Dynamics of Magma Supply, Storage and Migration at Basaltic Volcanoes: Geophysical Studies of the Galápagos and Hawaiian Volcanoes

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DYNAMICS OF MAGMA SUPPLY, STORAGE AND MIGRATION AT BASALTIC VOLCANOES: GEOPHYSICAL STUDIES OF THE GALÁPAGOS AND HAWAIIAN VOLCANOES

By
Marco Bagnardi

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of the University of Miami
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DYNAMICS OF MAGMA SUPPLY, STORAGE AND MIGRATION AT BASALTIC VOLCANOES: GEOPHYSICAL STUDIES OF THE GALÁPAGOS AND HAWAIIAN VOLCANOES

Marco Bagnardi

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Basaltic shields forming ocean island volcanoes, in particular those of Hawai‘i and of the Galápagos Islands, constitute some of the largest volcanic features on Earth. Understanding subsurface processes such as those controlling magma supply, storage and migration at these volcanoes, is essential to any attempt to anticipate their future behavior. Because these processes are hidden beneath the surface, geophysical measurements may represent the best tool to study them. This dissertation presents a series of studies carried out at Hawaiian and Galápagos volcanoes where the dynamics of magma supply, storage and migration are investigated primarily using space-born interferometric synthetic aperture radar (InSAR) measurements of the surface displacement. Other ground-based geophysical methods, such as microgravity, are also used and combined with the InSAR data.

InSAR measurements acquired between 2003 and 2010 at Fernandina Volcano, Galápagos, are used to study the structure and the dynamics of the shallow magmatic system of the volcano (Chapter 3). Spatial and temporal variations in the measured displacements reveal the presence of two hydraulically connected areas of magma storage, and the modeling of the deformation data provides an estimate of their location.
and geometry. The same dataset also provides the first geodetic evidence for two subvolcanic sill intrusions (in 2006 and 2007) deep beneath the volcano’s flank. The lateral migration of magma from the reservoirs during these intrusions could provide an explanation for enigmatic volcanic events at Fernandina such as the 1968 caldera collapse without significant eruption.

Space-geodetic measurements of the surface deformation produced by the most recent eruptions at Fernandina, reveal that all have initiated with the intrusion of subhorizontal sills from the shallow magma reservoir (Chapter 4). A synthetic aperture radar (SAR) image acquired 1–2 h before the start of a radial fissure eruption in 2009 captures one of these sills in the midst of its propagation toward the surface. Galápagos eruptive fissures of all orientations have previously been presumed to be fed by vertical dikes, but these new findings allow a reinterpretation of the internal structure and evolution of Galápagos volcanoes and of similar basaltic shields elsewhere on Earth and on other planets.

A joint analysis of InSAR and ground-based microgravity data acquired at Kīlauea volcano, Hawai’i, between 2009 and the end of 2012 (Chapter 5), allows us to infer the location of a shallow area of magma storage beneath the summit caldera and detect a process of mass increase within the reservoir. This mass accumulation, however, occurred without a significant uplift of the surface and the volume change inferred from the modeling of the InSAR deformation data can account for only a small portion (<10%) of the mass addition responsible for a gravity increase. We propose that this discrepancy between gravity change and deformation could be explained by the replacement of gas-rich magma within the shallow reservoir with denser, outgassed magma. In fact, since
2008, the opening of a new vent within Kilauea’s summit caldera allows magma to convect up to the surface, loose its volatiles content and sink back into the reservoir.

Finally, in Chapter 6 we use InSAR and GPS time-series of the surface displacement to characterize the storage system of the remaining five active volcanoes of the western Galápagos Islands and to estimate volumes and rates of magma supply to the archipelago during the past two decades (1992 – 2011). Together with Fernandina, four other volcanoes, Wolf, Darwin, Alcedo, and Sierra Negra, all have a shallow reservoir within 1-3 km depth, while at Cerro Azul magma is stored at greater depth (~6 km) and no evidence for shallower storage is found. Our results highlight that the rate of magma supply from the mantle hotspot to the Galápagos volcanoes may be an order of magnitude lower (~0.02 km³ yr⁻¹) than that inferred at the Hawaiian volcanoes (0.1-0.2 km³ yr⁻¹). The magma supply rate, however, largely varies through time and seems to be influenced by the occurrence of eruptive and intrusive activity at the volcanoes. On the other hand, eruptions during the past two decades have only occurred at those volcanoes showing the highest rates of magma supply (Sierra Negra, Fernandina and Cerro Azul). A positive feedback between the two processes is therefore possible.
Dedicated to Laura.
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Publications Note

- The content of Chapter 3 was published in *J. of Geoph. Res. – Solid Earth* (Bagnardi and Amelung, 2012).
- The content of Chapter 4 was published in *Earth and Plan. Sci. Lett.* (Bagnardi et al., 2013)
- The content of Chapter 5 is in revision after peer-review. Submitted for publication in *J. of Geoph. Res. – Solid Earth*
Chapter 1: Introduction

Basaltic shields constitute some of the largest volcanic features on Earth and other planets. Understanding how these volcanoes are built and how they evolve is essential to any attempt to anticipate their future behavior. The lack of direct observations of the subsurface processes controlling magma supply, storage and migration, leaves these fundamental aspects far from being sufficiently characterized. Geophysical measurements may represent the best tool to shed light on these hidden processes, especially if different techniques are integrated or if they are combined with information from geological and geochemical studies.

Ocean island volcanoes, in particular those of Hawai‘i and of the Galápagos Islands, are the most famous examples of basaltic shields, and because of their frequent activity represent natural laboratories for the study of this type of volcanism. Despite some similarities and the fact that both the Hawaiian and the Galápagos archipelagos are formed by mantle hotspots, there are important differences in the dynamics of magma supply, storage and migration, between volcanoes at these two locations. Such differences are sometimes large enough that volcanoes of the two archipelagos may represent the end members in the tectonic setting of oceanic intraplate-volcanism.

This manuscript presents a series of studies of Hawaiian and Galápagos volcanoes extending from the analysis of small-scale eruptive and intrusive events to an archipelago-wide study of the dynamics of magma storage and supply. These aspects are investigated primarily using space-born Interferometric Synthetic Aperture Radar
(InSAR) measurements of surface displacements. Other ground-based geophysical methods, such as microgravity, are also used and combined with the InSAR data.

1.1 Magma supply

The amount of magma coming from the mantle and supplied to crustal reservoirs is a fundamental factor in controlling how the volcano grows and behaves in terms of intrusive and eruptive activity [Dvorak and Dzurisin, 1993]. The way magma is supplied, continuously or episodically, can for example control the frequency and style of eruptions. Long-lasting eruptions, such as the one started in 1983 and continuing today at the East Rift Zone (ERZ) of Kīlauea, Hawai‘i [Swanson, 1972; Dvorak and Dzurisin, 1993; Cayol et al., 2000] may only occur if supply is almost continuous. If instead magma arrives in batches, it can trigger eruptions such as that of Mt. Pinatubo, Philippines, in 1991, when the input of fresh basaltic magma into a highly differentiated magma body may have caused the volcano to erupt [Pallister et al., 1992].

The rate of magma supply can be inferred in different ways, but always indirectly. The supply rate over millions of years to an archipelago like the Galápagos Islands can be estimated by studying its isostatic crustal thickness [Ito et al., 1997]. Rates over thousands of years can be determined by dividing the volume of a volcanic edifice by its age, assuming that precise constraints on the ages are available [Naumann and Geist, 2000]. Models of heat flow measured at the surface and the amount of melt needed to maintain such flow can provide estimates for both short- and long-term rates of magma supply to a volcano [Bacon, 1982; Francis et al., 1993]. Measurements of the effusion rate of long-lasting eruptions, such as those at Kīlauea [Swanson et al., 1972], can be
used as a proxy for the amount of magma entering the plumbing system of a volcano.

Similar results can also be obtained using gas emissions. For example, *Lowestern and Hurwitz* (2008) used CO$_2$ degassing to infer the amount of mantle-derived basalt entering the storage system beneath Yellowstone Caldera.

Finally, surface deformation data can be extremely valuable in calculating magma supply rates at timescales of interest to humans [e.g., *Lu et al.*, 2005; *Fournier et al.*, 2009; *Poland et al.*, 2012], even at those volcanoes where eruptions are not frequent and/or have short duration (e.g., Galápagos [*Chadwick et al.*, 2006]). In this study, surface deformation data acquired at the six active volcanoes forming the islands of Fernandina and Isabela, Galápagos, are used to infer rates of magma supply to crustal reservoirs at each volcano. By summing the contribution of all six volcanoes, the amount of magma that was supplied to the western portion of the Galápagos archipelago during the past two decades (1992-2011) is determined. These results indicate that the current rate of magma supply to the Galápagos archipelago is an order of magnitude lower than that measured in Hawai‘i [Swanson *et al.*, 1972; *Poland et al.*, 2012], and also much lower than the rate inferred from studies of the isostatic crustal thickness [Ito *et al.*, 1997]. The detailed analysis can be found in Chapter 6.

### 1.2 Magma storage

Most volcanoes are characterized by the presence of crustal reservoirs (more commonly defined as magma chambers) where magma, coming from its mantle source, is stored before intruding the volcano’s flanks and erupting at the surface. Inside the reservoirs magma can cool, fractionate, degas, and undergo chemical and physical
transformations. The lack of direct observations of active magma chambers and the use of conceptual models based on the studies of plutons, erupted rocks and geophysical data, has led to a variety of definitions of their physical state, geometry, etc.

Geophysical measurements, especially geodetic and seismic, are the tools most commonly used to infer location, geometry and volume of areas where magma is stored. Earthquake hypocenters, and their absence in certain areas, can be used to outline regions at very-high temperatures likely representing zones of magma storage [e.g., Scandone and Malone, 1985]. Further inferences can be made by jointly analyzing earthquake locations and the variations in the velocity of seismic waves [e.g., Lin et al., 2014]. Deformation measurements can instead be analyzed using analytical [e.g., McTigue, 1987; Yang et al., 1988; Fialko et al., 2001] and numerical models [Masterlark et al., 2012] to infer the location, the geometry and changes in pressure within the reservoirs. Surface deformation data, however, are not sensitive to the reservoir volume. Microgravity measurements, which are sensitive to subsurface changes in density, can instead constrain mass flow and help detect areas where magma accumulation is not associated with deformation of the surface [e.g., Johnson et al., 2010].

Magma reservoirs are best characterized when multiple datasets are jointly analyzed [e.g., Peltier et al., 2009] or combined in physics based models that can estimate not only reservoir depth and location but also parameters such as reservoir volume and volatile content [Anderson and Segall, 2011].
1.2.1 The depth of magma reservoirs

The depth at which magma accumulates in reservoirs can largely vary depending on different factors. Buoyancy forces that allow magma to rise from a mantle source also control where its further ascent is prevented by the low density of the overlying crust [Ryan, 1987]. Temperature of the surrounding rocks, so that they are too cold to deform significantly, can also represent a mechanical barrier for the ascent of buoyant magma [Burov, 2003].

The depth of magma accumulation at basaltic volcanoes (Table 1.1) can vary between tens of kilometers beneath the surface, as in the case of Hekla, Iceland [Ofeigsson et al., 2011], to less than a kilometer, as modeled at Kīlauea [Cervelli and Miklius, 2003], suggesting a possible tectonic control on the depth of magma reservoirs. Within the same tectonic setting, a weak correlation is present between reservoir depths of ocean island volcanoes and the age of the lithosphere on which they grow (Figure 1.1). A stronger direct correlation exists instead between depths and the rate of magma supply (Figure 1.2), with the reservoirs becoming deeper when supply rates decrease [González et al., 2013]. Some volcanoes, as will be shown in this study (Chapter 3, 5 and 6), can also have multiple vertically stacked reservoirs, a consequence of density variations between the magmas they contain or because of the presence of thermo-mechanical barriers.
**Figure 1.1:** Depth of magma reservoirs (shallow in red and deep in blue) at different ocean island volcanoes compared to the age of the underlying oceanic lithosphere. A weak trend toward deeper magma reservoirs as the age of the oceanic lithosphere increases seems to be present with an exception for volcanoes of Hawai‘i [González et al., 2013].

![Figure 1.1](image1.png)

**Figure 1.2:** Depth of magma reservoirs (shallow in red and deep in blue) at different ocean island volcanoes compared to magma supply rate. Both shallow and deep reservoirs are deeper when magma supply rate is lower [González et al., 2013].

![Figure 1.2](image2.png)
Table 1.1. Depth and number of magma reservoirs at selected basaltic volcanoes.

<table>
<thead>
<tr>
<th>Location</th>
<th>Reservoir depth (km)</th>
<th>Number of reservoirs</th>
<th>Tectonic setting</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>Hawai‘i</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Kilauea</td>
<td>&lt;1.0–5.0</td>
<td>2+</td>
<td>Intraplate</td>
<td>Eaton, 1959, 1962; Dvorak et al., 1983; Davis, 1986; Delaney et al., 1990, 1993; Cervelli and Miklius, 2003; Poland et al., 2009; Montgomery-Brown et al., 2010; Baker and Amelung, 2012; Poland et al., 2012; Lundgren et al., 2013; this study.</td>
</tr>
<tr>
<td>Mauna Loa</td>
<td>3.0–8.0</td>
<td>1+</td>
<td>Intraplate</td>
<td>Decker et al., 1983; Lockwood et al., 1987; Johnson, 1995a; Miklius et al., 1995; Okubo, 1995; Miklius and Cervelli, 2003; Amelung et al., 2007.</td>
</tr>
<tr>
<td>Galápagos</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Fernandina</td>
<td>1.0–5.0</td>
<td>2</td>
<td>Intraplate</td>
<td>Amelung et al., 2000; Geist et al., 2006a; Chadwick et al., 2011; Bagnardi and Amelung, 2012; this study.</td>
</tr>
<tr>
<td>Sierra Negra</td>
<td>1.9–2.1</td>
<td>1</td>
<td>Intraplate</td>
<td>Amelung et al., 2000; Jónsson et al., 2005; Chadwick et al., 2006; Geist et al., 2006a; Yun et al., 2006; Geist et al., 2008; Jónsson, 2009.</td>
</tr>
<tr>
<td>Cerro Azul</td>
<td>5.0–6.0</td>
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<td>Intraplate</td>
<td>Amelung et al., 2000; Baker, 2012.</td>
</tr>
<tr>
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<td>Amelung et al., 2000.</td>
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<td>Alcedo</td>
<td>2.2</td>
<td>1</td>
<td>Intraplate</td>
<td>Hooper et al., 2007.</td>
</tr>
<tr>
<td>Wolf</td>
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<td>1</td>
<td>Intraplate</td>
<td>Amelung et al., 2000; Geist et al., 2005.</td>
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<tr>
<td>Piton de la Fournaise</td>
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<td>Peltier et al., 2009</td>
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<tr>
<td>El Hierro</td>
<td>4.0–9.5</td>
<td>2</td>
<td>Intraplate</td>
<td>González et al., 2013.</td>
</tr>
<tr>
<td>Gran Canaria</td>
<td>2.0–4.0</td>
<td>1</td>
<td>Intraplate</td>
<td>Hansteen et al., 1998.</td>
</tr>
<tr>
<td>Teide</td>
<td>4.0–6.0</td>
<td>1</td>
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<td>Andújar et al., 2010.</td>
</tr>
<tr>
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<td>Divergent boundary</td>
<td>Oféigsson et al., 2011.</td>
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<tr>
<td>Askja</td>
<td>14.0–22.0</td>
<td>1</td>
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<td>Sturkell et al., 2006b; Hooper et al., 2008.</td>
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<tr>
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<td>2+</td>
<td>Divergent boundary</td>
<td>Pedersen et al., 2006; Sturkell et al., 2003; Sigmundsson et al., 2010.</td>
</tr>
</tbody>
</table>

(Continues in the following page)
Table 1.1 Depth and number of magma reservoirs at selected basaltic volcanoes. (Continues from previous page)

<table>
<thead>
<tr>
<th>Reservoir depth (km)</th>
<th>Number of reservoirs</th>
<th>Tectonic setting</th>
<th>References</th>
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<td></td>
<td></td>
</tr>
<tr>
<td>Okmok 2.6–3.5</td>
<td>1</td>
<td>Convergent boundary</td>
<td>Lu et al., 2005, 2010.</td>
</tr>
<tr>
<td>Seguam 2.5–6.0</td>
<td>1</td>
<td>Convergent boundary</td>
<td>Lee et al., 2013</td>
</tr>
</tbody>
</table>

1.2.2 The geometry of magma reservoirs

Given the ambiguity in the definition of magma chamber, several geometries have been proposed based on different observations. Most reservoirs presumably develop from repeated intrusions of sills, and follows that the initial shape is sill-like (Figure 1.3a). If magma volume and pressure are large enough, the layers above and below the sill can bend transforming the sill-like reservoir geometry into a thick, oblate-shaped chamber. Through repeated and sustained intrusions, some of these reservoirs eventually become spherical (Figure 1.3b) or even prolate-ellipsoidal (Figure 1.3c) [Gudmundsson, 2012 and references therein].

Spherical and ellipsoidal magma reservoirs are favored in geophysical models, primarily because such shapes are simple to visualize and model [e.g., Davis, 1986; McTigue, 1987; Yang et al., 1988]. Even if reservoirs have more complex geometries, as for example inferred at Seguam, Aleutian Islands [Masterlark and Lu, 2004], the stress field induced by a plexus of dikes and sills resembles that of an ellipsoidal source [Kühn and Dahm, 2008]). Also, a magma chamber that originally has irregular boundaries is thermally (and mechanically) unstable (Figure 1.3d) [Gudmundsson, 2012]. Asperities in
the shape of the reservoir that project into the magma tend to melt, and the magma-filled notches that project into the host rock tend to solidify, resulting in a smoother geometry.

**Figure 1.3:** Schematic illustration of the various possible shapes of magma chambers. (a) Roughly oblate ellipsoidal or sill-like chambers are presumably the most common chamber geometry. They are particularly common at mid-ocean ridges, but have been detected seismically beneath many volcanoes. (b) Roughly spherical magma chambers may be common, particularly at the later stages of the chamber evolution. (c) Roughly prolate ellipsoidal chambers may exist beneath some volcanic edifices, particularly comparatively narrow cones with steep slopes. (d) Chambers with very irregular boundaries (surfaces) are thermally and mechanically unstable and tend to smooth out the irregularities (modified after Gudmundsson, 2012).

1.2.3 New insights on magma storage at Hawaiian and Galápagos volcanoes

Throughout this study, deformation data acquired at volcanoes of the Galápagos Islands and Hawai‘i, are used to estimate magma reservoir locations, geometries and
magma volume variations through time. In Chapter 3 we characterize the storage system of Fernandina volcano, Galápagos, and infer the presence of two hydraulically connected reservoirs at crustal depths, which undergo periods of both inflation and deflation directly correlated with the eruptive and intrusive activity. In Chapter 5 deformation data measured at the summit of Kīlauea volcano, Hawai‘i, are combined with microgravity measurements acquired in the same area. The microgravity data reveal a hidden process of significant mass accumulation within a shallow reservoir that is not reflected in the deformation signal measured at the surface. Finally, InSAR measurement acquired at the six active volcanoes of the western Galápagos Islands are used in Chapter 6 to characterize the magma storage system of each volcano, a step also necessary to infer the rate of magma supply to the archipelago (see Chapter 1.1). All six volcanoes show evidence of volume changes within shallow crustal reservoirs located beneath the summit calderas and with depths ranging between ~1 and ~6 km.

1.3 Magma migration

The movement of magma away from the crustal reservoirs towards the surface of basaltic shields usually occurs in the form of tabular intrusions (dikes), as suggested by their abundance in eroded volcanoes [Gudmundsson, 1983, 2006; Walker, 1987] or by geophysical studies [e.g., Pollard et al., 1983]. Dike propagation can occur both vertically and horizontally and is controlled by numerous factors such as magma pressure, the preexisting stress field, magma viscosity, and host rock properties [Rubin, 1995; Taisne and Jaupart, 2009; Traversa et al., 2010].
When a dike is emplaced, it characteristically propagates perpendicular to the least compressive stress in the crust [Nakamura, 1977; Rubin and Pollard, 1988]. Several physical processes can control the characteristics of the stress field within and around the volcanic edifice and control the orientation of dikes (Figure 1.4). More than one of these factors may also act simultaneously, producing complex dike patterns [Acocella and Neri, 2009 and reference therein].

Figure 1.4: Main features, seen as end-members, controlling dike patterns at volcanoes. Dike patterns at given volcanoes result from the features listed in the upper part of the figure, in order of importance [Acocella and Neri, 2009].

The emplacement of dikes is commonly associated with deformation of the surface and seismicity. The high resolution of InSAR data offers a unique opportunity to characterize the geometry of intrusions as demonstrated by numerous recent studies [e.g.,
The analysis of deformation measurements at Fernandina volcano, Galápagos, provides new and important insights into the dynamics of magma migration at the Galápagos volcanoes. In Chapter 3 we present the first geodetic evidence for deep subvolcanic intrusions in the Galápagos that may provide an explanation for enigmatic events such as the rapid uplift at Urvina Bay in 1954 [Couffer, 1956; Richards, 1957] and the caldera collapse without a significant eruption at Fernandina in 1968 [Simkin and Howard, 1970; Filson et al., 1973]. InSAR measurements spanning the three most recent eruptions at Fernandina provide instead a mean to study paths of magma migration prior to and during eruptive events. Chapter 4 presents a detailed analysis of these eruptions leading to a new and important interpretation of how the Galápagos volcanoes grow and evolve. Previous models of magma migration at Galápagos volcanoes [e.g., Chadwick and Dieterich, 1995] are based on the assumptions that fissure eruptions are fed by subvertical dikes. We instead show evidence that all types of eruptions, from either circumferential or radial fissures, are initiated by the intrusion of subhorizontal sills propagating from the shallow storage system of Fernandina volcano. These results can also explain the characteristic pattern of eruptive fissures common to all the western Galápagos volcanoes and offer the opportunity to forecast type and location of a future eruption at Fernandina volcano.
1.4 Study areas

1.4.1 Galápagos Islands

The Galápagos Islands owe their origin to a mantle hotspot now centered at the western edge of the archipelago, beneath Fernandina and Isabela islands (Figure 1.5). The Galápagos lie ~100 km south of the Cocos-Nazca spreading center on the Nazca plate, which moves eastward at 5.1 cm/yr. Volcanic activity related to the hotspot is currently focused at the seven shield volcanoes forming Isabela and Fernandina islands.

More than sixty eruptions have been recorded in the western Galápagos since the end of the eighteenth century [Simkin and Siebert, 1994], but the actual number of events is likely to be twice as large given that a much higher rate of activity has been recorded since the islands were permanently inhabited in the 1950s. Historical eruptions have mainly occurred at Fernandina, Sierra Negra, Cerro Azul and Wolf, while only one event has been recorded at Alcedo and Darwin, and none at Ecuador.

A great contribution to the understanding of the volcanic activity in the Galápagos Islands has been provided by the application of space-geodetic measurements. InSAR data have revealed that all but one of the seven volcanoes on the islands of Isabela and Fernandina experienced surface deformation during 1992-99 (Figure 1.5 [Amelung et al., 2000]). Other studies that used InSAR and GPS data have added important constraints on specific eruptive and volcano-tectonic events [e.g., Jónsson et al., 1999; Jónsson et al., 2005; Yun et al., 2006; Chadwick et al., 2006; Geist et al., 2006a; Hooper et al., 2007; Chadwick et al., 2011; Baker, 2012].
Figure 1.5: Observed deformation at Isabela and Fernandina (main image), the westernmost islands in the Galápagos archipelago (inset). The map shows line-of-sight (LOS) ground displacement towards the ERS-1 and ERS-2 radar satellites measured in interferograms spanning 5 years from 1992 to 1997 (Ecuador, Wolf, Fernandina) and 1992 to 1998 (Darwin, Alcedo, Sierra Negra and Cerro Azul). Each color cycle represents 20 cm displacement. Areas where the interferometric coherence is lost are shown in gray [from Amelung et al., 2000].
1.4.2 Kīlauea Volcano, Hawai‘i

Kīlauea lies at the southeastern edge of the Hawaiian-Emperor chain of volcanoes and seamounts (Figure 1.6), which stretches for more than 6,000 km across the Pacific Ocean and is formed by the passage of the Pacific Plate (at a current rate of ~9 cm/yr) over a hotspot. Kīlauea and Mauna Loa on the island of Hawai‘i, the southernmost island of the Hawaiian archipelago, offer the best opportunity to study active volcanism at ocean islands given their accessibility, frequent activity and long-term record of observation and instrumental monitoring [Kauahikaua and Poland, 2012].

Figure 1.6: Bathymetric map showing the Hawaiian and Emperor seamount chains. Inset shows the Island of Hawai‘i and surface deformation measured by InSAR (Radarsat-1 satellite) between 2002 and 2005 (modified after Amelung et al. [2007]).
A general model of the magmatic system of Kilaeua was already proposed in 1960 [Eaton and Murata, 1960] and remains largely valid today (Figure 1.7a). Magma is generated in the underlying mantle at a depth greater than 80 km and buoyantly ascends in a subvertical conduit to areas of storage that are one to a few kilometers beneath the summit. From here magma can be erupted at the summit or can migrate into two rift zones that radiate to the east and southwest, and possibly be erupted there, tens of kilometers from the summit [e.g., Poland et al., in press] (Figure 1.7b).

Figure 1.7: a) Schematic cross section through an idealized Hawaiian volcano [Eaton and Murata, 1960; Tilling and Dvorak, 1993]. b) Structure of Kilaeua’s subsurface magma plumbing system based on geophysical data, as proposed by Poland et al., in press; H=Halema‘uma‘u reservoir, K=Keanakāko‘i reservoir, SC=south caldera reservoir, SWRZ=southwest rift zone. Plan view gives the relations of magma pathways to surface features and topography in the vicinity of Kilaeua Caldera.
Nearly all geophysical, geochemical and geological techniques have been applied at Kīlauea. Among these, geodetic studies have been critical to identifying zones of magma storage, magma supply to the volcano, relationships between volcanic activity and tectonics and changes in behavior preceding and accompanying intrusions and eruptions [Poland et al., in press, and reference therein]. In particular, studies of InSAR data have recently provided clear evidence of distinct magma storage areas beneath the summit [Baker and Amelung, 2012; Poland et al., in press]. They also offered unprecedented spatial resolution in studies of small-scale deformation [Richter et al., 2012], allowed a precise characterization of intrusions in the ERZ [Lundgren et al., 2013], and recorded displacements associated with volcano-tectonic activity [Wauthier et al., 2013].

Finally, integrated analysis of deformation and microgravity data have yielded important insights into mass change within the volcano [Dzurisin et al., 1980; Jachens and Eaton, 1980; Johnson, 1992; Kauahikaua and Miklius, 2003; Johnson et al., 2010; Carbone and Poland, 2012; Carbone et al., 2013], as will also be demonstrated in this study (Chapter 5).
Chapter 2: Methods and data

Detailed descriptions of data and methods are provided in each of the subsequent chapters, however, here follows a brief introduction to the InSAR methodology and dataset, to microgravimetry, and to the techniques used to process, analyze and interpret the data.

2.1 Interferometric synthetic aperture radar (InSAR)

2.1.1 InSAR method

InSAR is a remote sensing technique used to measure ground displacements between two epochs. Synthetic Aperture Radar (SAR) images record the electromagnetic wave travel time between the emitting/receiving source mounted on an orbiting satellite, or aircraft, and the ground. The phase difference—usually referred to as “interferogram”—between two SAR images of approximately the same area and acquired at two different times, provides a measurement of the surface displacement in the time spanned by the images [Hanssen, 2001; Massonnet and Feigl, 1998]. This displacement is measured along the line-of-sight (LOS) direction, between the satellite and the ground. Other sources other than ground deformation can produce phase differences between two epochs (e.g., variations in atmospheric water vapor, incomplete removal of the topographic component to the signal, error in the estimation of the satellite position, thermal effects, etc.) but different techniques are available to estimate and remove them from the interferograms [e.g., Fattahi and Amelung, 2013].
2.1.2 InSAR data

The studies presented here utilize space-borne SAR data acquired by multiple satellites mounting different sensors (X-, C- and L-band, wavelengths: 3.1 cm, 5.6 cm and 23.6 cm respectively). Data is from both ascending and descending orbital passes of the satellites with different look angles. SAR images are from the European Space Agency (ESA) ERS-1, ERS-2, and Envisat satellites, the Japanese Aerospace Exploration Agency (JAXA) JERS-1 and ALOS satellites, the German Space Agency (DLR) TerraSAR-X satellite and the Italian Space Agency (ASI) Cosmo-SkyMed satellite constellation. The complete dataset consists of 535 acquisitions from 17 tracks for studies of the Galápagos Islands and of 176 acquisitions from 4 tracks for the study of Hawai‘i (see Table 2.1). The number of SAR images for each track does not reflect the total number of available acquisitions, but the number of images used here. In fact, some of the SAR acquisitions were discarded because of high noise. In each chapter a detailed description of the specific portion of the dataset used is provided.

Interferograms are generated using different software: Gamma SAR Processor and Interferometry [Werner et al., 2000], ROI_PAC SAR Software [Rosen et al., 2004] and GMTSAR InSAR processing system [Sandwell et al., 2001].

2.1.3 InSAR time series analysis

In this study, InSAR time series analysis is performed when the temporal evolution of the surface displacement is used to gather information on the dynamics of magma storage and supply. The Small Baseline Subset (SBAS) method [Berardino et al., 2002; Lanari et al., 2007; Fattahi and Amelung, 2013], proven to be successful in the studies of volcano
deformation [e.g., Lundgren et al., 2004; Casu et al., 2006; Baker and Amelung, 2012], is applied to InSAR data from different satellites. The algorithm inverts a large number of interferograms and relies on the redundancy of data to determine the surface displacement through time.

The selection of the most suitable interferometric pairs for the InSAR time series generation is performed following a multistep approach based on the key assumption of the SBAS method, for which small spatial and temporal separation between SAR acquisitions minimize decorrelation and maximize the temporal coherence of pixels [Pepe and Lanari, 2006]. For the specific steps and threshold values used in this selection we refer to the specific studies in which the SBAS technique is used (Chapter 3, 5 and 6).

<table>
<thead>
<tr>
<th>Area of coverage</th>
<th>Satellite</th>
<th>Pass</th>
<th>Track</th>
<th>Beam</th>
<th>Date Span</th>
<th>SAR images</th>
</tr>
</thead>
<tbody>
<tr>
<td>Galápagos Islands</td>
<td>ERS-1/2</td>
<td>D</td>
<td>140</td>
<td>S2</td>
<td>1992/06/15 – 2007/10/04</td>
<td>17</td>
</tr>
<tr>
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<td>ERS-1/2</td>
<td>D</td>
<td>412</td>
<td>S2</td>
<td>1992/09/12 – 2009/10/27</td>
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<td>2003/01/23 - 2010/06/10</td>
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<tr>
<td>Galápagos Islands</td>
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<td>2007/08/05 - 2010/08/13</td>
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<td>N/A</td>
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</table>

D = Descending; A = Ascending.
Once an optimized, high quality (high signal-to-noise ratio) connected network of interferometric pairs is formed, we use the University of Miami-RSMAS InSAR time series software package [Gourmelen et al., 2010; Fattahi and Amelung, 2013] to retrieve the temporal evolution of surface displacement. Each LOS displacement time series is relative to a single pixel that exhibits high coherence and is located in an area where surface deformation is minimum throughout the entire studied-period.

2.2 Microgravimetry

Gravity measurements are an important component in the study of active volcanoes since changes in earth’s gravity reflect mass-transport processes at depth [Dzurisin, 2003; Battaglia et al., 2008]. The ability of gravity measurements to detect mass flows is greatly enhanced if gravity is analyzed and modeled together with ground deformation data. In fact, numerous studies have demonstrated that combined time-dependent geodetic and gravimetric measurements can detect subsurface mass redistribution long before other eruption precursors appear [e.g., Dzurisin, 2003; Rymer and Williams-Jones, 2000; Gottsmann and Rymer, 2002; Johnson et al., 2010; this study].

2.2.1 Microgravity data acquisition and reduction

Relative microgravity readings obtained at individual benchmarks and repeated in time-lapse surveys can be used to build gravity time series of measurements relative to a reference point located outside the area of interest. Individual readings at a benchmark need to be adjusted for the effect of solid-earth tides, ocean-loading and gravimeter drift [Battaglia et al., 2008 and reference therein].
To extract the gravity signal produced by a subsurface mass and/or density change, gravity residuals must be quantified [Eggers, 1987]. The residual gravity change at each benchmark \( \Delta g_r \) is given by

\[
\Delta g_r = \Delta g_{obs} - \gamma u_z - \Delta g_{def} - \Delta g_{wt}
\]

where \( \Delta g \) represents a gravity difference over time at a specific site [for example, \( \Delta g_{obs} = \Delta g_{obs}(t_2) - \Delta g_{obs}(t_1) \)], \( \gamma \) is the free-air gravity gradient (theoretical value=-308.6 \( \mu \)Gal m\(^{-1} \)), \( u_z \) is the vertical displacement (positive for uplift and negative for subsidence), \( \Delta g_{def} \) is the Bouguer effect of deformation, and \( \Delta g_{wt} \) is the effect of groundwater-table variation [Battaglia et al., 2008].

### 2.3 Analytical modeling

All the studies presented here use analytical solutions that offer a closed-form description (expressed in terms of a finite number of known functions) of sources of deformation and gravity changes related to volcanic activity [Mogi, 1958; Okada, 1985, 1992; McTigue, 1987; Yang et al., 1988; Fialko et al., 2001; Battaglia et al., 2008]. These solutions are used to fit the data and interpret the origin of the measured changes. Although these models are based on a number of simplifications (e.g., the assumption that the crust is homogenous, isotropic, elastic and flat) that make the set of differential equations describing the problem tractable, their use together with high quality datasets can yield valuable insights into the nature of the volcanic activity [Battaglia et al., 2013].

We characterize the magmatic sources in terms of their location, geometry and variations in strength (e.g., pressure, volume changes) using MATLAB functions and scripts part of the Miami InSAR Modeling and Interpretation Code geodmod. The
software implements solutions for surface deformation due to an expanding or contracting point source [Mogi, 1958], a finite pressurized spherical source [McTigue, 1987], a dipping prolate ellipsoid [Yang et al., 1988], a penny-shaped crack [Fialko et al., 2001a], and a rectangular strike-slip, dip-slip and tensile dislocation (and their combination) source [Okada, 1985, 1992]. A graphic representation of these geometries and of the source parameters is shown in Figure 2.1.

Figure 2.1. Geometry and parameters for source models. (a) A spherical source is described by four parameters: two for location $(x_0, y_0)$, depth $d$, and volume change $\Delta V$. (b) Prolate spheroids need seven parameters: two for location $(x_0, y_0)$, depth $d$, volume change $\Delta V$, aspect ratio $A$, dip angle $\phi$, and strike angle $\theta$. (c) A horizontal sill-like source is determined by five parameters: two for location $(x_0, y_0)$, depth $d$, volume change $\Delta V$, and radius $b$. (d) A dike is represented by six parameters: two for location $(x_0, y_0)$, depth $d$, rectangular dike length $L$ and width $W$, and dip angle $\phi$. We can represent the strike-slip along the fault by the dislocation vector $U_1$, the dip-slip by $U_2$ and the tensile opening component by $U_3$ (after Okubo, 1992; Battaglia et al., 2008).
2.3.1 Inversion method

Sources of surface deformation are inferred from the non-linear inversion of InSAR LOS displacements measured in either single interferograms or SBAS timeseries-analysis results. To decimate the number of data points and reduce their redundancy, the subsampling of each LOS-displacement data vector is performed using the *quadtree* algorithm [Jónsson et al., 2002]. For all points is then assumed unit variance and the quality of the fit is assessed using the normalized root-mean-square error (RMSE) between observed and modeled displacements, and defined as follows:

\[
RMSE = \sqrt{\frac{1}{N} \sum_{i=1}^{N} (d_i - m_i)^2}
\]

where \(d\) is the data, \(m\) is the model prediction, and \(N\) is the number of data points.

The optimal solutions are estimated together with the posterior probability density distributions of the parameters, as indicators of resolution and uncertainties, using the Monte-Carlo based Gibbs sampling (GS) algorithm [Brooks and Frazer, 2005].

2.4 Finite element modeling

To investigate stress and strain variations caused by the intrusion of planar magma bodies (sills) at Fernandina Volcano, Galápagos, during the 1995 and 2009 radial fissure eruptions (see chapter 4), we use the finite element method (FEM). With this approach we can overcome the flat half-space limitation imposed by analytical solutions and consider a more realistic topography of the volcano. Although our model is homogeneous, FEM models can also allow the implementation of mechanical
heterogeneities within the medium when information on their distribution is available, for example from geological data or seismic tomography [e.g., Currenti et al., 2008].

The basis of the finite element method lies in the approximation of a structure by an assembly composed of a finite number of structural components. These components, or elements, are joined at nodal points [Dieterich and Decker, 1975]. In our study we build a three-dimensional model of the volcano using the Structural Mechanics module of the commercial FEM software COMSOL Multypysics (www.comsol.com), and discretize the entire volume into trapezoidal elements. A linear elastic and isotropic body is considered, with elastic properties set to values suitable for extrusive and intrusive volcanic rocks (see chapter 4 for specific values). A 3D visualization of the model geometry is presented in Figure 2.2.

Figure 2.2. Configuration of the finite element model of Fernandina Volcano, Galápagos. (a) A 100 km x 100 km x 100 km volume is considered with the following boundary conditions: the Earth’s surface is treated as a free surface (displacement in all directions is allowed); at all the other boundaries, displacement is prohibited normal to the boundaries (zero displacement in z direction at the bottom boundary, zero displacement in x direction at the eastern and western boundary, zero displacement in y direction at the northern and southern boundary). (b) Finite element mesh for the inner portion.
Chapter 3: Space-geodetic evidence for multiple magma reservoirs and subvolcanic lateral intrusion at Fernandina Volcano, Galápagos Islands

3.1 Summary

Using Interferometric Synthetic Aperture Radar (InSAR) measurements of the surface deformation at Fernandina Volcano, Galápagos (Ecuador), acquired between January 2003 and September 2010, we study the structure and the dynamics of the shallow magmatic system of the volcano. Through the analysis of spatial and temporal variations of the measured line-of-sight displacement we identify multiple sources of deformation beneath the summit and the southern flank. At least two sources are considered to represent permanent zones of magma storage given their persistent or recurrent activity. Elastic deformation models indicate the presence of a flat-topped magma reservoir at ~1.1 km below sea level and an oblate-spheroid cavity at ~4.9 km b.s.l. The two reservoirs are hydraulically connected. This inferred structure of the shallow storage system is in agreement with previous geodetic studies and previous petrological analysis of both subaerial and submarine lavas. The almost eight-year-long observation interval provides for the first time geodetic evidence for two subvolcanic lateral intrusions from the central storage system (in December 2006 and August 2007). Subvolcanic lateral intrusions could provide the explanation for enigmatic volcanic events at Fernandina such as the rapid uplift at Punta Espinoza in 1927 and the 1968 caldera collapse without significant eruption.
3.2 Overview

Six volcanoes in the western Galápagos Islands of Isabela and Fernandina have shown clear signs of active deformation since their geodetic monitoring began [Amelung et al., 2000; Geist et al., 2006a] and three of them have erupted since 2005. Among these, Fernandina can be considered the most active volcano in the archipelago, having experienced 25 eruptions in the past two centuries and three eruptions since 1995 [Jónsson et al., 1999; Rowland et al., 2003; Chadwick et al., 2011; Smithsonian Institution, Bulletin of the Global Volcanism Network, monthly reports for Fernandina volcano, 1995–2012, available at http://www.volcano.si.edu]. However, the eruptive activity involves only a fraction of the magma coming from the mantle. Large portions of the fluid are in fact stored in crustal reservoirs or intruded laterally within the volcanic edifice, and knowing where and how the un-erupted magma is stored or intruded become fundamental aspects in the study of the volcanic activity. For example, Fernandina is the site of one of the largest caldera collapses on a basaltic volcano in recorded history (June 1968, ~350 m over a 12-day period), but the cause of this event remains enigmatic [Simkin and Howard, 1970; Filson et al., 1973; Rowland and Munro, 1992; Howard, 2010]. For a caldera collapse to occur magma has to be removed from a reservoir, usually through a volcanic eruption. Although the 1968 collapse at Fernandina was preceded by two eruptions, the volume of lava and ash erupted during these events was more than 100 times smaller than the estimated volume of the collapse. Therefore, another process for magma withdrawal in addition to the eruptions must be taken into account to explain the volume of the collapse.
The presence of a large caldera is the evidence that each of the seven volcanoes of the western Galápagos Islands has a shallow magma reservoir beneath the summit, however, the remaining portion of their storage system is not fully understood. From integrative geophysical and petrologic studies the Galápagos volcanoes can be divided based on their evolutionary stage [Geist, 2011]. Fernandina, together with Sierra Negra and Wolf, is considered a mature, monotonic volcano erupting strongly evolved tholeiites. Through time, magma evolution and crystallization would create a growing mush pile several kilometers thick and magmas would transit through this pile before residing in the shallow subcaldera reservoir [Geist et al., 2006b]. Petrologic models also suggest the presence of deeper zones of crustal storage where magma undergoes cooling and fractionation [Geist et al., 1998] and evidence for at least a second, discrete magma reservoir is provided by geodetic measurements [Chadwick et al., 2011]. A multiple magma reservoir system would make Fernandina different from Sierra Negra where only one subcaldera reservoir is inferred from the measured deformation [Amelung et al., 2000; Yun et al., 2006; Jónsson, 2009] and no evidence for deeper sources has been found.

In this chapter, the different aspects of the volcanic activity at Fernandina are studied by looking at their surface expression as ground deformation. Favorable ground conditions (scarce vegetation, surface mostly covered by lava) provide a good opportunity for the use of detailed Interferometric Synthetic Aperture Radar (InSAR) measurements when studying the surface deformation. We use Synthetic Aperture Radar (SAR) data acquired by the European Space Agency's ENVISAT satellite between January 2003 and September 2010, a longer time-interval than any previous geodetic
studies at Fernandina. This period spans two eruptions and 18 major earthquakes (Mw 3.8–5.4), with epicenters located within 100 km from the summit of the volcano. Most of this seismicity is clustered in two seismic swarms that accompany broad and rapid deformation of the volcanic edifice. Temporal and spatial variations in the surface displacement are studied using InSAR time series and through the analysis of single interferograms. Data acquired from different viewing geometries and frequent repeat passes of the satellite provide the opportunity to observe the same events from different perspectives and better study the timing of each event.

Our findings suggest that the un-erupted magma is stored in multiple crustal reservoirs, hydraulically connected and centered at different depths below the summit caldera. Surface deformation recorded during periods of inflation or deflation is used to study variations in the excess magma pressure within the magma plumbing system. We also identify the occurrence of subvolcanic lateral intrusions accompanied by seismic swarms, and provide for the first time geodetic evidence for this type of activity in the Galápagos. These intrusions cause the withdrawal of large quantities of magma from the storage system and may play an important role for the development of the Galápagos volcanic system.

3.3 Geologic setting

The Galápagos Archipelago is a cluster of 13 major (>100 ha) basaltic volcanic islands located near the equator on the eastward-moving Nazca plate, 1000 km west of Ecuador (Figure 3.1a). These islands sit on a shallow submarine platform that rises for more than 2000 m above the surrounding ocean floor. It is believed that the volcanoes
Figure 3.1: (a) Map showing the location of the Galápagos Islands relatively to South America and the Galápagos Spreading Center (GSC). Black arrows indicate the motion of the Nazca Plate (91°) relative to the global hot spot reference frame. (b) Map of the western islands of the Galápagos Archipelago (Landsat 7 shaded-relief color composite image; bathymetry: GEBCO_08 Grid, version 20100927, http://www.gebco.net; topography: hole-filled seamless SRTM data V4). Six volcanoes on Isabela and Fernandina islands have been actively deforming during the last decade, and three of them erupted. For each volcano, dates of the latest eruptions are reported. The white square marks the area covered by subsequent figures in this paper.
have grown above a hot spot now placed on the southern side of the E-W trending Galápagos Spreading Center (GSC), and centered under Fernandina and Isabela Islands [Wilson, 1963; Morgan, 1971; Hey, 1977; Villagómez et al., 2007].

Fernandina and the other western Galápagos volcanoes (Figure 3.1b) represent the type locality for the “Galápagos shield volcano”: edifices are characterized by steep upper flanks, proportionately large calderas, arcuate summit fissures, and radial flank fissures [Simkin, 1984; Chadwick and Howard, 1991].

Fernandina Island, with a diameter of about 30 km and maximum elevation of 1476 m, is the westernmost eruptive center in the archipelago, and has a very well developed subaerial circumferential and radial fissure system (blue lines in Figure 3.2a), probably the most evident of all the Galápagos volcanoes [Chadwick and Howard, 1991]. While the subaerial portion of the volcano lacks well-defined rift zones, the submarine part of Fernandina shows three well-developed rifting zones [Geist et al., 2006b] that extend from the western side of the island (red dashed lines in Figure 3.2a).

The recent eruptive activity at Fernandina includes: an eruption in 1982 from a circumferential fissure at the southern caldera rim; intracaldera eruptions in 1984, 1988 and 1991 [Rowland and Munro, 1992; Smithsonian Institution, Bulletin of the Global Volcanism Network, monthly reports for Fernandina volcano, 1982–2012, available at http://www.volcano.si.edu]; an eruption on the southwestern flank from a radial fissure in 1995 [Jónsson et al., 1999]; an eruption from circumferential fissures at the southern caldera rim in May 2005 [Chadwick et al., 2011] and a new eruption from radial fissures in April 2009 [Bourquin et al., 2009]. The trace of the eruptive fissures and the extension of the lava flows produced during the latest three eruptions are presented in Figure 3.2b.
**Figure 3.2:** (a) Shaded relief map of Fernandina Island and bathymetry (topography: hole-filled seamless SRTM data V4; bathymetry: multibeam data from NGDC-NOAA, http://map.ngdc.noaa.gov/). Blue lines represent mapped eruptive fissures from Chadwick and Howard [1991]; Red-dashed lines mark submarine rift zones identified by Geist et al. [2006b]. (b) Blow-up of the area represented in Figure 3.2a with a dashed-white square and covering the location of the last three eruptions. Red solid lines mark the eruptive fissures associated with the 1995, 2005 and 2009 eruptions. The image also shows the extension of lava flows produced by these latest eruptions (yellow = 1995; purple = 2005 and green = 2009) from Bourquin et al. [2009] and Chadwick et al. [2011].
The elliptical summit caldera, \( \sim 21 \text{ km}^2 \) in area, with the \( \sim 6.2 \text{ km} \) long major axis elongated in NW-SE direction and the \( \sim 4.3 \text{ km} \) long minor axis directed NE-SW, is today \( \sim 900 \text{ m} \) deep at its maximum, with walls sloping inward at 30–50°. The present morphology is the result of repeated cycles of partial filling and collapse, and of the down drop of the southeastern portion by as much as 350 m during the June 1968 event. A flank eruption on 21 May 1968, minor earthquake activity (mb 3.9–4.6) through the first week of June and a large hydromagmatic explosion from the caldera wall occurred prior to this collapse, which started on 12 June. However, the estimated combined volume of the two eruptions is less than 0.2 km\(^3\) while the volume of the collapse is equivalent to 1–2 km\(^3\). No evidence of submarine eruptions during this event has been found and Simkin and Howard [1970] concluded that magma was probably withdrawn from the reservoir through lateral intrusions that didn't reach the surface.

No geodetic evidence of large flank intrusions at Fernandina has ever been provided but their occurrence is suggested by the rapid uplift observed at Punta Espinoza in 1927. Here, a fishing boat anchored for the night was stranded by an uplift of “several feet” occurred in a few hours [Cullen et al., 1987]. A similar event was recorded in May 1954 at the nearby Urvina Bay where the coastline of Darwin Volcano was uplifted by as much as 4.6 m and moved inland for more than a kilometer [Couffer, 1956; Richards, 1957]. The uplift occurred in less than an hour and so rapidly that fish were stranded in pools.

### 3.4 SAR data and processing methods

It is well recognized that InSAR is a successful geodetic technique used to measure surface deformation associated with different sources, such as earthquakes, volcanoes or
anthropogenic activity [Massonnet et al., 1994; Amelung et al., 2000]. The phase difference (interferogram) of SAR image pairs for the same area, acquired at different times, provides measurements of the surface displacement along the radar line-of-sight (LOS) with centimeter to millimeter accuracy.

We processed 330 SAR images acquired by the European Space Agency's ENVISAT satellite from four ascending and three descending tracks (Table 3.1), with a 35-day repeat pass. Compared to other satellites that acquired SAR images for the same area, ENVISAT provides full spatial coverage of Fernandina Island from four different viewing geometries, partial coverage from other three and the largest number of acquisitions. The sensor is a C-band SAR with an operating wavelength of 56.3 mm. Data from the three IS2 tracks are part of the ENVISAT's background mission and span the entire studied period (2003–2010). In 2005 we tasked the acquisition of data from another five tracks covering Fernandina and the neighboring Isabela Island.

<table>
<thead>
<tr>
<th>Track (Beam, Inc. Angle)</th>
<th>Pass</th>
<th>Start Date–End Date</th>
<th>Acquisitions</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Full Spatial Coverage</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>54 (IS7, 42.5°–45.2°)</td>
<td>Descending</td>
<td>10 June 2005–11 Dec. 2009</td>
<td>37</td>
</tr>
<tr>
<td>147 (IS6, 39.1°–42.8°)</td>
<td>Ascending</td>
<td>20 October 2006–20 Aug. 2010</td>
<td>42</td>
</tr>
<tr>
<td><strong>Partial Spatial Coverage</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>376 (IS5, 35.8°–39.4°)</td>
<td>Ascending</td>
<td>3 Jul. 2005–3 May 2009</td>
<td>34</td>
</tr>
<tr>
<td><strong>Total</strong></td>
<td></td>
<td></td>
<td><strong>330</strong></td>
</tr>
</tbody>
</table>
We use Gamma SAR Processor and Interferometry to focus the raw radar images and the JPL/Caltech ROI_PAC SAR software [Rosen et al., 2004] to form interferograms. We use precise DORIS orbits provided by ESA and remove the topographic contribution to the interferometric phase using a 90 m resolution digital elevation model (DEM), re-sampled to 30 m, generated by the NASA Shuttle Radar Topography Mission (SRTM). The interferograms are phase-unwrapped using the Snaphu unwrapper [Chen and Zebker, 2001].

3.4.1 InSAR pairs selection and time series generation

To resolve the temporal evolution of surface deformation we use the small baseline subset (SBAS) method [Berardino et al., 2002; Lanari et al., 2004] and apply it to InSAR data from the descending (T412) and the ascending (T61) IS2 tracks, which entirely cover Fernandina Island and provide the longest temporal coverage. This method is based on the inversion of a large number of phase-unwrapped interferograms (e.g., 213 for the descending pass) to retrieve the LOS displacement history for each pixel at the epoch of each SAR acquisition.

The selection of the most suitable interferometric pairs for the InSAR time series generation is performed following a multistep approach based on the key assumption of the SBAS method, for which small spatial and temporal separation between SAR acquisitions minimize decorrelation and maximize the temporal coherence of pixels [Pepe and Lanari, 2006]. For each pass we first execute a Delaunay triangulation for all possible pairs in the temporal/perpendicular baseline plane to create an interconnected network of interferograms. We remove from this selection all pairs with a perpendicular
baseline larger than 400 m and successively add all possible interferograms spanning less than 300 days and with perpendicular baseline smaller than 300 m. Once all the selected interferograms are generated we set a rectangular area of interest (AOI) that encloses most of Fernandina Island and remove those images that have more than 50% of pixels with a coherence value lower than 0.4 within the AOI. With this approach we assess the overall quality of each interferogram and, by removing the images where pixels with low or null coherence prevail, we increase the likelihood for each pixel of maintaining good coherence through time. Several SAR acquisitions are affected by a varying degree of atmospheric water vapor delay, particularly in the summit area, and a few acquisitions present significant noise level due to ionosphere perturbations, frequent in equatorial areas. We therefore evaluate each interferogram and exclude from our selection those SAR acquisitions affected by strong atmospheric and ionospheric perturbations, which could strongly interfere with the deformation signal and generate ambiguity on the source of the phase delay. In a last step we check for missing interconnections in the network and we remove the isolated pairs.

We finally obtain an optimized, high quality (high signal-to-noise ratio) connected network of interferometric pairs (descending IS2, Figure 3.3a; ascending IS2, Figure S1.1 in the auxiliary material) and we use the University of Miami-RSMAS InSAR time series software package \cite{Gourmelen et al., 2010} to retrieve the temporal evolution of surface displacement. Each LOS displacement time series is relative to a single pixel that exhibits high coherence and is located in an area where surface deformation is minimum throughout the entire studied-period (descending IS2, R1 in Figure 3.3b). LOS
displacement, along the ground-satellite direction, is considered positive when the offset is toward the satellite and negative when directed away from it.

3.5 Volcanic activity and surface deformation at Fernandina

3.5.1 Overview

The analysis of the InSAR time series and of individual interferograms reveals that surface deformation at Fernandina varies both in time and in space. LOS displacement history for the descending pass is presented in Figure 3.3b. The highest rates of deformation are always recorded at the center of the summit caldera in periods of both inflation and deflation. At this location (D1, red squares) we recognize four major events (E1 through E4) that produce large and rapid (<35 days) displacement and that are directly ascribable to volcanic eruptions or synchronous to local seismic activity (2005 and 2009 eruptions; 2006 and 2007 seismic swarms; gray vertical lines in Figure 3.3b, details in Table 3.2). These events divide the entire study period into five intervals that represent pre- or post-eruptive/seismic activity (marked with white or gray background in Figure 3.3b). Because of spatial variations of the deformation pattern within each interval we further subdivide them into a total of eleven periods (P1 through P11, details in Table 3), limited by blue-dashed lines in Figure 3.3b. Since the acquisition of SAR data is not continuous in time but limited by the repeat-pass cycle of the satellite, we are not able to determine the exact occurrence of each change in the deformation pattern. Therefore, we use the time of the closest SAR acquisition to define the start-date and the end-date of each period, with an uncertainty ranging from 13 h to 10 days.
Figure 3.3: (a) Triangulated network of interferometric pairs used to generate the SBAS time series. Each square represents a SAR acquisition. Solid lines show the selected pairs, dotted lines are pairs that did not meet the selection criteria (see section 3.1). (b) LOS displacement times series relative to R1 for two pixels: (D1) at the center of the summit caldera and (D2) on the northeastern upper flank. Dark gray solid lines represent the occurrence of eruptions or local seismic activity associated to rapid displacement (Event E1–E4). Pre- and post-eruptive/seismic intervals are shown with different background colors. Blue dotted lines represent the occurrence of spatial variations in the deformation pattern and divide the studied interval in eleven time periods (P1 through P11). Red squares and green diamonds mark LOS surface displacement for each location at the time of SAR acquisitions. The intracaldera displacement associated with the 2009 eruption cannot be fully measured using the ENVISAT data set. The total LOS displacement shown here (black star) is obtained from the analysis of SAR data acquired by the L-band ALOS satellite [Baker, 2012].
Table 3.2. Major Magmatic and Seismic Events Occurring at Fernandina During the Studied Period, Duration and Characteristics

<table>
<thead>
<tr>
<th>Event</th>
<th>Start Date</th>
<th>End Date</th>
<th>Notes</th>
</tr>
</thead>
<tbody>
<tr>
<td>May 2005 eruption</td>
<td>13 May 2005</td>
<td>29 May 2005</td>
<td>Circumferential fissures on the southern caldera rim</td>
</tr>
<tr>
<td>Aug. 2007 seismic swarm</td>
<td>27 Aug. 2007</td>
<td>30 Aug. 2007</td>
<td>8 major shocks, max magnitude Mw 5.4</td>
</tr>
<tr>
<td>April 2009 eruption</td>
<td>10 April 2009</td>
<td>28 April 2009</td>
<td>Radial fissures on the southwestern flank</td>
</tr>
</tbody>
</table>

We observe that while intracaldera LOS displacement is always present, there are also short-lived episodes of deformation (<7 months) involving a broader area that widely extends outside the caldera. The LOS displacement time series for a second pixel located on the northeastern upper flank and outside the caldera (D2, green diamonds in Figure 3.3b) shows the deformation history for this second area.

From the ENVISAT data set we are not able to fully retrieve the intracaldera displacement associated with the April 2009 eruption. In fact, the large deformation results in high fringes rates within the caldera (LOS displacement of half or more the signal wavelength between adjacent pixels) leading to difficulties in the phase unwrapping step and impacting the usability of the differential interferograms in the displacement time series [Casu et al., 2011]. However, a maximum LOS displacement of $-0.88$ m has been successfully measured by an independent study that uses SAR data acquired by the L-band (operating wavelength of 236.2 mm) Japanese Space Agency's ALOS PALSAR satellite [Baker, 2012]. Therefore, for the entire studied period the net LOS displacement is negative both within the caldera ($-0.24$ m at its maximum) and in the remaining area around the summit. Rates of intracaldera inflation are variable and
maximum values are recorded during P7 (52.1 cm/yr), when deformation occurs both within and outside the caldera.

Table 3.3. Time Periods of Surface Deformation at Fernandina Volcano Separated by the Date of the Nearest SAR Acquisition and Type of Activity

<table>
<thead>
<tr>
<th>Time Period/Event</th>
<th>SAR acquis. Date</th>
<th>Deformation Type</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Pre-2005 eruption</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Period 2</td>
<td>28 Sept. 2004–7 May 2005</td>
<td>Intracaldera and edifice-wide inflation</td>
</tr>
<tr>
<td><strong>2005 eruption</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Event 1</td>
<td>7 May 2005–31 May 2005</td>
<td>Intracaldera and edifice-wide deformation</td>
</tr>
<tr>
<td><strong>Post-2005 eruption</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Period 3</td>
<td>31 May 2005–22 Nov. 2005</td>
<td>Intracaldera inflation</td>
</tr>
<tr>
<td><strong>December 2006 seismic swarm</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Event 2</td>
<td>22 Dec. 2006–16 Jan. 2007</td>
<td>Intracaldera deflation and minimum edifice-wide deflation; southeastern flank uplift</td>
</tr>
<tr>
<td><strong>Post-2006 seismic swarm</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>August 2007 seismic swarm</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Event 3</td>
<td>25 Aug. 2007–18 Sep. 2007</td>
<td>Intracaldera and edifice-wide deflation; southeastern flank uplift</td>
</tr>
<tr>
<td><strong>Post-seismic/pre-2009 eruption</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>April 2009 eruption</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Event 4</td>
<td>10 Apr. 2008–5 May 2009</td>
<td>Intracaldera and edifice-wide deflation; uplift of southern caldera rim and eastern side of the eruptive fissures</td>
</tr>
<tr>
<td><strong>Post-2009 eruption</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Period 9</td>
<td>5 May 2009–3 Oct. 2009</td>
<td>Intracaldera inflation</td>
</tr>
<tr>
<td>Period 10</td>
<td>3 Oct. 2009–1 May 2010</td>
<td>Intracaldera and edifice-wide inflation</td>
</tr>
<tr>
<td>Period 11</td>
<td>1 May 2010–18 Sep. 2010</td>
<td>Intracaldera inflation</td>
</tr>
</tbody>
</table>

LOS displacement maps obtained from the SBAS analysis are not shown here due to the large number of pixels that do not maintain coherence during the entire studied
period. The occurrence of two eruptions with large displacements, the presence of steep slopes and of patches of vegetation, generate loss of phase coherence in the upper part of the volcano where most of the deformation occurs. We instead present a selection of high signal-to-noise ratio interferograms, from the descending or the ascending IS2 passes, spanning the four major events (Figure 3.4) and each period (Figure 3.5). A detailed description of each period and of each major event is given below in chronological order. When available, we provide information relative to the known volcanic activity and to the recorded local seismicity, and correlate such activities to the observed displacement. Earthquake locations (Figure 3.6) are derived from the U.S. Geological Survey – National Earthquake Information Center (USGS-NEIC) database and based on data recorded by the Global Seismic Network (GSN). Epicenter locations are affected by varying uncertainties (5–100 km) depending on the earthquake magnitude and the number of stations used to locate the earthquake. The lack of stations close to the western Galápagos Islands results in large location uncertainties (>10 km, Figures S1.3 and S1.4 in the auxiliary material) for earthquake of magnitude less than 5.

3.5.2 Pre-2005 eruption (P1 and P2)

Our study begins with the first available full-coverage ENVISAT SAR image acquired on 11 February 2003 and the first interval ends with the May 2005 eruption. The first 20 months of this interval form P1, characterized by small LOS displacements (<5 cm) mostly occurring in an area limited by the caldera rim (Figure 3.5a). Starting from the end of September 2004, intracaldera deformation is associated to displacement of a broader area that extends outside of it and covers the entire summit of the volcano
(Figure 3.5b). Positive LOS displacement characterizes both areas (P2). This pattern of edifice-wide deformation is recorded until the last pre-eruption acquisition.

**Figure 3.4:** Selection of ENVISAT SAR interferograms showing the surface deformation at Fernandina associated with the four eruptive/seismic events (E1-E4). Detailed information for each event is given in Section 3.4. Satellite flight direction (ascending T61, descending T412) and radar look direction are presented with arrows. Both tracks are standard beam IS2 (look angle 19.2°–26.7°). Each fringe (full color cycle) represents 2.8 cm of range change between the ground and the satellite, or LOS (line-of-sight) displacement. Areas with low interferometric coherence (<0.3) are uncolored.
Figure 3.5: Selection of ENVISAT SAR interferograms showing spatial variations in the surface deformation at Fernandina associated with periods of inflation (P1-P11). (left and right) Periods of deformation within the caldera only and (middle) periods of edifice-wide displacement. Detailed information for each period is given in Section 3.4. Satellite flight direction (ascending T61, descending T412) and radar look direction are presented with arrows. Both tracks are standard beam IS2 (look angle 19.2°–26.7°). Each fringe (full color cycle) represents 2.8 cm of range change between the ground and the satellite, or LOS (line-of-sight) displacement. Areas with low interferometric coherence (<0.3) are uncolored.
Figure 3.6: Map of the seismicity in the western Galápagos Islands during the studied period. Squares represent the location of major seismic events recorded by the global seismic network, retrieved from the USGS/NEC PDE Catalog. For the two seismic swarms the numbers indicate the days of the month.

3.5.3 The May 2005 eruption (E1)

On the morning of 13 May 2005, a circumferential eruptive fissure, formed by a set of 5 right-stepping en echelon segments, opened on the southern summit plateau of Fernandina (Figure 3.2b). The eruption that lasted until 29 May, was preceded by a mb5.0 earthquake on 11 May, whose epicenter has been located (USGS-NEIC) ~30 km east of Fernandina's summit (red square in Figure 3.6).
The interferograms spanning the eruption show: (i) a local area of positive LOS displacement centered on the southern caldera rim, where the circumferential fissure opened, superimposed to (ii) a broad area of negative LOS displacement, that covers a large part of the island (Figure 3.4a). It is also evident how the emplacement of the lava flows on the southern upper flank causes loss of phase coherence for this area.

3.5.4 Post-2005 eruption (P3 through P5)

Interferograms spanning the post-2005 eruption time interval record a first period characterized by positive LOS displacement only within the caldera, lasting until November 2005 (P3, Figure 3.5c). From this time through late January 2006, persistent displacement within the caldera is associated with deformation of a larger area similar to the one displaced during the co-eruptive phase (P4, Figure 3.5d), with both areas showing positive LOS displacement. In February 2006 the edifice-wide deformation ceases and only the caldera continues to show positive LOS displacement until late December 2006 (P5, Figure 3.5e). Only two months of this eighteen-month-long time interval show edifice-wide deformation, which started with a delay of about six months after the 2005 eruption ended.

3.5.5 December 2006 seismic swarm, co-seismic (E2) and post-seismic (P6)

The general trend of re-inflation that characterizes the post-2005 eruption interval is suddenly interrupted in correspondence with the occurrence of seismic activity close to Fernandina Island on 22–23 December 2006 (E2). The USGS-NEIC located two major earthquakes (mb 4.0 on 22 December and N/A on 23 December) within ~85 km of the
summit of Fernandina (green squares in Figure 3.6). Interferograms spanning this event (Figure 3.4b) show rapid (<35 days) negative LOS displacement of the caldera and of the summit area. While the signal within the caldera is well defined (~0.09 m at its maximum) it is subtler for the rest of summit area (<0.03 m). The same interferograms also reveal a broad area of positive LOS displacement on the southern flank of the volcano. No eruptive activity was detected.

Information on the timing between the occurrence of the seismicity and the associated surface deformation can be obtained from a SAR image acquired between the two main shocks, 6 h after the occurrence of the first earthquake. In interferograms generated using this image there is no evidence of the deformation observed in later acquisitions. The total displacement is however present in the following SAR image acquired five days later, on 28 December 2006 (see Table S1.1 and Figures S1.2a and S1.2b in the auxiliary material).

Surface deformation following the December 2006 seismic swarm is characterized by intracaldera positive LOS displacement only (~0.15 m at its maximum) with a pattern similar to previous periods of intracaldera uplift (P6, Figure 3.5f). Further seismic activity occurred on 5 February 2007, in the same area of the previous seismic events (mb 4.1, blue square in Figure 3.6), but in this case there is no clear evidence of perturbations in the pattern of deformation at Fernandina. The positive LOS displacement within the caldera is detected until late August 2007.
3.5.6 August 2007 seismic swarm (E3)

Between 27 August and 30 August 2007, the GSN recorded 8 major earthquakes in the vicinity of the Galápagos archipelago with magnitudes (Mw or mb) between 3.8 and 5.4 (USGS-NEIC). Epicenters for five main shocks are located within 35 km of the summit of Fernandina and the rest of the events within 60 km (yellow squares in Figure 3.6). A large location uncertainty is associated with the lower magnitude earthquakes and the linear pattern formed by the epicenters suggests the presence of artifacts in the estimated locations.

All the interferograms that entirely span the August 2007 seismic swarm show a very distinctive pattern of deformation: (i) the caldera and the summit area show up to −0.36 m of LOS displacement, while (ii) a broad area located on the southern flank of Fernandina, similar to the one displaced during the December 2006 seismic swarm, shows up to 0.06 m of positive LOS displacement (Figure 3.4c). By looking at the geometry of the interferometric fringes on the southern flank we can infer that a large portion of the uplifting area lies below sea level, where InSAR does not provide any measurements. However, no deformation is detected on the nearby Isabela Island and no surface or submarine eruptions have been reported.

Two SAR images acquired during the seismic swarm provide further information on the timing between seismicity and surface deformation. On 28 August 2007 (SAR image from T104), ~8 h after the first earthquake and between the second and the third major shocks, there is already evidence of surface deformation involving large part of the island, but its magnitude is only a portion of the total displacement recorded by later acquisitions. Interferograms formed using the following SAR image acquired on 30
August (T61), ∼9 h after the last recorded earthquake, show the entire displacement associated with this event (see Table S1.2 and Figures S1.2c and S1.2d in the auxiliary material).

3.5.7 Post-2007 seismic swarm and pre-2009 eruption (P7 and P8)

Interferograms spanning the early post-seismic interval reveal positive LOS displacement in both areas, within the caldera and across the broad area that showed subsidence during the previous event (P7, Figure 3.5g). Also in this case the LOS displacement rate is larger within the caldera. No displacement is detected for the southern lower flank of Fernandina that showed uplift during the seismic swarm. This pattern of deformation is recorded for about six months, until March 2008. Interferograms generated using SAR images acquired during or after April 2008 do not show any displacement outside the caldera (P8, Figure 3.5h).

3.5.8 The April 2009 eruption (E4)

During the night between 10 and 11 April 2009, a new eruption started at Fernandina Volcano: three main radial fissures (Figure 3.2b) opened on the southwestern flank and produced lava flows that reached the ocean [Bourquin et al., 2009]. The eruption ended on 28 April. Two areas of positive LOS displacement, one centered on the southwestern caldera rim and one on the eastern side of the eruptive fissures, represent the surface expression of the propagation of the feeding dike and the opening of the eruptive fissures (Figure 3.4d). As for the previous eruption and for the deformation associated with the seismic swarms, the caldera and a large portion of the island show negative LOS
displacement. The subsiding area outside the caldera is similar to the one displaced during previous episodes as for shape and gradient variations of the interferometric fringes.

3.5.9 Post-2009 eruption (P9 through P11)

The last interval studied here spans sixteen months after the April 2009 eruption, with the last SAR acquisition acquired on 18 September 2010. From November 2010 the ENVISAT satellite changed its orbit and interferograms between images acquired before and after this date are no longer possible.

As during the post-eruptive interval that followed the 2005 eruption, the early post-2009 eruption interval (P9) is characterized by rapid positive LOS displacement only within the caldera (Figure 3.5i). After five months from the end of the eruption, beginning in October 2009, the InSAR data record edifice-wide positive LOS displacement (P10, Figure 3.5j). This pattern of deformation persists for more than seven months until May 2010, when positive LOS displacement only within the caldera characterizes the remaining four months of the study period (P11, Figure 3.5k).

3.6 Sources of deformation and modeling

3.6.1 Active sources of deformation

The analysis of the InSAR data reveals the presence of multiple sources of deformation active at different times and locations, implying the existence of a complex magmatic plexus rather than a system formed by a single magma chamber. Two recurrent patterns of deformation are centered at the summit caldera but have different areal
extensions. They are interpreted as the expression of variations in the excess fluid pressure in two long-term magma reservoirs located at different depths. A relatively shallow source produces almost continuous displacement within the caldera. A second, deeper one is intermittently active and produces edifice-wide displacement. Additional sources of deformation are active during the two seismic swarms (E2 and E3) and the two eruptions (E1 and E4). During the two seismic swarms the deformation on the southern flank is generated by a relatively deep source (given the low gradient of the interferometric fringes), likely representing the lateral intrusion of magmatic bodies. The two eruptions show deformation associated with the shallow intrusion of the eruptive dikes together with deflation of the entire summit of the volcano [Chadwick et al., 2011].

3.6.2 Modeling approach

Among the different sources of deformation active during the studied period we model the two interpreted as magma reservoirs, and the intrusion associated with the August 2007 seismic swarm (E3). The model for E3 can be representative for the other seismic swarm/intrusion (E2) for which the measured surface displacement is too small for a robust estimation of the source parameters. The geometry of eruptive dikes feeding the eruptions (E1 and E4) is very complex [e.g., Chadwick et al., 2011] and since it is not considered fundamental for this study is not modeled here. Our modeling strategy is as follow: we identify, for the three sources, the interferograms with the highest signal-to-noise ratio within periods for which contemporary ascending and descending interferograms (descending SAR image acquired ∼13 h before the ascending one) are available. We then perform a nonlinear inversion of the InSAR data in a homogeneous
isotropic elastic half-space. Magma bodies are modeled as rectangular dislocation sources with uniform opening [Okada, 1992] or as fluid-pressurized ellipsoidal cavities [Yang et al., 1988]. We also tested for point sources [Mogi, 1958] and other radially symmetric cavities [e.g., McTigue, 1987; Fialko et al., 2001] but the modeled surface deformation of these sources did not fit the data well, indicating the necessity for more complex geometries. In all models we assume a Poisson's ratio \( \nu = 0.25 \) for the elastic half-space.

Since each full-resolution SAR interferogram subset for Fernandina consists of about 107 data points, we perform spatial averaging using the Quadtree algorithm [Jónsson et al., 2002]. The reduced data sets consist of a minimum of 503 and a maximum of 920 data points. For each data set we determine best fit source parameters such as location \((x, y, z)\), geometry \((l, w, a, b/a)\), strength \((\Delta P/\mu)\) from expressions given by Okada [1992] and Yang et al. [1988] respectively. Unit variance is assumed for all data points. To evaluate the quality of the predicted deformation we use the normalized root-mean square error (RMSE) between observed and modeled interferograms, defined as:

\[
\text{RMSE} = \sqrt{\frac{1}{N} \sum_{i=1}^{N} (d_i - m_i)^2}
\]

where \(d\) is the data, \(m\) is the model prediction, and \(N\) is the number of data points. We estimate the optimal solution together with the posterior probability density distributions of the parameters, as indicators of resolution and uncertainties, using the Monte-Carlo based Gibbs sampling (GS) algorithm [Brooks and Frazer, 2005].

The effect of topography on the surface deformation signal is taken into account using the approach proposed by Williams and Wadge [1998]. The varying depth model
corrects the changing distance between the source and the surface by varying its depth with topography.

3.6.3 Results

We first model the most active source of deformation, which generates displacement within the caldera. The most suitable data set spans 175 days within P6 (specific dates are reported in Table 3.4). We invert for a horizontal, uniform opening dislocation source, representing the top surface of a sill-like shallow magma reservoir. We fix the sill to be horizontal while all the other parameters are allowed to float within geologically realistic values. The best fitting model centers the sill beneath the caldera at a depth of ~1.1 km below sea level (~1.6 km below the caldera floor), within a 95% confidence interval ranging from 0.8 to 1.3 km. The inferred sill is ~2.8 km long and ~2.0 km wide, and oriented NW-SE. The comparison between the observed InSAR data and the predicted LOS displacement shows good agreement for both viewing geometries (Figures 3.7a–3.7f).
Figure 3.7: Modeling results for the shallower source: (a and b) ascending IS2 and (d and e) descending IS7, comparison between observed data and model predictions. (c and f) Comparison between topography (in gray), observed (in blue) and modeled (in green) surface displacement along the A-A’ trace (see Figure 3.8). Surface deformation is modeled using a rectangular horizontal sill (blue rectangle). For source parameters see Table 3.4. The solid and the dashed gray lines represent the summit caldera rim and the caldera floor outline respectively.
Table 3.4. Results From Nonlinear Inversion of InSAR Data: Estimated Source Parameters for the Shallower Source (Horizontal Sill), the Deeper Source (Oblate Spheroid) and the August 2007 Intrusion (Horizontal Sill)

<table>
<thead>
<tr>
<th>Length (km)</th>
<th>Width (km)</th>
<th>Depth (km)</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Strike (deg)</th>
<th>Opening (m)</th>
<th>Volume Change (10^6 m^3)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Shallower Source</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>2.8</td>
<td>2.03</td>
<td>1.08</td>
<td>0.3795 S</td>
<td>91.5504 W</td>
<td>121°</td>
<td>0.2</td>
<td>1.1</td>
</tr>
<tr>
<td>[2.43–3.15]</td>
<td>[1.60–2.45]</td>
<td>[0.81–1.34]</td>
<td></td>
<td></td>
<td></td>
<td>[0.14–0.26]</td>
<td></td>
</tr>
<tr>
<td>Deeper Source</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>2.73</td>
<td>0.18</td>
<td>4.93</td>
<td>0.3684 S</td>
<td>91.5466 W</td>
<td>48°</td>
<td>7.45*10^{-3e}</td>
<td></td>
</tr>
<tr>
<td>[1.96–3.70]</td>
<td>[0.11–0.30]</td>
<td>[4.32–5.58]</td>
<td></td>
<td></td>
<td></td>
<td>[3.30<em>10^{-3}–9.89</em>10^{-3}]</td>
<td></td>
</tr>
<tr>
<td>August 2007 Intrusion</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>17.55</td>
<td>21.7</td>
<td>4.49</td>
<td>0.4227 S</td>
<td>91.4974 W</td>
<td>222°</td>
<td>0.05°</td>
<td>19</td>
</tr>
<tr>
<td>[15.86–19.39]</td>
<td>[14.24–29.57]</td>
<td>[3.60–5.52]</td>
<td></td>
<td></td>
<td></td>
<td>[0.04–0.06]</td>
<td></td>
</tr>
</tbody>
</table>

a The 95% confidence intervals are shown in brackets. RMS – root mean square. Sill dip angle is fixed to 0°.
Oblate spheroid dip angle is fixed to 0°.
b Ascending IS2 (27/01/2007–21/07/2007), descending IS7 (26/01/2007–20/07/2007), and RMS = 9.52 mm.
c Mean.
e For μ = 5 GPa ΔV = 15.4 10^6 m^3 [Tiampo et al., 2000].
We next model the second, deeper source, for which deformation extends outside the caldera. The LOS displacement during periods of edifice-wide deformation shows only one maximum at the center of the caldera and decreases radially away from the summit. However, the radial decrease is not regular and shows a significant gradient contrast between the area inside the caldera and the rest of the summit. Therefore, the displacement field cannot be explained by a single source. We model it as a stack of two sources but given the difficulty of modeling their contemporary contributions, we mask out the deformation inside the caldera and invert the remaining data for a deeper source only. We find that an oblate spheroid cavity provides a better fit than a uniform opening dislocation source. The inversion is performed using two interferograms spanning 210 days within P7, when edifice-wide deformation is recorded and there is no evidence for further active sources other than the two centered below the summit. The best fitting oblate spheroid is centered beneath the summit at a depth of \( \sim 4.9 \) km b.s.l. (95% confidence interval: 4.3–5.6 km) with the \( \sim 2.7 \) km long major axes oriented in the NE-SW direction (azimuth of 48 degrees from north). We then reintroduce the data points previously masked and, by fixing the geometries of both sources and the inferred normalized excess pressure for the deeper one, we invert for the opening of the shallower source only. Our model is able to well reproduce the LOS displacement, in particular for the ascending IS2 viewing geometry (Figures 3.8a–3.8f).

A third model is generated to reproduce the uplift of the southeastern flank associated with E3. In this case, we invert ascending and descending 35-day interferograms spanning the event. For simplicity we fix the geometries and the positions of the two sources previously inferred and invert for a further planar dislocation source with uniform
Figure 3.8: Modeling results: deeper source, (a and b) ascending IS2 and (d and e) descending IS7, and August 2007 lateral sill intrusion, (g and h) ascending IS2 and (j and k) descending IS7, comparison between observed data and model predictions. (c, f, i, and l) Comparison between topography (in gray), observed (in blue) and modeled (in green) surface displacement along the A-A’ and the B’B’ traces. Surface deformation is modeled using an oblate spheroid for the deeper source (black star), and rectangular sills for the shallower source and the August 2007 intrusion (gray and blue rectangle respectively). For source parameters see Table 3.4. The solid and the dashed gray lines represent the coastline of Fernandina Island, the summit caldera rim and the caldera floor outline respectively.
opening. Best fit is obtained for a $\sim 17.5 \times \sim 21.7$ km horizontal rectangular sill, located at a depth of $\sim 4.5$ km b.s.l. (95% confidence interval: 3.6–5.5 km) and a total volume of the intrusion of $\sim 19.0 \times 10^6$ m$^3$. Also in this case, the inferred model is able to largely explain the observed LOS displacement, in particular for the ascending data set (Figures 3.8g–3.8l).

**Figure 3.9:** Normalized posterior probability distributions for the depth parameter of each source obtained using the Gibbs Sampling algorithm (100,000 samples). Red lines represent the depth of the best fitting models and green lines represent the 95% confidence interval of the modeled parameter.
3.7 Discussion

3.7.1 Implications for the shallow magma storage system

3.7.1.1 Two magma reservoir

Our analysis indicates the presence of two stacked crustal magma reservoirs beneath the summit of the volcano. We model a shallow flat-topped reservoir at \( \sim 1.1 \) km b.s.l. overlaying a second, deeper area of magma storage centered at \( \sim 4.9 \) km b.s.l. Figure 10 shows a schematic cross section of the volcanic edifice and of the upper crust below Fernandina. The total crustal thickness at Fernandina is \( \sim 12–14 \) km [Feighner and Richards, 1994]. The lower 6 km represent the pre-existing oceanic crust, locally flexed by the weight of the volcanic edifice, while the upper 6–8 km are the product of the hot spot volcanic activity. The modeled source depths are well constrained and center both reservoirs within the volcanic pile. However, inhomogeneity in the medium elastic parameters, different rheologies (e.g., viscoelasticity) and the presence of discontinuities in the volcanic edifice can significantly affect ground deformation [De Natale and Pingue, 1996] and cause bias of the modeled parameters. In particular, homogeneous elastic models underestimate magma reservoirs depths and volume changes [Montgomery-Brown et al., 2009; Foroozan et al., 2010]. Therefore, the inferred depths would be larger if a layered heterogeneous medium (e.g., with variable Young's modulus) is considered [Manconi et al., 2007] and would center the deeper reservoir at greater depth, likely at the boundary between the volcanic pile and the underlying oceanic crust. Furthermore, if we consider the effect of viscoelastic relaxation within the regions surrounding the magma reservoirs, the estimated source size could be larger [Segall, 2010].
Figure 3.10: Schematic cross-section across Fernandina Islands and the underlying oceanic crust showing the inferred structure of the shallow magmatic system. Source positions are inferred from the analysis of the InSAR data. The inferred source depths could be larger if a layered heterogeneous medium is considered [Manconi et al., 2007]. For example, the deeper reservoir could likely be centered at the boundary between the volcanic edifice and the underlying oceanic crust. Similarly, the August 2007 intrusion could have propagated along the same boundary.

Source modeling is presented here only for specific periods. The remaining periods have similar patterns of deformation, indicating that the same sources are activated during different times and therefore represent permanent areas of magma storage. The surface deformation observed at Fernandina has clear magmatic origin. Evidence is provided by the high eruptive frequency during the recent history and the observation of inter-eruptive uplift and co/post-eruptive subsidence. In Figure 3.11 we summarize the activity of the two magma reservoirs during each event and period (E1-E4, P1-P11). Excess magma pressure decrease within the reservoirs (expressed by subsidence at the surface) is marked in blue. Its increase (uplift) is marked in red.
Figure 3.11: Summary of the activity of the two magma reservoirs during the studied period. The red color represents inflation of the source and uplift at the surface, in blue deflation and subsidence. Dotted blue lines separate the entire interval into 11 time periods of inflation (P1-P11) and 4 events of deflation (E1-E4).

Our finding of two shallow magmatic sources is consistent with the model of the 2005 eruption and its pre- and post-eruptive phases of Chadwick et al. [2011], who inferred a shallow sill at ~1 km below the surface and a deeper point source at ~5 km depth. Geist et al. [2006a] also modeled GPS and micro-gravity measurements carried out at Fernandina between 2000 and 2002. They inferred the presence of a single source centered under the caldera at a depth of 1.0 km below the surface, which is likely the same as our shallower source. The presence of the second reservoir was probably missed because of the limited extension of the GPS network.

3.7.1.2 Comparison with petrologic results

Support for our model of two different magma reservoirs at crustal depths is also provided by the petrology of the erupted lavas. Subaerial lavas erupted at Fernandina are monotonous, evolved (plagioclase-dominated mineralogy), and incompatible-element enriched tholeiites. The phase assemblage suggests eruptive temperatures of ~1150°C and is stable only at low pressure (<0.1 GPa), corresponding to depths of a few
kilometers or less [Allan and Simkin, 2000; Geist et al., 2006b]. This is consistent with what we infer for the shallower source (~1.1 km b.s.l.). The lack of primitive melts within the subaerial lavas suggests extensive mixing and homogenization of evolved melts in a shallow reservoir, testified by the narrow compositional range of the lavas. Allan and Simkin [2000] proposed that this reservoir would act as a buffer for magma coming from depth and prevent the eruption of any more primitive melts.

Even if most of submarine lavas erupted at Fernandina are similar in composition to the subaerial suite, Geist et al. [2006b] identified two other suites of lava in the submarine record. Evolved basalts and icelandites have been recovered from the SW rift. It is believed that these lavas crystallize and fractionate at pressures between 0.3 and 0.5 GPa, and depths of 10–15 km, greater than any inferred depth for the sources generating the observed surface displacement. These lavas are thought to represent extensive fractional crystallization in the upper mantle and lower crust that would bypass the shallow plumbing system. A second suite, defined as “High-K” lavas and also recovered from the SW rift, is considered as hybridization between the most common series basalts and the evolved series magma. Results from our models could suggest that magma from this last suite is temporarily stored in the deeper reservoir (~4.9 km b.s.l.) and, when subvolcanic lateral intrusions occur and intersect the submarine surface of the volcano, it erupts at the SW rift. In fact, in our model for E3, the sill intrudes toward the southern portion of the volcano at a depth of ~4.5 km b.s.l., close to the area of magma storage inferred for the deeper source.
3.7.2 Implications for the shallow magma plumbing system

3.7.2.1 Hydraulically connected system

We observe that while the shallower source is overall always active (Figure 3.11), the deeper one generates displacement only during the four events (E1-E4) and during some periods (P2, P4, P7 and P10). During these events and periods, ground deformation and excess pressure have always the same signature in both, the shallower and the deeper reservoirs, either positive or negative. This can be interpreted as evidence for a hydraulically connected system between the two reservoirs. Magma seems to easily migrate from the deeper to the shallower reservoir and vice versa, as shown by contemporary rapid excess pressure variations within both reservoirs.

3.7.2.2 Dynamics of magma migration

A previous study has shown that the dike feeding the 2005 eruption likely originated from the shallower reservoir [Chadwick et al., 2011]. The measured LOS displacement for the 2009 eruption is very similar to the 1995 one for which a shallow dipping dike, extending from the surface to a maximum depth of ~1.3 km, has been inferred [Jónsson et al., 1999]. It is likely that during most eruptions, magma mainly withdraws from the shallower portion of the magma storage system, as also demonstrated by the monotonous petrology of the erupted lavas. An opposite mechanism, with magma withdrawing primarily from greater depth, is suggested by our model for the 2007 subvolcanic intrusion.

The periods of reservoir re-filling and excess magma pressure increase provide further insight into magma migration dynamics. The temporal sequence of surface
deformation is different following eruptions than following subvolcanic lateral intrusions. Following eruptions (E1 and E4) first the shallower reservoir inflates for a period of four to six months (P3 and P9), then the two reservoirs inflate contemporarily (P4 and P10) and finally only the shallower source inflates (P5 and P11). The pattern is different after subvolcanic lateral intrusions. Following the larger intrusion (E3), both sources inflate contemporarily (P7) for a seven-month period. Following the smaller intrusion (E2), the shallower reservoir clearly inflates, whereas changes of the deeper source are not resolved (P6).

We explain these observations in terms of (i) temporal variations in excess pressure in the two reservoirs as consequence of eruptions and subvolcanic intrusions (Figure 3.12), and (ii) differences in magma source stiffness and compliance. Magma reservoirs shaped as ellipsoidal cavities are relatively stiff, whereas cracks (dikes and sills) are highly compliant [Rivalta and Segall, 2008; Rivalta, 2010]. For the same amount of excess pressure a compliant source will expand more than a stiff source. However, our interpretation does not include effects from the compressibility of the fluid contained in the reservoirs. We also limit our analysis to fully elastic models, although the presence of viscoelastic rinds surrounding active magma chambers [e.g., Del Negro et al., 2009; Masterlark et al., 2010] or viscous resistance within the conduit connecting the reservoirs could represent further controlling factors on the observed deformation.

Prior to eruptions, we assume equilibrium between the shallower and the deeper reservoirs with identical magma excess pressures, \( ps = pd \), where \( ps \) and \( pd \) are the excess pressures within the shallower and deeper reservoirs, respectively. An eruption primarily fed by the shallower reservoir is associated with a larger decrease in excess pressure of
the shallower than of the deeper reservoir, and after the eruption we have \( ps < pd \). This pressure gradient would drive newly injected magma toward the shallower reservoir, and inflate it. When excess magma pressure equilibrium is reached \( (ps = pd) \), the pressure gradient to the deep roots of the plumbing system would drive additional magma into the shallow system and both sources would inflate contemporarily. When a threshold is reached, inflation of the deeper reservoir ceases. We interpret this cessation as the consequence of limited source compliance (ellipsoidal cavity). The deeper source cannot further expand and excess pressure increase within the system is primarily accommodated by the inflation of the shallower reservoir. Similarly, a subvolcanic lateral intrusion primarily fed by the deeper reservoir is associated with a larger decrease in excess pressure of the deeper than of the shallower reservoir.

\[ \text{Figure 3.12.} \] Schematic representation of the excess pressure within the two reservoirs after (a) eruptions and (b) deep intrusions. \( ps \) – excess pressure in the shallower reservoir. \( pd \) – excess pressure in the deeper reservoir.

After the intrusion we have \( pd < ps \). In this case the pressure gradient would drive new magma primarily into the deeper reservoir. We would expect inflation of the deeper
reservoir only, but we observe contemporaneous inflation of both sources. We interpret the inflation of the shallower reservoir as the result of the high compliance of this sill-type reservoir. A small increase in excess pressure results in significant inflation of this reservoir. In the same way as following eruptions, once the threshold is reached inflation of the deeper reservoir ceases.

3.7.3 Subvolcanic lateral intrusions and implications for caldera dynamics

The December 2006 (E2) and August 2007 (E3) events represent evidence for subvolcanic lateral intrusions at Fernandina. Even if the 1927 Punta Espinoza rapid uplift and the 1954 Urvina Bay event suggested the occurrence of subvolcanic intrusions at Fernandina and at the other Galápagos volcanoes, such deformation has never been directly measured. Our data provide evidence for the rapid intrusion of relatively large volumes of magma at depth (e.g., ~0.02 km$^3$ in August 2007), comparable with intruded and erupted volumes associated with the previous eruptions (e.g., 1995 eruption, ~0.05 km$^3$ [Jónsson et al., 1999; Rowland et al. 2003] and 2005 eruption, ~0.026 km$^3$ [Chadwick et al., 2011]). Furthermore, we can speculate that similar events might have occurred in the recent history of the volcano, but gone unnoticed because of less dramatic uplift. These intrusions could also represent an important mechanism for the growth of the volcanic edifice and the surrounding platform and be complementary to the effusive activity.

We also observe that seismicity accompanies both intrusions, but the exact temporal relationship to the intrusions is unclear. While for E2, surface deformation is not yet recorded after 6 h from the first earthquake, for E3 a partial amount of the total
displacement is already recorded after 8 h from the first major shock. We also observe that, at least for E3, seismicity continues during the intrusion and stops once it is completed. If these earthquakes have tectonic origin they could represent the trigger mechanism for the subvolcanic intrusions. The same seismicity could also be a consequence of the intrusion itself and represent the brittle response of the crust to a volcanic intrusion, or be the combination of both mechanisms.

3.7.3.1 Caldera subsidence and collapse

An important aspect of the inferred intrusions is their effect on the summit caldera. We have shown that rapid subsidence of the caldera floor occurred during both events (−0.10 m during E2 and −0.31 m during E3). If this subsidence is considered as volumes of magma withdrawn from the storage system, we demonstrate that comparable volumes of magma are erupted at the surface or intruded a depth.

For the two intrusions studied here the measured deformation represents the elastic response of the overlying portion of the edifice to the pressure decrease within the reservoir but, if larger volumes of magma are withdrawn, the pressure within the system could drop enough to satisfy the failure criteria and cause the onset of a piston-like collapse. Furthermore, the transfer of magma to the flanks during the intrusions expands the volcanic edifice and widens the fractures bounding the block overlying the reservoir, facilitating its downward sliding. If freed to descend, the piston can accelerate the process by pushing more magma toward the deeper reservoir. This mechanisms could explain the 1968 caldera collapse and the discrepancy between the volume of magma erupted at the surface prior to the collapse and the volume of the collapse itself. The estimated 1–2 km3
of magma (total volume of the collapse) could have withdrawn from the shallow reservoir from a combination of both eruptive (0.02 km3) and intrusive activity. Multiple intrusions could have occurred and continued during the 12 days of the collapse. Evidence could be found in the incremental collapse inferred from the seismicity that accompanied this event [Stix and Kobayashi, 2008]. A feedback mechanism with the flank intrusions widening the edifice and enabling the downward motion of the piston, and the piston increasing the pressure within the system and pushing more magma toward the intrusions could have generated a large portion of the total volume of the collapse.

3.8 Conclusions

We use single interferograms, LOS displacement time series and the modeling of the observed deformation to study the volcanic activity at Fernandina Volcano. We interpret the surface displacement as the expression of magmatic sources embedded within the volcanic edifice.

1. The shallow magma storage system is composed of two reservoirs at different depths, modeled as a shallower flat-topped body at ~1.1 km b.s.l. and a deeper ellipsoidal cavity at ~4.9 km b.s.l. Our findings are in agreement with previous geodetic and petrologic studies. Similar and recurrent patterns of deformation during the eight-year-long study period are interpreted as the activity of long-term magma reservoirs.

2. The two magma reservoirs are hydraulically connected given the rapid and contemporary response to pressure release events such as eruptions and subvolcanic lateral intrusions.

3. Further insights into the magma migration and reservoir dynamics are provided by
post-eruptive/post-intrusive phases. Re-filling of the shallower reservoir first characterizes post-eruptive phases, while contemporary re-filling of both reservoirs occurs after deep lateral intrusions. We propose that pressure gradients within the plumbing system, together with different stiffness/compliance of the inferred sources, control the source activity when pressure increases.

4. In two occasions (December 2006 and August 2007), sills departing from the deeper reservoir intruded under the southern flank of the volcano generating broad uplift. A similar mechanism could explain the rapid deformation observed at Punta Espinoza in 1927. Earthquake swarms are associated with both intrusions. This seismicity could have tectonic origin and act as a trigger mechanism for the intrusion. It also could simply be the consequence of magma movement through the brittle crust or a combination of both.

5. Magma withdrawal from the shallow storage system and pressure decrease during intrusions also cause the rapid subsidence of the summit and the caldera floor. This aspect is important to understand the dynamics of the summit caldera and can be used to explain the discrepancy between the volume of the 1968 collapse and the volume of the magma erupted prior to the event. During this event magma likely migrated downward from the shallow reservoir and largely intruded at depth causing the removal of the necessary support to the overlaying block and triggered an incremental caldera collapse.

3.9 Acknowledgments

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Chapter 4: A new model for the growth of basaltic shields based on deformation of Fernandina Volcano, Galápagos Islands

4.1 Summary

Space-geodetic measurements of surface deformation produced by the most recent eruptions at Fernandina – the most frequently erupting volcano in the Galápagos Archipelago – reveal that all have initiated with the intrusion of subhorizontal sills from a shallow magma reservoir. This includes eruptions from fissures that are oriented both radially and circumferentially with respect to the summit caldera. A Synthetic Aperture Radar (SAR) image acquired 1–2 h before the start of a radial fissure eruption in 2009 captures one of these sills in the midst of its propagation toward the surface. Galápagos eruptive fissures of all orientations have previously been presumed to be fed by vertical dikes, and this assumption has guided models of the origin of the eruptive fissure geometry and overall development of the volcanoes. Our findings allow us to reinterpret the internal structure and evolution of Galápagos volcanoes and of similar basaltic shields. Furthermore, we note that stress changes generated by the emplacement of subhorizontal sills feeding one type of eruption may control the geometry of subsequent eruptive fissures. Specifically, circumferential fissures tend to open within areas uplifted by sill intrusions that initiated previous radial fissure eruptions. This mechanism provides a possible explanation for the pattern of eruptive fissures that characterizes all the western Galápagos volcanoes, as well as the alternation between radial and circumferential fissure eruptions at Fernandina. The same model suggests that the next
eruption of Fernandina will be from a circumferential fissure in the area uplifted by the 2009 sill intrusion, just southwest of the caldera rim.

### 4.2 Overview

Current understanding of the construction and evolution of basaltic shield volcanoes is largely based on studies of Hawai‘i. The great attention paid to Hawaiian volcanoes, particularly Kīlauea and Mauna Loa, is justified given their accessibility and long-term record of observation and instrumental monitoring [Kauahikaua and Poland, 2012]. Intrusive/effusive activity at these locations often occurs along narrow radial rift zones that are a product of magmatic and tectonic processes within the volcanic edifices [Dieterich, 1988, Fiske and Jackson, 1972 and Tilling and Dvorak, 1993]. Both geologic studies of eroded volcanoes [Walker, 1987] and geological/geophysical monitoring of active volcanoes [Pollard et al., 1983] indicate that intrusions within these rift zones occur as subvertical dikes, even proximal to the summit region. Such a conceptual model, however, represents an end member that is not necessarily representative of basaltic shields found elsewhere on Earth and other planets [Batiza et al., 1984, Jaggar, 1931, Macdonald, 1948, Montési, 2001 and Simkin, 1972].

Volcanoes of the western Galápagos Islands generally lack well-developed rift zones and are characterized instead by eruptive fissures oriented both circumferentially and radially with respect to the summit calderas [Chadwick and Howard, 1991, Figure 4.1]. These fissures are the surface expression of underlying subvolcanic intrusions that propagated from the magma reservoirs beneath the calderas. The conditions and mode of
emplacement of these bodies is therefore fundamental to understanding the internal growth of the volcanoes.

The cause of the great difference between patterns of eruptive fissures at Hawaiian volcanoes versus those of the western Galápagos Islands has long been a source of speculation, especially given the similar intra-plate hotspot origin for the two archipelagos [Chadwick and Howard, 1991 and Rowland and Munro, 1992]. Motion of volcanic flanks has been tied to rift zone evolution in Hawai‘i; however, flank instability in the Galápagos is not widespread, perhaps because the oceanic crust on which Galápagos volcanoes grow is young and lacks the sediment cover that may act to promote flank motion at other locations [Dieterich, 1988 and Nakamura, 1980]. Sequential growth of volcanoes in Hawai‘i has also been suggested to control rift zone geometry through buttressing of newer volcanoes against older ones [Dieterich, 1988 and Fiske and Jackson, 1972], but the western Galápagos volcanoes formed concurrently, limiting the buttressing effect between adjacent volcanoes [Naumann et al., 2002 and Rowland and Munro, 1992]. Most previous studies have attempted to explain the Galápagos radial-circumferential fissure pattern as due to topographic effects [Munro and Rowland, 1996, Simkin, 1984 and Simkin, 1972], a response to caldera collapse [Nordlie, 1973], or magma reservoir geometry [Chadwick and Dieterich, 1995]. All proposed models, however, make the assumption that both radial and circumferential eruptive fissures in the Galápagos are fed by subvertical dikes.

Here, we study surface deformation of Fernandina (the most frequently erupting volcano in the Galápagos) associated with the 2009 eruption and reexamine displacements associated with two previous eruptions (in 1995 and 2005). Our findings
lead us to propose a new model for the internal growth of Galápagos volcanoes, as well as other basaltic shields that do not follow the Hawaiian example. Moreover, we observe that circumferential fissures in 2005 opened within an area that was deformed during the previous radial fissure eruption in 1995. We investigate the relation between the two modes of eruption by calculating the stress changes generated by the intrusions that feed radial fissures. These models suggest that deformation associated with one eruption can be used to project the probable location and orientation of future eruptive fissures [e.g., Walter, 2008].

### 4.3 Fernandina Volcano, Galápagos

The application of space-geodetic techniques, such as Interferometric Synthetic Aperture Radar (InSAR) [Amelung et al., 2000], has provided an unprecedented opportunity over the past 20 years to study the subsurface structure of active volcanoes [e.g., Baker and Amelung, 2012]. At Fernandina, InSAR data spanning both eruptive and inter-eruptive time periods are the basis for models of the magma storage system of the volcano. Fernandina is characterized by at least two hydraulically connected magma reservoirs, at ~1 km and ~5 km b.s.l. [Bagnardi and Amelung, 2012, Chadwick et al., 2011 and Geist et al., 2006a]. The deeper reservoir appears to be the source of large sill-like intrusions in 2006 and 2007 that are indicated by broad uplift of the southern flank of the volcano [Bagnardi and Amelung, 2012], while the shallower reservoir primarily feeds summit and flank eruptions [Chadwick et al., 2011].

Among Galápagos volcanoes, Fernandina best represents the characteristic pattern of radial and circumferential eruptive fissures (Figure 4.1). InSAR is particularly instructive
with regard to the origin of the fissure pattern, since data span both circumferential (2005) and

Figure 4.1: Topography of Fernandina Island and its pattern of eruptive fissures. (a) Shaded relief map of Fernandina and surrounding ocean floor. Topography from SRTM data V4, bathymetry from NGDC-NOAA multi-beam data. In dark gray, areas where bathymetric data are not available. Solid lines represent circumferential (in blue) and radial (in red) eruptive fissures mapped by Chadwick and Howard [1991]. Purple-dashed lines mark submarine ridges identified by Geist et al. [2006b]. Inset: Location map of the Galápagos Islands. Black rectangle indicates area of part (b). (b) Close-up covering the location of the last three eruptions on Fernandina. Dashed lines mark eruptive fissures and filled polygons represent the surface covered by lava flows produced by each eruption.
Figure 4.2: Deformation at Fernandina. (a)–(e) Interferograms and modeling results. In gray are the deformation sources used to model magma reservoirs (rectangle: horizontal sill-like source at \(\sim 1\) km depth; ellipse and circle: deeper source at \(\sim 5\) km depth); yellow and orange rectangles outline planar sources used to model the intrusions feeding the eruptions (in orange, the subhorizontal segments). For inclined sources, the thicker line represents the shallower edge. (a) January–April 1995 eruption. (b) May 2005 eruption (model from Chadwick et al., 2011). (c) April 2009 pre-eruptive sill intrusion; interferogram formed using a SAR image acquired 1–2 h prior to the opening of the first eruptive fissure. (d) Typical pattern of deformation observed during several periods of intra-caldera uplift and due to the inflation of the \(\sim 1\)-km-depth magma reservoir (Envisat interferogram, T61, ascending, 16/01/2010–14/08/2010, model from Bagnardi and Amelung [2012]). (e) Total displacement associated with the April 2009 eruption.
radial (1995 and 2009) fissure eruptions, as shown in Figure 4.2. Jónsson et al. [1999], studied deformation associated with the 1995 radial fissure eruption (Figure 4.1b) and modeled the displacement recorded in a five-year differential interferogram (from the ERS-1/2 satellites) as due to a NE-SW-trending dike with a gentle dip (34° from horizontal) to the SE (yellow rectangle in Figure 4.2a). These results argued, for the first time, that dikes feeding radial fissure eruptions in the Galápagos may not be vertical.

Chadwick et al. [2011] found that Global Positioning System (GPS) and InSAR data spanning the May 2005 circumferential eruption (Figure 4.1b) could be fit by a dike that is shaped like a concave shell (curving both vertically and horizontally). The dike was steeply dipping near the surface (∼60°) but more gently dipping at its origin at ∼1 km depth (<15°), where it intersects the shallower magma reservoir (yellow, orange and gray rectangles in Figure 4.2(b)).

On 10 April 2009, between 23:30 and 00:00 local time, a new eruption started at Fernandina. The start time of the eruption is inferred from thermal data acquired by the Geostationary Operational Environmental Satellites (GOES; Figure 4.3). No thermal anomalies were apparent prior to 23:30 on April 10 (April 11, 05:30 UTC), while high temperatures were recorded on the southwestern part of Fernandina Island in a subsequent image, acquired at 00:00 on April 11 (06:00 UTC). The eruption was characterized by three fissures that formed a left-stepping en echelon set between an elevation of 400 m and 1100 m a.s.l., oriented radially to the summit caldera and in the same sector of the volcano as the eruptions of 1995 and 2005 (Figure 1b). Serendipitously, the Envisat satellite acquired a SAR image at 22:15 on April 10, 1–2 h before the opening of the first eruptive fissure, which captures the intrusion that fed the
eruption in the midst of its emplacement. Subsequent SAR images span the entire duration of the eruption and allow a detailed analysis of the surface deformation associated with the event.

Figure 4.3: GOES satellite thermal images showing the start time of the April 2009 eruption – Channel 2, 3.78–4.03 μm – yellow circle represents Fernandina volcano. The first appearance of a thermal anomaly (black pixel), representing the first evidence of a surface eruption, is in the 00:00 (06:00 UTC) image (black arrow).

4.4 Data and methods

4.4.1 InSAR data processing and modeling

We used SAR data from the European Space Agency’s ERS-1/2 and Envisat satellites and from the Japanese Aerospace Exploration Agency's JERS-1 satellite to construct interferograms showing deformation of Fernandina. All of the interferograms used in this study were processed using the Gamma SAR Processor and Interferometry software to focus the raw SAR images and the JPL/Caltech ROI_PAC SAR Software [Rosen et al., 2004] to form interferograms. We used precise DORIS orbits provided by the European Space Agency (ESA), and we removed the topographic contribution to the
interferometric phase using a 90-m-resolution (resampled to 30 m) digital elevation model (DEM) generated by the NASA Shuttle Radar Topography Mission [SRTM; Farr et al., 2007]. The interferograms were then phase-unwrapped using the SNAPHU algorithm [Chen and Zebker, 2001].

Deformation for both the 1995 and 2009 eruptions at Fernandina was modeled using analytical solutions for sources embedded in a homogeneous elastic half-space. For all models, we assumed a Poisson's ratio \( \nu = 0.25 \) and took into account the effect of topography on the surface deformation using the approach proposed by Williams and Wadge [1998]. The optimal solutions and their probability density distributions were estimated through the nonlinear inversion of the InSAR data, which was performed using the Monte-Carlo-based Gibbs Sampling (GS) algorithm [Brooks and Frazer, 2005]. The fit of the predicted deformation to the measured displacements was assessed using the normalized root-mean square error (RMSE) between observed and modeled interferograms. To reduce the number of data points (for which unit variance was assumed), a spatial averaging of full-resolution SAR interferograms was performed using a Quadtree algorithm [Jónsson et al., 2002].

4.4.2 Stress change modeling

Stress changes generated by the intrusion of planar magmatic bodies within the volcanic edifice were calculated using the Structural Mechanics module of the commercial Finite Element Modeling (FEM) program COMSOL Multiphysics. A three-dimensional model of the volcano (width, length and depth 100×100×100 km) was constructed using topographic data from the SRTM DEM and assuming a linear elastic and isotropic body with typical elastic properties for basaltic
extrusive and intrusive rocks (Poisson’s ratio $\nu=0.25$, shear modulus $\mu=10$ GPa) and density $\rho=2700$ kg/m$^3$ [Jónsson, 2009 and Rubin and Pollard, 1987]. The boundary conditions were set as free surface at the top and free to move horizontally and vertically at the bottom and sides of the model, respectively. We simulated the opening of planar bodies using inclined thin rectangular cavities with normal displacement (geometry, location and displacement inferred from the nonlinear inversion of the InSAR data). A convergence test was performed by moving the boundaries and changing the size of the mesh, which confirmed that those parameters do not significantly affect the numerical solutions.

To predict the probable location and orientation of future eruptive fissures, the stress changes caused by magmatic intrusions should be superimposed on the reference stress state within the volcanic edifice [Muller et al., 2001, Grosfils, 2007 and Bistacchi et al., 2012]; however, this reference stress is the result of multiple and complex factors (e.g., growth and evolution of the volcano, regional tectonics, etc.) and difficult to characterize. In our model, we made the assumption that, prior to each intrusion, the reference stress state is isotropic (or hydrostatic; e.g., Chadwick and Dieterich, 1995). Neither regional tectonic stress nor gravitational load were applied, and we only examine the perturbation to a stress-free host rock generated by magmatic intrusions that fed eruptions.

### 4.5 Deformation observations and modeling results

#### 4.5.1 April 2009 pre-eruptive intrusion

Differential interferograms formed using the Envisat SAR image acquired 1–2 h before the opening of the first eruptive fissure in 2009 show an area about 5.5 km in
diameter near the southwestern rim of the summit caldera that uplifted by a maximum of 0.50 m in the satellite's line-of-sight (LOS) direction (Figure 4.2c). The deformation clearly differs from the typical pattern of intra-caldera uplift that is caused by inflation of the shallower magma reservoir between eruptions, which is centered on, and generally confined within, the caldera (Figure 4.2d). Interferograms formed using a SAR image acquired ~13 h earlier (at 09:41 on April 10) do not show any evidence for similar displacements (see Figure S2.1 in Supplementary Material), implying that the anomalous pre-eruptive deformation occurred sometime between 09:41 and 22:15 on April 10.

We modeled the LOS displacements from the immediate pre-eruptive interferogram (Envisat, track 61, beam mode IS2, ascending, 31/01/2009–10/04/2009) using a planar source with uniform opening and no dip- or strike-slip motion [Okada, 1985] and for which all source parameters were set free to float within geologically realistic bounds during the inversion. Weak subsidence around the uplifting area was modeled using a prolate spheroidal cavity [Yang et al., 1988], approximating Fernandina's deeper magma reservoir, with source parameters (except for the normalized pressure change) fixed to values from Bagnardi and Amelung [2012]. The best-fitting model is a rectangular sill that dips gently (~26°) toward the center of the volcano, originates below the caldera at ~2.3 km b.s.l. and extends to ~0.80 km b.s.l. (orange rectangle in Figure 4.2c). Best-fitting planar source parameters are reported in Table S2.1 (Supplementary Material), and a comparison between the observed InSAR data, predicted LOS displacement, and the residual (difference between observed and predicted) is shown in Figure 4.4.
Figure 4.4: Modeling of deformation for the April 2009 pre-eruptive sill intrusion. Comparison between (a) measured, (b) modeled, and (c) residual LOS displacement recorded by an Envisat interferogram (track 61, beam model IS2, ascending, 31/01/2009–10/04/2009). Each fringe (full color cycle) represents 2.8 cm of LOS displacement. The green rectangle outlines the geometry of the best-fitting planar source (sill) while the green ellipse represents the center of the deeper magma reservoir beneath the caldera (modeled as an ellipsoidal cavity at ~5 km depth).

4.5.2 April 2009 eruption – total displacement

The total displacement produced by the 2009 eruption was recorded by two nearly synchronous interferograms (each spanning 105 days) from the Envisat satellite, one from an ascending orbit (track 61, beam mode IS2) and one from a descending orbit (track 54, beam mode IS7). The measured displacements (Figure 4.2e) reveal a pattern very similar to that associated with the previous eruption of a radial fissure, in 1995 (Figure 4.2a). Broad, semi-circular, positive LOS displacement (uplift) is present on the east side of the eruptive fissures and is superimposed on negative displacement (subsidence) ranging across the entire volcanic edifice.

Given the complexity of the measured surface displacement, the use of multiple sources of deformation (five, in our preferred model) is necessary. Edifice-wide and intra-caldera subsidence is interpreted as deflation of the two subcaldera magma reservoirs [Bagnardi and Amelung, 2012] and was modeled using a prolate spheroidal cavity (Yang et al., 1988, source r1) and a planar source with uniform opening [Okada,
For these two sources, all parameters that define position and geometry were fixed to values determined by Bagnardi and Amelung [2012], and we inverted only for the normalized pressure change (for r1) and opening (for r2). A third source of deformation (s) was used to reproduce the displacement generated by the pre-eruptive sill intrusion modeled in Section 4.1, and all parameters were fixed to the values of that best-fitting model. To explain the remaining displacement observed on the southwestern flank of the volcano and to best represent the surface expression of the intrusion feeding the eruptive fissures, we followed a multi-step approach. We first ran several forward models to test different combinations (one or more rectangular dislocation sources) and geometries. A reasonable fit to the observed data can only be achieved by using a minimum of two planar sources with uniform opening, oriented parallel to the eruptive fissures, dipping to the SE, and with a dip angle that becomes progressively more vertical with distance from the summit. Successively, we fixed the orientation, location, and size of these two planar sources (d1 and d2) to values inferred from the forward models (see Table S2.2 in Supplementary Material) and inverted for dip angle and source opening. The best fit is obtained for a dip angle of 33° from horizontal for d1, which is located closer to the summit, and 50° for d2, which is located beneath the flank of the volcano.

A comparison between the observed InSAR data, predicted LOS displacement, and residual is shown in Figure 4.5. Our model approximations and assumptions result in significant residual deformation for some areas. A clear overestimation of the predicted deformation is present in both viewing geometries where the sill intrusion (s) and the radial intrusion (d1) overlap, suggesting that our model does not do a good job of accounting for the complex geometry of the intrusion in this area. For the ascending pass
(Figure 4.5d–f), residual interferometric fringes are also present around the summit as result of an overestimation of the negative LOS displacement (subsidence) in this area, which we modeled as deflation of both magma reservoirs. It is probable that the simple geometry of r1 (prolate spheroid) is not adequate to explain the observed displacement and a more complex geometry is needed. A similar effect is probably responsible for the overestimated subsidence shown by residual fringes south and east of the radial intrusion. Also, the use of analytical solutions for planar sources having dimensions comparable to, or greater than, the source depth can lead to imprecise estimates of the surface deformation. Despite their limitations and the residuals in some areas, however, these models are instructive for interpreting the geometries of the pre-eruptive sill intrusion, as well as the intrusion that fed the eruption.

4.5.3 New insights into the January–April 1995 eruption

On the basis of our findings for the 2009 eruption, we reexamine the deformation associated with the similar 1995 eruption. As was the case in 2009, the 1995 eruption was also characterized by deformation of the summit plateau. Uplift of a sub-circular area centered near the southern caldera rim is recorded by two independent interferograms, one from the ERS-1/2 satellites (track 412, descending, 12/09/1992–30/09/1997, analyzed by Jónsson et al., 1999) and one from the JERS-1 satellite that spans a shorter time interval (Figure 4.2a; track 473, descending, 16/10/1993–11/05/1995). Although the quality of the interferometric signal is higher for the JERS-1 image, we only modeled the ERS-1/2 interferogram because of the absence of significant pre- and post-eruptive
deformation in that dataset (the JERS-1 interferogram probably records edifice-wide subsidence due to deflation of the deeper magma reservoir beneath the summit caldera).

**Figure 4.5:** Modeling of deformation for the April 2009 eruption. Comparison between ((a) and (d)) measured, ((b) and (e)) modeled and ((c) and (f)) residual LOS displacements recorded by two Envisat interferograms (Desc. – 30/01/2009–15/05/2009; Asc. – 31/01/2009–16/15/2009). Each fringe (full color cycle) represents 2.8 cm of LOS displacement. The green ellipse represents the center of the deeper deflation source (r1), and the green rectangles outline the geometries of the planar source models (r2, s, d1 and d2).
Our best fit to the observed deformation is obtained using two planar sources [Okada, 1985], one centered beneath the southwestern caldera rim and dipping (34°) toward the center of the volcano, and one parallel to the radial eruptive fissures and gently dipping (25°) to the SE (Figure 4.6, remaining source parameters in Table S2.3 – Supplementary Material). The geometry and location of the latter source are similar to that of Jónsson et al. (1999), who did not model the remaining deformation at the summit because of decorrelation of the interferometric signal in that area (possibly caused by their use of a different DEM to remove the topographic contribution to the interferometric phase). The deformation model for the 1995 radial eruption therefore strongly resembles that for the 2009 radial eruption, with a shallowly dipping dike on the flank and an inward-dipping sill beneath the summit plateau.

Figure 4.6: Modeling of deformation for the January–April 1995 eruption. Comparison between (a) measured, (b) modeled and (c) residual LOS displacement recorded by the ERS-1/2 interferogram (12/09/1992–30/09/1997). Each fringe (full color cycle) represents 2.8 cm of LOS displacement. The green rectangles outline the geometry of the planar source models.
4.6 Discussion

4.6.1 Geometry and origin of subvolcanic intrusions

Our deformation modeling, together with inferences from previous studies [Chadwick et al., 2011 and Jónsson et al., 1999], reveals the surprising result that, despite their radically different orientations, the 1995 and 2009 radial and the 2005 circumferential eruptions at Fernandina were all fed by shallowly dipping sill intrusions that initiated from, or just beneath, the ∼1-km b.s.l. magma reservoir. In 2005, the intrusion curved to become a steeply dipping dike as it rose toward the surface and erupted as a circumferential fissure near the caldera margin (Figure 4.7a). This rotation can perhaps be explained by mechanical models indicating that when sills reach a length-to-depth ratio of the order of unity, they are likely to interact with the free surface and rise to shallower levels [Fialko, 2001]. In 1995 and 2009, however, the intrusions twisted about an axis oriented radial to the caldera as they propagated toward the flank of the volcano (Figure 4.7b), becoming shallow-dipping dikes by the time they intersected the surface to feed eruptions from radial fissures. These two radial fissure eruptions were apparently fed by magmatic intrusions with nearly identical geometries but slightly different orientation (directed SSW in 1995 and SW in 2009).

Figure 4.7: Three-dimensional representation of circumferential (a) and radial (b) intrusions; in purple the ∼1-km-depth magma reservoir, in yellow the intrusions feeding fissure eruptions (circumferential intrusions as in Chadwick et al., 2011).
A transition from a sill-like intrusion to a radially aligned dike at some distance from the reservoir is supported by numerical models of interactions between the magma reservoir and volcanic edifice [Hurwitz et al., 2009]. Furthermore, the submarine morphology of Fernandina suggests that intrusions that propagate more than ~10 km beneath the flanks of the volcano may eventually twist into subvertical orientations. Three linear volcanic ridges are present below sea level on the western side of the island (Figure 4.1a), ~13–14 km from the summit, and could represent volcanic rift zones similar to those of Hawai‘i [Geist et al., 2006b]. By analogy, these rift zones are probably underlain by subvertical dikes, leading us to propose that a complete 90-degree twist, from subhorizontal to subvertical, is possible only if the intrusion propagates far enough from the summit.

Although slightly inclined sills initiated both radial and circumferential eruptions, modeling results suggest that they may originate at different depths. In 2005, a subhorizontal intrusion propagated from a modeled depth of ~1 km b.s.l. [Chadwick et al., 2011], which coincides with the top of the shallower magma reservoir (inferred from independent data; e.g., Bagnardi and Amelung, 2012 and Chadwick et al., 2011). Sill intrusions in 1995 and 2009, however, were modeled to have originated at ~2.0 and ~2.3 km b.s.l. respectively (the difference between the two is probably not significant). Support for these findings can be found in the strikingly different petrologic signatures between the lavas erupted in 1995 and 2005, with the former bearing abundant phenocrysts (plagioclase, clinopyroxene, and olivine) and the latter containing only few small phenocrysts grown from the liquid shortly before eruption. To explain this difference, Chadwick et al. [2011] proposed that the intrusion that fed the 1995 eruption
eroded plagioclase-rich crystal mush from the margin of the magma reservoir. In contrast, the 2005 dike propagated from the top of the reservoir and did not encounter the deeper mush zone.

On a broader scale, the characteristic pattern of radial and circumferential fissures shown at Fernandina is evident to varying degrees at all volcanoes in the western Galápagos [Chadwick and Howard, 1991], suggesting that the processes controlling intrusive and eruptive dynamics at Fernandina are common to all the edifices in the archipelago. Similar patterns of eruptive fissures are also present outside the Galápagos, including many seamounts [Batiza et al., 1984 and Simkin, 1972], other ocean islands [Jaggar, 1931 and Macdonald, 1948], and even on Mars [Montési, 2001]. Our results and interpretations are consistent with the numerical modeling of Grosfils [2007], who demonstrated that elastic models of magma-reservoir rupture cannot produce laterally propagating, subvertical dikes under any geological and geometrical conditions. Feeding radial fissures in the Galápagos and elsewhere via rotated subhorizontal sills, therefore, is consistent with modeled stress conditions within a volcanic edifice. Magma transport by mean of subhorizontal sills has also been recognized as an important element controlling the growth and evolution of shield volcanoes of the Canary Islands [Staudigel et al., 1986 and Ancochea et al., 2008] and La Réunion Island [Famin and Michon, 2010]. Our model for magma transport at the Galápagos may therefore be broadly applicable to other subaerial, submarine, and extraterrestrial volcanoes, and provides an alternative to interpretations based on the example of Hawai‘i.
4.6.2 Type and location of the next Fernandina eruption

Our observations and models of Fernandina may also be used to forecast the location and orientation of future eruptive fissures [e.g., Walter, 2008]. Previous studies [Chadwick and Dieterich, 1995 and Chadwick et al., 2011] have suggested a general feedback between radial and circumferential fissure eruptions because the stresses created by a radial dike would favor future intrusions with a circumferential orientation, and vice versa. This is consistent with the recent alternation between fissure eruption trends (circumferential in 1982, radial in 1995, circumferential in 2005, and radial in 2009, all in the same sector of the volcano).

Modeling of stress changes produced by the intrusion of a sill beneath the summit plateau in 1995 indicates a positive change of the least compressive stress in the crust, \( \sigma_3 \), beneath the area in which the 2005 circumferential eruptive fissures opened (Figure 4.8a). This change is directed radially with respect to the summit, favoring the intrusion of a circumferential dike in this area. We therefore suggest that the orientation and location of the 2005 eruptive fissures was controlled by the stress perturbation imposed by the sill intrusion that was emplaced at the start of the 1995 eruption. Based on this example, the model of deformation during the April 2009 eruption could be used to forecast the type and location of future eruptive fissures. The stress change generated by the 2009 pre-eruptive sill intrusion indicates that the greatest positive change in \( \sigma_3 \) occurred along the SW rim of the summit caldera (Figure 4.8b). We therefore project that the next eruption at Fernandina will be from circumferential fissures within the area uplifted by the 2009 sill intrusion, near the caldera rim and northwest of the 2005 eruption site.
Figure 4.8: Models of stress change due to sill intrusions. Magnitude (background) and orientation (blue arrowheads) of the changes in the minimum compressive stress after the (a) 1995 and (b) 2009 sill intrusions (black rectangles) at 0 km depth (sea level). Arrowhead length is proportional to change in $\sigma_3$ and to its plunge (maximum when horizontal). White-dashed lines mark the location of the May 2005 eruptive fissures in part (a) and of the forecasted location of future fissures in part (b). Based on this example and the stress change due to the 2009 sill, we forecast that the next eruption will be from a circumferential fissure within the area uplifted by the 2009 sill intrusion.
4.7 Conclusions

Results from the analysis and modeling of space-geodetic data spanning both radial and circumferential fissure eruptions at Fernandina volcano have led us to a new interpretation of the dynamics of magma migration and internal growth at Galápagos volcanoes. Contrary to the assumption of magma transport through vertical dikes, we have demonstrated that both orientations of eruptive fissures are initiated by the intrusion of subhorizontal sills. These intrusions curve upward to become steeply dipping dikes when feeding circumferential fissures near the caldera margin or twist about a radially oriented axis to feed fissure eruptions on the flanks of the volcano. While the mechanism for the stress field that causes these unique geometries remains ambiguous, our model for the development of intrusions that feed eruptive activity at Fernandina does provide insights into the subsurface structure of Galápagos and similar volcanoes, as well as the process by which they grow.

The intrusion of subhorizontal sills during the initial stages of a radial fissure eruption could explain the apparent alternation between eruption types, with circumferential eruptions occurring as a consequence of perturbations in the stress field generated by the preceding radial fissure eruption. Such a model provides a means of forecasting the style and location of future eruptions, although their timing cannot be predicted at this point. The 2005 circumferential eruption occurred in the area that experienced the most uplift in 1995 due to emplacement of the sill that fed that radial fissure eruption. By analogy, and with the support of stress change modeling of the 2009 pre-eruptive sill intrusion, we anticipate that the next eruption at Fernandina will probably occur from circumferential fissures opening within the area uplifted in 2009.
4.8 Acknowledgements

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Chapter 5: Gravity changes and deformation at Kīlauea Volcano, Hawai‘i, associated with summit eruptive activity, 2009–12

5.1 Summary

Analysis of microgravity data collected at the summit of Kīlauea Volcano, Hawai‘i (United States), between December 2009 and November 2012, reveals a net mass accumulation at ~1.5 km depth beneath the northeast margin of Halema‘uma‘u Crater, within Kīlauea Caldera. Although residual gravity increases and decreases are accompanied by periods of uplift and subsidence of the surface, respectively, the volume change inferred from the modeling of interferometric synthetic aperture radar (InSAR) deformation data can account for only a small portion (as low as 8%) of the mass addition responsible for the gravity increase. We propose that, since the opening of a new eruptive vent at the summit of Kīlauea in 2008, magma rising to the surface of the lava lake outgasses, becomes denser, and sinks to deeper levels, replacing less-dense gas-rich magma stored in the Halema‘uma‘u magma reservoir. In fact, a modest density increase (<200 kg m$^{-3}$) of a portion of the reservoir can produce the positive residual gravity change measured during the period with the largest mass increase, between March 2011 and November 2012. Other mechanisms may also play a role in the gravity increase without producing significant uplift of the surface, including compressibility of magma and filling of void space by magma. The rate of gravity increase, higher than during previous decades, varies through time and seems to be directly correlated with the volcanic activity occurring at both the summit and the east rift zone of the volcano.
5.2 Overview

Microgravity measurements have been collected at Kīlauea Volcano, on the Island of Hawai‘i (United States), since 1975 and, when combined with deformation measurements, have yielded important insights into mass change within the volcano [Dzurisin et al., 1980; Jachens and Eaton, 1980; Johnson, 1992; Kauahikaua and Miklius, 2003; Johnson et al., 2010; Carbone and Poland, 2012; Carbone et al., 2013]. Campaign gravity and leveling data spanning the 29 November 1975 Mw 7.7 Kalapana earthquake revealed that mass loss beneath the summit due to drainage of magma into Kīlauea’s rift zones was larger than expected given the measured deformation, implying the creation of void space [Dzurisin et al., 1980; Jachens and Eaton, 1980]. Subsequent surveys, extending through 2008, indicated a steady gravity increase (up to 450 µGal during 1975–2008) accompanied by subsidence (maximum of almost 2 m) near Halema‘uma‘u Crater within Kīlauea Caldera, which suggests filling of void space by magma [Johnson et al., 2010] or accommodation of additional magma volume by rifting of the summit [Zurek and Williams-Jones, 2013]. Since March 2008, a new long-term eruption has been occurring at Kīlauea’s summit, and a fissure eruption interrupted the ongoing (since 1983) east rift zone (ERZ) eruption in March 2011 (Figure 5.1).

We completed 5 gravity surveys of the Kīlauea summit network between December 2009 and November 2012 to assess whether the gravity increase measured during 1975–2008 continued and also how recent volcanic activity impacted the distribution of mass within the summit reservoir system. The new surveys have a higher temporal resolution (5-15 months) than previous measurements, which were carried out only once every several years prior to 2008. The 2009–12 time interval includes periods of both summit
inflation and deflation, as deduced from Interferometric Synthetic Aperture Radar (InSAR) data, and also spans variations in summit and ERZ eruptive activity.

**Figure 5.1** – Shaded relief map of the summit of Kilauea Volcano. Yellow circles indicate the locations of gravity stations that were measured in 2012. The location of the reference station P1 is marked by a black circle. HMM = Halema‘uma‘u Crater. Red ellipse marks the location of the summit eruptive vent, which hosts an actively circulating lava lake. Black lines outline major faults and craters. Insets show the location of Kilauea’s summit with respect to Hawai‘i Island and other features of the volcano. ERZ = east rift zone. SWRZ = southwest rift zone.

We modeled the gravity variation and deformation using analytical solutions to infer mass and volume changes over the three years spanned by the surveys. Our results indicate that mass continued to accumulate beneath the summit of Kilauea through November 2012. Although surface uplift was also measured, the increase in reservoir volume inferred from deformation is an order-of-magnitude smaller than that needed to
produce the measured gravity changes, assuming the volume was filled by basaltic magma. This suggests that the mass increase has occurred through mechanisms that do not produce significant uplift of the surface.

### 5.3 Kilauea Volcano

Kilauea has been erupting continuously since 1983 from vents on the volcano’s ERZ (Figure 5.1) [Heliker and Mattox, 2003]. The first 20 years of the eruption were characterized by ~2 m of subsidence of the south part of the caldera [Cervelli and Miklius, 2003; Johnson et al., 2010], beneath which lies the volcano’s main magma storage area centered at ~2.5–5-km depth [Eaton, 1959, 1962; Dvorak et al., 1983; Davis, 1986; Delaney et al., 1990, 1993; Cervelli and Miklius, 2003; Baker and Amelung, 2012]. Gravity change associated with the south caldera source during this time period consisted of a small decrease, suggesting that most of the magma that intruded the volcano was transported to the ERZ and that summit subsidence was a combination of several processes, including extension due to seaward motion of the volcano’s south flank [Johnson, 1987, 1992; Kauahikaua and Miklius, 2003; Plattner et al., 2013]. Subsidence switched to uplift during 2003–07 due to a surge in magma supply to the volcano [Poland et al., 2012]. The sources of uplift were not only the south caldera magma reservoir, but also small reservoirs ~1–2 km beneath the east margin of Halema’uma’u Crater [Cervelli and Miklius, 2003; Poland et al., 2009; Montgomery-Brown et al., 2010; Baker and Amelung, 2012; Lundgren et al., 2013] and ~2.5–5-km depth near Keanakāko’i Crater [Baker and Amelung, 2012; Poland et al., 2012]. Unlike the south caldera source, the shallow Halema’uma’u reservoir experienced a large mass increase during 1975–2008, but the lack of coincident inflation led Johnson et al. [2010] to propose that magma was
filling subsurface void space, and Zurek and Williams-Jones [2013] to speculate that rifting of the summit led to additional magma accumulation without uplift. Summit deformation fluctuated after 2007, with inflation and deflation tied to changes in ERZ eruptive activity [Poland et al., 2008, 2012; Lundgren et al., 2013].

Coincident with the ongoing ERZ activity, an eruption started at the summit of Kīlauea with the opening of a pit crater along the southeastern margin of Halemaʻumaʻu Crater (HMM in Figure 5.1) on 19 March 2008 [Wilson et al., 2008; Houghton et al., 2011; Orr et al., 2013]. The new vent is located within a few hundred meters of the positive gravity change measured during 1975–2008. Since its formation, rim and wall collapses have enlarged the opening of the pit from an initial diameter of 35 m to over 220 m by November 2012 [Richter et al., 2013; Orr et al., 2013]. A lava lake was observed within the vent in September 2008, and since then the level of its surface has experienced multiple cycles of rise and fall [Patrick et al., 2011; Orr et al., 2013], reaching a maximum height of 1006 m a.s.l. (~22 m beneath the floor of the Halemaʻumaʻu Crater) in October 2012. These cycles follow the general pattern of surface deformation, with the lava-lake level rising during periods of summit inflation and falling during deflation—a mechanism that suggests a direct coupling to pressure variations within the plumbing system of the volcano [Patrick and Orr, in review].

Two continuous gravimeters were installed in 2010 at the summit of Kīlauea to track gravity change associated with the new eruptive vent; these installations proved excellent at characterizing short-term changes in gravity and eruptive activity. For example, in May–June 2010, the instruments detected a gravity oscillation with a period of 2–5 minutes that had a source in the shallow Halemaʻumaʻu magma reservoir and may have
been related to rapid magma convection [Carbone and Poland, 2012]. The gravity signal during draining of the summit lava lake associated with the March 2011 ERZ Kamoamoa fissure eruption [Lundgren et al., 2013] was used to derive the density of the upper ~120 m of the lava lake, which was found to be 950 (+/- 300) kg m$^{-3}$ [Carbone et al., 2013]. Although continuous gravity measurements offer excellent temporal resolution to characterize mass movement on the order of minutes to days, they usually do not provide information on longer-term (weeks to years) processes because of instrumental effects [Carbone et al., 2003]. The elevated cost of gravimeters also prevents the deployment of large arrays at single locations, limiting the spatial resolution of continuous gravity measurements. Campaign gravity surveys across a wide network of stations, on the other hand, can help in resolving long-term gravity changes and constraining the spatial distribution of subsurface mass flow, although the temporal resolution is limited by the repeat time of the campaigns.

5.4 Data

5.4.1 Microgravity data

A network of fixed benchmarks across Kīlauea’s summit region was reoccupied five times between December 2009 and November 2012 (Figure 5.1, see Table 5.1 for specific information about each measurement campaign) using, simultaneously, two Scintrex CG-5 gravimeters (CG-578 and CG-579). The two instruments were calibrated against one another by repeated occupations of three calibration lines: Mauna Kea (Hawaii), Mt Hood (Oregon) and Mt Hamilton (California) [Barnes, 1968; Oliver and Barnes, 1968]. If the meter has a linear response, its calibration can be represented by a
single scale factor [Valiant, 1991]. These factors are 0.999736±0.000104 for CG-578 and 0.999343±0.000147 for CG-579.

### Table 5.1. Gravity surveys at Kīlauea’s summit 2009–12

<table>
<thead>
<tr>
<th>Date of survey</th>
<th>Measured stations</th>
<th>Average standard deviation (µGal)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2-18 December, 2009</td>
<td>47</td>
<td>13.1</td>
</tr>
<tr>
<td>25 June- 16 July, 2010</td>
<td>49</td>
<td>12.8</td>
</tr>
<tr>
<td>15-25 March, 2011</td>
<td>53</td>
<td>11.9</td>
</tr>
<tr>
<td>1-22 June, 2012</td>
<td>55</td>
<td>14.9</td>
</tr>
<tr>
<td>23 October - 28 November, 2012</td>
<td>55</td>
<td>12.0</td>
</tr>
</tbody>
</table>

Data were collected following a daily double-looping procedure consisting of three occupations of the reference station (at the start and at the end of each loop) and two occupations of selected sites (once during each loop). All gravity measurements are relative to benchmark P1, located 4 km northwest of the caldera center (black circle in Figure 5.1). P1 is assumed to be stable, although the area in which it is located did experience surface deformation (< 0.02 m) coincident with summit deflation and inflation that occurred during and after the March 2011 ERZ eruption, respectively. The assumption of stability can be justified because the measured deformation would have contributed a gravity signal of no more than 7 µGal—a value within the overall uncertainty of our measurements and two orders-of-magnitude smaller than the maximum gravity change we measured.

We reduced the gravity measurements using the GTOOLS software [Battaglia et al., 2012]. The code first adjusts gravity measurements for solid Earth tides, ocean loading
and instrument drift, then computes the weighted least-square-adjusted gravity values and their standard deviations. Specifically,

1. Earth tides are estimated using an improved version of Longman’s [1959] model that includes: (a) the original formulas by Bartels [1957, p. 747] for the Moon longitude, (b) updated values for the astronomical constants from USNO [2011] and (c) the gravimetric factor for an anelastic Earth [Agnew, 2007];

2. ocean loading is computed using the HARDISP code [Agnew, 2010] and ocean loading harmonics from the TPXO7.2 ocean tide model [Bos and Scherneck, 2012];

3. instrument drift is corrected using a linear function.

All corrections are performed up to microGal (µGal) precision, in accordance with the specifications of high-resolution surveys. The average standard deviation of all measurements collected during individual surveys varies between 12 and 15 µGal (Table 5.1).

Residual gravity changes at each station ($\Delta g_r$, in µGal) were then calculated for each time interval as $\Delta g_r = \Delta g - 308.6 \times \Delta h$, where $\Delta g$ is the gravity change in µGal at one station between two surveys, $\Delta h$ is the vertical displacement in meters at that station during the same interval, and -308.6 µGal/m is the theoretical free-air gradient [LaFehr, 1991]. The vertical displacement was calculated by combining ascending and descending InSAR measurements (see Section 3.2) that span the same time interval as the gravity measurements. Although variations in the height of the water table can produce changes in the measured gravity [e.g., Battaglia and Hill, 2009], at Kīlauea the water table is ~500
m beneath the surface [Kauahikaua, 1993] and experiences only minor fluctuations [J. Kauahikaua, personal communication, 2010]. We therefore follow the approach of previous studies [Johnson, 1992; Kauahikaua and Miklius, 2003; Johnson et al., 2010] and do not consider water table effects to be a significant source of gravity change [Battaglia et al., 2003].

5.4.2 InSAR data

We used synthetic aperture radar (SAR) data acquired along both ascending and descending orbital passes by the German Space Agency (DLR) TerraSAR-X satellite (from December 2009 to June 2012) and the Italian Space Agency (ASI) Cosmo-SkyMed satellite constellation (from June to November 2012). The TerraSAR-X dataset includes 74 images from track 24 (descending, beam mode strip_007, incidence angle 31°) and 80 images from track 32 (ascending, beam mode strip_008, incidence angle 33°). The Cosmo-SkyMed dataset is composed of 10 images from an ascending track (incidence angle 39°) and 12 images from a descending track (incidence angle 41°).

The TerraSAR-X interferograms were processed using the JPL/Caltech ROI_PAC SAR Software [Rosen et al., 2004], while the Cosmo-SkyMed data were processed using the GMTSAR InSAR processing system [Sandwell et al., 2011]. We removed the topographic contribution to the interferometric phase using a 30-m-resolution digital elevation model (DEM) generated by the NASA Shuttle Radar Topography Mission (SRTM) [Farr et al., 2007]. The interferograms were then phase-unwrapped using the SNAPHU algorithm [Chen and Zebker, 2001].
To resolve the temporal evolution of surface deformation, we used the small-baseline subset (SBAS) method [Berardino et al., 2002; Lanari et al., 2004; Fattahi and Amelung, 2013] and generated InSAR time series with inversions done independently for each orbital track. InSAR displacement time series are measured along the radar line-of-sight (LOS), but measurements of the vertical component of the deformation at the time of each gravity survey are necessary to correct the microgravity data for the free-air effect. We therefore combined the InSAR time series results from ascending and descending orbital passes to calculate the vertical component of motion following the method of Wright et al. [2004] (for a detailed description of the approach see Baker and Amelung [2012] and Baker [2012]).

5.5 Residual gravity changes and deformation at Kīlauea’s summit

Time series of residual gravity changes and vertical deformation during 2009–12 for selected stations representative of key areas of Kīlauea’s summit are presented in Figure 5.2. The maximum change throughout the entire time interval occurred at station HOVL-G (Figure 5.2a), which is located 20 m SE of the rim of Halema‘uma‘u Crater, 80 m above the crater floor, and 150 m east of the center of the summit eruptive vent. This station, which was first measured in July 2010 after installation of a continuously recording gravimeter at the site [Carbone and Poland, 2012; Carbone et al., 2013], shows a moderate negative change (-39±9 µGal) by the time of its second occupation in March 2011, after the summit of the volcano had subsided in response to an ERZ fissure eruption during 5–9 March 2011 [Lundgren et al., 2013]. From March 2011 through November 2012, however, a large positive change of 370±35 µGal is measured at the site
Figure 5.2 – Residual gravity change (error bars) and vertical deformation (filled circles, TSX = TerraSAR-X; open circles, CSK= Cosmo-SkyMed) for selected areas around Kīlauea’s summit. (a) Sites located near the SE rim of Halema’uma’u Crater (adjacent to the summit eruptive vent), which show the maximum residual gravity changes over the course of the measurements. (b) Sites located in the central part of the summit caldera. (c) Sites located in the south part of the summit caldera. (d) Sites located in the upper part of the east rift zone. Maps show the location of each station (plot colors correspond to station colors on associated maps). Error bars indicate one standard deviation of uncertainty for the residual gravity measurements. Vertical deformation and residual gravity change scales are the same in all plots to highlight differences in the magnitudes of changes between areas. Vertical deformation is obtained from InSAR data (combined LOS displacements from ascending and descending orbits) and is shown at the times of all SAR images acquired during the studied time interval.
during a period of modest uplift (~0.15 m). Two other stations located within 250 m of HOVL-G—HVO41 and 205YY (Figure 5.2a)—show the same trend in residual gravity but, despite their close proximity, the magnitude of the March 2011–November 2012 change is lower (226 and 231 µGal at HVO41 and 205YY, respectively).

Significant changes in residual gravity are also observed in the central portion of Kīlauea Caldera (Figure 5.2b). Stations in this region show similar trends to those located closer to the rim of Halemaʻumaʻu Crater, but gravity changes are smaller in magnitude during March 2011–November 2012 (80-123 µGal) despite similar ground uplift. Residual gravity changes and deformation at stations located in other areas of Kīlauea’s summit, including those over other important features of the volcano’s magma plumbing system such as the south caldera (Figure 5.2c) and the upper ERZ (Figure 5.2d), are within the uncertainty of the measurements and indicate no significant change in subsurface mass.

To better resolve the spatial characteristics of changes in gravity and deformation, we divided the three-year time interval (Dec. 2009 – Nov. 2012) into four periods defined by the epochs of the gravity surveys. For all periods, we present the residual gravity changes measured at each station together with their uncertainty (one standard deviation, Figure 5.3) and the surface deformation in the LOS direction measured by InSAR time series (Figure 5.4).

During the first time period, between December 2009 and July 2010, a positive residual gravity anomaly is centered on and limited to the summit caldera (Figure 5.3a). The magnitude of this anomaly is moderate (53±20 µGal at its maximum), but its spatial distribution matches the area of uplift also centered on the summit caldera, which is
**Figure 5.3** – Maps of residual gravity change at Kīlauea’s summit. The entire time interval is divided into four periods (1 to 4, panels (a) to (d)) based on the dates of the gravity surveys. Each station is marked with a filled circle that is color-coded according to the magnitude of the gravity change (warm colors indicate positive changes and cold colors indicate negative changes). For each station, the calculated residual gravity change and, in parenthesis, one standard deviation of uncertainty are also indicated. Gray lines outline major faults and craters.
Figure 5.4 – Surface deformation at Kīlauea’s summit measured by InSAR time-series. The entire time interval is divided into four periods (1 to 4, panels (a) to (d)) based on the dates of the SAR acquisitions closest to those of the gravity surveys (dates given in upper left of each image). Black circles mark locations of gravity stations and black lines outline major faults and craters. Satellite flight direction, look direction, and name (TSX = TerraSAR-X; CSK = Cosmo-SkyMed) are reported in each panel. Each fringe (full color cycle) represents 1.55 cm of LOS displacement (positive for uplift, negative for subsidence). Images show mean line-of-sight deformation for each time interval as deduced from time-series analysis. (a) Subsidence is centered on the south caldera and upper SWRZ, while uplift is centered near Halema‘uma‘u Crater. (b) Subsidence characterizes both areas and is at a maximum east of Halema‘uma‘u Crater. (c) Subsidence is centered on the upper SWRZ, and uplift is centered east of Halema‘uma‘u Crater. (d) Uplift is centered near Halema‘uma‘u Crater. In the southern portion of the image, localized deformation is apparent due to normal faulting during an earthquake swarm (max magnitude M=3.7) that occurred in the Koa‘e fault zone on 4–5 June 2012.
visible in the InSAR-derived surface displacement map (Figure 5.4a). This same period is characterized by broad subsidence south of the caldera and along the southwest rift zone (SWRZ) of the volcano, but stations located in these areas do not show any consistent residual gravity change.

The second period, starting in July 2010 and ending two weeks after the end of the March 2011 ERZ fissure eruption, is not characterized by large gravity changes, although negative values are present across the summit (Figure 5.3b). A few outliers are also present but, given their large discrepancy with measurements at nearby stations, can be disregarded. Subsidence of the ground (net maximum deformation -0.06 m) is centered east of Halema‘uma‘u Crater and merges with the broader subsiding area encompassing the south caldera and the SWRZ (Figure 5.4b). The net deformation during this period, however, does not reflect the complexity of the changes that took place. During the 5–9 March 2011, ERZ fissure eruption, the area east of Halema‘uma‘u Crater subsided by a maximum of 0.15 m [Lundgren et al., 2013]. Rapid uplift preceded and followed the eruption, reducing the magnitude of the net ground displacement measured between the epochs of the two gravity surveys.

The third and the fourth periods, from March 2011 to June 2012 and from June to November 2012, share similar characteristics to one another. During both time intervals, residual gravity changes are mostly positive, and maximum values occur along the southeast rim of Halema‘uma‘u Crater (Figure 5.3c-d), with positive changes extending outward from the center of the caldera. Further minor positive changes are present northeast of the caldera during the third period, but when summed with variations measured during the fourth period they become negligible. Also, these stations do not
overlie any known portion of the plumbing system of the volcano and, since they were all measured during the same day of the June 2012 campaign, we believe that these anomalous readings have been possibly caused by systematic errors. InSAR data for both time periods indicate uplift of the caldera centered east of Halema‘uma‘u Crater, in the same location as the center of subsidence during period 2 (Figure 5.4c-d). Most of the caldera uplift occurs during the months immediately following the March 2011 ERZ eruption. As during the previous time periods, subsidence characterizes the area south of the caldera and along the SWRZ, but no significant residual gravity changes are measured in these areas.

5.6 Lava-lake effect

Continuous gravity measurements have shown that significant gravity changes (tens of µGal) are associated with fluctuations in the level of the lava lake within Kīlauea’s summit eruptive vent [Carbone et al., 2013]. Data recorded at the rim of Halema‘uma‘u Crater during the 14-hour-long and 120-meter drainage of the lava lake, which occurred on 5 March 2011 coincident with the opening of a fissure on the ERZ, showed a gravity decrease of more than 100 µGal. This change was modeled as due to a mass removal from the pit of 2.5x10⁶ m³ of low-density magma (<1000 kg m⁻³). These results indicate that variations in the lava-lake level can also influence campaign measurements carried out in the vicinity of the summit eruptive vent. In fact, the maximum residual gravity changes (both positive and negative) calculated from campaign data are in most cases measured at the stations closest to the summit eruptive vent, and values decrease rapidly with distance from the vent.
We assessed the contribution of changes in lava-lake level to campaign gravity data using a numerical model that takes into account the pit geometry and the lava height inside the pit at the time of each gravity survey [Carbone et al., 2013]. The geometry of the conduit and of the lava lake surface is inferred from visual observations and ground-based LIDAR data. While the pit opening at the surface has progressively enlarged through episodic rim and vent wall collapses [Orr et al., 2013; Richter et al., 2013], no significant variations have been observed in the geometry of the vent at the level of the lava lake since the start of our campaign gravity data in December 2009. We therefore assume a constant model geometry for the summit vent for the entire time interval of our gravity measurements.

Our model for the gravity changes due to lava-level variations is the same of Carbone et al. [2013]. The pit hosting the lava lake is approximated by a cylinder with an elliptical section that abruptly widens at the top (Figure 5.5). The bottom is located at 800 m a.s.l. (228 m below the floor of Halema‘uma‘u Crater, which is lower than the minimum height reached by the lava lake during the time of our study) and has a 160-m-long major axis oriented NW-SE, and a 140-m-long minor axis oriented SW-NE. The top portion, forming a “ledge” that is only occasionally flooded by magma, has a major axis of 220 m and the same minor axis as the bottom part. The entire pit is discretized into 252 vertical square-based (10 x 10 m) parallelepipeds with changeable height. The height of the elements represents the lava level within the pit at a specific time (Table 5.2) for which we assume a density of 1000 kg m$^{-3}$ [Carbone et al., 2013]. The gravity effect produced by each parallelepiped is calculated [Talwani, 1973], and the total effect is obtained by summing the contribution of each element. Since we are interested in the
lava-lake contribution to the measured residual gravity changes, its effect is calculated at the distance of each measurement site and subtracted from the measured gravity changes between surveys.

Figure 5.5 – Geometry of summit eruptive vent and model of lava lake, taken from Carbone et al. [2013]. (a) Schematic cross section (redrawn from Orr et al., 2013) through Halema'uma'u Crater and the summit eruptive vent and showing vent shape as deduced from visual and LIDAR observations. (b) Model geometry consisting of 252 vertical square-based (10×10 m) parallelepipeds with changeable height to simulate varying lava levels. The shape of the model is intended to reproduce the asymmetric shape of the eruptive vent.

The calculated lava-lake effect is negligible at all campaign gravity stations for the first two time periods (December 2009 to June–July 2010 and June–July 2010 to March
2011) because the lava-level variations between the epochs of the gravity surveys are less than 20 m (resulting in less than 10 µGal of gravity change at the closest stations). For the two subsequent time periods (March 2011 to June 2012 and June 2012 to October–November 2012), however, the lava level is much higher and the level changes are much more significant—the lava-lake level rose by up to 167 m between March 2011 and November 2012. In Figure 5.6a, we show the gravity changes induced by variations in lava-lake level calculated for selected stations on the caldera floor. The maximum effect is at HOVL-G, and the effect decreases rapidly with radial distance from the summit vent, becoming almost zero at station HVO48, ~1000 m from the center of the lava lake.

We apply the calculated adjustment to the residual gravity changes at all those stations that are influenced by the lava-lake level change contribution (> 3 µGal), obtaining new datasets for periods 3 and 4 (Figure 5.6b-c). After this adjustment, the residual values at stations near the rim of Halema‘mau‘u Crater are similar to those measured in the central portion of the summit caldera (64-86 µGal during period 3 and 65-68 µGal during period 4).

<table>
<thead>
<tr>
<th>Date of survey</th>
<th>Height a.s.l. (m)</th>
<th>Height below the floor of Halema‘uma‘u Crater (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>December 2009</td>
<td>835</td>
<td>193</td>
</tr>
<tr>
<td>June-July 2010</td>
<td>829 - 854</td>
<td>199 - 174</td>
</tr>
<tr>
<td>March 2011</td>
<td>827 - 840</td>
<td>201 - 188</td>
</tr>
<tr>
<td>June 2012</td>
<td>936 - 971</td>
<td>92 - 57</td>
</tr>
<tr>
<td>October-November 2012</td>
<td>949 - 994</td>
<td>79 - 34</td>
</tr>
</tbody>
</table>
Deformation and gravity change modeling

Modeling of deformation and residual gravity was carried out in four steps. First, we performed a non-linear inversion of the measured LOS displacements, from ascending and descending orbital tracks (spanning approximately the same time intervals), to infer the characteristics of deformation sources that were active between December 2009 and
November 2012. These models were then used to track subsurface volume change during each time period spanned by gravity measurements. Successively, we inverted the residual gravity datasets to infer the source of mass change responsible for the observed residual gravity variations. Finally, we compared source depths, locations, and volume/mass changes obtained from the deformation and gravity data.

5.7.1 Deformation modeling

Two sources of deformation were active beneath Kīlauea’s summit during the 2009–12 time interval: one caused displacement of the ground just east of Halema‘uma‘u Crater (hereafter, HMM source), and the second produced broad subsidence along the SWRZ (hereafter, SWRZ source). To constrain these sources, we inverted InSAR data that span time periods when only an individual source is active (Table 5.3). This approach can reduce the ambiguity caused by modeling two or more overlapping sources of deformation. Also, while the SWRZ source seems to only be associated with subsidence, the HMM source produced both uplift and subsidence. To confirm that both types of deformation are generated by the same source, we inverted for data spanning intervals of both ground uplift and subsidence and then compared the results.

Both sources have been identified and characterized by previous studies [e.g., Poland et al., 2012; Baker and Amelung, 2012]. We therefore adopted the same approach and used the analytical solution for a finite spherical magma body [McTigue, 1987] to model the HMM source and that for a rectangular dislocation source with uniform opening [Okada, 1985] to reproduce ground displacement in the SWRZ. Both solutions are for sources embedded in a flat, isotropic, homogeneous, elastic half-space (Poisson’s
ratio $\nu = 0.25$). Since each original dataset consists of $10^6$ data points, we performed spatial averaging using the \textit{Quadtree} algorithm [Jónsson et al., 2002] to generate the data vectors used in the inversions. The optimal solutions and their probability density distributions were then estimated using the Monte-Carlo-based Gibbs Sampling (GS) algorithm [Brooks and Frazer, 2005]. The quality of the fit of the predicted deformation to the measured displacement was assessed using the normalized root-mean-square (RMS) error between observed and modeled InSAR LOS displacements.

<table>
<thead>
<tr>
<th>Deformation Source</th>
<th>Satellite</th>
<th>Date 1 - Date 2 (satellite pass)</th>
</tr>
</thead>
<tbody>
<tr>
<td>South Caldera - SWRZ</td>
<td>TerraSAR-X</td>
<td>15/12/2009 - 12/07/2010 (descending)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>16/12/2009 - 02/07/2010 (ascending)</td>
</tr>
<tr>
<td>HMM deflation</td>
<td>TerraSAR-X</td>
<td>17/02/2011 - 02/04/2011 (descending)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>01/03/2011 - 03/04/2011 (ascending)</td>
</tr>
<tr>
<td>HMM inflation</td>
<td>TerraSAR-X</td>
<td>24/04/2011 - 21/07/2011 (descending)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>25/04/2011 - 22/07/2011 (ascending)</td>
</tr>
</tbody>
</table>

LOS displacements measured during December 2009–July 2010 (Table 5.3) were used to constrain the SWRZ source, since deformation within the caldera was minimal. The best-fitting model is a $\sim 10.8 \times 0.9$ km rectangular source oriented NE-SW and centered at $\sim 3.9$ km depth beneath the upper SWRZ (see Supplementary materials). This geometry approximates a contracting sill, although processes other than magma withdrawal may contribute to, or be the predominant cause of, the observed subsidence (e.g., extension due to seaward motion of the volcano’s south flank [e.g., Plattner et al., 2013]). The lack of gravity change in this area means that our model assumption of a contracting sill will not influence our interpretations of gravity change in the caldera. It
is, however, necessary to account for the superposition of the broad subsidence with the deformation measured inside the caldera which, if not accounted for, will lead to biased estimates of the volume changes for the HMM source.

Surface deformation measured by InSAR data spanning the March 2011 ERZ eruption was used to characterize the HMM source during periods of subsidence, while data spanning the post-eruptive summit re-inflation, from April to July 2011, provided a means of characterizing the same source during periods of uplift (Table 5.3). Source depth (Figure 5.7a) and location (Figure 5.7b) are best characterized by the deflationary data because of the larger displacement (Figure 5.7c), but both inversions provide similar results: a spherical source located at ~1.5-km depth (1.2–1.7 km, 95% confidence interval) and centered just east of Halema‘uma‘u Crater, beneath the area of maximum surface displacement (for detailed results see Supplementary materials). The overlapping results suggest that the same source is responsible for both subsidence and uplift of Kīlauea’s summit caldera during the time interval spanned by our gravity surveys. Our results are also very similar to those of an independent study that modeled the summit deflation measured by InSAR (CosmoSkyMed satellite) and GPS during the March 2011 ERZ eruption [Lundgren et al., 2013].

The geometries of the SWRZ and HMM sources were then fixed based on the best fits obtained above. Using these parameters, the volume changes during each time period were estimated through non-linear inversion of the InSAR data (Table 5.4; see Supplementary material for further details). Minor volume fluctuations are inferred for the HMM source during periods 1 and 2, while a cumulative volume increase of 1.71 × 10^6 m^3 is obtained from the inversion of deformation data spanning periods 3 and 4.
Figure 5.7 – Deformation modeling results for the Halema‘uma‘u source (HMM). (a) Normalized posterior probability distributions for the depth parameter obtained using the Gibbs Sampling algorithm (25,000 samples). In red are results for a period of inflation, and in blue for a period of deflation. Dotted lines represent the depth of the best fitting models. (b) Two-dimensional scatterplot of the latitude and longitude positions for the source of inflation (red) and deflation (blue). (c) Profiles showing the fit of each model (dotted lines) to the data (solid lines). Profile location is given by the black dashed line in panel (b).

Table 5.4. Volume changes during each time period

<table>
<thead>
<tr>
<th>Period</th>
<th>HMM source $\Delta V$ in $x \times 10^6$ cubic meters</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.16 [0.01 – 0.37]</td>
</tr>
<tr>
<td>2</td>
<td>-0.33 [-0.26 – -0.40]</td>
</tr>
<tr>
<td>3</td>
<td>1.06 [1.00 – 1.11]</td>
</tr>
<tr>
<td>4</td>
<td>0.65 [0.54 – 0.77]</td>
</tr>
<tr>
<td>Total</td>
<td>1.54 [1.29 – 1.85]</td>
</tr>
<tr>
<td>3+4</td>
<td>1.71 [1.54 1.88]</td>
</tr>
</tbody>
</table>

5.7.2 Modeling residual gravity changes

Residual gravity changes can be inverted to constrain their source location and associated mass variation. The gravitational attraction of a spherical body of finite size and mass, where mass $m = \rho \Delta V$ with $\rho$ as the density of the mass and $\Delta V$ the change in volume, is identical to that of a point source with the same mass $m$:

$$\Delta g_r = G \frac{m}{(r^2 + d^2)^{3/2}}$$
where $G = 6.67 \times 10^{-11}$ N m$^2$ kg$^{-2}$ is the universal gravity constant, $d$ is depth of the point source, and $r$ the radial distance from the surface projection of the source [Battaglia and Hill, 2009]. The best-fit to the measured gravity changes was calculated using MATLAB’s built-in `fminsearch` function (Nelder-Mead method), which performs an unconstrained nonlinear minimization of the sum of squared residuals (SSR) with respect to the various parameters. The nonlinear minimization requires starting estimates for the fit parameters. To avoid convergence on local minima, we tested 128 random realizations of the initial parameters [Bergstra and Bengio, 2012] and used the solution providing the smallest value for the SSR. The goodness of the fit is provided by the value of the coefficient of determination $R^2$: if $R^2$=1, the model is able to explain all variations in the observed data; if $R^2$=0, the model is not able to explain any of the observed data. 95% confidence limits were computed by a bootstrap percentile method [Efron and Tibshirani, 1986].

Gravity changes during periods 1 and 2 are too small compared to the background noise to estimate any significant mass change. Larger variations, on the other hand, characterize periods 3 and 4. Given the similarities in the spatial distribution of the residual gravity changes during these two periods (Figure 5.6b and 5.6c), we inverted for the total variations measured between March 2011 and November 2012 (period 3 + period 4). This approach allows us to maximize the signal-to-noise ratio as well. The best fitting source is a point source centered just east of Halemaʻumaʻu Crater (Figure 5.8)—approximately the same location as the HMM deformation source. The inferred depth of the gravity source, very similar to that of the HMM source of deformation, is 1.59 km (1.35−1.96, 95% confidence interval), and the mass addition is modeled as $0.56 \times 10^{11}$ kg
(0.45 – 0.71 x 10^{11} \text{ kg}, 95\% \text{ confidence interval}). Assuming a density for basaltic magma of \( \rho = 2500 \text{ kg m}^{-3} \), the associated volume change \( \Delta V \) would be 22.4 x 10^6 m^3 (18.0 – 28.4 x 10^6 m^3, 95\% \text{ confidence interval}), which is an order-of-magnitude larger than the volume change inferred from the deformation data during the same time interval (1.71 x 10^6 m^3).

**Figure 5.8** – Residual gravity change modeling results. The solid black line gives best fit to residual gravity change recorded between March 2011 and November 2012 (gray errorbars). Black squares indicate residual gravity values adjusted for the lava-lake effect. The best-fitting source is centered at 1.59-km depth beneath the NE margin of Halema‘uma‘u Crater (white star in the inset) and has a mass addition of 0.56 x 10^{11} \text{ kg}. The \( R^2 \) value is 0.89.

**5.8 Discussion**

Significant residual gravity changes (10-10^2 \mu\text{Gal}) were measured between December 2009 and November 2012 at the summit of Kīlauea. When adjusted for the effect of variations in lava-lake level, these changes indicate a significant mass increase in the shallow magma reservoir beneath the northeast margin of Halema‘uma‘u Crater. InSAR data spanning the same time interval indicate moderate deformation (both uplift and subsidence, <0.20 m) accompanying the gravity changes, also centered northeast of the Halema‘uma‘u Crater. This location is similar to the loci of gravity increase measured
between 1975 and 2008 that was interpreted as the result of magma accumulation in a void space at ~1 km depth [Johnson et al., 2010] or within a volume created by the rifting of the summit [Zurek and Williams-Jones, 2013]. Also, as in 1975–2008, during 2009-12 there were no significant gravity changes south of the caldera, implying no resolvable mass change in the south caldera reservoir. Surface deformation in this area is likewise minor. Subsidence measured in the upper part of the SWRZ during 2009–12 is not associated with any significant change in residual gravity (although we must keep in mind that this could be a reflection of the poor distribution of gravity stations in this area), suggesting that the process causing the deformation is not related to subsurface mass flow.

5.8.1 Rapid mass increase

Although residual gravity changes during 2009–12 were centered in the same area as the positive anomaly measured during previous decades (1975–2008), the rate at which the more recent changes occurred is much faster. A positive gravity change of 132 µGal (after adjustment for the lava-lake effect) was measured at station 205YY between March 2011 and November 2012, corresponding to a rate of increase of ~83 µGal/yr. The rate of gravity increase at the same station during 1975–2008 was just ~11 µGal/yr. Furthermore, the shorter time interval between surveys of our measurements highlights that gravity changes can fluctuate significantly over time. Gravity increased between December 2009 and June 2010 (period 1) and between March 2011 and November 2012 (periods 3 and 4), but decreased slightly between June 2010 and March 2011 (the time period that includes the March 2011 ERZ Kamoamoa fissure eruption). Although the net
change over the 3-year interval is positive, gravity changes were not linear in time and reflect variations due to volcanic activity.

5.8.2 Accounting for the “missing” volume

Residual gravity changes measured between 2009 and 2012 at stations located inside Kīlauea Caldera were accompanied by deformation of the same area—positive changes were associated with uplift, while negative gravity changes occurred during subsidence. Our modeling results indicate, however, that even though the source of deformation and gravity change coincide—both centered at ~1.5 km beneath the northeast margin of Halema‘uma‘u Crater—the volume change inferred from the deformation is only 8-9\% of a conservative estimate of the volume changes inferred from residual gravity variations between March 2011 and November 2012.

Our new results raise again the question of how mass can accumulate beneath the surface of Kīlauea without generating the expected uplift of the ground, as originally considered by Johnson et al. [2010]. Possible mechanisms include: (i) formation of cumulates through partial replacement of magma in the reservoir by olivine; (ii) upward migration of the magma reservoir by assimilation of host rock; (iii) filling of void space by magma, whether the space is already present (e.g., drainage of the magma reservoir during volcanic/tectonic events, see Johnson et al. [2010]) or forms progressively (e.g., secular extension of the summit due to rifting and “stretching” of the reservoir [Zurek and Williams-Jones, 2013]); (iv) bulk compression of gas-rich magma [Johnson, 1992; Rivalta and Segall, 2008] and consequent increase in density; and (v) progressive
densification of the reservoir through the replacement of gas-rich magma by denser, degassed magma.

The first two mechanisms were effectively ruled out by Johnson et al. [2010]. Replacement of magma (2500 kg/m³) by denser olivine cumulates (3300 kg/m³) is unlikely because accumulation of olivine would gradually fill the shallow reservoir, causing magma to move towards the surface and resulting in significant deformation. In addition, about $17.0 \times 10^6$ m³ of olivine should have accumulated (given the mass increase of $0.56 \times 10^{11}$ kg), but there is no evidence of cumulates in rocks erupted historically from Kilauea’s summit (including the current eruption) or in lava flows and pyroclastic deposits exposed in the walls of the caldera, which argues against the presence of shallow cumulate bodies [e.g., Casadevall and Dzurisin, 1987]. Upward stoping of magma is also unlikely without some sort of major deformation signal. Furthermore, both processes are even less likely during 2011–12 than in previous decades because the inferred mass increase indicated by our surveys occurred in less than two years.

Filling of void space by magma is the preferred mechanism of previous studies, although different processes were invoked for the formation of the space [Johnson et al., 2010; Zurek and Williams-Jones, 2013]. Johnson et al. [2010] proposed that, between 1975 and 2008, magma filled space that was created by drainage of the summit reservoir during the Mw 7.5 Kalapana earthquake in November 1975. Continuation of this process through 2012 implies a large volume of void space beneath Kilauea Caldera or persistent formation and filling of void space at shallow levels. Orr et al. [2013] demonstrated that the amount of lithic material erupted during the formation of the presently active summit
eruptive vent in 2008 was <0.01% of the $4 \times 10^6$ m$^3$ crater volume. Such a large “missing” volume is consistent with the presence of void space beneath the summit.

Zurek and Williams-Jones [2013] proposed that rifting of the summit may also allow magma to accumulate without an increase in reservoir pressure and consequent surface uplift. In their model, an increase of 3 cm yr$^{-1}$ (which is the rate of summit extension due to south flank motion [e.g., Delaney et al., 1998]) in the radius of a 1-km$^3$ sphere at a depth of 1 km would result in $1.45 \times 10^5$ m$^3$ yr$^{-1}$ of magma to be stored without producing uplift of the surface. Such a process could explain about 16% of the measured gravity increase during 1975–2008 (the calculation in Zurek and Williams-Jones [2013] contains an error that led them to a value of 59% [J. Zurek, written comm., 2013], which we correct here), demonstrating that it may be a contributor to the mass increase without causing significant vertical deformation. If we use the same approach, assuming a 1-km$^3$-volume reservoir at a depth of 1.5 km, magma density of 2500 kg m$^{-3}$, and annual volume increase of $1.45 \times 10^5$ m$^3$, the rate of increase in residual gravity would be 2.4 µGal yr$^{-1}$, resulting in 3.8 µGal of total increase between March 2011 and November 2012—only $\sim$3% of the measured gravity increase.

Mass increase in the absence of significant inflation may also occur due to bulk compression of gas present within the reservoir as magma accumulates [Johnson, 1992]. Studies of geodetic measurements spanning a dike intrusion into Kīlauea’s ERZ in 1997 [Owen et al., 2000; Rivalta and Segall, 2008] have shown that the volume of the intrusion was $\sim$4 times larger than that of the deflating source reservoirs—a discrepancy that can be partially explained by compressibility of magma [Rivalta and Segall, 2008]. The compressibility of gas-rich magma inside the HMM reservoir during periods of surface
uplift could therefore account for perhaps 32% of the mass accumulation (given that volume change alone inferred from deformation models can account for 8%) without producing the uplift that would be expected from models of pressure increase in a reservoir that is filled with incompressible magma. This process should occur during all periods of magma accumulation and might explain the moderate gravity increase between December 2009 and June 2010, which was associated with minimal uplift, and also a portion of the large positive change measured after March 2011.

Finally, ongoing summit eruptive activity may provide a means of allowing mass accumulation without significant surface uplift. The opening of the summit eruptive vent in March 2008 was associated with an order-of-magnitude increase in gas emissions from the summit of Kīlauea [Elias and Sutton, 2012]. The heightened emissions are the result of a convecting lava lake in the summit vent, which allows magma to rise to the surface, outgas, and then sink to deeper levels [Carey et al., 2013]. Outgassed magma could therefore progressively replace the mass of gas-rich magma that is stored within the HMM reservoir, assuming the accumulation of outgassed magma outpaces the influx of gas-rich magma from below. If the HMM reservoir is approximated by a 1-km$^3$ sphere ($10^9$ m$^3$), an increase in density of 200 kg m$^{-3}$ for the entire reservoir would produce a mass increase of $2 \times 10^{11}$ kg—over 3 times the mass change needed to produce the residual gravity signal measured between March 2011 and November 2012 ($\sim 0.6 \times 10^{11}$ kg). Densification of $\sim 30\%$ of the reservoir by 200 kg m$^{-3}$, or of the entire reservoir by $\sim 65$ kg m$^{-3}$, would therefore explain the measured residual gravity changes in the absence of surface uplift.
5.8.3 Gravity changes and eruptive activity

The rates of the residual gravity changes over time provide further insights into the mechanism of mass variation beneath Kīlauea. The first and last time periods (December 2009 – June 2010 and June 2012 – November 2012) show the highest rates of gravity increase (68 µGal yr\(^{-1}\) and 144 µGal yr\(^{-1}\), respectively) but are not associated with the pre-, co-, or post-eruptive phases of the March 2011 ERZ Kamoamoa fissure eruption. A slight gravity decrease was recorded during the second period (June 2010 – March 2011), which spans the Kamoamoa eruption, implying magma withdrawal from the HMM reservoir (also indicated by deflation of the reservoir). Finally, a gravity increase at lower rate (51 µGal yr\(^{-1}\)) characterizes the third period (March 2011 – June 2012), which covers the ~year that followed the Kamoamoa eruption. From this pattern, we conclude that the Kamoamoa eruption drew magma from the HMM reservoir, effectively “flushing” the denser, degassed magma that had been accumulating since the start of the summit eruption in 2008 and that was causing the rapid gravity increase measured during the first period. The co-eruptive emptying of the reservoir was followed by rapid refilling with fresh, gas-rich magma from depth—a process that is also suggested by rapid post-eruptive uplift. This resulted in a gravity increase, but at a lower rate than the first and fourth time periods (because of the lower density of the accumulating magma). During the last time period, which followed refilling of the HMM reservoir, the process of persistent degassing, convection of the lava lake, and densification of the reservoir magma resumed, accelerating the rate of gravity increase.
5.8.4 A combination of different processes

With respect to the gravity increase measured between March 2011 and November 2012, it is likely that a combination of several of the processes discussed above is the ultimate source of the measured mass accumulation. During that time period, only ~8% of the mass increase inferred from the gravity measurements can be explained by the increase in reservoir volume inferred from modeling deformation data. Accounting for magma compressibility, which can increase the volume of stored magma by up to four times without causing additional deformation [Rivalta and Segall, 2008], still only explains up to ~32% of the mass accumulation. Filling of void space with magma can easily explain the gravity increase and was the mechanism used to explain the mass accumulation measured between 1975 and 2008. This process would require the presence or continued formation of void space beneath the surface—a condition that may exist given the discrepancy in the volume of ejecta versus that of the summit vent [Orr et al., 2013]. Growth of the reservoir by rifting of the summit, however, does not significantly contribute to the mass accumulation, since the rate of gravity increase during the 20-month period far exceeds that expected from rifting [Zurek and Williams-Jones, 2013], although that process may be a more significant factor over longer time periods. Finally, replacement of gas-rich magma by denser, outgassed magma that lost its volatiles when it reached the surface of the lava lake via convection could explain a large portion—and potentially all—of the measured gravity increase. We suspect that all three processes play a role in gravity change at Kīlauea. Continued frequent (every 6–12 months) surveys of the summit gravity network in combination with deformation studies and in the context
of the volcano’s eruptive and intrusive activity should help to further distinguish between the relative importance of the various mechanisms.

5.9 Conclusions

Microgravity data collected at the summit of Kīlauea Volcano between December 2009 and November 2012 reveal significant residual gravity changes centered near Halema‘uma‘u Crater—a location similar to the center of the positive residual gravity anomaly measured between 1975 and 2008. Gravity changes over the 3-year period indicate a net mass accumulation at a depth of ~1.5 km, which coincides with the source of surface deformation inferred from InSAR data spanning the same time interval. This source has been previously identified as a small magma storage zone beneath the northeast margin of Halema‘uma‘u Crater. The rate of gravity increase, which was much higher during 2009–12 than between 1975 and 2008, varies through time and seems to be directly correlated with the volcanic activity occurring at both the summit and the ERZ of the volcano.

InSAR data show that uplift occurred during periods of gravity increase and that subsidence characterizes a time period of slight gravity decrease—which includes the 5-9 March 2011 ERZ eruption. Despite this connection between gravity change and deformation, the volume change inferred from the modeling of the InSAR data between March 2011 and November 2012—the period of greatest gravity change—can only account for ~8% of the gravity increase (assuming that the volume increase is due to magma with a density typical of basalt).
Given the discrepancy between gravity change and surface deformation, mechanisms beyond simple filling of a magma reservoir must have occurred at Kīlauea. The replacement of gas-rich magma within the Halemaʻumaʻu reservoir by denser, outgassed magma that had convected up to the surface within the summit eruptive vent and lost its volatiles can account for the entire residual gravity increase measured during March 2011–December 2012. Other mechanisms, such as the compressibility of magma and the filling of void space by magma, may also contribute to the mass addition beneath the summit of Kīlauea without producing significant deformation.

5.10 Acknowledgements

Part of this research was supported by the National Aeronautics and Space Administration (NASA, NESSF11 graduate assistantship for Marco Bagnardi). The purchase of the gravimeters by the Hawaiian Volcano Observatory was made possible by the American Reinvestment and Recovery Act. The InSAR data are courtesy of the Hawaiʻi Supersite (CSK from ASI and TSX from DLR). We would like to thank M. Patrick and T. Orr for providing data on lava-lake geometry and levels. We also thank USGS-HVO employees and volunteers that helped with the gravity surveys, including K. Anderson, S. Brantley, I. Johanson, J. Johnson, A. Learner, S. Mordensky, A. Pitty, M. Sako, and S. Wilkinson.
Chapter 6: Magma supply to the western Galápagos volcanoes inferred from InSAR and GPS time series (1992–2011).

6.1 Summary

We use space–geodetic measurements of the surface displacement at six volcanoes of the western Galápagos Islands (Ecuador) to estimate volumes and rates of magma supply to the archipelago during 1992–2011. Surface deformation is measured using interferometric synthetic aperture radar (InSAR) and global positioning system (GPS) data. The storage system of each deforming volcano is characterized through the non–linear inversion of the InSAR data, and vertical displacement time series are used to track volume changes within the magma reservoirs through time. We calculate that a cumulative volume of $0.350\pm0.100 \text{ km}^3$ was supplied at an average rate of $\approx0.02\pm0.005 \text{ km}^3 \text{ yr}^{-1}$ during 1992–2011 to the Galápagos volcanoes and that more than half of it was directed towards Sierra Negra, the most actively deforming volcano in the archipelago. The rate at which magma is supplied, however, varied through time in correlation with eruptive and intrusive activity at the volcanoes. Our results provide the first archipelago–wide magma supply rate estimate in the Galápagos on a time–scale of interest to humans.

6.2 Overview

In the western Galápagos Islands, Ecuador, space–geodetic measurements have revealed that the surface of at least six volcanoes, Wolf, Darwin, Alcedo, Sierra Negra, Cerro Azul and Fernandina, uplifted between 1992 and 1999 [Amelung et al., 2000]. This deformation can be interpreted as the evidence of magma entering the storage system of the volcanoes [e.g., Yun et al., 2006; Hooper et al., 2007] and implies the occurrence of a
competition for magma coming from the common hotspot mantle–source. The amount of magma supplied to each volcano from a deep source region and the rate at which it enters crustal reservoirs are perhaps the most important factors controlling the eruptive and intrusive activity at a volcano [Dvorak and Dzurisin, 1993]. The total amount of magma supplied to the entire archipelago can instead provide information on deeper processes such as the temperature and the productivity of the mantle hot spot [e.g., Ito et al., 1997].

Magma supply volumes and rates can be inferred in different ways, but always indirectly. In Hawai‘i, for example, measurements of the effusion rate during long–lasting eruptions, such as those at Kīlauea, have been used to quantify the amount of magma entering the plumbing system of the volcano [Swanson et al., 1972]. Historical eruptions in the Galápagos, however, have been of shorter duration compared to those at Kīlauea, so using effusion rates to infer magma supply to the volcanoes is not as efficient.

Detailed measurements of the surface deformation can thus provide the best means to track volume variations within the storage system of the Galápagos volcanoes and quantify the amount of magma supplied to the archipelago. Furthermore, the high temporal frequency of space–geodetic measurements can provide insights on short–term variations in the rate of magma supply [e.g., Poland et al., 2012].

Currently available estimates of the magma supply rate to the Galápagos archipelago are limited to studies of the isostatic crustal thickness [Ito et al., 1997] that propose values ranging between 0.03 and 0.3 km$^3$ yr$^{-1}$ over the past 8 Ma (average of 0.130 km$^3$ yr$^{-1}$). This long–term average rate of magma supply is similar to that inferred for Kīlauea Volcano in Hawai‘i over the past decades—approximately 0.1–0.2 km$^3$ yr$^{-1}$ [Swanson et al., 1972; Dzurisin et al., 1984; Dvorak and Dzurisin, 1993; Poland et al., 2012].
In this study we present the analysis of surface deformation measurements from Interferometric Synthetic Aperture Radar (InSAR) and Global Positioning System (GPS) data acquired in the western Galápagos Islands of Isabela and Fernandina between 1992 and 2011 (Figure 6.1). After characterizing the geometry and location of areas of magma storage beneath the summit of each volcano through the non-linear inversion of the InSAR data, we estimated volume changes within the storage systems throughout the nineteen–year–long time interval. By summing the contribution of each volcano in terms
of stored magma, we determined the cumulative volume and rate of supply to the entire archipelago, and studied variations in this rate through time. Our results provide the first archipelago–wide magma supply rate estimate on a time–scale of interest to humans. We infer that the average volume of magma injected each year into the crust beneath the Galápagos volcanoes is lower than that inferred from the isostatic crustal thickness. The high temporal frequency of the deformation measurements also shows significant variations in the rate of supply and in the amount of magma distributed between the six actively deforming volcanoes.

6.3 Data and methods

6.3.1 InSAR data and time series analysis

InSAR deformation measurements used in this study were obtained from synthetic aperture radar (SAR) images acquired by five satellites, the European Space Agency’s ERS–1, ERS–2 and Envisat satellites, the Canadian Space Agency’s Radarsat–1 satellite, and the Japanese Aerospace Exploration Agency’s ALOS satellite. Data from ALOS was acquired by an L–band sensor (wavelength 23.6 cm) while all the other satellites mounted a C–band sensor (wavelength 5.6 cm). The entire dataset is composed of 542 SAR images from 17 satellite–tracks along both ascending and descending orbital passes, and in different beam modes (incidence angles varying between 23° and 47°). The dataset spans a total of 18.6 years, from 15 June 1992 to 12 February 2011 (see Table 6.1 for further details). The SAR data was processed using the JPL/Caltech ROI_PAC SAR Software [Rosen et al., 2004] to form the Radarsat–1 interferograms and the GMTSAR InSAR processing system [Sandwell et al., 2011] for the other satellites. We used precise DORIS orbits information provided by the European Space Agency (ESA), and removed
the topographic contribution to the interferometric phase using a 90–m–resolution digital elevation model (DEM) generated by the NASA Shuttle Radar Topography Mission (SRTM) [Farr et al., 2007]. The interferograms were then phase–unwrapped using the SNAPHU algorithm [Chen and Zebker, 2001].

For each satellite track, tens to hundreds of interferograms (depending on the number of SAR acquisitions available for a specific track) were used to generate time series of the surface displacement using the small–baseline subset (SBAS) method [Berardino et al., 2002; Lanari et al., 2004; Fattahi and Amelung, 2013]. InSAR time series provide measurements of the surface displacement along the radar line–of–sight (LOS), but measurements of the vertical and horizontal components become more useful when trying to characterize magmatic sources of deformation. Converting LOS displacements from different viewing geometries into a common reference frame (e.g., into vertical displacements) also allows us to align them into a continuous time series, spanning the entire study period (Figure 6.2). We therefore combined the InSAR time series results from ascending and descending orbital passes to calculate the two components of motion (east–west and vertical) following the method of Wright et al. [2004] (for a detailed description of the approach also see Baker and Amelung [2012] and Baker [2012]). When contemporary data from both orbital passes is not available (i.e., Radarsat–1 data before 2000 is available from the ascending orbit only and ERS–1/2 data was rarely acquired before 1999) estimates of the vertical component of the surface displacement were obtained from the direct conversion of the LOS displacement. This approach assumes that the deformation of the surface is mostly vertical, which could be the case for displacements measured at the center of the summit calderas, and divides the LOS
displacement by the vertical component of the satellite range look vector [e.g., Baker, 2012].

<table>
<thead>
<tr>
<th>Volcanoes covered</th>
<th>Satellite</th>
<th>Pass</th>
<th>Track</th>
<th>Beam (Inc. angle)</th>
<th>Date Span</th>
<th>SAR images</th>
</tr>
</thead>
<tbody>
<tr>
<td>D, A, SN, CA</td>
<td>ERS–1/2</td>
<td>Desc.</td>
<td>140</td>
<td>S2 (~23°)</td>
<td>1992/06/15 – 2007/10/04</td>
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<tr>
<td>W, D, F</td>
<td>ERS–1/2</td>
<td>Desc.</td>
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<td>S2 (~23°)</td>
<td>1992/09/12 – 2009/10/27</td>
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<td>Asc.</td>
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<td>S5 (~39°)</td>
<td>1998/09/13 – 2007/03/06</td>
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<tr>
<td>W, D, A, SN, CA, F</td>
<td>Radarsat–1</td>
<td>Desc.</td>
<td>29</td>
<td>S7 (~47°)</td>
<td>2000/09/22 – 2006/12/20</td>
<td>48</td>
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<tr>
<td>W, D, A, SN, CA</td>
<td>Envisat</td>
<td>Desc.</td>
<td>140</td>
<td>S2 (~23°)</td>
<td>2003/01/23 – 2010/06/10</td>
<td>51</td>
</tr>
<tr>
<td>W, D, F</td>
<td>Envisat</td>
<td>Desc.</td>
<td>412</td>
<td>S2 (~23°)</td>
<td>2003/02/11 – 2010/09/07</td>
<td>60</td>
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<tr>
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<td>2003/07/12 – 2010/09/18</td>
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<tr>
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<td>2005/06/10 – 2009/12/11</td>
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<td>Desc.</td>
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<td>S4 (~34°)</td>
<td>2005/06/13 – 2008/11/24</td>
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<tr>
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<td>W, D, A, SN</td>
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<td>S5 (~38°)</td>
<td>2005/07/03 – 2009/05/03</td>
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<td>W, D, A, SN, CA, F</td>
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<td>Asc.</td>
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<td>S6 (~41°)</td>
<td>2006/10/20 – 2010/08/20</td>
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<tr>
<td>A, SN</td>
<td>Alos</td>
<td>Desc.</td>
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<td>7 (~34°)</td>
<td>2007/01/16 – 2010/09/11</td>
<td>18</td>
</tr>
<tr>
<td>W, D, A, SN</td>
<td>Alos</td>
<td>Asc.</td>
<td>133</td>
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<td>2007/03/02 – 2011/01/26</td>
<td>18</td>
</tr>
<tr>
<td>CA, F</td>
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<td>Asc.</td>
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<td>7 (~34°)</td>
<td>2007/02/01 – 2010/11/12</td>
<td>17</td>
</tr>
<tr>
<td>W, D, CA, F</td>
<td>Alos</td>
<td>Desc.</td>
<td>475</td>
<td>7 (~34°)</td>
<td>2007/08/05 – 2010/08/13</td>
<td>9</td>
</tr>
</tbody>
</table>

**TOTAL** 535

W = Wolf; D = Darwin; A = Alcedo; SN = Sierra Negra; CA = Cerro Azul; F = Fernandina.

6.3.2 GPS data

Continuous GPS data from two stations (GV02 and GV04) located within Sierra Negra’s summit caldera (Figure 6.1c) were used to better characterize the temporal evolution of the vertical displacement between 2002 and 2011. Sierra Negra is the only volcano in the Galápagos Island for which GPS data is available. The GPS dataset was obtained through the NAVSTAR Consortium (UNAVCO) and processed using the JPL–
GIPSY–OASIS software [Stephen et al., 1996]. The results provided daily solutions of the three components of displacement (north–south, east–west and vertical) in the terrestrial reference frame (ITRF2005). Vertical displacements measured at station GV02, located approximately at the center of Sierra Negra’s caldera, were aligned with vertical displacements measured by InSAR data at the same location. Due to instrument failure at station GV02 in 2005, data from the closest station GV04 was used as a proxy for the vertical displacement occurred until GV02 was re-established in 2006.

6.3.3 Source modeling

In order to constrain the volume of magma that was supplied to each volcano during the study period, the areas where magma is stored must be first characterized in terms of their geometry and location. Models of the storage system at the six volcanoes showing surface deformation were inferred through the non–linear inversion of LOS displacements from the time series analysis, measured during specific time periods characterized by maximum signal–to–noise ratio (see Table 6.2). Best–fitting models were obtained using analytical solutions for magmatic sources of deformation embedded in an elastic isotropic half–space (Poisson’s ratio ν = 0.25), specifically that for a finite spherical magma body [McTigue, 1987], for a finite spheroidal cavity [Yang et al., 1988], and for a rectangular dislocation source with uniform opening [Okada, 1985] to simulate flat–topped/sill–like reservoirs.

Since each original dataset consists of $10^6$ data points, we performed spatial averaging using the Quadtree algorithm [Jónsson et al., 2002] to generate the data vectors used in the inversions. The optimal solutions and their probability density
distributions were then estimated using the Monte–Carlo–based Gibbs Sampling (GS) algorithm [Brooks and Frazer, 2005]. The GS algorithm inverts for a large number of models (in this case 25,000) and the results provide a proxy for the probability distribution of each parameter (i.e., Figure 3.9 in Chapter 3). The quality of the fit of the predicted deformation to the measured displacement was assessed using the normalized root–mean–square error (RMSE) between observed and modeled InSAR LOS displacements (see Section 2.3.1).

6.3.4. Volume change time series

The amount of magma supplied to a volcano, once the reservoir location and geometry is determined, can be estimated by calculating the volume change necessary to produce the displacement measured at the surface. This approach assumes that magma filling the reservoirs is incompressible. To generate time series of the cumulative volume at each volcano we used the optimal source locations and geometries and inverted for the volume change only, and linearly propagated the uncertainty of the source parameters to the volume estimates.

Volume changes during periods of rapid subsidence associated with magmatic events, such as eruptions and intrusions, were removed from the cumulative volume time series and the first post-event data point was aligned with the last pre-event one. In fact, the scope of the study is to estimate the volume of magma supplied to the volcanoes from their source deep in the mantle. These events represent, instead, episodes of withdrawal of magma already present in the reservoir and its emplacement within the volcanic
edifice or at the surface. If and how much magma is supplied from depth during these events, however, is not known and represents a limitation of the approach.

Finally, we calculated monthly average volume changes at times when these values are available at all the volcanoes and, by summing each contribution, we estimated the relative archipelago–wide supply volume. This approach provides a mean to study the total amount of magma supplied to the western Galápagos volcanoes, and the rates at which it is supplied, with an unprecedented temporal resolution.

6.4. 1992-2011 deformation

Surface displacements measured between 1992 and 2011 at each of the six actively deforming volcanoes forming the islands of Isabela and Fernandina are described here and presented in Figure 6.2. Deformation between 1992 and 2000 is poorly constrained in terms of its temporal evolution because of the limited number of SAR acquisitions during this time interval. In particular, it is not possible to establish the timing of episodes of deformation that started between 1992 and 1997–98 since no SAR images were acquired during this time period. After 2000, however, the temporal resolution was largely improved by a more frequent acquisition of SAR data that reached its maximum between 2005 and 2009, when more than seven images per month (Envisat + ALOS) were acquired in the western Galápagos Islands.
Figure 6.2 (continues in the next page)
Figure 6.2 (continues from previous page) – Vertical displacement time series for the Galápagos volcanoes measured by InSAR and GPS (Sierra Negra only). Time-series are for points located at the center of the summit calderas (black squares in Figure 6.1). Radarsat–1 vertical displacement time series (yellow triangles) for Wolf, Darwin, Alcedo and Fernandina are LOS converted, for Sierra Negra and Cerro Azul are computed by combining ascending and descending LOS displacements. Gray bars mark the occurrence of eruptive, intrusive or faulting events.
Wolf, the northernmost volcano on Isabela Island, shows the steadiest behavior among the six active volcanoes (Figure 6.2a). Between 1992 and the end of 1997 the summit caldera uplifted up to ~0.10 m but when the inflation actually started cannot be determined. Steady linear inflation continued between 1998 and early 2008 when the inflation rate progressively decreased from ~0.045 m yr\(^{-1}\) to zero by the end of 2009, producing a total vertical displacement at the center of the caldera of ~0.59 m since 1992. No significant deformation is measured between 2009 and 2011. Displacements are limited to an area bounded by the summit caldera rim and maximum values are measured at its center.

At Darwin volcano, uplift is recorded between 1992 and 1998 (Figure 6.2b). Inflation continued at a mean rate of ~0.025 m yr\(^{-1}\) and produced a total uplift of ~0.30 m by early 2003 when it switched to slight deflation, which continued until the end of 2010 and produced up to ~0.06 m of subsidence (< –0.01 m yr\(^{-1}\)). The recorded deformation is centered on the summit caldera but largely extends to the flanks of the volcano forming a wide circular area of displacement.

At Alcedo most of the flanks of the volcano are densely covered by vegetation, which causes decorrelation of the InSAR data when this is acquired by C–band radar sensors. Data from the ALOS satellite (L–band) instead, provides good interferometric correlation for most of the volcano but only spans the last four years of the study period (2007–2011). Good correlation, however, characterizes InSAR data within the summit caldera and allows us to track its deformation for the entire time period (Figure 6.2c). Up to ~0.85 m of uplift are recorded between 1992 and 1999, a time interval during which a small phreatic eruption may have occurred (in late 1993, but this event is poorly
documented [\textit{Green}, 1994]). The uplift continued until 2001, when the summit caldera started to subside. Subsidence at a mean rate of \( \sim -0.022 \text{ m yr}^{-1} \) continued until early 2007 when it switched again to uplift. At a fairly constant rate of \( \sim -0.077 \text{ m yr}^{-1} \) the summit caldera uplifted for \( \sim 0.23 \text{ m} \) until early 2010 when rapid subsidence occurred for a few weeks and switched back to uplift by June 2010. Rapid uplift at a rate of \( \sim 0.12 \text{ m yr}^{-1} \) characterized the last seven months of the study period, until January 2011.

Sierra Negra is the most actively deforming volcano in the archipelago (Figure 6.2d). Between 1992 and 1999 up to \( \sim 2.70 \text{ m} \) of uplift were recorded at the center of the summit caldera. In 2000 inflation switched to deflation at an approximately steady rate of \( \sim -0.08 \text{ m yr}^{-1} \) until April 2003 when deformation of the caldera floor changed again from deflation to inflation. Uplift continued at an exponentially increasing rate and reached \( \sim 1.05 \text{ m} \) of displacement in April 2005. On 16 April 2005 an episode of inelastic trapdoor faulting associated to a \( m_b 4.6 \) earthquake caused a very rapid uplift of the southern portion of the caldera (up to \( \sim 0.84 \text{ m} \)) [\textit{Chadwick et al.}, 2006; \textit{Jonsson}, 2009]. Inflation continued at high rates even after this event and approached a maximum of \( \sim 1 \text{ cm/day} \) by October 2005 (reaching \( \sim 2.20 \text{ m} \) of uplift since April 2003), when a fissure eruption started at the NE rim of the summit caldera. The eruption caused very rapid subsidence of the caldera floor (not precisely resolved but likely \( >5 \text{ m} \)) [\textit{Yun et al.}, 2006]). A few days after the eruption ended, inflation restarted at high rates and continued until the end of our study period in 2011. A maximum uplift of \( \sim 2.83 \text{ m} \) is recorded between the end of the 2005 eruption and early 2011, and a cumulative uplift of \( \sim 6.5 \text{ m} \) occurred between 1992 and 2011 if the co-eruptive subsidence is removed.
At Cerro Azul, surface deformation measured during periods of uplift or subsidence extends to almost the entire subaerial portion of the volcanic edifice and is centered on its summit caldera. Between 1992 and 1998 the volcano uplifted to a maximum of ~0.13 m at the summit (Figure 6.2e) before erupting from radial fissures on its eastern flank in September–October 1998. Co–eruptive subsidence is rapidly followed by re–inflation of the volcanic edifice, which continued until mid–2001 following a gradual decay. No significant deformation is recorded between 2001 and May 2008, when a second eruption occurred in the same region of the 1998 event and caused deflation throughout the volcanic edifice. Differently from the previous eruptive event, no rapid re–inflation occurred after the eruption ended and only ~0.05 m of uplift is recorded by the end of 2010.

Finally, a detailed analysis of the surface deformation recorded at Fernandina is presented in Chapter 3 of this manuscript and by Bagnardi and Amelung [2012]. The volcano is characterized by almost continuous uplift for the entire study period, briefly interrupted by rapid co–eruptive (January 1995, May 2005 and April 2009) and co–intrusive (December 2006 and August 2007) subsidence (Figure 6.2f). If these periods of subsidence are removed from the deformation time series, the summit caldera cumulatively uplifted up to ~2.27 m between 1992 and 2011, the second largest amount of deformation after that measured at Sierra Negra. Rates at which uplift occurs are however variable, with a maximum value of ~0.36 m yr$^{-1}$ recorded during the six months after the August 2007 subvolcanic intrusion.


6.5. Sources of deformation

Here follows a brief description of the most significant results from the source modeling (e.g., geometry and depth, Table 6.2). Further information on the remaining source parameters is available in Appendix S4.

At Wolf, the modeling of intracaldera inflation measured during a three–year time interval (12/2006–12/2009) provides best fit for a rectangular source with uniform opening [Okada, 1985] centered beneath the summit caldera at a depth of 1.4 km (95% confidence interval: 0.9–1.9 km). This source is elongated in the NW–SE direction as the caldera overlying it. Reservoir geometry and depth are different from those inferred by Amelung et al. [2000] who modeled it as a point source [Mogi, 1958] at ~2 km depth.

Surface deformation measured at Darwin during a period of inflation between 1992 and the end of 2000 offers the best opportunity to infer the source of uplift. Best fit is obtained using a spherical source [McTigue, 1987] centered beneath the summit caldera at a depth of 3.1 km (95% confidence interval: 2.1–4.0 km), similarly to what inferred by previous studies [Amelung et al., 2000; Manconi et al., 2007].

The spatial distribution of the surface deformation at Alcedo is more complex than at other Galápagos volcanoes and is likely the result of the combined effect of magmatic and volcano–tectonic structures. For the scope of our study it is important to characterize the magma reservoir that produces the long–term deformation. The ALOS dataset provides the best opportunity to do so, and data spanning a period of steady inflation between 2007 and 2009 was inverted. The deformation pattern cannot be explained using a single source. Several combinations between magmatic sources and fault planes were tested and best fit was obtained using a spherical source centered beneath the caldera at a
depth of 3.0 km (95% confidence interval: 2.3–3.7 km) and a SSW–NNE striking reverse
fault, dipping towards the center of the volcano and approximately reaching the depth of
the reservoir. The reservoir depth is similar to that inferred by Hooper et al. [2007] who,
however, modeled it as an extremely elongated pipe–like body.

Several models of the source of surface deformation at Sierra Negra are available in
the literature [Amelung et al., 2000; Yun et al., 2006; Chadwick et al., 2009; Jonsson,
2009], and all agree that uplift of the summit caldera is generated by a flat–topped source
located at 1.9–2.1 km beneath the surface. Using the same approach, we inverted for
InSAR data spanning a period of inflation between 2006 and 2008 and inferred that the
deformation is best explained by the opening of a ~3 x 5 km rectangular horizontal
source (for simplicity we only consider uniform opening) located at 2.3 km depth (95%
confidence interval: 2.0–2.6 km).

For Cerro Azul, InSAR data spanning the period of inflation that followed the 1998
eruption was used to constrain the source of deformation. The data is well explained by
the inflation of a spherical source centered beneath the summit of the volcano at a depth
of 6.1 km (95% confidence interval: 5.8–6.4 km), slightly deeper than the point source
modeled by Amelung et al. [2000] at ~5 km depth. No evidence for shallower reservoirs
similar to those inferred at the other volcanoes was found.

Finally, for Fernandina we adopt the model of the storage system proposed in
Chapter 3 and by Bagnardi and Amelung [2012], which includes a shallow flat–topped
reservoir located at 1.1 km depth and a second stacked ellipsoidal reservoir centered at
4.9 km depth (95% confidence intervals: 0.8–1.3 km and 4.3–5.6 km respectively).
Table 6.2: Reservoir geometry and depth at each volcano (95% confidence interval in brackets).

<table>
<thead>
<tr>
<th>Volcano</th>
<th>Depth (km)</th>
<th>Source geometry</th>
<th>Modeled time-period</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wolf</td>
<td>1.4 [0.9–1.9]</td>
<td>Rectangular (1.7 x 1.0 km)</td>
<td>12/2006 – 12/2009</td>
</tr>
<tr>
<td>Sierra Negra</td>
<td>2.3 [2.0–2.6]</td>
<td>Rectangular (2.7 x 4.8 km)</td>
<td>12/2006 – 1/2008</td>
</tr>
<tr>
<td>Fernandina (1) shallow</td>
<td>1.1 [0.8–1.3]</td>
<td>Rectangular</td>
<td>1/2007 – 7/2007</td>
</tr>
<tr>
<td>Fernandina (2) deep</td>
<td>4.9 [4.3–5.6]</td>
<td>Prolate spheroid (aspect ratio: 0.2)</td>
<td>9/2007 – 4/2008</td>
</tr>
</tbody>
</table>

Note: all the other source parameters are presented in Appendix S4. Rectangular sources [Okada, 1985]; spherical sources [McTigue, 1987]; prolate spheroid [Yang et al., 1988].

6.6 Magma supply

6.6.1 Supply to single volcanoes

In Figure 6.3 we present time series of cumulative volume change inferred from the deformation data. Periods of surface uplift represent phases of magma accumulation in the reservoirs, while periods of subsidence likely track magma reservoir emptying not related to eruptive or shallow intrusive activity (subsidence associated with these events is removed from the time series).

At Wolf and Darwin very limited volumes of magma were supplied and stored during the study period, with a total of ~0.005 and ~0.010 km$^3$ respectively (Figure 6.3a and 6.3b). At Alcedo ~ 0.032 km$^3$ of magma had accumulated by the time it started to subside in 2003, and ~ 0.004 km$^3$ were loss until inflation started again in 2007 and magma accumulation resumed, reaching a total of ~0.039 km$^3$ between 1992 and 2011 (Figure 6.3c).

Larger volumes of magma were supplied to the remaining three volcanoes, Cerro Azul, Fernandina and Sierra Negra. At Cerro Azul the reservoir accumulated ~0.062 km$^3$ of magma during the entire study period and erupted twice, in 1998 and 2008 (Figure 3d).
(Continues in the next page)
Figure 6.3 (Continues from previous page). – Cumulative volume change time series (right axis) inferred from the linear inversion of the vertical displacement (left axis) measured at each volcano. Red lines mark the occurrence of event associated with rapid deformation, removed from the time series. At Sierra Negra (panel e) gray and green dots indicate GPS measurements at station GV02 and GV04 respectively.
At Fernandina a total volume addition of ~0.051 km$^3$ occurred between 1992 and the end of 2010 (Figure 3f). This volume was stored for > 60% in the deeper reservoir and < 40% into the shallower zone of magma storage, despite the latter continuously inflated at least between 1998 and 2010 (except for brief episodes of deflation during the two eruptions and the two intrusions, [Chapter 3; Bagnardi and Amelung, 2012]).

At Sierra Negra > 0.11 km$^3$ accumulated within its 2–km–deep reservoir from 1992 until it erupted in 2005 (Figure 6.3e). From the end of the eruption, ~0.067 km$^3$ of magma were supplied to the volcano and stored in its shallow reservoir. The total cumulative magma addition between 1992 and 2011 to Sierra Negra’s storage system is of ~ 0.185 km$^3$, more than the total volume of all the other volcanoes together (~0.164 km$^3$).

6.6.2 Supply to the archipelago

In Figure 6.4 we present the archipelago-wide magma supply time series obtained by summing the relative contribution of each volcano (the different colors correspond to the different volcanoes). A total of 0.350±0.100 km$^3$ of magma was supplied to the western Galápagos volcanoes between 1992 and the end of 2010, at an average rate of ~0.02 km$^3$ yr$^{-1}$. During this time period, 53% of the magma supplied to the Galápagos volcanoes was directed toward Sierra Negra, 18% to Cerro Azul, 14% to Fernandina, 11% to Alcedo, 3% to Darwin and only 1% to Wolf (also see right column in Table 6.3 for total volumes and uncertainties).

The rate of supply, however, was not constant but largely varied from values as low as 0.002 km$^3$ yr$^{-1}$ between 2000 and 2003 during a period of subsidence at Sierra Negra,
to 0.06–0.07 km\(^3\) yr\(^{-1}\) prior to the 2005 eruption at Sierra Negra and after the 2007 intrusion at Fernandina (average rates are represented by dashed lines in Figure 6.4). No inferences on rate variability can be made for the years prior to 2000 because of the limited number of SAR acquisitions during this time period, which however shows an average rate of 0.019 km\(^3\) yr\(^{-1}\), in accordance with the longer-term one (0.020 km\(^3\) yr\(^{-1}\)). An average higher rate of supply, ~0.028 km\(^3\) yr\(^{-1}\), is instead inferred starting from 2003, when inflation at Sierra Negra resumed, through 2010.

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**Figure 6.4** – a) Cumulative volume change time series for the western Galápagos volcanoes. Average rates of magma supply for discrete periods (dashed lines) are reported. Only the total volume between 1992 and 2000 is shown because of the limited number of SAR acquisitions during this time period.

### 6.6.3 Magma distribution between volcanoes

As for the rate of supply, the amount of magma distributed between the six actively deforming volcanoes varied through time (Table 6.3). For example, no magma entered the storage system of Sierra Negra between 2000 and 2003 but mostly that of Cerro Azul
Sierra Negra, however, received 40% and 68% of the total supply during the years prior to 2000 and after 2003 respectively.

Table 6.3: Volume changes and magma distribution between volcanoes.

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
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<tbody>
<tr>
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<td>0.188 x10^3 km^3</td>
<td>0.350 x10^3 km^3</td>
<td>0.013 x10^3 km^3</td>
</tr>
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<td>128 [0.54–2.94]</td>
<td>2.72 [1.15 – 6.27]</td>
<td>4.86 [2.06–11.17]</td>
<td>2.72 [1.15 – 6.27]</td>
</tr>
<tr>
<td>6%</td>
<td>0%</td>
<td>0%</td>
<td>3%</td>
<td>6%</td>
</tr>
<tr>
<td>20%</td>
<td>0%</td>
<td>0%</td>
<td>3%</td>
<td>11%</td>
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<td>68%</td>
<td>53%</td>
<td>40%</td>
</tr>
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<td>27%</td>
<td>0%</td>
<td>6%</td>
<td>18%</td>
<td>27%</td>
</tr>
<tr>
<td>F</td>
<td>8.54 [5.66–12.05]</td>
<td>1.16 [0.77 – 1.64]</td>
<td>41.12 [38.15 – 44.70]</td>
<td>50.83 [31.75–73.03]</td>
</tr>
<tr>
<td>6%</td>
<td>9%</td>
<td>22%</td>
<td>14%</td>
<td>6%</td>
</tr>
</tbody>
</table>

W = Wolf; D = Darwin; A = Alcedo; SN = Sierra Negra; CA = Cerro Azul; F = Fernandina. 95% confidence intervals are obtained by linear propagation of the uncertainty in the estimate of the source parameters.

6.7. Discussion and conclusions

The average magma supply rate to the Galápagos volcanoes between 1992 and 2010, based on the analysis of deformation data, was ~0.020±0.005 km^3 yr^-1. This rate, however, appears to be variable as demonstrated by the occurrence of periods with very low supply (< 0.002 km^3 yr^-1) and periods, although brief, of much higher magma supply (> 0.060 km^3 yr^-1). If the 2003–2010 time period is considered, that which coincides with a time of inflation at Sierra Negra volcano, the average rate increases to ~0.028 km^3 yr^-1. Except for brief time intervals (< 6 months) during which magma supply is very high, the inferred rates are much lower than that calculated from the isostatic crustal thickness (0.03–0.3 km^3 yr^-1; [Ito et al., 1997]).
A limitation of our approach, and of most studies based on deformation data alone, is that the compressibility of magma is not considered. In a more realistic scenario where magma containing exsolved volatiles is supplied, a volume 3–4 times larger could be accommodated and generate the same amount of displacement at the surface [Johnson, 1992; Rivalta and Segall, 2008]. If we therefore multiply the supply rate inferred assuming incompressible magma, by a factor of 3–4 to account for the effect of its compressibility, the average rate of supply to the Galápagos volcanoes could be as high as 0.06–0.08 km³ yr⁻¹. If this same approach is applied to the higher rate of supply measured between 2003 and 2010, the value could increase up to 0.1 km³ yr⁻¹, similar to that inferred at Kīlauea volcano, Hawai‘i, using the effusion rate of long–lasting eruptions [Swanson et al., 1972].

Over the entire study period magma supply to the Galápagos volcanoes largely favored Sierra Negra (>50%), Cerro Azul and Fernandina (18% and 14% respectively). Short–term variations in the distribution, however, seem to occur and are likely to be related to the effect of eruptions, intrusions and of pressure imbalance that these events cause. In fact, magma is likely to be directed toward rapidly emptied reservoirs to re–establish the pressure equilibrium with their deeper source [e.g., Chapter 3; Bagnardi and Amelung, 2012]. On the other hand, eruptions during the past two decades have only occurred at the three volcanoes showing the highest rates of magma supply (Sierra Negra, Cerro Azul and Fernandina). A positive feedback between the two processes is therefore possible.
Chapter 7: Conclusions

The work presented in this dissertation relied on the analysis of geophysical data, in particular of space-born Interferometric Synthetic Aperture Radar (InSAR) measurements of surface displacement, to study the dynamics of magma supply, storage and migration at volcanoes forming the Galápagos and Hawaiian Islands. We overall demonstrated the value of geodetic measurements in characterizing phenomena related to the volcanic activity and highlighted the importance of combining deformation data with other geophysical techniques when studying active volcanoes.

7.1 Magma supply

We provided the first estimate of recent rates of magma supply to the Galápagos volcanoes and carried out the most detailed geodetic study of the entire archipelago. These results are based on the analysis of InSAR data acquired for almost two decades (1992-2011) at the six actively deforming volcanoes forming the islands of Isabela and Fernandina, where seven eruptions and at least two subvolcanic intrusions occurred during the study period. Our results highlight that the rate of magma supply from the mantle hotspot to the Galápagos volcanoes may be an order of magnitude lower (~ 0.02 km³ yr⁻¹) than that inferred at the Hawaiian volcanoes (0.1-0.2 km³ yr⁻¹). We also observed that, on average during the time spanned by the dataset, supply favored Sierra Negra volcano, which is not located at the leading edge of the Galápagos hotspot. Magma supply rates, however, seem to be largely influenced by the occurrence of eruptive and intrusive activity at the volcanoes, although eruptions during the past two decades have
only occurred at those volcanoes showing the highest rates of magma supply. A positive feedback between the two processes is therefore possible.

7.2 Magma storage

The storage system of all six actively deforming volcanoes in the Galápagos Islands has been characterized by studying InSAR measurements of the surface deformation. We inferred that in the Galápagos, magma is stored in crustal reservoirs with depth ranging between 1 and 6 km. Five volcanoes, Wolf, Darwin, Alcedo, Sierra Negra and Fernandina, have a shallow reservoir within 1-3 km depth, while at Cerro Azul magma is stored at greater depth (~6 km) and no evidence for shallower storage was found. A very detailed analysis of the surface deformation at Fernandina volcano has shown evidence of the presence of a second distinct area of magma storage at greater depth (~5 km). The two reservoirs are hydraulically connected as they both rapidly respond to pressure release events such as eruptions and intrusions.

At Kīlauea volcano, Hawai‘i, the joint analysis of deformation InSAR data and ground-based microgravity measurements collected between 2009 and the end of 2012, allowed us to not only infer the location of a shallow area of magma storage beneath the summit caldera, but also to detect a process of mass increase within the reservoir. This mass accumulation, however, did not produce significant displacement of the surface, implying that mechanisms beyond simple filling of a magma reservoir must have occurred at Kīlauea during the study period. Although the InSAR data shows that the ground uplifted during periods of gravity increase and that subsidence characterizes a time period of slight gravity decrease, the volume change inferred from the modeling of
the InSAR data can only account for a small portion (<10%) of the gravity increase. This discrepancy between gravity change and deformation could be explained through the replacement of gas-rich magma within the shallow reservoir by denser, outgassed magma. In fact, since 2008, the opening of a new vent within Kilauea’s summit caldera allows magma to convect up to the surface, loose its volatiles content and sink back into the reservoir.

### 7.3 Magma migration

Surface deformation data have provided an unprecedented opportunity to study processes of magma migration at the Galápagos volcanoes. At Fernandina, three eruptions and two subvolcanic intrusions are imaged by InSAR data. By modeling the measured displacements we inferred the geometry of these intrusive bodies.

The analysis and modeling of InSAR data spanning both radial and circumferential fissure eruptions at Fernandina volcano have led us to a new interpretation of the dynamics of magma migration and internal growth of the Galápagos volcanoes. Contrary to the assumption of magma transport through vertical dikes, we have demonstrated that both orientations of eruptive fissures are initiated by the intrusion of subhorizontal sills. More specifically, the intrusion of sills during the initial stages of a radial fissure eruption could explain the apparent alternation between eruption types, with circumferential eruptions occurring as a consequence of perturbations in the stress field generated by the preceding radial fissure eruption. Such a model provides a means of forecasting the style and location of a future eruption at Fernandina, which we anticipate will probably occur from circumferential fissures opening on the southwestern summit plateau of the volcano.
In two occasions, in 2006 and 2007, sills departing from the deeper reservoir at Fernandina volcano intruded under its southern flank generating broad uplift at the surface. These intrusions could provide an explanation for enigmatic episodes of rapid deformation observed in other coastal sectors of the islands (e.g., Punta Espinoza in 1927 and Urvina Bay in 1954). Furthermore, these events are associated to earthquake swarms with shocks of magnitude $M_w>5$. This seismicity could have tectonic origin and act as a trigger mechanism for the intrusions or, vice-versa, it could simply be the consequence of magma movement through the brittle crust or a combination of both.

Magma withdrawal from the shallow storage system during the subvolcanic intrusions also caused the rapid subsidence of the summit and of the caldera floor. This mechanism can be used to explain the discrepancy between the large volume of a caldera collapse that occurred in 1968 at Fernandina and the small volume of the magma erupted prior to the event. During this event magma likely migrated downward from the shallow reservoir and largely intruded at depth, causing the removal of the necessary support to the overlaying block and triggering an incremental caldera collapse.

### 7.4 Galápagos vs. Hawai‘i

This work has provided multiple elements to compare the dynamics of magma supply, storage and migration at the basaltic volcanoes forming the Galápagos and the Hawaiian Islands. The most evident differences or similarities are:

- Over the past decades, magma supply to the Galápagos volcanoes is up to an order of magnitude lower than that inferred for Hawai‘i. Magma is also
distributed between at least six volcanoes in the Galápagos, while it is only
directed toward two/three edifices in Hawai‘i.

- Volcanoes at both locations display similar subcaldera magma-storage
characteristics, in terms of reservoir depth range and shape. Volcanoes in the
Galápagos, however, do not show evidence for magma storage beneath the
volcanoes flanks, which represent important areas of storage at Hawaiian
volcanoes.

- Magma migration in the Galápagos seems to be dominated by subhorizontal
intrusions departing from both, shallow and deeper reservoirs, opposed to the
predominance of vertical dikes in Hawai‘i. This difference results in significant
dissimilarities in volcano growth and evolution.
Bibliography


Carbone, D., and M. P. Poland (2012), Gravity fluctuations induced by magma convection at Kilauea Volcano, Hawai‘i, *Geology*, 40(9), 803-806, doi:10.1130/G33060.1


Ito, G., J. Lin, and C. Gable (1997), Interaction of mantle plumes and migrating mid-ocean ridges: Implications for the Galápagos plume-ridge system,


Appendix S1: Supplementary material for Chapter 3

This appendix contains supplementary information for the two seismic swarms occurred in December 2006 (E2) and August 2007 (E3). It includes tables with the chronology of the single earthquakes recorded by the Global Seismic Network (GSN) and of the SAR images acquired during these events (Table S1.1 and Table S1.2).

Table S1.1. Chronological sequence of events during the 22-23 December 2006 seismic swarm

<table>
<thead>
<tr>
<th>Date</th>
<th>Time (UTC)</th>
<th>Type of event</th>
<th>Notes</th>
</tr>
</thead>
<tbody>
<tr>
<td>22/12/2006</td>
<td>15:41</td>
<td>SAR acquisition T54</td>
<td>Figure S2a</td>
</tr>
<tr>
<td>22/12/2006</td>
<td>22:27</td>
<td>mb 4.4 earthquake</td>
<td>Purple star in Figure S3</td>
</tr>
<tr>
<td>23/12/2006</td>
<td>04:15</td>
<td>SAR acquisition T61</td>
<td>Figure S2b</td>
</tr>
<tr>
<td>23/12/2006</td>
<td>10:40</td>
<td>mb N/A earthquake</td>
<td>Orange star in Figure S3</td>
</tr>
</tbody>
</table>

For total displacement associated to this event see Figure 3.3b. Earthquake information from USGS-NEIC.

Table S1.2. Chronological sequence of events during the 27-30 August seismic swarm

<table>
<thead>
<tr>
<th>Date</th>
<th>Time (UTC)</th>
<th>Type of event</th>
<th>Notes</th>
</tr>
</thead>
<tbody>
<tr>
<td>27/08/2007</td>
<td>18:11</td>
<td>mb 4.6 earthquake</td>
<td>Red star in Figure S4</td>
</tr>
<tr>
<td>27/08/2007</td>
<td>20:16</td>
<td>mb 5.0 earthquake</td>
<td>Dark green star in Figure S4</td>
</tr>
<tr>
<td>28/08/2007</td>
<td>04:21</td>
<td>SAR acquisition T104</td>
<td>Figure S2c</td>
</tr>
<tr>
<td>28/08/2007</td>
<td>22:46</td>
<td>Mw 5.2 earthquake</td>
<td>Orange star in Figure S4</td>
</tr>
<tr>
<td>29/08/2007</td>
<td>05:26</td>
<td>mb 3.8 earthquake</td>
<td>Brown star in Figure S4</td>
</tr>
<tr>
<td>29/08/2007</td>
<td>21:58</td>
<td>mb 4.7 earthquake</td>
<td>Light green star in Figure S4</td>
</tr>
<tr>
<td>29/08/2007</td>
<td>22:01</td>
<td>mb 4.3 earthquake</td>
<td>Blue star in Figure S4</td>
</tr>
<tr>
<td>29/08/2007</td>
<td>22:10</td>
<td>Mw 5.4 earthquake</td>
<td>Purple star in Figure S4</td>
</tr>
<tr>
<td>30/08/2007</td>
<td>04:29</td>
<td>mb 4.3 earthquake</td>
<td>Black star in Figure S4</td>
</tr>
<tr>
<td>30/08/2007</td>
<td>15:52</td>
<td>SAR acquisition T140</td>
<td>Figure S2d</td>
</tr>
</tbody>
</table>

For total displacement associated to this event see Figure 3.3c. Earthquake information from USGS-NEIC.
In Figure S1.1 we present four SAR interferograms formed using the SAR images acquired during the two events. The data is provided to better observe the evolution of the surface deformation during each event.

Figure S1.1: Envisat SAR interferograms partially spanning: a-b) the 22-23 December 2006 seismic swarm (E2), details in Table S1.1; c-d) the 27-30 August 2007 seismic swarm (E3), details in Table S1.2. Satellite flight direction and radar look direction are presented with arrows. Each fringe (full color cycle) represents 2.8 cm of range change between the ground and the satellite, or LOS (line-of-sight) displacement.
In Figure S1.2 and Figure S1.3 we present two maps showing location and 90% error ellipse for each earthquake recorded during the two seismic swarms.

**Figure S1.2:** Earthquake locations and 90% error ellipses for the December 2006 seismic swarm (E2). Information retrieved from the USGS-NEIC catalog (ftp://hazards.cr.usgs.gov/edr/mchedr/).
Finally, we include InSAR time-series results for the ascending pass (track T61) to compare with results from the descending pass (track T412) presented in Figure 3.3.
Figure S1.4: a) Triangulated network of interferometric pairs used to generate the SBAS time series. Each square represents a SAR acquisition. Solid lines show the selected pairs, dotted lines are pairs that did not meet the selection criteria (see 3.4.1). b) LOS displacement times series relative to R2 for two pixels: (A1) at the center of the summit caldera and (A2) on the southwestern upper flank. Dark gray solid lines represent the occurrence of eruptions or local seismic activity associated to rapid displacement (Event E1 - E4). Pre- and post- eruptive/seismic intervals are shown with different background colors. Blue dotted lines represent the occurrence of spatial variations in the deformation pattern and divide the studied interval in eleven time periods (P1 through P11). Red squares and green diamonds mark LOS surface displacement for each location at the time of SAR acquisitions.
Appendix S2: Supplementary material for Chapter 4

This appendix contains supplementary information for data presented in Chapter 4.

Figure S2.1: Deformation preceding the April 2009 pre-eruptive sill intrusion at Fernandina. Envisat interferogram (track 54, beam mode IS7, descending, spanning 70 days) that shows deformation up to 09:41 on April 10, 2009 (local time), ~14–15 hours prior to the start of the April 2009 radial fissure eruption and ~13 hours prior to the subsequent SAR acquisition for the same area (which shows evidence for the sill-intrusion; Figure 4.2c). The interferogram shows typical weak intra-caldera uplift comparable to other periods of inflation of the shallow magma reservoir (Figure 4.2d) [Bagnardi and Amelung, 2012]; there is no significant displacement extending outside the caldera. Each fringe (full color cycle) represents 2.8 cm of LOS displacement. The white line outlines the summit caldera rim.
Table S2.1: 2009 pre-eruptive sill intrusion, results from non-linear inversion of InSAR data\(^a\):
estimated source parameters. The 95% confidence intervals are shown in brackets. RMS – root mean
square = 15.38 mm

<table>
<thead>
<tr>
<th>Sill-intrusion</th>
<th>Width (km): 3.46 [3.32 – 3.60]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Length (km): 2.48 [2.26 – 2.70]</td>
<td>Min. depth (km): 0.80 [0.67 – 0.91]</td>
</tr>
<tr>
<td>X_UTM(^b) (m): 659746 [659926 – 659566]</td>
<td>Opening (m): 0.85 [0.71 – 0.99]</td>
</tr>
<tr>
<td>Y_UTM(^b) (m): 9957310 [9957130– 9957490]</td>
<td>Volume change (10^6 m^3): 7.29</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Deeper reservoir</th>
<th>Aspect ratio b/a: 0.18 [fixed]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Maj. Axis a (km): 2.73 [fixed]</td>
<td>Dip angle (degrees): 0 [fixed]</td>
</tr>
<tr>
<td>Depth (km): 4.93 [fixed]</td>
<td>Normalized pressure change (μ <em>Pa): -2.2</em>10^{-3}</td>
</tr>
<tr>
<td>Strike (degrees): 48 [fixed]</td>
<td>X_UTM(^b) (m): 663436 [fixed]</td>
</tr>
<tr>
<td>Y_UTM(^b): 9960010 [fixed]</td>
<td></td>
</tr>
</tbody>
</table>

\(^a\)Envisat, track 61, beam mode IS2, ascending (31/01/2009–10/04/2009).

\(^b\)UTM zone 15M, X = easting, Y = northing.
Table S2.2: 2009 eruption, modeling of InSAR data\(^a\): estimated source parameters. RMS – root mean square = 30.93 mm

<table>
<thead>
<tr>
<th>Deeper reservoir (r1) – for the other parameters (fixed) see Table S1.</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Normalized pressure change (µ*Pa):</strong> -16.50*10(^{-3})</td>
</tr>
<tr>
<td><strong>Shallow reservoir (r2)</strong></td>
</tr>
<tr>
<td>Length (km): 2.80 [fixed]</td>
</tr>
<tr>
<td>Depth (km): 1.08 [fixed]</td>
</tr>
<tr>
<td>Strike (degrees): 121 [fixed]</td>
</tr>
<tr>
<td>X(_{\text{UTM}}^b) (m): 661613 [fixed]</td>
</tr>
<tr>
<td>Volume change (10(^6) m(^3)): - 4.8</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Sill-intrusion (s) – for all parameters (fixed) see Table 1</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Radial intrusion (d1)</strong></td>
</tr>
<tr>
<td>Length (km): 3.91 [fixed]</td>
</tr>
<tr>
<td>Depth (km): 0.65 [fixed]</td>
</tr>
<tr>
<td>Strike (degrees): 210 [fixed]</td>
</tr>
<tr>
<td>X(_{\text{UTM}}^b) (m): 658623 [fixed]</td>
</tr>
<tr>
<td>Volume change (10(^6) m(^3)): 6.92</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th><strong>Radial intrusion (d2)</strong></th>
</tr>
</thead>
<tbody>
<tr>
<td>Length (km): 2.50 [fixed]</td>
</tr>
<tr>
<td>Depth (km): 0.0 [fixed]</td>
</tr>
<tr>
<td>Strike (degrees): 210 [fixed]</td>
</tr>
<tr>
<td>X(_{\text{UTM}}^b) (m): 647953 [fixed]</td>
</tr>
<tr>
<td>Volume change (10(^6) m(^3)): 2.81</td>
</tr>
</tbody>
</table>

\(^a\)Envisat, track 61, beam mode IS2, ascending (31/01/2009–16/05/2009) – Envisat, track 54, beam mode IS7, descending (30/01/2009–15/05/2009)

\(^b\)UTM zone 15M, X = easting, Y = northing.
Table S2.3: 1995 eruption, results from non-linear inversion of InSAR data\(^a\): estimated source parameters. The 95% confidence intervals are shown in brackets. RMS – root mean square = 16.28 mm

<table>
<thead>
<tr>
<th></th>
<th>Sill intrusion</th>
<th>Radial intrusion</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Length (km):</strong></td>
<td>2.35 [1.97 – 2.73]</td>
<td>4.21 [4.03 – 4.39]</td>
</tr>
<tr>
<td><strong>Depth (km):</strong></td>
<td>0.64 [0.42 - 0.86]</td>
<td>0.0 [fixed]</td>
</tr>
<tr>
<td><strong>Strike (degrees):</strong></td>
<td>104 [98 - 110]</td>
<td>230 [225 - 230]</td>
</tr>
<tr>
<td><strong>X_UTM(^b) (m):</strong></td>
<td>660823 [660663 - 660983]</td>
<td>657803 [657603 - 658003]</td>
</tr>
<tr>
<td><strong>Volume change (10(^6) m(^3)):</strong></td>
<td>3.47</td>
<td>6.09</td>
</tr>
</tbody>
</table>

\(^a\)Envisat, track 61, beam mode IS2, ascending (12/09/1992–30/09/1997)

\(^b\)UTM zone 15M, X = easting, Y = northing.
Appendix S3: Supplementary material for Chapter 5

This appendix contains deformation-modeling results for the Halema‘uma‘u (HMM) source, the southwest rift zone (SWRZ) source and for each time period (period 1 through period 4). A comparison between data, model and residual is given together with the estimated source parameters (and their 95% confidence interval). See Figure S3.1-S3.5.

Figure S3.1 – Modeling results for the HMM source during a period of deflation. Comparison between DATA (from TerraSAR-X time series: TSXD = descending; TSXA= ascending), MODEL and RESIDUAL. Surface deformation is modeled using a finite spherical source [McTigue, 1987]. The + sign indicates the surface projection of the source.
Figure S3.2 – Modeling results for the HMM source during a period of inflation. Comparison between DATA (from TerraSAR-X time series: TSXD = descending; TSXA= ascending), MODEL and RESIDUAL. Surface deformation is modeled using a finite spherical source [McTigue, 1987]. The + sign indicates the surface projection of the source.
Figure S3.3 – Modeling results for the SWRZ source. Comparison between DATA (from TerraSAR-X time series: TSXD = descending; TSXA= ascending), MODEL and RESIDUAL. Surface deformation is modeled using a rectangular planar source [Okada, 1985]. The red rectangle outlines the source location and geometry.

RMS: 13.84 mm
N. of data points: TSXD 1330, TSXA 4182

Dislocation source (Okada, 1985):
Length[km]: 10.82 [ 9.38 – 12.26]
Width[km]: 0.91 [ 0.44 – 1.38]
X[km]*: 7.31 [6.79 – 7.83]
Y[km]*: 4.88 [4.36 – 5.40]
Z[km]: 3.88 [3.42 – 4.34]
Strike[deg.]: 225 [222 – 228]
Opening[m]: -0.21 [-0.19 – -0.23]

*XY origin UTM coordinates:
50 249934 m E 2137954 m N
**Figure S3.4** - Modeling results for the periods 1 and 2. Comparison between DATA (from TerraSAR-X time series: TSXD = descending), MODEL and RESIDUAL. The SWRZ source (red rectangle) is modeled using a rectangular planar source [Okada, 1985]. The red rectangle outlines the source location and geometry. The HMM source (+ sign) is modeled using a finite spherical source [McTigue, 1987].

**Period 1**

- RMS: 5.85 mm
- N. of data points: TSXD 1964

**Spherical source (McTigue, 1987):**
- \(X_[km] = 10.94\) [fixed]
- \(Y_[km] = 9.89\) [fixed]
- \(Z_[km] = 1.48\) [fixed]
- \(\Delta V [x 10^6 m^3]: 0.16 [0.01 - 0.37]\)

**Dislocation source (Okada, 1985):**
- Length[km]: 10.82 [fixed]
- Width[km]: 0.91 [fixed]
- \(X_[km] = 7.31\) [fixed]
- \(Y_[km] = 4.88\) [fixed]
- \(Z_[km] = 3.88\) [fixed]
- Strike[deg]: 225 [fixed]
- Opening[m]: -0.21 [-0.19 - 0.23]

**Period 2**

- RMS: 7.63 mm
- N. of data points: TSXD 3350

**Spherical source (McTigue, 1987):**
- \(X_[km] = 10.94\) [fixed]
- \(Y_[km] = 9.89\) [fixed]
- \(Z_[km] = 1.48\) [fixed]
- \(\Delta V [x 10^6 m^3]: -0.33 [-0.26 - 0.40]\)

**Dislocation source (Okada, 1985):**
- Length[km]: 10.82 [fixed]
- Width[km]: 0.91 [fixed]
- \(X_[km] = 7.31\) [fixed]
- \(Y_[km] = 4.88\) [fixed]
- \(Z_[km] = 3.88\) [fixed]
- Strike[deg]: 225 [fixed]
- Opening[m]: -0.10 [-0.06 - 0.14]

*XY origin UTM coordinates: 5G 249934 m E 2137954 m N
Figure S3.5 – Modeling results for the periods 3 and 4. Comparison between DATA (from TerraSAR-X time series: TSXD = descending), MODEL and RESIDUAL. The SWRZ source (red rectangle) is modeled using a rectangular planar source [Okada, 1985]. The red rectangle outlines the source location and geometry. The HMM source (+ sign) is modeled using a finite spherical source [McTigue, 1987].

**Period 3**
- RMS: 8.81 mm
- N. of data points: TSXD 6805

**Spherical source (McTigue, 1987):**
- $X[km]^*: 10.94$ [fixed]
- $Y[km]^*: 9.89$ [fixed]
- $Z[km]^*: 1.48$ [fixed]
- $\Delta V \times 10^6 m^3$: 1.06 [1.00 – 1.11]

**Dislocation source (Okada, 1985):**
- Length[km]: 10.82 [fixed]
- Width[km]: 0.91 [fixed]
- $X[km]^*: 7.31$ [fixed]
- $Y[km]^*: 4.88$ [fixed]
- $Z[km]: 3.88$ [fixed]
- Strike[deg.]: 225 [fixed]
- Opening[m]: -0.14 [-0.10 – -0.18]

**Period 4**
- RMS: 11.84 mm
- N. of data points: CSKD 5646

**Spherical source (McTigue, 1987):**
- $X[km]^*: 10.94$ [fixed]
- $Y[km]^*: 9.89$ [fixed]
- $Z[km]^*: 1.48$ [fixed]
- $\Delta V \times 10^6 m^3$: 0.65 [0.54 – 0.77]

**Dislocation source (Okada, 1985):**
- Length[km]: 10.82 [fixed]
- Width[km]: 0.91 [fixed]
- $X[km]^*: 7.31$ [fixed]
- $Y[km]^*: 4.88$ [fixed]
- $Z[km]: 3.88$ [fixed]
- Strike[deg.]: 225 [fixed]
- Opening[m]: -0.16 [-0.12 – -0.20]

*XY origin UTM coordinates:
5G 249934 m E 2137954 m N
It also contains tables with the gravity data used in this study (Table S3.1-S3.5).

### Table S3.1. Gravity data for Period 1, December 2009 – July 2010

<table>
<thead>
<tr>
<th>Site</th>
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<th>Lon</th>
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“Lon”, degrees, longitude of the gravity station.
“G@t1”, mGal, relative gravity at the time of the first gravity survey (relative to P1)
“Stdev”, mGal, standard deviation of G@t1
“H@t1”, m, elevation at the time of the first gravity survey
“Stdev”, m, standard deviation of H@t1
“G@t2”, mGal, relative gravity at the time of the second gravity survey (relative to P1)
“Stdev”, mGal, standard deviation of G@t2
“H@t2”, m, elevation at the time of the second gravity survey
“Stdev”, m, standard deviation of H@t2
“DG”, mGal, gravity change (G@t2 – G@t1)
“DH”, m, elevation change (H@t2 – H@t1)
“DGR”, mGal, residual gravity change after free-air correction (DG-DH*0.308)
“Stdev”, mGal, standard deviation of DGR
“LL500”, mGal, lava-lake adjustment for a magma density of 500 kg m^-3.
“LL1000”, mGal, lava-lake adjustment for a magma density of 1000 kg m^-3 (used in this study).
“LL1500”, mGal, lava-lake adjustment for a magma density of 1500 kg m^-3.
Table S3.2. Gravity data for Period 2, July 2010 – March 2011

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**“Site”, gravity station name.**
**“Lat”, degrees, latitude of the gravity station.**
**“Lon”, degrees, longitude of the gravity station.**
**“G@t1”, mGal, relative gravity at the time of the first gravity survey (relative to P1)**
**“Stdev”, mGal, standard deviation of G@t1**
**“H@t1”, m, elevation at the time of the first gravity survey**
**“Stdev”, m, standard deviation of H@t1**
**“G@t2”, mGal, relative gravity at the time of the second gravity survey (relative to P1)**
**“Stdev”, mGal, standard deviation of G@t2**
**“H@t2”, m, elevation at the time of the second gravity survey**
**“Stdev”, m, standard deviation of H@t2**
**“DG”, mGal, gravity change (G@t2 – G@t1)**
**“DH”, m, elevation change (H@t2 – H@t1)**
**“DGR”, mGal, residual gravity change after free-air correction (DG-DH*-0.308)**
**“Stdev”, mGal, standard deviation of DGR**
**“LL500”, mGal, lava-lake adjustment for a magma density of 500 kg m^-3.**
**“LL1000”, mGal, lava-lake adjustment for a magma density of 1000 kg m^-3. (used in this study).”**
**“LL1500”, mGal, lava-lake adjustment for a magma density of 1500 kg m^-3.**
**“LL2000”, mGal, lava-lake adjustment for a magma density of 2000 kg m^-3.**
Table S3.3. Gravity data for Period 3, March 2011 – June 2012

| Site | Lat  | Lon  | G@t1 | Stdev | H@t1 | Stdev | G@t2 | Stdev | DG  | Stdev | DH  | Stdev | DGR  | Stdev | LL500 | LL1000 | LL1500 | LL2000 |
|------|------|------|------|-------|------|-------|------|-------|------|-------|------|-------|-------|-------|-------|--------|--------|--------|--------|
| FS   | -155.3835508 | -205.282200 | 0.409 | 0.000 | -0.410 | 0.015 | 0.000 | 0.000 | 0.007 | 0.010 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 |
| L1   | -155.3565008 | -205.282200 | 0.403 | 0.000 | -0.413 | 0.014 | 0.000 | 0.000 | 0.007 | 0.012 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 |
| L2   | -155.3575008 | -205.282200 | 0.413 | 0.000 | -0.422 | 0.010 | 0.000 | 0.000 | 0.004 | 0.007 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 |
| L3   | -155.3675008 | -205.282200 | 0.423 | 0.000 | -0.432 | 0.010 | 0.000 | 0.000 | 0.004 | 0.007 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 |
| L4   | -155.3775008 | -205.282200 | 0.433 | 0.000 | -0.443 | 0.010 | 0.000 | 0.000 | 0.004 | 0.007 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 |
| L5   | -155.3875008 | -205.282200 | 0.443 | 0.000 | -0.453 | 0.010 | 0.000 | 0.000 | 0.004 | 0.007 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 |
| L6   | -155.3975008 | -205.282200 | 0.453 | 0.000 | -0.463 | 0.010 | 0.000 | 0.000 | 0.004 | 0.007 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 |

“Site”, gravity station name.
“Lat”, degrees, latitude of the gravity station.
“Lon”, degrees, longitude of the gravity station.
“G@t1”, mGal, relative gravity at the time of the first gravity survey (relative to P1)
“Stdev”, mGal, standard deviation of G@t1
“H@t1”, m, elevation at the time of the first gravity survey
“Stdev”, m, standard deviation of H@t1
“G@t2”, mGal, relative gravity at the time of the second gravity survey (relative to P1)
“Stdev”, mGal, standard deviation of G@t2
“H@t2”, m, elevation at the time of the second gravity survey
“Stdev”, m, standard deviation of H@t2
“DG”, mGal, gravity change (G@t2 – G@t1)
“DH”, m, elevation change (H@t2 – H@t1)
“DGR”, mGal, residual gravity change after free-air correction (DG-DH*-0.308)
“Stdev”, mGal, standard deviation of DGR
“LL500”, mGal, lava-lake adjustment for a magma density of 500 kg m^-3.
“LL1000”, mGal, lava-lake adjustment for a magma density of 1000 kg m^-3 (used in this study).
“LL1500”, mGal, lava-lake adjustment for a magma density of 1500 kg m^-3.
Table S3.4. Gravity data for Period 4, June 2012 – November 2012

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“Stdev”, mGal, standard deviation of G@t1
“H@t1”, m, elevation at the time of the first gravity survey
“Stdev”, m, standard deviation of H@t1
“G@t2”, mGal, relative gravity at the time of the second gravity survey (relative to P1)
“Stdev”, mGal, standard deviation of G@t2
“H@t2”, m, elevation at the time of the second gravity survey
“Stdev”, m, standard deviation of H@t2
“DG”, mGal, gravity change (G@t2 – G@t1)
“DH”, m, elevation change (H@t2 – H@t1)
“DGR”, mGal, residual gravity change after free-air correction (DG-DH*0.308)
“Stdev”, mGal, standard deviation of DGR
“LL500”, mGal, lava-lake adjustment for a magma density of 500 kg m\(^{-3}\).
“LL1000”, mGal, lava-lake adjustment for a magma density of 1000 kg m\(^{-3}\) (used in this study).
“LL1500”, mGal, lava-lake adjustment for a magma density of 1500 kg m\(^{-3}\).
“LL2000”, mGal, lava-lake adjustment for a magma density of 2000 kg m\(^{-3}\).
Table S3.5. Gravity data for Period 3+4, March 2011 – November 2012

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“Stdev”, mGal, standard deviation of G@t1
“H@t1”, m, elevation at the time of the first gravity survey
“Stdev”, m, standard deviation of H@t1
“G@t2”, mGal, relative gravity at the time of the second gravity survey (relative to P1)
“Stdev”, m, standard deviation of G@t2
“H@t2”, m, elevation at the time of the second gravity survey
“Stdev”, m, standard deviation of H@t2
“DG”, mGal, gravity change (G@t2 – G@t1)
“DGR”, mGal, residual gravity change after free-air correction (DG-DH*-0.308)
“Stdev”, mGal, standard deviation of DGR
“LL500”, mGal, lava-lake adjustment for a magma density of 500 kg m^-3.
“LL1000”, mGal, lava-lake adjustment for a magma density of 1000 kg m^-3 (used in this study).
“LL1500”, mGal, lava-lake adjustment for a magma density of 1500 kg m^-3.
Appendix S4: Supplementary material for Chapter 6

This appendix contains deformation-modeling results for Wolf, Darwin, Alcedo, Sierra Negra, and Cerro Azul volcanoes, Galápagos Islands. For Fernandina see Chapter 3.

Table S4.1: Wolf, magma reservoir modeling results (95% confidence interval in brackets).

<table>
<thead>
<tr>
<th>Length (km)</th>
<th>Width (km)</th>
<th>Depth (km)</th>
<th>Strike (deg.)</th>
<th>X (km)</th>
<th>Y (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.67</td>
<td>1.01</td>
<td>1.43</td>
<td>150</td>
<td>10.49</td>
<td>8.77</td>
</tr>
<tr>
<td>[0.47–2.87]</td>
<td>[0.45–1.57]</td>
<td>[0.93–1.93]</td>
<td>[88–202]</td>
<td>[9.95–11.03]</td>
<td>[8.07–9.47]</td>
</tr>
</tbody>
</table>

$X_0$: 91.44 degrees (longitude); $Y_0$: -0.06 degrees (latitude).

Table S4.2: Darwin, magma reservoir modeling results (95% confidence interval in brackets).

<table>
<thead>
<tr>
<th>X (km)</th>
<th>Y (km)</th>
<th>Depth (km)</th>
<th>Radius (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>8.65</td>
<td>8.87</td>
<td>3.07</td>
<td>0.2</td>
</tr>
</tbody>
</table>

$X_0$: 91.37 degrees (longitude); $Y_0$: -0.27 degrees (latitude).

Table S4.3: Alcedo, magma reservoir modeling results (95% confidence interval in brackets).

<table>
<thead>
<tr>
<th>X (km)</th>
<th>Y (km)</th>
<th>Depth (km)</th>
<th>Radius (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>11.57</td>
<td>7.27</td>
<td>3.04</td>
<td>0.2</td>
</tr>
</tbody>
</table>

$X_0$: 91.22 degrees (longitude); $Y_0$: -0.50 degrees (latitude).

Table S4.4: Sierra Negra, magma reservoir modeling results (95% confidence interval in brackets).

<table>
<thead>
<tr>
<th>Length (km)</th>
<th>Width (km)</th>
<th>Depth (km)</th>
<th>Strike (deg.)</th>
<th>X (km)</th>
<th>Y (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2.68</td>
<td>4.83</td>
<td>2.27</td>
<td>160</td>
<td>10.91</td>
<td>10.11</td>
</tr>
</tbody>
</table>

$X_0$: 91.25 degrees (longitude); $Y_0$: -0.915 degrees (latitude).
Table S4.5: Cerro Azul, magma reservoir modeling results (95% confidence interval in brackets).

<table>
<thead>
<tr>
<th>X (km)</th>
<th>Y (km)</th>
<th>Depth (km)</th>
<th>Radius (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>13.77</td>
<td>11.13</td>
<td>6.08</td>
<td>0.2</td>
</tr>
<tr>
<td>[13.59–13.95]</td>
<td>[10.95–11.31]</td>
<td>[5.78–6.38]</td>
<td>Fixed</td>
</tr>
</tbody>
</table>

\[X_0 = 91.51 \text{ degrees (longitude); } Y_0 = -1.02 \text{ degrees (latitude).}\]