dike and are no longer coming from the deep source beneath the summit.

The time series at the summit and on the east flank at the location of the intrusion show the deformation as the eruption progresses (Fig. 6.3). From the beginning of the eruption to day 2, the summit subsided 10-15 cm. Another 15 cm of subsidence occurred from day 2 to day 5, and 6 cm of subsidence from day 5 to day 8 for a total of over 30 cm during the eruption (Fig. 6.3a). All deformation at the summit is done by day 8 of the eruption. Following the eruption, there is very little inflation at the summit with no more than 5 cm total. On the flank, the time series shows ~1 cm of subsidence in the first 2 days of the eruption, another 4 cm of subsidence occurred from day 2 to day 5, and ~6-7 cm from day 5 to day 8 for a total of 12 cm of subsidence during the first eight days of the eruption (Fig. 6.3b). Following the eruption, this area experienced an additional 15 cm of subsidence over the next year. The rate of deflation decreased exponentially evidenced by the gradually changing slope of the time series (Fig. 6.3b).
Deformation Source Modeling

Elastic half-space models for the summit source and flank intrusion sources were run to determine the depth, location, and geometry for each. A Mogi source is used for the deep source beneath the summit and a uniform dislocation source (Okada) is used for the dike intrusions on the flank. For the 1998 eruption, single interferograms were used for the modeling, and for the 2008 eruption models, SBAS time series LOS displacements were used. For the 1998 eruption, a pair of ascending-descending interferograms from ERS was modeled and shows a depth of 6.1±1.3 km for a source on the northern edge of the caldera (Fig. 6.5).
Phase 1 in 2008 was modeled with both a uniform dislocation (circumferential dikes on the flank) and a Mogi point source (deep deflation source beneath summit). The deep source is just to the south of the caldera at a depth of 6.2±0.5 km, and a dike dipping at 70 degree on the edge of the caldera at 3.2±0.4 km depth (Fig. 6.6). The length of the modeled dike is 5.5±1.0 km, which puts the bottom of the dike close to the depth of the deep deflation source. The opening volume for the dike is 4.6x10^7 m³.

The first part of Phase 2 is modeled with a uniform dislocation and a Mogi source as well. The deep deflation source is located at a depth of 5.7±0.3 km along the northern part of the caldera (Fig. 6.7). The dislocation source extends away from the summit and is located at a depth of 2.1±0.2 km, dipping 70 degrees, with a length of 8.8 km and width of 3.0 km, and volume of 7.2x10^7 m³. The width and depth of the source places the bottom edge overlapping with the top edge of the modeled dike from phase 1.
For the second part of phase 2, again a combination of a uniform dislocation and Mogi source was used (Fig. 6.8). The deep deflation source is in roughly the same horizontal location as the first part of phase 2, but slightly deeper at 6.4±1.0 km (closer to the depth in Phase 1). The amount of deflation is much less than the previously two modeled periods. The dislocation source extends radially away from the summit at a depth of 1.3±0.4 km, near vertical orientation (78 degree dip), a length of 3.5 km and
width of 3.2 km, and volume of $3.1 \times 10^7$ m$^3$. Again, given the width and depth of the source, the bottom edge extends down to a depth close to the that found for the dike in part one of phase 2.

The PPD for 2008 phase 1 and the first part of phase 2 are not as broad as the 1998 and second part of phase 2. This is likely a result of the level of noise in the data. For the 1998 eruption, individual interferograms were modeled and not SBAS time series displacements. For the second part of phase 2 in 2008, only one look direction (descending) was used for modeling. For the other two periods modeled, SBAS time series displacements with ascending-descending pairs provided better constraints on the modeling parameters resulting in narrow PPDs.

**Discussion and Conclusions**

Figure 6.9 is a schematic cross section across the summit and flanks of Cerro Azul showing the generalized orientation and positions of the models sources. For the 1998 eruption, less frequent SAR acquisitions only captured the overall deformation and not the incremental sequence as for 2008. The deformation consisted of a deep deflation source beneath the summit and a dike on the flank (Fig. 6.9a). For the 2008 eruption, the deformation is separated out into two separate phases. Deflation of the deep source observed during the 1998 eruption was also observed during 2008. This suggests that this deformation source is the primary contribution for magmas during eruptions. For phase 1 of the eruption from 29 May to 1 Jun 2008, a deep dike intrusion was confined close to the summit, which produced lava flows inside the caldera and down the flank (Fig. 6.9b). For phase 2, InSAR observations show a deep dike orientated radially away
from the summit on the lower flank from 2 Jun to 3 Jun 2008 (Fig. 6.9c) followed by a shallower dike on 3 Jun (Fig. 6.9d).

The morphology and composition of Cerro Azul is unique compared to the other Galapagos volcanoes. Cerro Azul is situated on the leading edge of the Galapagos Platform, and the summit caldera is the smallest of all the active Galapagos volcanoes. The erupted basalts have a wide range of major- and trace-element compositions yet are less evolved than those of neighboring volcanoes [Naumann et al., 2002]. The caldera is similar to those of Fernandina and Wolf with a larger depth-to-diameter ratio [Munro and Rowland, 1996], but unlike the others, there is little evidence from the InSAR data for localized deformation within the caldera. The deformation due to the deep magma reservoir is seen as a broad area of inflation and deflation surrounding the summit. This edifice wide deformation and lack of confined caldera deformation makes Cerro Azul unique from all the others.
Figure 6.9: Cross-sectional schematic sequence for 1998 and 2008 Cerro Azul eruptions (not to scale). a.) 1998 eruption showing location of flank dike and deep deflation source beneath summit, b.) Phase 1 (day 1-2) of the 2008 eruption with deep deflation source beneath the summit and dike on upper flank, c.) Phase 2 (day 4-5) with deep deflation source and deep dike on lower flank, d.) Phase 2 (day 5-6) with deep deflation source and shallow dike on lower flank.
Table 6.1: Model parameters for the summit and flank sources.

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<th>Source</th>
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<th>Length/Width (km)</th>
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<td>Mogi</td>
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<tr>
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<td>2008 Phase 2: Day 5</td>
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<td>0.2</td>
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Chapter 7

Geodetic evidence for interaction between the Galapagos Islands volcanoes

Summary

The Galapagos Islands volcanoes are some of the most active in the world producing 7 eruptions at 4 different volcanoes in the last 20 years. Given the remote location and the difficulties involved with in situ measurements, multi-satellite interferometric synthetic aperture radar (InSAR) time series provide an ideal method for measuring surface displacements by providing both spatially and temporally continuous measurements at these volcanoes for monitoring past and ongoing activity. We present results from small baseline subset (SBAS) InSAR time series that show correlations between neighboring volcanoes and we provide evidence supporting the nature of the observed interactions.

Methods

The InSAR time series are from Radarsat-1, ERS, Envisat and ALOS Palsar data. The Envisat and ALOS interferograms were processed using a modified version of GMTSAR and the Radarsat-1 and ERS interferograms were processed with a modified version of ROI_PAC. Using the small baseline subset (SBAS) method [Berardino et al,
2002; Lanari et al., 2004], the interferograms were used to compute independent LOS time series for each track. In order to increase the temporal frequency, the individual LOS time series were combined by using ascending-descending pairs to calculate the vertical component of motion at the summit of each volcano (see Chapter 2 for details).

The GPS data were processed with GIPSY to obtain daily precision point position solutions for the four continuous stations at the summit of Sierra Negra. Previous studies of Sierra Negra GPS obtained a local reference frame using the closest IGNS GPS station (GALA, and later GLPS), located at the Charles Darwin Research Station in Puerto Ayora on Santa Cruz island [Chadwick et al., 2006]. However, this station was not operational during the time of the 2008 eruption at Cerro Azul, so another method was employed to obtain a local reference frame. To project the GPS time series into a stable...
Galapagos reference frame for this study, the Nazca plate motion was calculated using the UNAVCO plate motion utility and this motion was removed from the GPS displacements. The resulting displacements show the effects of the local changes at Sierra Negra.

**Geodetic Evidence for Volcanic Interactions**

Surface displacement measurements from InSAR time series data provide evidence for four interactions between the volcanoes related to earthquake and eruptive activity. First, in late 2006 and 2007, earthquakes located near Fernandina and Alcedo correlated with changes in the surface displacements at each of these volcanoes (Fig. 7.2a). The earthquake locations form a line running between Fernandina, Alcedo and Darwin and are associated with deflation followed by rapid inflation at the summit of Fernandina (Fig. 7.2c, e1: M4.0 on 22 Dec 2006, e3: M4.7 27 Aug 2007, M5.0 28 Aug 2007) and Alcedo (Fig. 7.2c, e2: M4.0 5 Feb 2007). Second, the 2009 eruption at Fernandina (Fig. 7.2b) marks the end of a period of rapid inflation at Alcedo (occurring since the 5 Feb 2007 earthquake) as well as an end to steady inflation at Wolf that had been happening for over 10 years (Fig. 7.2c, see Fig. 5.2a for extended time series). Third, a rapid deflation event at Alcedo in May 2010 correlates with a lull in rapid inflation occurring at Fernandina since the end of the 2009 eruption (Fig. 7.2c). Fourth, the 2008 eruption of Cerro Azul creates a lull in the rapid inflation of Sierra Negra that was occurring since it last erupted in 2005 (Fig. 7.3).
The nature of the earthquakes affecting Fernandina and Alcedo are largely unknown due to the poor coverage of seismic stations in the area. The earthquake locations are based on the global seismic network, and the uncertainty associated with

**Figure 7.2**: Displacement maps and time series for Wolf, Alcedo, and Fernandina volcanoes. a.) LOS displacement for Envisat track 61 from 14 Jun 2006 to 3 Nov 2007. Black stars mark the global earthquake catalog locations for events with displacements at Alcedo and Fernandina in c. One fringe represents 5 cm of LOS displacement. b.) LOS displacement covering 2009 Fernandina eruption for Envisat track 61 from 11 Apr 2009 to 16 May 2009. c.) Vertical InSAR time series at the summit of each volcano. Dashed blue lines mark earthquake event, blue solid line marks the time span of the 2009 eruption, and dashed-dotted blue line marks the Alcedo subsidence event.

**Earthquake-Fernandina and Earthquake-Alcedo Interactions**

The nature of the earthquakes affecting Fernandina and Alcedo are largely unknown due to the poor coverage of seismic stations in the area. The earthquake locations are based on the global seismic network, and the uncertainty associated with
these events relating to their exact location is high. Centroid moment tensor (CMT) solutions from the global CMT catalog are only available for two of the earthquakes in August 2007. The CMT locations do not match with the catalog locations and show predominately oblique reverse motion, so given these uncertainties, the CMTs are likely to be unreliable.

The surface displacements from InSAR time series provide a good indication of the effects that these earthquakes have on the volcanoes. There is a clear relationship between the earthquakes and displacements at the summits of each volcano. The December 2006 event (e1, Fig. 7.2a) and the August 2007 event (e3, Fig. 7.2a) cause rapid deflation inside the caldera at Fernandina followed by rapid re-inflation (Fig. 7.2c). The August 2007 event shows clear displacement on the flank of Fernandina as well, and an obvious question is whether magma intrusion on the flank causes the earthquake or whether the earthquake provides a scenario where magma can intrude. A similar scenario is present for the February 2007 earthquake (e2, Fig. 7.2a) that marks the change from deflation to rapid inflation within the caldera at Alcedo (Fig. 7.2c). The location of the earthquake is nearly the same as the December 2006 event (e1), but shows no evidence for deformation at Fernandina.

Given the uncertainty in the locations, we do not attempt to interpret the observed deformation based on the earthquake locations, rather we use the timing to help decipher the processes that are occurring at the respective volcanoes. The timing of the earthquakes along with the InSAR time series indicates that the summit is not deflating before the occurrence of the earthquakes. For the December 2006 earthquake, an acquisition on 22 Dec 2006 on Envisat track 54 does not show any sign of subsidence,
but the 23 Dec 2006 acquisition of Envisat track 61 does indicate subsidence. The 28 August 2007 has a similar scenario, the track 61 acquisition on 25 Aug does not indicate subsidence, but the 28 Aug acquisition on track 104 does. The timings indicate that the deflation occurs immediately after the earthquakes, and suggests that the earthquakes are a cause and not an effect. The earthquake that is correlated with displacement at Alcedo (Feb 2007) could be related to intrusion near Alcedo, triggering the rapid inflation that we see occurring there until the eruption at Fernandina in 2009, discussed below.

**Fernandina-Alcedo-Wolf Eruptive Interaction**

The 2009 eruption of Fernandina was accompanied by over 1 m of subsidence in the summit caldera and dike intrusion on the southwest flank (Fig. 7.2b). The pre-eruption data point at Fernandina is related to a propagating sill (Bagnardi, personal communication). Prior to the eruption, the submit caldera at Alcedo and Wolf were inflating, but coincident with the Fernandina eruption, both unexpectedly stopped (Fig. 7.2c). The inflation at Alcedo was rapid following the 2007 earthquake, and the time series show an abrupt stop to this inflation. The inflation does not resume after the eruption, and this continued lull in inflation lasts until a subsidence event in May 2010 marks the beginning of another inflation period (discussed below). Wolf experienced a steady rate of inflation for over 10 years (see Fig 5.2a) that also abruptly stops, appearing as though a “valve” was shut off during the 2009 eruption. The inflation does not appear to resume after the end of the eruption, and the continued lack of inflation lasts for over 2 years up to the end of the time series.
**Alcedo-Fernandina 2010 Interaction**

In May 2010, Alcedo experienced rapid subsidence inside the caldera (dash-dotted blue line, Fig. 7.2c). The pattern of displacement shows that the deformation was concentrated in the southern portion of the caldera (Fig. 5.4f). Given the lack of eruptive activity, one possible explanation might be trapdoor faulting similar to event at Sierra Negra. The global earthquake catalog does not show any large earthquake events during this time period, but the surface displacement is correlated with a decrease of inflation at

---

**Figure 7.3:** Deformation related to the 2008 eruption of Cerro Azul. The days specified in parentheses at the top of a.-c. refers to the time span for GPS vectors and not InSAR. a.) Pre-eruption LOS displacement for Envisat track 61 from 12 Jan 2008 to 26 Apr 2008. One fringe represents 5 cm of LOS displacement (same for b. and c.). Arrows represent the horizontal GPS displacements (scale in lower left, same for in b. and c.). b.) Co-eruptive LOS displacement for Envisat track 61 from 26 Apr 2008 to 5 Jul 2008. c.) Post-eruption LOS displacement for Envisat track 61 from 5 Jul 2008 to 27 Dec 2008. d.) GPS and InSAR vertical time series. e.) GPS time series with a linear trend removed. Blue dashed lines mark the start and end of the eruption.
Fernandina. This lull in inflation is similar to that observed at Sierra Negra during the 2008 eruption at Cerro Azul (discussed below), with rapid inflation before and after the event.

Cerro Azul-Sierra Negra Eruptive Interaction

The 2008 eruption at Cerro Azul lasted 13 days, however, the InSAR time series show that the elastic deformation due to the eruption happened in the first 5 days. The horizontal GPS displacements at Sierra Negra show a clear response to the deformation at Cerro Azul during this time (Fig. 7.3b), but the vertical displacements reveal more details about the effects. At Sierra Negra, the horizontal displacement vectors of the continuous GPS stations all indicate motion toward Cerro Azul, a clear deviation from the roughly radially outward pattern observed in the periods before and after the eruption (Figs. 7.3a and 7.3c, respectively). The horizontal vectors do not point directly toward Cerro Azul and is likely an effect of the dike intrusion on the southeast flank.
The inflation and deflation of a magma body will create a predominantly vertical signal; therefore the vertical motions reveal relevant details about the nature of the interaction due to the underlying magmatic system. Station GV02 located within the summit caldera at Sierra Negra shows the largest vertical displacement of all the GPS stations. The location is closest to the center of a shallow magma body beneath the caldera [Amelung et al., 2000; Yun et al., 2006], and thus shows the best measurement of the effects induced by the eruption at Cerro Azul. The vertical motion shows that the lull in the rapid inflation does not resume immediately at the end of the eruption; rather there is a lag time of at least 1 month (Fig. 7.3e). The other stations (GV01, GV07, and GV08) are located along the edges of the caldera and show a lag in the vertical there as well, but this effect is most clear at GV02.

Table 7.1: Interaction events

<table>
<thead>
<tr>
<th>Event</th>
<th>Date</th>
<th>Type</th>
<th>Volcanoes*</th>
<th>Observation</th>
</tr>
</thead>
<tbody>
<tr>
<td>e1</td>
<td>22 Dec 2006</td>
<td>Earthquake</td>
<td>Fern.</td>
<td>Fern. caldera subsidence</td>
</tr>
<tr>
<td>e2</td>
<td>5 Feb 2007</td>
<td>Earthquake</td>
<td>Al.</td>
<td>Al. caldera subsidence</td>
</tr>
<tr>
<td>e3</td>
<td>28 Aug 2007</td>
<td>Earthquake</td>
<td>Fern.</td>
<td>Fern. caldera subsidence</td>
</tr>
<tr>
<td>Cerro Azul 2008</td>
<td>30 May-17 Jun 2008</td>
<td>Eruption</td>
<td>CA-SN</td>
<td>SN inflation slows</td>
</tr>
<tr>
<td>Fernandina 2009</td>
<td>10 Apr-28 Apr 2009</td>
<td>Eruption</td>
<td>Fern.,Al.,Wo.</td>
<td>Al. and Wo. inflation stops</td>
</tr>
<tr>
<td>Alcedo 2010</td>
<td>May 2010</td>
<td>Unknown</td>
<td>Al., Fern.</td>
<td>Fern inflation slows</td>
</tr>
</tbody>
</table>

*Abbreviations as follows: Fern.=Fernandina, Al.=Alcedo, CA=Cerro Azul, SN=Sierra Negra, Wo.=Wolf

Discussion

The observed displacements discussed above provide evidence for four types of interaction. The first is related to three earthquakes in 2006 and 2007 located near Fernandina and Alcedo. The occurrence of earthquakes close to these two volcanoes results in rapid deflation followed by rapid inflation within the summit calderas. The earthquake locations are close to one another, but only one volcano shows a response
during each. The second interaction is the 2008 Cerro Azul eruption that resulted in a decrease of the ongoing rapid inflation at Sierra Negra. The rapid inflation did not resume until weeks after the end of the eruption. The third interaction is the 2009 eruption at Fernandina that effected nearby Wolf and Alcedo volcanoes. Rapid inflation at Alcedo stops concurrent with the eruption as well as the steady inflation that had been occurring at Wolf for over 10 years. The last interaction is a non-eruptive event at Alcedo in 2010 the resulted in a lag of rapid inflation at Fernandina. This lag is similar to that at Sierra Negra in 2008, with a short period of decrease inflation around the time of the event.

We propose two possible hypotheses that explain the observed interactions. Each hypothesis has different implications about volcanic activity and the underlying magmatic systems. The first hypothesis is based on the volcanoes having a common magma source (most likely in the shallow mantle just beneath the Moho) and the occurrence of eruptive activity or rapid inflation at one volcano will reduce the amount of magma that can be supplied to the other(s) (Fig 7.5a). The second hypothesis is related to magma mush equilibration from changes in stress and the resulting effects on shallow magmatic systems of surrounding volcanoes (Fig 7.5b).

The case for a common magma source is most evident during the 2008 Cerro Azul eruption and the 2009 Fernandina eruption. During both eruptions, the inflation occurring at neighboring volcanoes slowed or stopped. For the 2008 Cerro Azul eruption, we consider that Cerro Azul and Sierra Negra share a common magma source (mantle depths), and magma is able to move from this source into the shallow systems beneath either Cerro Azul or Sierra Negra. We consider that magma is being supplied
from the mantle source at a near constant rate, and before the 2008 eruption, Sierra Negra is receiving the bulk of the magma from the source, evident from the rapid inflation (Fig. 7.5). When the eruption at Cerro Azul occurs, the supply of magma to Sierra Negra is altered, and this causes a lag in the rapid inflation at the summit. Eventually, magma starts going back into Sierra Negra and rapid inflation continues.

For the 2009 eruption at Fernandina, a common magma source can also be considered. Prior to the eruption, Wolf, Alcedo, and Fernandina were inflating, but the eruption brought on changes at Wolf and Alcedo following the eruption. During the eruption, Fernandina subsided rapidly followed by rapid re-inflation, while inflation at Wolf and Alcedo ceased and did not return shortly after. In this scenario, magma is being supplied to all three prior to the eruption, but the eruption at Fernandina emptied the reservoir beneath the volcano, and following the eruption, the magma is going into Fernandina and not the others.

One disadvantage for this hypothesis is the lack of petrologic evidence that supports a shallow common magma source for these volcanoes. The erupted lavas have varying compositions, which suggests that the systems are not as interconnected as would be necessary for these interactions to occur. Evidence has been presented for the presence of a common mantle source of magmas (in the transition zone region) [Hanan and Graham, 1996] and others have shown possible models to explain complex geochemical signatures in the region [Harpp and White, 2001; Kurz and Geist, 1999], but these sources are too deep to explain to the dynamic interactions that we observed.

The magma mush equilibration hypothesis does not necessarily require the interacting volcanoes to have a common magma source in the shallow upper mantle (Fig.
The elastic deformation from the eruption affected the magmatic system at Sierra Negra by causing an extension of the shallow magma bodies, resulting in a drop in pressure. In this scenario, the supply of magma to Sierra Negra may remain steady, but there is a lag in the observed rapid inflation for the system as the pressure rebuilds following the eruption.

The cause of the lull in 2010 at Fernandina is likely the same (Fig. 7.2c). The stress from the elastic deformation at Alcedo affects the shallow magmatic system at Fernandina, causing a drop in pressure that is seen as a lull in the inflation that quickly returns similar to Sierra Negra. In both cases, we observe one volcano subsiding and the neighboring volcano reacting, which suggests that this process is not unique to just a single pair of volcanoes, but is possibly a regular phenomena affecting volcanic activity.

The magma mush equilibration hypothesis is also more applicable to the interactions and changes observed during the earthquake events at Fernandina and Alcedo. These events correlate with changes at one volcano and not necessarily the other. It is possible that the earthquakes could be a result of magmatic processes evident from the uplift on Fernandina in August 2007 (e3, Fig. 7.2), and if there was a common magma source, then one would expect the neighboring volcano to show subsidence, but this is not the case. Rather, these earthquake events (albeit poorly located) have an effect on only one of the volcanoes at a time. This is evidence that stress changes play an important role in how these volcanoes behave.

The magma mush equilibration model is the preferred hypothesis. This model relies on changes within the shallow subsurface of each volcano and does not rely heavily
on mantle depth processes. Changes in local and regional stresses have an effect on the individual magmatic system at each volcano.

**Conclusions**

The Galapagos Islands volcanoes provide a unique opportunity to study many actively deforming volcanoes in close proximity to one another. This setting provides the opportunity to study how volcanoes interact with one another and what effects can influence volcanic behavior. Based on geodetic evidence alone, the stress change hypothesis is the preferred explanation for the observed volcanic interactions. We observed changes during eruptions and also during earthquakes that are difficult to explain only with a common magma source. Further evidence from petrology and other geophysical measurements (i.e. seismic and gravity) would be valuable contributions for distinguishing between the two hypotheses. Improved seismic networks would allow for better locations for earthquake events, and gravity measurements would provide evidence for migration of magma in the shallow subsurface, helping to distinguish between magmatic and hydrothermal deformation. Geochemical analyses could provide evidence of the origin of erupted lavas and help to support or refute the notion of a common magma source for neighboring volcanoes.
Figure 7.5: Schematic cross-section (not to scale) for south Isabella Island showing examples of the two proposed explanations for the volcanic interaction between Cerro Azul and Sierra Negra. Red bodies represent areas of magma (both liquid and mush) and not just liquid magma.
Chapter 8

Conclusion

Space-based geodetic measurements provide a way to measure and analyze the three-dimensional, time-varying deformation at active volcanoes. Combined with other geophysical measurements (i.e. from seismic, gravity, and electromagnetic methods), geodesy provides a way to study the dynamic processes at active volcanoes and track the movement of magma beneath the surface. Geochemical studies of erupted lavas provide details of magma compositions, which are useful for inferring source generation, transport and storage, as well as mixing and differentiation processes. Maintaining ground-based networks on volcanoes in remote areas (such as the Galapagos) has proven to be both difficult and costly, and InSAR provides the only measurements of deformation for many of these locations. Given the easy accessibility and frequency of eruptions, Kilauea provides an ideal location for combining ground-based studies with remote sensing measurements to learn more about the potentials and pitfalls that may arise when only remotely sensed measurements are available. What is learned at Kilauea from applying these techniques and verifying the results with ground-based measurements can be applied to studies in more difficult areas, ultimately providing a higher level of confidence in the measurements.
Kilauea Volcano, Hawaii

We can determine when and where magma intruded or migrated at Kilauea using InSAR, GPS, and seismic data. We used InSAR time series analysis over a 12-year period to describe the activity related to migration and storage of magma beneath the summit of Kilauea. The shallow magmatic system beneath the summit of Kilauea Volcano is best described as an interconnected plexus of magma bodies and pathways. These bodies act as short-term storage reservoirs with a time varying pattern of activity before magma either migrates down the rifts or feeds summit eruptions. In addition to the multiple magma bodies beneath the summit, there is evidence for pockets of magma storage within the east rift zone. Magma from the summit supplies upper east rift intrusions evidenced from rapid deflation of shallow Halemaumau source/

The timing of source activity at the summit before and after the 2007 intrusion revealed a top-down (shallow to deep) process of inflation and deflation. Previous studies have noted this general pattern of top-down inflation before, and the same sequence for deflation is clearly seen in the InSAR time series analysis following the 2007 intrusion. Beginning in late 2003, the shallow source to the NE of Halemaumau Crater inflated first, followed by progressively deeper sill bodies to the south of the caldera leading up to the 2007 intrusion. Inflation of the deep sills (3-4 km) is coincident with seismic swarms in the upper rift zones, providing details about the conduit leading from the summit. The location of the earthquake swarms outlines the area of magma pathways leading away from the summit. The swarms occur as magma enters the rift, and in 2006, magma was not able to propagate further down the east rift and instead
inflated the deep sills to the south of the caldera. This top-down process of inflation and deflation was not observed for the 2011 intrusion.

During the 2007 and 2011 intrusions, the shallow source NE of Halemaumau responded with rapid deflation, evidence that magma from the summit contributed to both intrusions. However, both intrusions have varying pre- and post-intrusion deformation. Inflation of the summit occurred prior to each intrusion, but the amount and extent of inflation was different for both. Prior to 2007, the summit revealed a top-down inflation involving multiple magma bodies with a total of over 35 cm in 3.5 years, and in 2011 only the shallow Halemaumau source inflated a total of 5-10 cm in 1.5 years. In both cases, there was an increase in the rate of inflation and elevated seismicity in the upper east rift zone leading up to the intrusion. There is a striking difference in the deformation at the summit after the intrusions. Following the 2007 intrusion, the shallow caldera source inflated for a very short period. Afterwards, the entire summit area entered a prolonged period of subsidence that lasted until late 2009, when the summit area began to inflate prior to the 2011 intrusion. Following the 2011 intrusion, the shallow caldera source sustained inflation for a period of several months before the inflation waned in later 2011. This same rapid inflation was observed following the 1997 intrusion as the summit sustained inflation for several months until it returned to near pre-intrusion levels. This suggests that the 2007 post-intrusion deformation is abnormal, possibly related to the higher volume of magma that accumulated prior to the intrusion.

After the 2011 intrusion the area around Napau Crater experienced a period of sustained inflation. With the Napau Crater source, we have for the first time a measurement of input into an area of secondary magma storage in the upper east rift
zone. The existence of this deformation source helps explain the observations of highly differentiated lava that have been analyzed by geochemical studies. Previous studies have only seen deflation due to these sources, and knowing how much and how long magma was stored in these bodies is important for mixing models. The measurement of inflation in the east rift is important because it can help determine the size of magma bodies and provide estimates for residence times. It is unclear the pathway that magma followed to this secondary reservoir. It is possible that magma from the summit migrated down the rift to fill this source, or it is possible that magma migrated up from the deep rift. Magmas from the summit and magmas from the deep rift will have different amounts of degassing, so gas measurements in the area could help to constrain the source of the magma.

**Galapagos Islands Volcanoes**

The Galapagos Islands volcanoes provide a unique opportunity to study many actively deforming volcanoes in close proximity to one another. Changes in activity at individual volcanoes are easily seen in the time series, as well as variable patterns of deformation from the displacement maps (Chapter 5). By combining multiple satellites and tracks to generate vertical time series, the temporal frequency of observations is increased which can lead to improved detection of changes at any of the volcanoes. The pattern of deformation is highly non-linear through time for many of the volcanoes, further justifying the importance of frequent observations. For the 2008 eruption at Cerro Azul and the 2009 eruption at Fernandina, acquisition just prior to the eruption and co-eruptive acquisitions provided the means to track the progress of the eruptions. At Cerro
Azul, the frequency of the SAR acquisitions captured the progression of the eruption between two eruptive phases and showed that the deformation due to dike intrusion was over by the day 7 of the eruption.

Identifying the normal behavior of the volcanoes is easily realized with the combined InSAR time series. There are differences in behavior between these seemingly similar volcanoes where Sierra Negra and Alcedo show quite a lot of variation through time, Wolf and Fernandina have an overall increase or inflation; yet only Fernandina and Sierra Negra experienced eruptions. Because the activity at each volcano is so different, it is not possible to use the activity at one as a model for another. Identifying these subtleties leads to a better awareness of unusual activity and assessment of associated hazards.

Interactions between volcanoes are also realized with InSAR time series. There is evidence from the time series that inflation at Sierra Negra wanes during the 2008 eruption at Cerro Azul and continues shortly after. Likewise, the 2009 eruption at Fernandina marks an abrupt end to the rapid inflation at Alcedo since 2007, and the steady inflation at Wolf flattens out to nearly nothing as well. These clear changes in the behavior at neighboring volcanoes means that small stress changes at one volcano will have an effect at others that are in close proximity. From December 2006 to August 2007, a series of earthquakes located between Fernandina and Alcedo coincide with changes and increases in the deformation at each volcano. The earthquakes mark the beginning of rapid inflation at Alcedo in February 2007, and rapid subsidence and re-inflation occur at Fernandina in December 2006 and again in August 2007. Due to the
sparse seismic network in the Galapagos, the nature of the earthquakes in this area is largely unknown, but surface displacements show a clear correlation with these events.

**The Future of InSAR and Volcano Geodesy**

Many active volcanoes are situated in remote, hard-to-reach locations, making it difficult to perform ground-based studies and maintain sensor networks. The spatial and temporal frequency of deformation observations at volcanoes has seen an increase due to the recent advances in InSAR time series analysis. InSAR provides remote measurements of deformation, and in volcanic settings, combining ascending-descending data provides a novel approach to increase frequency of observation and track changes at the surface. This study demonstrated the novel approach of combining InSAR time series from multiple satellites and multiple geometries, or tracks, for monitoring surface displacements in volcanic regions.

The use of InSAR as a method for monitoring volcanic hazards has not been widely used given the low temporal frequency. Historic SAR missions had repeat pass time of 24-46 days (ERS-1/2, Radarsat-1, Envisat, ALOS), but more recent missions have much shorter repeat times (11 days for TerraSAR-X/TanDEM-X and 1-16 days for the COSMO-SkyMed constellation). During the 2008 Cerro Azul eruption (see Ch. 6), a very favorable timing of the eruption with acquisitions of Envisat provided near daily observations for the first 8 days of the eruption using a combination of multiple tracks (or relative orbits) and swaths. Given the lower repeat-pass times of current satellite missions, it is conceivable to have daily acquisitions for any location by combining missions as well as different relative orbits and swaths.
With the increase of current and planned SAR missions, combining multi-satellite time series into a single continuous time series brings InSAR into the realm of monitoring tools, providing an opportunity for volcanic monitoring at unprecedented scales. With the combined time series, the next repeat-pass of the satellite is no longer the limiting factor, and each acquisition (regardless of satellite, track, or swath) can extend the time series. In volcanic island settings such as the Galapagos and Hawaii, each pass of the satellite is more likely to acquire data because there are fewer conflicts when it comes to tasking the satellite. By combining time series from multiple satellites and look geometries, the analyses provide a more frequent measurement of the surface displacement at any one volcano. This approach also emphasizes the need for better acquisition strategies from individual satellite missions to facilitate better temporal coverage.
BIBLIOGRAPHY


Appendix

The Appendix is arranged into two sections. The first section, InSAR Time Series Software, describes software and data used for generating the time series and gives examples for creating publication quality figures using Python. A list of the time series files generated with data from Table 2.1 is given. The second section, Supplemental InSAR Time Series, provides additional time series plots for Kilauea Volcano and the Galapagos Islands Volcanoes.

InSAR Time Series Software

The following is a description of the software and code used to generate InSAR time series and plotting/visualization. The multi-satellite time series are generated using the data and methods detailed in Chapter 2. DEM error correction is done following the work of Fattahi and Amelung [2012]. The HDF5 format is used to store the time series data, and for viewing and exploring the individual LOS time series, a point-and-click Python GUI client (timeseriesviewer.py) was created that allows a user to display displacement maps and time series. The GUI client and a time series dataset (ALOS Path 601 at Kilauea Volcano covering the 2007 and 2011 intrusions) can be downloaded from http://insar.rsmas.miami.edu/users/sbaker/timeseries/.

Installation and Setup for the Python InSAR Time Series Toolkit (pitstk)

The following are the dependencies for running pitstk for SBAS time series analysis:

Python 2.6 or above
Numpy ([http://www.scipy.org](http://www.scipy.org))

Scipy ([http://www.scipy.org](http://www.scipy.org))


h5py ([http://code.google.com/p/h5py/](http://code.google.com/p/h5py/))

If you are using Linux, Mac OS X, or Windows, you can download and compile everything yourself (tedious) or use the Enthought Python Distribution (EPD) ([http://www.enthought.com/products/epd.php](http://www.enthought.com/products/epd.php)) which is free for academic use. EPD is a “kitchen sink” python distribution that already has all of the above modules and more. If you are using Mac OS X, Macports ([http://www.macports.org/](http://www.macports.org/)) provides another convenient method to install the above dependencies.

The following is a list of files used for processing ROI_PAC generated interferograms using the SBAS method (see below for detailed descriptions of each):

- `readfile.py`
- `pitsutils.py`
- `makehdf5.py`
- `sbas.py`
- `timeseriesviewer.py`

To begin, a few environment variable need to be set first:

```
setenv PROCESSDIR /path/to/processing/directory/
setenv TSSARDIR /path/to/timeseries/output/directory/
```

These environment variables specify the locations of the processed interferograms and the output of the SBAS analysis.
These programs work from a template file with the naming convention:

```
ProjectName.template
```

The following descriptions will assume a ProjectName of HawaiiT322EnvA1, PROCESSDIR of /home/sbaker/PROCESS, and TSSARDIR od /home/sbaker/TSSAR.

The data used for the InSAR time series are the geo-coded unwrapped interferograms from ROI_PAC. `makehdf5.py` expects the processed data to be in the directory:

```
$PROCESSDIR/HawaiiT322EnvA1/DONE/
```

with each interferogram in its own directory with the naming convention:

```
IFGRAM_HawaiiT322EnvA1_date1-date2_tbase_pbase
```

where tbase is the time span in days, pbase is the perpendicular baseline, and date format is YYMMDD. The naming convention for the datafiles (and .rsc files) read by `makehdf5.py` are as follows:

```
geo_date1-date2_tbase_pbase.unw
geo_date1-date2_tbase_pbase.cor
geo_date1-date2_tbase_pbase.int
geo_incidence.unw
```

**Example Directory Trees for PROCESSDIR and TSSARDIR:**

```
/home/sbaker/PROCESS/
|-- HawaiiT322EnvA1/
 |  |-- HawaiiT322EnvA1.template
 |  `-- DONE/
 |      |-- IFGRAM_HawaiiT322EnvA1_051221-060125_0035_-0088
 |          |-- geo_051221-060125_0035_00108.cor
 |          `-- geo_051221-060125_0035_00108.cor.rsc
 |               `-- geo_051221-060125_0035_00108.int
```
Detailed File Descriptions:

/readfile.py /pitsutils.py

These are utility libraries/modules that provide methods to the programs below for reading ROI_PAC processed data and some of the plotting routines (shaded relief, colorbars, etc). These are not called directly from the command line, but need to be located in the same directory as the other programs below.

/makehdf5.py

This is used to create a single HDF5 format file with all the interferograms used as the input for sbas.py. It will contain the geocoded cor, int, unw, and incidence files for each interferogram (coregistered radar coordinate files should work as well). The input
for the programs is the template file, which at the moment does not contain any input variables for the program, rather it is used to extract the name of the project and where the data will be located (ie. $PROCESSDIR/HawaiiT322EnvA1/DONE/). The data files will be written out to:

$TSSARDIR/HawaiiT322EnvA1/geo_HawaiiT322EnvA1.h5

Usage: makehdf5.py HawaiiT322EnvA1.template

sbas.py
This program reads in $TSSARDIR/ProjectName/geo_ProjectName.h5 and then flattens each interferogram with a quadratic plane. The flattened interferograms are output to:

$TSSARDIR/HawaiiT322EnvA1/flat_HawaiiT322EnvA1.h5

After flattening, the program computes the time series, ampcormean, and incidencemean, and writes the output to:

$TSSARDIR/HawaiiT322EnvA1/timeseries_HawaiiT322EnvA1.h5

Usage: sbas.py HawaiiT322EnvA1.template

timeseriesview.py
This is a point-and-click GUI viewer for the time series generated by sbas.py:

$TSSARDIR/HawaiiT322EnvA1/timeseries_HawaiiT322EnvA1.h5

It will overlay the time series on a shaded relief using the DEM from the ROI_PAC processing (optional, specified with -d). The figure below shows what the viewer looks like. The icons in the bottom left provide zooming and panning tools to manipulate the displacement map on the left.

Usage: timeseries.py -f timeseries_HawaiiT322EnvA1.h5 -d DEMFILENAME.dem
Figure A1: Point-and-click time series viewer GUI. The image on the left is the total displacement between time1 and time2 and the two on the right are the time series for the points you clicked on. The top is the current point and the lat/lon is displayed and the bottom one is the previous point you clicked.

The following Python code shows how to read the data, plot a time series at a specific lat/lon, and save a PDF:

```python
import datetime, h5py
import numpy as np
import matplotlib.pyplot as plt

lon = -155.2625
lat = 19.4147
h5 = h5py.File('timeseries_HawaiiT601AlosD.h5')
keys = h5['timeseries'].keys()
length, width = np.shape(h5['timeseries'].get(keys[0]))

xtsdict = {'xfirst': float(h5['timeseries'].attrs['xfirst']), 'xstep': float(h5['timeseries'].attrs['xstep']), 'yfirst': float(h5['timeseries'].attrs['yfirst']), 'ystep': float(h5['timeseries'].attrs['ystep']), 'length': length, 'width': width}
tsd = [abs((tsdict['xfirst'] - lon) / tsdict['xstep']) * width for d in keys]
yd = [abs((tsdict['yfirst'] - lat) / tsdict['ystep']) for d in keys]
y = [h5['timeseries'].get(d)[Ydx, Xndx] for d in keys]
plt.plot(dates, y, '^')
plt.ylabel("LOS (m)"")
plt.xlabel("Date")
plt.savefig("timeseries_%f_%f.pdf" % (lat, lon))
```

The code above can be copied into a file (see example.py from website listed above) and run from the command line to facilitate time series figure generation.
List of InSAR Time Series Files

1. timeseries_HawaiiT093EnvA2.h5
2. timeseries_HawaiiT114EnvD5.h5
3. timeseries_HawaiiT136EnvA4.h5
4. timeseries_HawaiiT157EnvD3.h5
5. timeseries_HawaiiT179EnvA6.h5
6. timeseries_HawaiiT200EnvD2.h5
7. timeseries_HawaiiT284AlosA0.h5
8. timeseries_HawaiiT287AlosA3.h5
9. timeseries_HawaiiT291AlosA.h5
10. timeseries_HawaiiT294AlosA10.h5
11. timeseries_HawaiiT299AlosA17.h5
12. timeseries_HawaiiT322EnvA1.h5
13. timeseries_HawaiiT365EnvA3.h5
14. timeseries_HawaiiT386EnvD4.h5
15. timeseries_HawaiiT408EnvA4.h5
16. timeseries_HawaiiT429EnvD2.h5
17. timeseries_HawaiiT451EnvA7.h5
18. timeseries_HawaiiT593AlosD17.h5
19. timeseries_HawaiiT598AlosD10.h5
20. timeseries_HawaiiT599AlosD10.h5
21. timeseries_HawaiiT601AlosD.h5
22. timeseries_HawaiiT602AlosD.h5
23. timeseries_HawaiiT605AlosD3.h5
24. timeseries_HawaiiT608AlosD0.h5
25. timeseries_KilaueaTN24TsxDSM007.h5
26. timeseries_KilaueaTN32TsxASM008.h5
27. timeseries_GalapagosRsatA5.h5
28. timeseries_GalapagosRsatD7.h5
29. timeseries_GalapagosT054EnvD7.h5
30. timeseries_GalapagosT061EnvA2.h5
31. timeseries_GalapagosT061ErsA.h5
32. timeseries_GalapagosT097EnvD4.h5
33. timeseries_GalapagosT104EnvA4.h5
34. timeseries_GalapagosT133AlosA.h5
35. timeseries_GalapagosT134AlosA.h5
36. timeseries_GalapagosT140EnvD2.h5
37. timeseries_GalapagosT140ErsD.h5
38. timeseries_GalapagosT147EnvA6.h5
39. timeseries_GalapagosT376EnvA5.h5
40. timeseries_GalapagosT412EnvD2.h5
41. timeseries_GalapagosT474AlosD.h5
42. timeseries_GalapagosT475AlosD.h5
Supplemental InSAR Time Series

The following is a collection of various InSAR time series at Kilauea Volcano and the Galapagos Islands volcanoes. These figures were generated using the methods described above in the Appendix and also in Chapter 2.

Figure A2: Co-located InSAR and GPS time series at the summit of Kilauea Volcano, Hawaii. The GPS station name is indicated in the lower left corner (see Fig. 4.1 for locations)
Figure A3: Co-located InSAR and GPS time series around Puu Oo at Kilauea Volcano, Hawaii. The GPS station name is indicated in the lower left corner (see Fig. 4.1 for locations). Details for the wiggles in the Radarsat time series (green triangles) are shown on Fig. A4.
Figure A4: Radarsat-1 LOS time series and perpendicular baseline plots at Napau Crater, Kilauea Volcano, Hawaii. a.) descending and b.) ascending pair used for computing vertical shown in Fig. 4.4b. DEM error can be excluded for explaining the wiggles because there is no correlation with the perpendicular baseline history. Phase-unwrap errors related to the Napau crater depression can be excluded as points outside Napau show similar time-series (Fig. A3). As the time-series from 2006 onward (Fig. 4.4b) does not show such variation and there were no major changes with the satellite it is likely that this signal is real.
Figure A5: East-West Galapagos InSAR time series. a., b., c., and e. are plotted at the same vertical scale.
Figure A6: Vertical time series generated by assuming all vertical motion and converting LOS time series (same method as Figs. 2.3b, 2.4b).