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Geodynamic Signals Detected by Geodetic Methods in Iceland

Páll Einarsson[1], Freysteinn Sigmundsson[2], Erik Sturkell[2], Þóra Árnadóttir[2], Rikke Pedersen[2], Carolina Pagli[2], Halldór Geirsson[3]

[1] Institute of Earth Sciences, University of Iceland
[2] Nordic Volcanological Center, Institute of Earth Sciences, University of Iceland
[3] Icelandic Meteorological Office, Reykjavík

Abstract

The geodynamics laboratory provided by Iceland's position on an active mid-ocean ridge has been recognized for several decades. Geodetic experiments have been designed and carried out in Iceland since 1938 to verify various global geodynamic theories, such as Wegener's theory of continental drift, the sea floor spreading hypothesis, plate tectonics, mantle plumes etc. State-of-the-art techniques have been used to obtain data on crustal displacements with ever increasing accuracy to constrain the theories. Triangulation and optical levelling were used in the beginning, later EDM-trilateration. Network GPS surveying began in 1986 and has been used extensively since then to study crustal movements. With the addition of InSAR and continuous GPS in the last decade we have made a significant stride towards the goal of giving a continuous representation of the displacement field in time and space. The largest and most persistent signal is that of the plate movements. Geodetic points in East and West Iceland move with the Eurasia and North America Plates, respectively, and the vectors are consistent with global models of plate movements. The plate boundary zones are a few tens of kilometers wide, within which strain accumulates. This strain is released in rifting events or earthquakes that have a characteristic displacement field associated with them. In the Krafla rifting episode in 1975-1984 a 100 km long section of the plate boundary in North Iceland was affected and divergent movement as large as 8-9 m was measured. The June 2000 earthquakes in the South Iceland Seismic Zone were the most significant seismic events in the last decades. Two magnitude 6.5 earthquakes and several magnitude 5 events were associated with strike-slip faulting on several parallel faults along the transform-type plate boundary. Slow post-rifting and post-seismic displacements were detected in the months and years following these events, caused by coupling of the elastic part of the crust with the visco-elastic substratum. Viscosities in the range 0.3-30 x 10^{18} Pa s have been estimated from the time-decay of these fields. Similar values are obtained from crustal uplift measured around the Vatnajökull glacier due to the reduced load of the glacier in the last century. Magma movements in the roots of volcanoes are reflected by deformation fields measureable around them. The volcanoes inflate or deflate in response to pressure increase or decrease in magma chambers, and intrusive bodies are revealed by bulging of the crust above them. The most active volcanoes in Iceland, Katla, Hekla, and Grímsvötn, appear to be inflating at the present time, whereas Krafla and Askja are slowly deflating. An intrusion episode was documented near the Hengill volcano in 1994-1998 and two intrusion events occurred in the Eyjafjallajökull volcano in 1994 and 1999, all of which were accompanied by characteristic deformation fields.

1 Introduction

Iceland provides rare opportunities to study a multitude of geodynamic phenomena. It is a platform of dimensions 300 km x 500 km situated astride a divergent plate boundary and on top of a hotspot presumed to be fed by a deep mantle plume (Einarsson 1991a, 2001). In addition glaciers provide a changing load on the Earth´s crust providing a test of its response (Sigmundsson and Einarsson 1992). Evidence of recent crustal movements are common and widespread, such as fractures, faults, tilted strata, raised terraces and changing coastlines. Volcanic structures and volcanic events are plentiful. The potential to verify important geodynamic hypotheses by direct measurements in Iceland was acknowledged early on, e.g. when a German geodetic-geological expedition came to Iceland in 1938 to verify the Wegener hypothesis of continental drift (Niemczyk 1943). This effort was interrupted by World War II but was later continued when new theories of sea floor spreading and plate tectonics became established in the late sixties and early seventies (e.g. Gerke 1974). German geodesists have ever since played a key role in the study of the kinematics of the crust in Iceland.

In this paper we give a short overview of geodynamic processes that are expected to give a detectable signal in geodetic data and give examples that have already been established in the recent literature. For further and deeper treatise on this topic the reader is referred to the book by Sigmundsson (2006).

2 Short History of Methods

The initial German geodetic effort in 1938 (Niemczyk 1943) was based on conventional geodetic techniques, triangulation, optical levelling and gravimetry. Part of the network was destroyed during WW II because of suspicion of espionage. The advent of electronic distance measurement techniques (EDM) in the sixties allowed more precise measurements and new projects were initiated by German, USA, and English universities (e.g. Gerke 1974, Torge and Drewes 1977, Decker et al. 1971, Brander et al. 1976). These techniques were applied to study the extensive crustal movements that occurred during the Krafla rifting episode in the Northern Volcanic Zone (Figure 1) that started in 1975 (Tryggvason 1984). Optical leveling, gravity and tilt measurements were also used extensively to monitor plate boundary deformation and volcano inflation and deflation (Sigurdsson 1980, Tryggvason 1980, Johnsen et al. 1980, Torge 1989). The GPS technology opened up new possibilities and the first GPS-project was conducted in 1986. It was a large cooperative project with many participants and the main target was the South Iceland Seismic Zone (Foulger et al. 1987, 1993). The second cooperative project was conducted in 1987, mainly around the Northern Volcanic Zone, this time with a major participation of the University of Hannover. This network was remeasured in 1990 and 1995 (e.g. Seeber 1989) and provided data for two Doctoral Theses by Jahn (1992) and Völksen (2000). Geodetic GPS-instruments were acquired by Iceland in 1992 and since then numerous networks have been installed in the active zones, with or without participation of foreign groups.

A large progressive step was taken in 1995-1997 when continuous GPS-stations were set up in Reykjavík and Höfn, straddling the plate boundary. The Continuous GPS-network (CGPS) has been expanding since then (Geirsson et al. 2005 submitted), giving valuable information on the temporal evolution of the crustal deformation fields. Continuous GPS-stations are operated at 18 points at the present time providing location every day. The largest build-up phases of the CGPS-network occurred in 1999 in SW-Iceland following the inflation period of the Hrómundartindur volcanic system near the Hengill triple junction (Sigmundsson et al. 1997), in 2000 after the large earthquakes in the South Iceland Seismic Zone, and in 2001 around the Katla volcano in S-Iceland following a re-awakening of the volcano in 1999 (Sturkell et al. 2003b).

Interferometric analysis of synthetic aperture radar images (InSAR) acquired by satellites has further expanded the possibilities to detect geodynamic signals in the surface deformation field (e.g., Massonnet and Sigmundsson 2000). By comparing interferometrically two synthetic aperture radar images of the same area, taken at some time interval by the SAR satellites, one can determine surface displacement in the direction of the line-of-sight to the satellite. If the signals of the two images are sufficiently coherent one obtains a two-dimensional map of the field. The InSAR technique has been successfully applied in Iceland since 1997 to a range of geodynamic problems (e.g. Vadon and Sigmundsson 1997, Sigmundsson et al. 1997, Pedersen et al. 2003, Pagli et al. 2005).

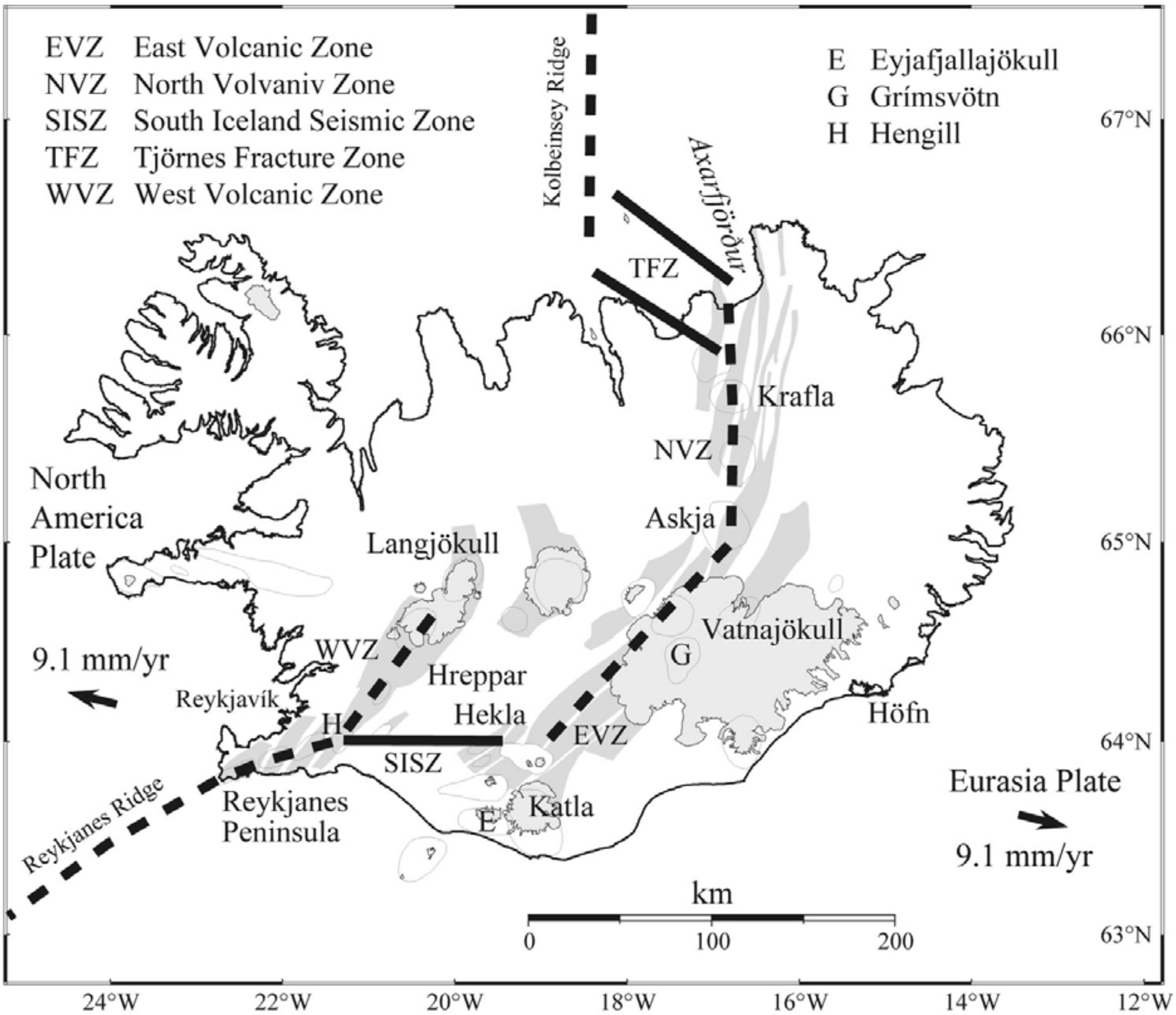

Figure 1: Index map of Iceland showing place names, lithospere plates, and plate boundary structures (after Einarsson and Sæmundsson 1987). Divergent plate boundaries are shown with fat, broken lines, seismic zones and transforms with fat lines. Fissure swarms are shown as gray stripes, central volcanoes with closed, thin lines. The divergent plate boundary zones are the Reykjanes Peninsula, the Western Volcanic Zone (WVZ), the Eastern Volcanic Zone (EVZ) and the Northern Volcanic Zone (NVZ). The transform zones are the South Iceland Seismic Zone (SISZ) and the Tjörnes Fracture Zone (TFZ). The central volcanoes are Hengill and Hrómundartindur (H), Hekla, Eyjafjallajökull (E), Katla, Grímsvötn (G), Askja, and Krafla. Vatnajökull and Langjökull are ice caps.

The long-term aim of crustal deformation measurements is to obtain a complete, four-dimensional picture of the deformation field, i. e. both in space and time. As a step toward this goal the combined use of network GPS, continuous GPS and InSAR has turned out to be quite successful. In areas where InSAR works, it gives an areal coverage of the deformation field in one dimension, i.e. in the direction of line-of-sight to the SAR satellite. A network of GPS-points in the same area allows the determination of the three-dimensional displacement vector, but only at discrete points. The gaps between points can be bridged using InSAR. Continuously recording GPS-stations give the time-history of displacements at a few discrete points, which is very important when interpreting the processes responsible for the time changes in the displacement field. Facilities for the use of these methods have been systematically built up in Iceland in the last decades. The geodetic reference network, ISNET, of about 120 GPS-points covers the country evenly and is resurveyed about every decade (Geirsson et al. 2005). Denser networks are available in the plate boundary zones and around most of the active volcanoes, in most cases with mesh-size of about 5-10 km. Conditions for InSAR analysis of crustal movements are rather favourable in Iceland as lava surfaces and rock outcrops in combination with low vegetation provide stable ground reflectors. Snow cover limits the use to summer time image acquisition.

3 Case Histories of Geodynamic Processes

Plate movements

The mid-Atlantic plate boundary runs through Iceland, from the tip of the Reykjanes Peninsula in the SW to the Axarfjörður Bay in the NE (Figure 1). Western Iceland is thus mostly a part of the North America Plate and Eastern Iceland is sitting firmly on the Eurasia Plate. The pole of relative rotation between these plates is located in NE-Siberia at 62.4°N and 135.8°E, and the relative rotation speed is 0.21° per million years according to the Nuvel-1A model of plate motions (DeMets et al. 1994). Holding the North America Plate fixed this gives a plate velocity vector of 18.2 mm/year in a direction of 105° for Central Iceland, slightly faster and more easterly for South Iceland, slightly slower and more southerly for North Iceland. This velocity is valid for the last few millions of years, the time scale of the magnetic and structural data used to constrain the Nuvel-1A model. GPS-data from the continuously recording stations (Figure 2) give results that are consistent with the Nuvel-1A velocity (Geirsson et al., submitted 2005), also preliminary results of measurements of the country-wide ISNET network in 1993 and 2004 (Geirsson et al. 2005). This demonstrates that the plate movements are consistent on time scales ranging between years and million years. An example of the constant rate of movements is shown in Figure 2, the time series 1999-2005 for the continuous GPS-station at Höfn in SE-Iceland. When the annual cycle of uncertain origin and the co-seismic effect of the June 2000 earthquakes in South Iceland on the reference station in Reykjavík have been removed, the graphs shows virtually straight lines. The slopes of the lines give an eastward component of 22.1 mm/year and a soutward rate of 3.9 mm/year. The vector therefore has a magnitude of 22.4 mm/year and a directon of 100° in reasonable agreement with the Nuvel-1A values.

Plate boundary deformation

Between the major plates there is a zone of deformation where the crustal movements are different from that of the plates. The width of this deformation zone is somewhat variable. In Northern Iceland it is about 100 km wide and coincides more or less with the zone of Holocene volcanism and fissuring. In Southern Iceland the plate boundary has two branches and a micro-plate can be defined between them, the Hreppar Microplate. The southern boundary of the Hreppar Plate is marked by the South Iceland Seismic Zone where large, strike-slip earthquakes occur.

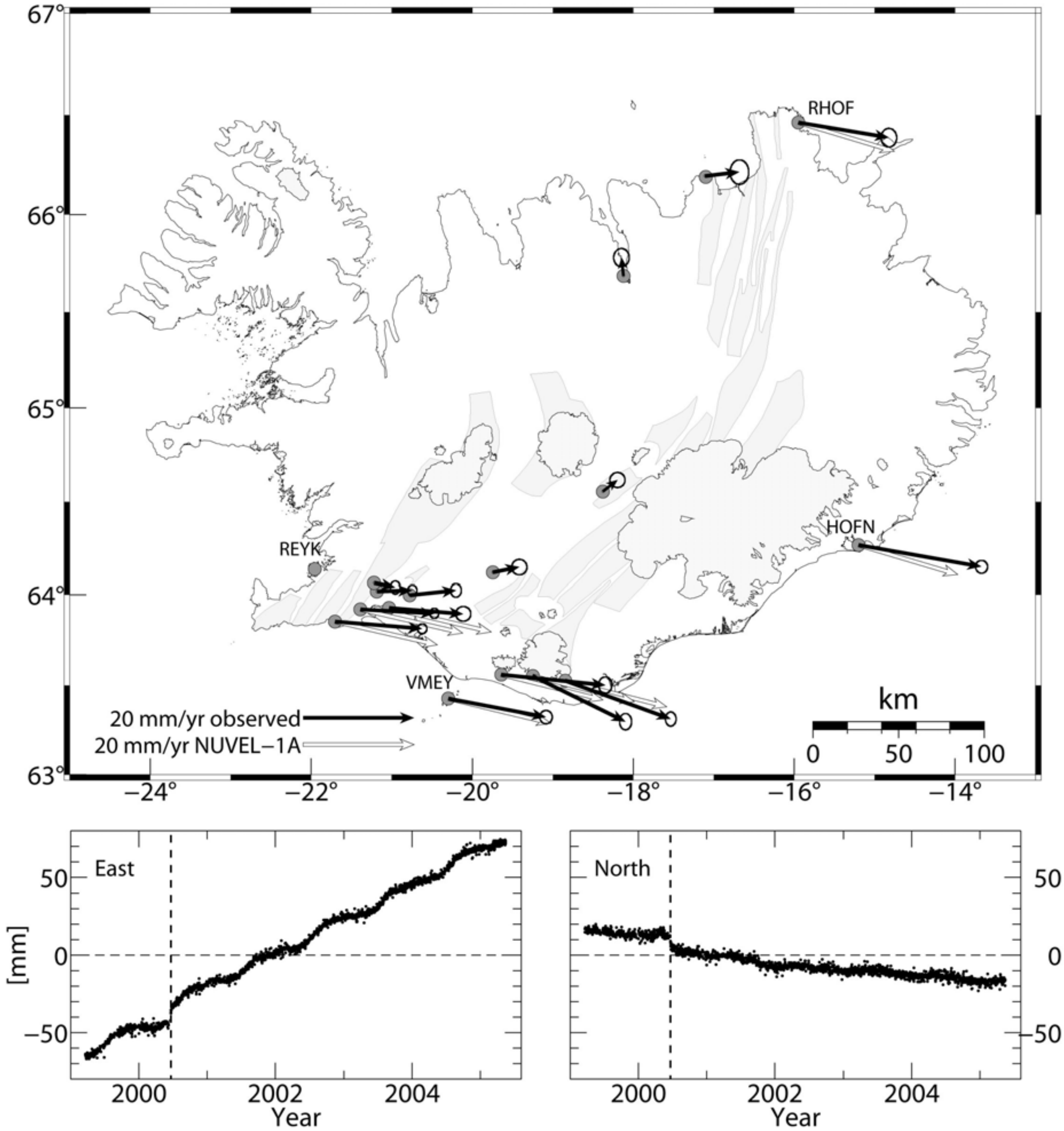

Figure 2: Results of continuous GPS-stations shown as average horizontal displacement vectors with respect to a reference station in Reykjavík. Also shown are the time-series for horizontal components of the station at Höfn in SE-Iceland. Note the good fit to the Nuvel-1A model for the stations located on the Eurasia Plate, particularly RHOF and VMEY.

It has been a matter of considerable debate how the plate movements in South Iceland are partitioned between the two parallel rift zones, the Western Volcanic Zone and the Eastern Volcanic Zone. It is generally assumed that the two zones are the expression of a ridge jump, i.e. that the WVZ is a dying rift that is being replaced by the currently much more active EVZ (e.g. Einarsson 1991a). The question is whether the ridge jump occurs by rift propagation, i.e. the EVZ propagating towards the SW while the WVZ recedes, or by activity alternating between the rifts (Sigmundsson et al. 1995) and the whole WVZ gradually becoming less active. The lack of evidence for rotated structures within the Hreppar Plate seems to support the latter hypothesis. Recent GPS-surveys, however, appear to support rotational movements of the Hreppar Plate (La Femina et al. 2005), which is in favour of the propagating rift hypothesis. The measurements suggest that near the Hengill triple junction as much as 35% of the plate movements is taken up by the WVZ. This proportion dies out towards the NE and is less than 10% in the Langjökull region. This must indicate a counter-clockwise rotation of the Hreppar Microplate, considering the lack of evidence for significant internal deformation of that plate. The plate boundary deformation zone accumulates strain during time intervals between significant failure events such as rifting episodes or larger earthquakes. Such gradual accumulation has been documented for the EVZ by Jónsson et al. (1997), across the SISZ (Sigmundsson et al. 1995, Alex et al. 1999, Perlt and Heinert 2000), and along the Reykjanes Peninsula oblique rift by Sturkell et al. (1994) and Hreinsdóttir et al. (2001).

Rifting in the divergent zones

The Krafla rifting episode provided a dramatic demonstration of crustal deformation along a divergent plate boundary where strain had been accumulating for more than two centuries. It was a sequence of magmatic and tectonic events along the plate boundary in N-Iceland, beginning in 1974 and lasting until 1989. It was accompanied by the largest earthquake sequence so far recorded along the divergent plate boundaries of the Atlantic (Einarsson 1986). The events took place mainly within the Krafla volcanic system between latitudes of 65º34'N and 66º18'N (Figure 1). The volcanic system consists of a central volcano with associated fissure swarms that extend along the plate boundary perpendicular to the plate separation vector. During most of the episode, magma apparently ascended from depth and accumulated in the magma chamber at about 3 km depth beneath the central volcano (e.g. Tryggvason 1980; Ewart et al. 1991). The inflation periods were punctuated by sudden deflation events lasting from several hours to 3 months when the walls of the chamber were breached and magma was injected laterally into the adjacent fissure swarm where subsequently large-scale rifting took place. Rifting, fissuring and graben subsidence took place in the fissure swarm but the flanks were uplifted and compressed laterally (e.g. Sigurdsson 1980, Torge and Kanngiesser 1980, Kanngiesser 1983). A total of about 20 discrete rifting events were identified, each one affecting only a portion of the fissure system (Björnsson et al. 1979; Tryggvason 1980; Einarsson 1991a,b). Subsidence within the Krafla caldera was concurrent with rifting and widening of segments of the Krafla fissure swarm (Björnsson et al. 1977, 1979; Tryggvason 1980, 1984, 1994). Early events were primarily associated with subsurface movements of magma and little or no lava extrusion. Later in the sequence most of the magma removed from the magma chamber reached the surface in fissure eruptions lasting from 5 to 14 days. Maximum cumulative extension of 8-9 m was measured across the fissure swarm slightly north of the Krafla volcano. A large segment of the plate boundary was affected by the Krafla events, extending at least 20 km south of Krafla and 70 km north of the volcano, at least to the rift-transform intersection in Axarfjörður (Figure 1).

Co-seismic displacements

In the transform zones of South and North Iceland the plate boundary is sub-parallel to the plate movement vector. The accumulated strain in these zones is released in large, strike-slip earthquakes, as large as magnitude 7, that take place at intervals of decades to centuries (Einarsson 1991a). The South Iceland Seismic Zone was hit by a series of earthquakes in June 2000, two of which caused considerable damage (Stefánsson et al. 2003). The earthquakes follow a pattern of large historic earthquakes in this zone where sequences of large events have occurred at intervals ranging from 45 to 112 years (Einarsson et al. 1981). The sequence began on June 17 with a magnitude 6.5 event in the eastern part of the zone. This immediately triggered a flurry of activity along at least a 90 km-long stretch of the plate boundary to the west. This activity included three events with magnitudes larger than 5 on the Reykjanes Peninsula oblique rift (Clifton et al. 2003, Pagli et al. 2003, Árnadóttir et al. 2004). A second mainshock, also of magnitude 6.5, occurred about 20 km west of the first one on June 21.

The mainshocks of the sequence occurred on N-S striking faults, transverse to the zone itself. The sense of faulting was right-lateral strike-slip conforming to the model of "bookshelf faulting" for the South Iceland Seismic Zone (e.g. Einarsson et al. 1981). According to the model the left-lateral transform motion across the zone is accomplished by right-lateral motion along numerous parallel transverse faults and rotation of the blocks between them. It was furthermore demonstrated that bookshelf faulting continues to the west, along the Reykjanes Peninsula oblique rift (Árnadóttir et al. 2004). One of the events of the sequence has the characteristics of a "slow earthquake", i.e. the radiation of seismic waves is comparatively weak for the amount of faulting observed by InSAR or GPS.

The two largest events of the sequence occurred on pre-existing faults and were accompanied by surface ruptures consisting primarily of en echelon tension gashes and push-up structures (Clifton and Einarsson 2005). The main zones of rupture were about 15 km long, and coincided with the epicentral distributions of aftershocks. Fault displacements were of the order of 0.1-1 m at the surface. Faulting along conjugate, left-lateral strike-slip faults also occurred, but was less pronounced than that of the main rupture zones.

The co-seismic displacement field of the sequence of earthquakes in the SISZ was captured by InSAR and GPS-measurements (Pedersen et al. 2001, Árnadóttir et al. 2001). The geodetic data (Figure 3) were used to invert for the optimal fault geometries and slip distribution for the two main shocks (Pedersen et al. 2003). According to these models faulting extends from the surface to a depth of 10 km in both events. Maximum displacements are 2.6 m and 2.9 m, respectively.

Post-rifting and post-seismic displacements

The rifting during the Krafla volcano-tectonic events and the co-seismic displacements during the South Iceland earthquakes of June 2000 can be modeled as elastic reaction to the failure of the elastic part of the crust under stress that had accumulated in the plate boundary region during the previous decades and centuries. The elastic part of the crust appears to be about 10-15 km thick and lies on top of a viscous or visco-elastic material that comprises the lower crust and the upper mantle. The sudden stress change in the elastic layer leads to increased stress and viscous reaction in the underlying layer which again induces movements in the elastic surface layer. These movements decay with time in an exponential way. The time-constant of this decay is directly dependent on the viscosity of the underlying layer. Such post-rifting movements have been measured after the Krafla events (Jahn et al. 1994, Foulger et al. 1992, Heki et al. 1993, Völksen and Seeber 1998, Völksen 2000). Depending on the assumed model parameters such as thickness of the elastic and viscous layers, viscosity values in the range 0.3-30 x 10^{18} Pa s are obtained.

Post-seismic deformation was observed in the SISZ on two spatio-temporal scales following the June 2000 earthquake sequence. A rapidly decaying deformation transient, localized around the two main shock faults, was captured by several radar interferograms. This signal has been explained by poro-elastic rebound due to post-earthquake pore-pressure changes (Jónsson et al. 2003). In contrast, the year-scale deformation observed by campaign and continuous GPS can be explained by either afterslip at 8-14 km depth or visco-elastic relaxation of the lower crust and upper mantle in response to the co-seismic stress changes, suggesting viscosities of 0.3-1 x 10^{19} Pa s (Árnadóttir et al. 2005).

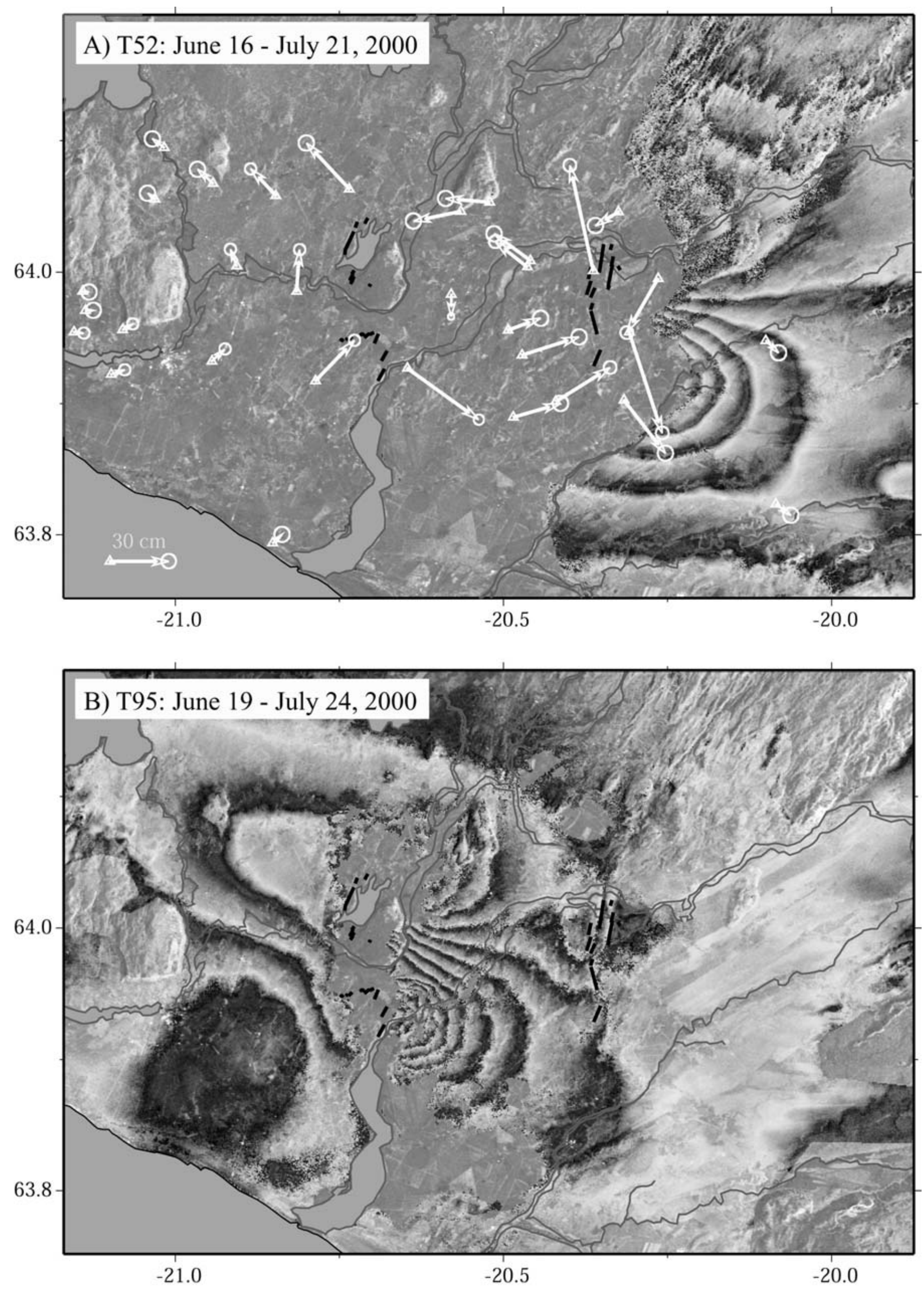

Figure 3: Horizontal GPS displacements and co-seismic interferograms draped on a radar amplitude image, showing ground deformation associated with the earthquakes in 2000, on June 17 and June 21. One interferometric fringe corresponds to 2.83 cm of range change in the line-of-sight direction. Incoherent areas are masked. The difference in areal InSAR coverage between Figure A and B is due to utilization of data from different track frames. Mapped ground ruptures (Clifton and Einarsson 2005) are shown in black, and rivers and seashore are outlined in grey. After Pedersen et al. (2003).

Volcano inflation and deflation

Many of the volcanoes in Iceland appear to be underlain by shallow-level magma chambers or semi-permanent bodies of molten material (Sturkell et al. 2006). Activity in the volcanoes is associated with pressure fluctuations in these bodies. Pressure change in a magma chamber leads to a characteristic displacement field at the surface above and around the chamber, frequently described by the so-called Mogi-model. The model was originally derived for a point source of pressure or a small sphere of pressure at a specified depth in a half-space. Later it was shown (McTigue 1987) that the model gave a good approximation for the field as long as $(\mathbf{a}/\mathbf{d})^5 \ll 1$, where **a** is the radius of the chamber and **d** is the depth below the surface. This implies that the deformation field is not very sensitive to the shape or the size of the chamber.

The deformation fields around volcanoes can be determined by many geodetic methods, e.g. levelling, tilt, gravity, GPS-geodesy and InSAR, and the results used to derive the depth to the magma chamber. An example is shown in Figure 4. Repeated GPS-surveys around the Katla volcano in South Iceland reveal pressure increase beneath the volcano. Interpreting this inflation in terms of the Mogi-model gives a depth of 5 km to the magma chamber of increasing pressure. The pressure center is located within the subglacial caldera of Katla. The inflation appears to have started in 1999 (Sturkell et al. 2003b) and was still continuing in 2005. This inflation is responsible for the deviation from the plate spreading vector apparent in the time series of the CGPS-stations south of Katla (Figure 2).

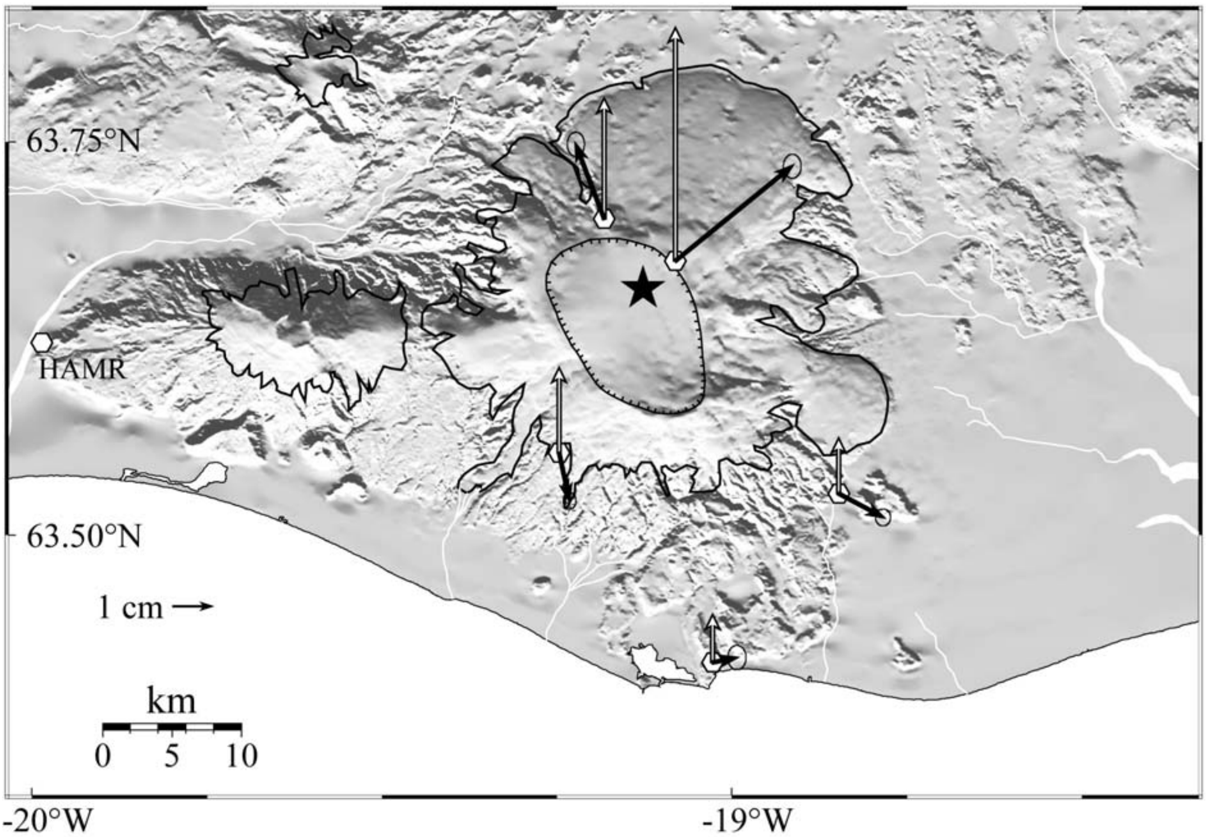

Figure 4: Inflation of the Katla volcano in South Iceland as shown by GPS measurements in 2000-2003. Black arrows show horizontal displacements, white arrows show vertical displacements. GPS-point near the volcano move up and outwards from a center in the northern part of the caldera, marked with a star, indicating a center of pressure increase at about 5 km depth. From Sturkell et al. (2006).

Many opportunities to determine displacement fields around a magma chamber were provided by the Krafla volcano-tectonic episode in 1975-1989, both inflationary and deflationary fields. The magma chamber appeared to be stationary throughout the period. Different inflation and deflation episodes gave depths in a narrow range around 3 km (e.g. Björnsson et al. 1979, Tryggvason 1980, 1994a, Ewart et al. 1991, Árnadóttir et al. 1998).

A large increase in seismic activity near the Hengill triple junction in SW-Iceland in 1994-1998 was associated with uplift of up to 10 centimeters over a wide area as shown by GPS-measurements, repeated leveling and InSAR. The uplift was interpreted as the result of magma injection into the crust at about 7 km depth below the Hrómundartindur volcanic system (Sigmundsson et al. 1997, Feigl et al. 2000). The inflation was associated with surface fracturing (Clifton et al. 2002), damaging earthquakes (about magnitude 5), and increased thermal activity, but it stopped without an eruption.

The Grímsvötn volcano is located near the center of the Iceland hotspot and is almost totally covered by the Vatnajökull ice cap, which limits the use of geodetic methods to monitor its activity. Repeated GPS-measurements on the only useable nunatak, Grímsfjall on the caldera rim, have given very characteristic time changes that can be interpreted with the help of the Mogi-model (Sturkell et al. 2003a, 2005a). The volcano deflated during the eruptions of 1998 and 2004, and inflated in the time period between the eruptions. The inflation rate was used to give a long-term forecast for the 2004 eruption (Sturkell et al. 2003a, Vogfjörð et al. 2005). Similarly, the present reinflation rate of the volcano indicates that an eruption is to be expected within a few years.

The Hekla volcano has erupted a few times within the time period of precise geodesy, i.e. 1970, 1980-81, 1991 and 2000. Several attempts have been made to determine a depth to a magma chamber feeding the eruptions but the results are vague and partly contradictory. Tilt measurements have indicated an inflation center west of the volcano (Tryggvason 1994b), EDM-trilateration in association with the 1980-81 activity gave a chamber at 8 km depth (Kjartansson and Grönvold 1983), and GPS-measurements during the 1991 eruption gave a rather uncertain depth between 2 and 11 km (Sigmundsson et al. 1992). A joint interpretation of GPS-data, InSAR, tilt, and volumetric strain changes associated with the 2000 eruption indicates a broad deformation field from a rather deep-seated magma chamber at 11 km depth, disturbed by irregular, local sources, most likely due to loading of the surface by lava flows (Sturkell et al. 2005). Current tilt changes indicate that the volcano is re-inflating and that it may have reached the pre-eruption inflation level already.

The Askja volcano in the Northern Volcanic Zone has been the target of many geodetic studies since it latest eruption in 1961. Eysteinn Tryggvason, the pioneer of volcano geodesy in Iceland, installed a levelling profile in 1966-68 which has been remeasured and extended many times since then. During the first years the volcano was inflating, but about 1973 the trend was reversed to deflation. The deflation was rapid in the beginning but the rate has been decreasing in an exponential manner with a decay constant of 39 years (Sturkell et al., submitted 2004). The geodetic data have been modelled and interpreted in terms of a single Mogi-type pressure source located close to the centre of the Askja main caldera (Tryggvason 1989; Rymer and Tryggvason 1993; Sturkell and Sigmundsson 2000). All these authors placed the point source at 1.5 to 3.5 km depth. This model accounts for most of the observed displacements in the main caldera and its immediate vicinity. At a greater distance, however, displacements observed with GPS do not show the same good fit. A more elaborate model is presented by Sturkell et al. (submitted 2004) who invoke two Mogi sources to account for the far field displacements. This model was also applied to the results of repeated micro-gravity studies for the period 1988-2003. A sub-surface mass decrease of 1.6×10^{11} kg is derived (de Zeeuw et al., submitted 2004) indicating that magma drainage is an important contributor to the sub-surface mass decrease. The geometry of the deeper pressure source is not well resolved but it is suggested that it is at 16 km depth and is elongated along the axis of the fissure swarm. This is confirmed in a general way by a study by Pagli et al. (2005), who add the constraints of InSAR data and use ellipsoidal models in an inversion for depth and chamber size (Figure 5). They explain most of the displacement field by an ellipsoidal pressure source at about 3 km depth, but an elongated zone of subsidence along the associated fissure swarm is ascribed to a deeper source of decreasing pressure.

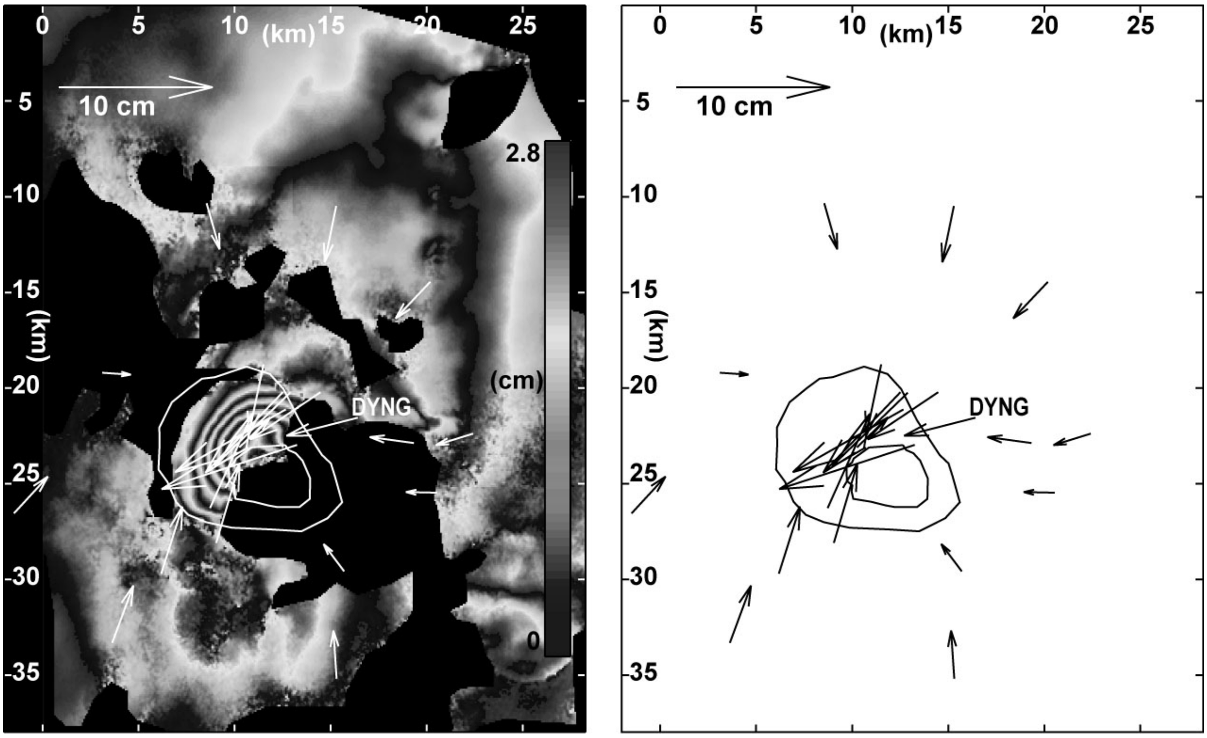

Figure 5: Displacements in the surrounding of the Askja volcano measured by InSAR (left panel) and GPS (both panels). Nested calderas of Askja are shown with thin lines. InSAR fringes in the caldera region show deflation from a shallow source beneath the caldera. Elongated fringes outside the caldera are extended along the fissure swarm of the volcano and show subsidence, possibly due to extension across the plate boundary. GPS-vectors show horizontal contraction towards the volcano. After Pagli et al. (2005).

Intrusions

Two separate intrusion events occurred beneath the Eyjafjallajökull volcano in South Iceland, the first one in 1994 and the second one in 1999. Both events began with increased earthquake activity beneath the NE flank of the volcano followed by inflation centered on the S flank. Ground deformation indicating inflation in 1994 was initially discovered by tilt and GPS-geodesy and appears to have occurred mostly during the peak of the seismic activity (Sturkell et al. 2003b). Inflation was further confirmed by InSAR, with a center beneath the S-flank at 4.5 km depth (Pedersen and Sigmundsson 2004). In 1999 the InSAR data showed that the inflation bulge was displaced slightly S with respect to the 1994 inflation bulge (Pedersen and Sigmundsson 2005), and a deeper source is required to fit the data (Figure 6). The InSAR data favour a sill-shaped intrusive body for both events. Estimated volume of intruded magma in 1994 is about 17×10^6 m^3 and in 1999 it is 30×10^6 m^3.

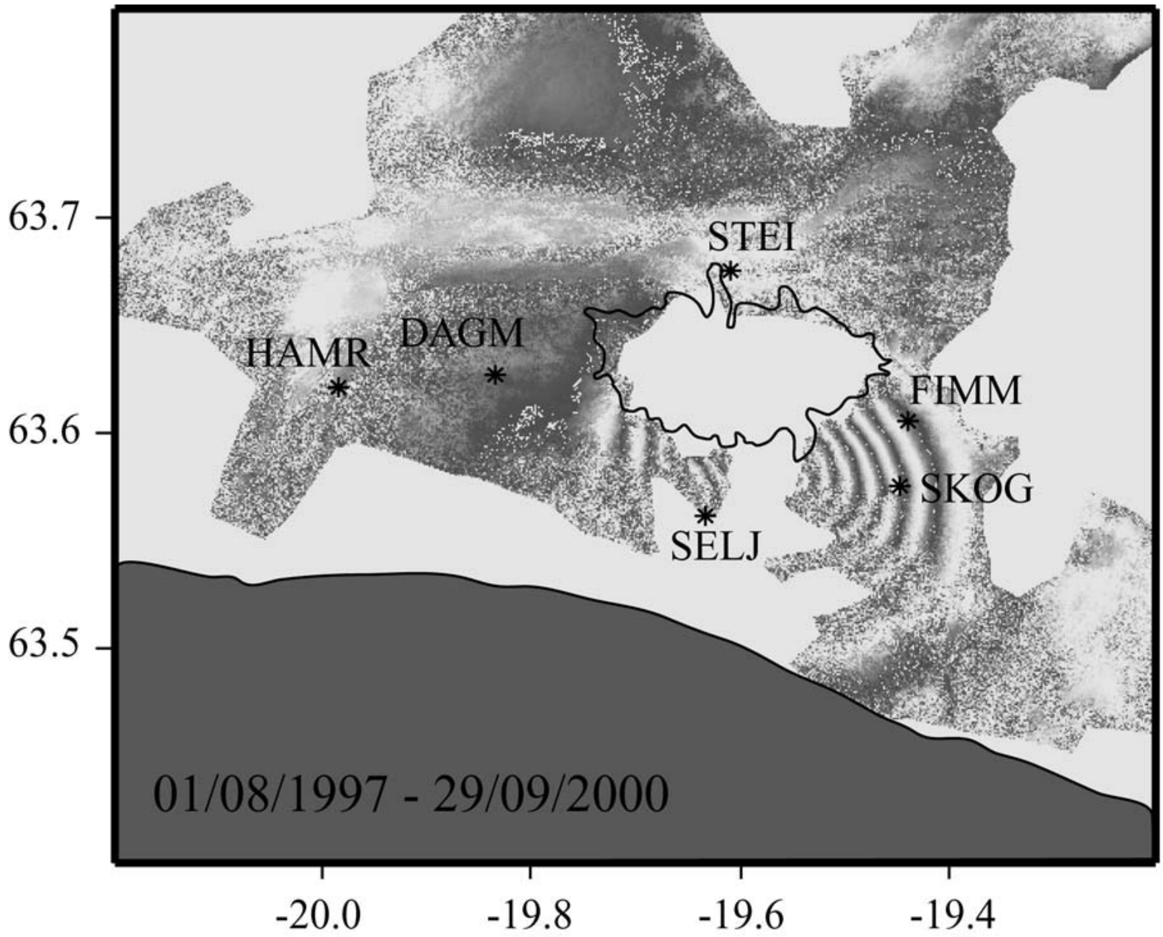

Figure 6: Ground displacements recorded by InSAR in the vicinity of the Eyjafjallajökull volcano between August 01, 1997 and Sept. 29, 2000 including the intrusion event of 1999. One interferometric fringe corresponds to 2.83 cm of range change in the line-of-sight direction. Incoherent areas are masked. The outline of the Eyjafjallajökull glacier covering the summit region of the volcano is shown with a black line, and locations of GPS stations in the area are shown with stars. After Pedersen and Sigmundsson (2005).

Glacial isostasy and uplift around Vatnajökull

The glaciers in Iceland provide variable loads on the crust that result in isostatic adjustments. By measuring the crustal response to the load changes one can estimate the thickness of the elastic upper crust and the viscosity in the underlying visco-elastic layer. Deglaciation in Iceland at the end of the Weichselian glaciation about 10000 years BP, was associated with rapid glacial rebound, being completed in only about 1000 years in coastal areas. This exceptionally fast postglacial rebound has been modeled to argue for viscosity under Iceland of the order of 10^{19} Pa s or less (Sigmundsson 1991). The low viscosity results in a rapid response of the Earth to contemporary changes in ice volume. In the last century, the ice volume of the Vatnajökull glacier has significantly decreased. Ongoing uplift around Vatnajökull is reported by several geodetic studies. Lake leveling measurements at Lake Langisjór at Western edge of Vatnajökull were performed in 1959-1991 (Sigmundsson and Einarsson 1992). Measurements show uplift rate of about 4 mm/yr between benchmarks spaced 15 km perpendicular to the ice edge. In 1991 a GPS network of ten points was first measured around the Southeastern edge of Vatnajökull (Einarsson et al. 1996). Gravity measurements were also conducted at all GPS stations and annually repeated until 2000, except in 1994 (Jacoby et al. 2001). Gravity changes between 1991 and 2000 for points close to the ice cap are consistent with uplift rates up to 20 mm/year. In 1992, 1996 and in 1999 the original 1991 GPS network was remeasured and eleven additional points were included in the network in 1992 (Sjöberg et al. 2000 and 2004). The uplift rate 1992-1999 was estimated to be about 5-19 mm/yr, decaying radially from the center of the ice cap (Sjöberg et al 2004).

Thoma and Wolf (2001) used the lake level measurements of Lake Langisjór from 1959 to 1991 and GPS measurements from 1992 to 1996 (Sjöberg et al. 2000) to constrain the rheology in Iceland. The authors use a compressible, self-gravitating, spherical Earth model with Maxwell viscoelasticity and an elliptic ice load. They consider two different ice thinning models. Modeling results suggest a lower crust/upper mantle viscosity between 7×10^{16}-3×10^{18} Pa s and a thickness of the elastic crust between 10-20 km. Similar results were reported by Sjöberg et al. (2004) who processed GPS measurements in 1992, 1996 and 1999. They conclude that the vertical GPS velocities can be fit by assuming an elastic thickness of the crust on the order of 10-20 km and a viscosity perhaps as low as 1×10^{17} Pa s.

Jacoby et al. (2001) measured gravity changes between 1991 and 2000 and compared them to the model by Sigmundsson and Einarsson (1992). Results suggest a lower crust/upper mantle viscosity on the order of 10^{18} Pa s and a thickness of the elastic crust of about 10 km.

In 1996 and in 2003 a GPS network of 15 points was measured around the south edge of Vatnajökull. GPS vertical velocities around the ice cap vary between 7-25 mm/yr (Figure 7). These GPS and lake leveling measurements have been modeled using the Finite Element Method (FEM). Results indicate a thickness of the elastic crust of 10-20 km and a viscosity of the lower crust/upper mantle of 3-8 $\times10^{18}$ Pa s (Pagli et al., in preparation). Knowledge of the Earth structure allows us to predict uplift around Vatnajökull in the next decades.

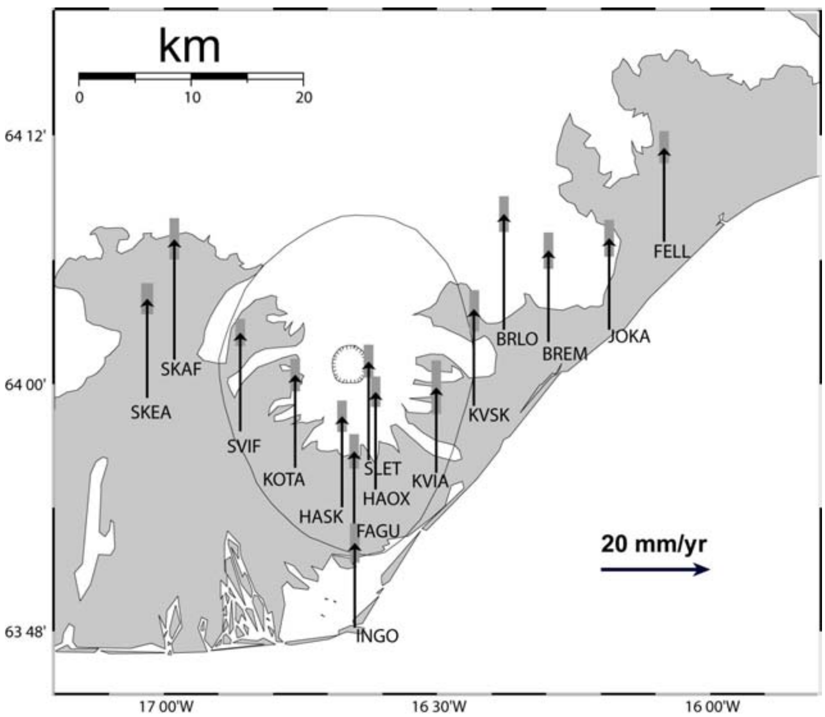

Figure 7: Vertical displacements of GPS-points near the edge of Vatnajökull glacier during 1996-2003 (in ITRF00 reference frame). The uplift rate is highest near the glacier and decreases with increasing distance from its edge. From Pagli et al. (in prep. 2005).

Mass movements suggested in the roots of the Iceland hotspot

Preliminary results of the 2004 re-measurements of the ISNET geodetic net of Iceland suggest that most points in the interior of Iceland are being uplifted at a rate of 10-20 mm/year (Geirsson et al. 2005). Part of this can be explained by isostatic uplift following the recent reduction in ice load of the present glaciers. The uplift appears to be too widespread, however, for all of it to be modeled by this process. The residual signal may have to be explained by additional processes taking place in the lower crust or the upper mantle. In fact, mass movements and pressure connection at these levels have been suggested to explain apparent correlation in time between activity at different volcanoes (e.g. Einarsson 1991, Tryggvason 1989, Sturkell et al. 2006). For further modeling of deep mass movements the importance of precise gravity measurements becomes apparent. Absolute gravity measurements were initiated in Iceland in 1987-1988 (Torge et al. 1992) at 5 points, one in Reykjavík and 4 in North Iceland. Additional measurements were done in 1997 by IfAG (now Bundesamt für Kartographie und Geodäsie) at 7 points, including Reykjavík and Höfn, and 2 points in the interior of Iceland. Further measurements were done at Reykjavík and Höfn by the Finnish Geodetic Institute. Repetition of these measurements and future expansion of this absolute gravimetry network is likely to contribute to better understanding of the dynamics of the Iceland hotspot.

4 Conclusions

Iceland has been, and still is, an important test ground for geodynamic models, both on a global and local scale. New geodetic techniques have been used and tested in Iceland for this purpose for decades.

The basic assumption of the plate tectonics theory has been verified, i.e. that the surface of the Earth is divided into plates with insignificant internal deformation separated by plate boundary deformation zones. It has been shown that the global model of plate movements applies well in Iceland. Measured plate movements of the last few years conform with the globally determined plate movements of the last few million years. Iceland sits on two of the major plates, the Eurasia and North America Plates. A small plate fragment, the Hreppar Microplate, is defined between the two volcanic rift zones in South Iceland. Further microplates or crustal blocks within the Tjörnes Fracture Zone remain to be defined.

Crustal dynamics models with an elastic plate on top of a viscous layer or half-space have been successfully used to explain crustal movements measured in conjunction with rifting events in the divergent plate boundary zones and earthquakes in the transform zones, both the immediate elastic response and the post-event movements. The thickness of the elastic plate is in the range 10-15 km and the viscosity in the underlying layer is in the range 0.3-30 x 10^{18} Pa s. Similar models have been used in interpreting glacio-isostatic movements observed around the Vatnajökull ice cap in response to reduced ice load in the last century. It is important to maintain some of the established time series of movements in order to refine these models, e.g. by introducing further layering.

Crustal deformation measurements provide some of the most useful data for the monitoring of active volcanoes. Pressure changes in the magma systems of volcanoes are reflected by crustal movements that can be measured. Mogi-type models, i.e. pressure point sources in an elastic half-space, have been very successfully applied to deformation fields measured around active volcanoes. These models are relatively insensitive to shape and size of the magma chambers where the pressure changes occur. They do, however, reveal whether the magma pressure is increasing or decreasing, and give an indication of the depth to the magma chamber. At the present time the magma pressure is increasing beneath the three most active Icelandic volcanoes, Grímsvötn, Hekla and Katla. Pressure is decreasing in the shallow-level magma chambers of Askja and Krafla.

Geodetic data are becoming available that have better continuity in space (InSAR) and in time (CGPS). These data demand refined models for their interpretation. Fault models with non-uniform slip distributions and magma chambers with ellipsoidal shapes are applied with considerable success in inversion algorithms.

A multitude of geodynamic problems are still unsolved and exciting geodetic experiments can be carried out in Iceland. These can benefit from ever more sophisticated geodetic measurements, with improved spatial and temporal resolution. Further densification of the network of continuous GPS stations may be particularly rewarding, augmented by broad-band seismic stations and absolute gravimetry. Multi-national cooperative geodetic projects have turned out to be very fruitful, such as collaborative work with German geodesists. Continued and improved German-Iceland collaboration in geophysical geodesy may therefore be a fruitful way to improve the understanding of Iceland geodynamics.

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