RESEARCH ARTICLE

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1983–2003 decaying rate of deflation at Askja caldera: Pressure decrease in an extensive magma plumbing system at a spreading plate boundary

Abstract New deformation data from the Askja volcano, Iceland, show that the volcano's caldera has been deflating continuously for over 20 years, and confirm that the rate of subsidence is slowing down. The decay in subsidence rate can be fitted with a function of the form $e^{-t/\tau}$, where τ is 39 years. Reanalysis of GPS data from 1993–1998 show that these data can be fitted with a model calling for two Mogi point sources, one shallow, and another one much deeper (16.2 km depth). Pressure decrease occurs in both sources. The deeper source is responsible for observed horizontal contraction towards Askja at distances that cannot be explained by the shallower source. Plate spreading of 19 mm/year distributed evenly over about 100-km-wide zone is also favoured by the data.

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Introduction

Askja volcano is situated in Mt. Dyngjufjöll at the divergent plate boundary in North Iceland (Fig. 1). It is transected by a 10–15 km wide and more than 100 km long fissure swarm. Askja hosts three calderas, the largest one of which is referred to as the main Askja caldera (Fig. 2). It was formed in the early Holocene and is 8 km in diameter (Sigvaldason 1979). A less pronounced older caldera, almost completely filled with lava, is located along the northern edge of the main caldera (Sigvaldason 1979). The youngest caldera, 4.5 km in diameter, was formed after a Plinian eruption in 1875 and is today occupied by Lake Öskjuvatn (Fig. 2). That event was a part of a rifting episode 1874–1876 with main fissure eruptions occurring in the northern part of the Askja fissure swarm in the Sveinagjá graben, ∼40 km north of the Askja volcanic center. The most recent eruption in Askja occurred 1961 when an 800-m long eruptive fissure opened along the northern rim of the main caldera (Fig. 2).

In the study period, the Askja area was not well covered by seismic networks, with nearest station at ∼45 km distance (Fig. 1). Seismological interpretation of available earthquake data in the area is therefore limited. Figure 3 shows the locations of 360 earthquakes from January 1997 to December 31, 2004 in the Askja area. The earthquakes occur in three main areas; the eastern part of the caldera, to the north-west of the caldera and to the north-east of the caldera. The area north-east of the caldera is the most active, with 249 earthquakes in the period, with most of them occurring in swarms.

Extensive geodetic studies have been conducted at the Askja caldera. A key set of observations is frequently repeated levelling measurements. Those measurements were started in 1966, and they provide the longest time-series of deformation at an Icelandic volcano. Observations up to 1998 were summarized by Sturkell and Sigmundsson (2000). The data set also includes results of GPS-measurements from 1993 to 1998 in a local network of more than 20 stations (Fig. 2).
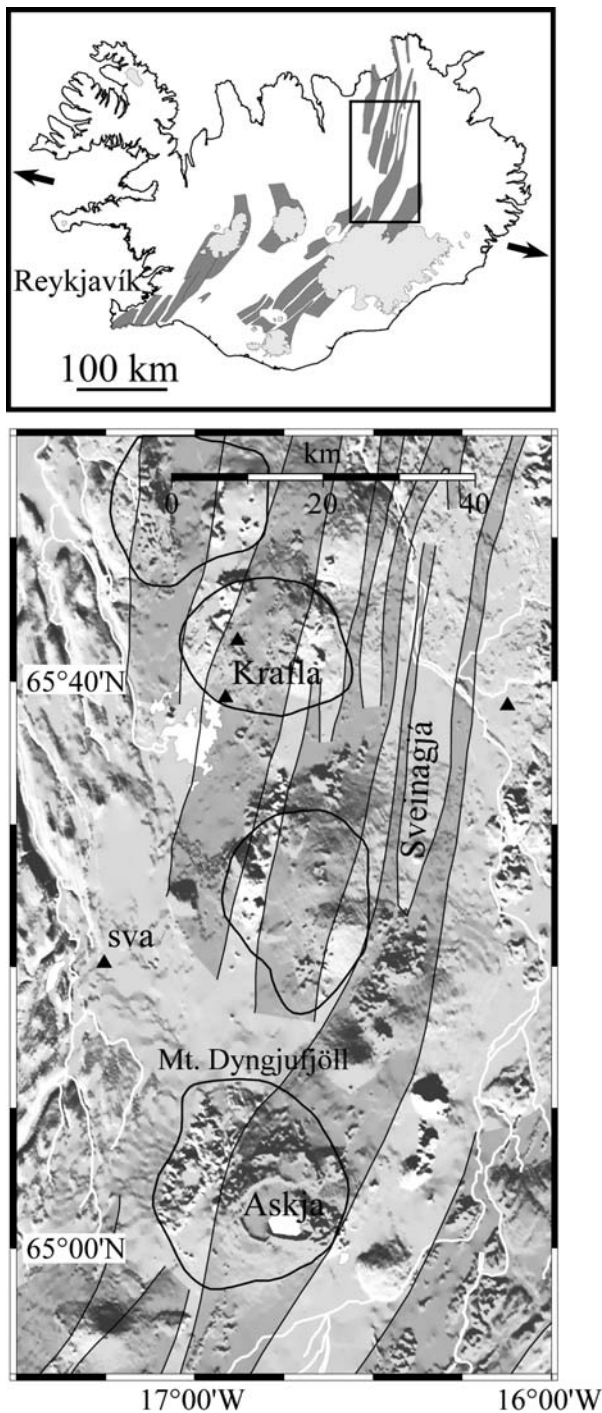
Fig. 1 Location of the Askja and Krafla volcanoes at the divergent plate boundary in North Iceland. Seismic stations are shown with triangles. The Svartárkot (sva) station is located at a distance of ∼45 km. The eruptive fissure Sveinagjá is located 40–70 km north of Askja

Fig. 2 Askja volcano, its calderas, the location of the 1961 eruptive fissure and shaded topography. The *thick dashed line* marks the main caldera. Lake Öskjuvatn fills the 1875 caldera. The location of the Askja levelling profile is shown in the enlarged box. GPS-stations are shown by hexagons. Names of stations discussed in the text are given

New measurements in the period 1998–2003 include yearly levelling along a 1700-m long segment of the levelling profile, and GPS observations at a few key sites (Fig. 2). These measurements are presented here and they clearly show that the caldera has continued to subside in this time p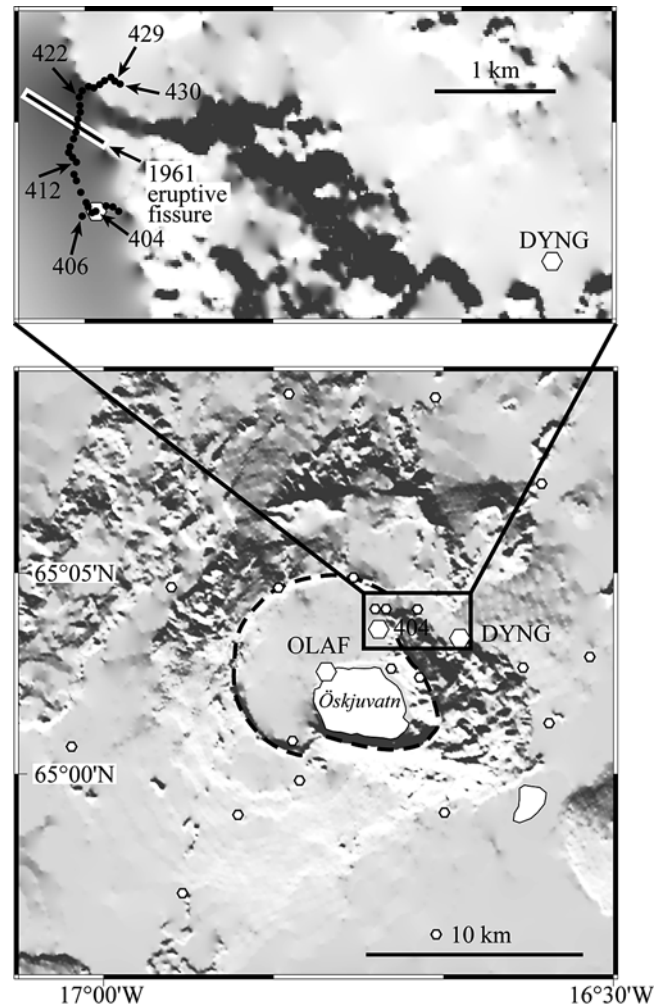eriod. The long time series from levelling obser-vations now confirms that the rate of subsidence is slowing down. Complementary observations include interferomet-ric analysis of synthetic aperture radar images from radar satellites (InSAR) that show clearly the ongoing subsidence at Askja. (Pagli et al. 2006). Repeated gravity measure-ments have also been used to constrain the nature of caldera subsidence (Rymer and Tryggvason 1993; de Zeeuw-van Dalfsen et al. 2004).

Much of the observed displacement can be modelled by a pressure decrease in a 2.8-km deep Mogi point source (Mogi 1958) in the centre of the main caldera. However, stations more than 8 km from the caldera centre show con-sistent contraction towards the caldera, which is not pre-dicted by one shallow point source. These residuals sug-gest that a more complete deformation model is warranted, calling for a deeper source in addition to a shallow one, in order to explain the contraction of the distant stations to-wards the caldera. The 1993–1998 GPS data are fitted here with a more complete model with two point sources (one shallow and one deep), used to represent a more extensive

Fig. 3 Earthquake epicentres in the Askja area from 01/01/1997 to 31/12/2004. All earthquakes larger than magnitude 1.5 that have been observed at four or more seismic stations are included. The background shows shaded topography, eruptive fissures (*thick lines*), faults (*hatched lines*) and rivers and lakes (*white*)
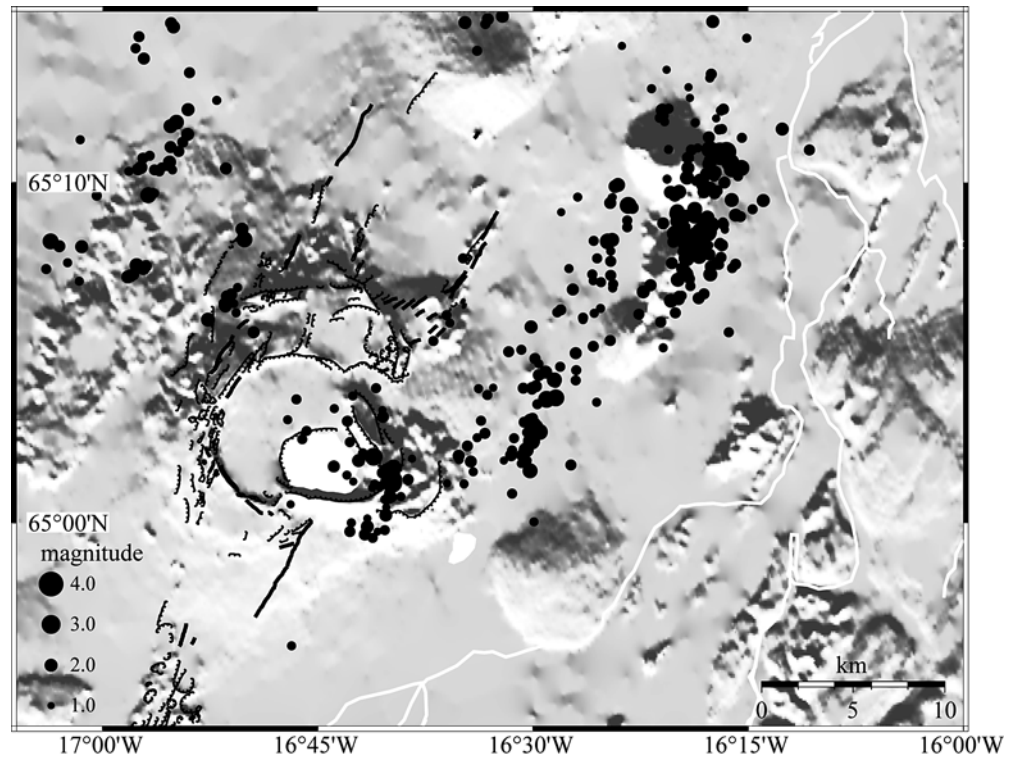
magma pluming system under Askja. Extensional deformation across the divergent plate boundary is also considered, in a similar manner as done by Camitz et al. (1995). Linear strain build up within a plate deformation zone is assumed. The geodetic observations and the revised interpretation are consistent with an extensive magma source under Askja, with pressure in the source eventually coupled to the conditions along the divergent plate boundary.

Geodetic observations

A network of 24 GPS benchmarks in Askja (Fig. 2) was initially measured in 1993 (Camitz et al. 1995). The whole network was re-measured 1998, after which yearly measurements have been performed at a few key stations (DYNG, OLAF and A404, see Fig. 2). Those yearly measurements complement the yearly levelling at Askja volcano, providing two independent time series. GPS data collected in 1994 and 1996 have also been included in this study.

The Askja levelling profile

The first 12 benchmarks of the Askja levelling profile were installed in 1966 by Eysteinn Tryggvason. They are located on a pahoehoe lava formed in the 1961 eruption. The profile was extended in 1968 to consist of 30 benchmarks. Results from the profile have been documented elsewhere (Tryggvason 1989; Rymer and Tryggvason 1993; Sturkell and Sigmundsson 2000). Previously, results have been interpreted using a single Mogi source close to the middle of

the Askja caldera. The benchmarks on the profile are located at varying distances from the caldera centre (Fig. 2), and show a characteristic relative change in response to inflation or deflation of the caldera (vertical deformation decreasing with distance from the caldera centre). The profile was levelled yearly from 1966 to 1972, and again every year since 1983. The profile that has been measured annually extends from benchmarks numbered A401 to A430 (benchmarks A401, A402 and A403 have sometimes been snow covered). The full length of the profile is located within the inferred deformation field at Askja, so only relative changes can be inferred. The A406 benchmark is closest to the centre of subsidence, and at the other end of the profile station A430 is furthest away from the subsidence centre. It has, however, a history of somewhat erratic movements, along with station A422. The levelling data are used to evaluate the time variation of deformation rates. Annual elevation differences between benchmark A406 and A429 are shown in Fig. 4a. The annual change is ∼1 cm, decaying somewhat with time, and includes some scatter. Cumulative height changes from a starting year in 1968 (Fig. 4b) are regular and show also a modest decay in rate after 1983. The minus sign on the elevation difference between A406 and A429 signifies caldera subsidence, and a positive sign indicates inflation. Early in the period 1983–2003 the rate of relative height change between benchmarks A406 and A429 was ∼12 mm/year, but decreased to ∼7 mm/year near the end of the period. The 1983–2003 changes fit an exponential function of the form

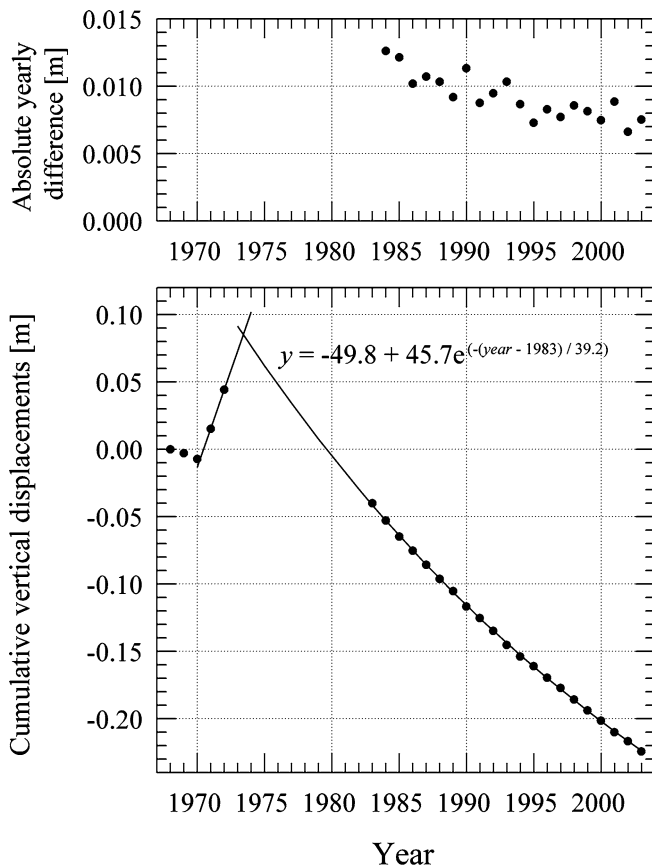$$y = -49.8 + 45.7 \times e^{(-(year-1983)/39.2)} \tag{1}$$

Fig. 4 The height differences between benchmarks A406 and A429 on the levelling profile. Benchmark A406 is located closest to the centre of subsidence and A429 farthest from it. **a** Annual height differences in the 1993–2003 period. **b** Cumulative vertical displacements between the benchmarks since 1968. The decay from 1983 to 2003 ($\Delta_{429-406}$=−0.1842 m) fits an exponential function. From back tracing of the subsidence curve and extrapolations of the inflation trend between 1970 and 1972, the change of regime from inflation to deflation is estimated to have occurred 1973

where the time (*year*) is the calendar year and height changes (*y*) are in centimetres. The subsidence rate appears to decay exponentially, being reduced by a factor of 1/*e* over a time period of 39.2 years.

Deformation in the earlier period of levelling measurements from 1968 to 1972 is very different from that observed after 1983, as alternating periods of inflation and deflation were observed. A time series of vertical change between benchmarks A429 and A406 (Fig. 4b) is therefore a good indicator of the state of the volcano. The period of no observations on levelling profile, 1972–1983, is bracketed by rapid inflation preceding the time interval, and steady deflation afterwards. It has been suggested by Tryggvason (1989) that the continuing subsidence started at least 7 years before the observations in 1983. After little change from 1968 to 1970, the highest deformation rates yet observed in Askja were observed from 1970 to 1972, when uplift of the caldera occurred. From 1971 to 1972 the $\Delta_{429-406}$ value was +28.8 mm, and the total $\Delta_{429-406}$ value for the 1970–1972 uplift period was +51.2 mm. Rate of uplift in the 1970–1972 period can be extrapolated into the time pe-

riod of no observations, and the rate of deflation observed after 1983 (equation 1) can extrapolated backwards in time as well. These extrapolations suggest that uplift continued to 1973 at Askja, but then subsidence started.

Deformation model for Askja

We present a revised deformation model for Askja that considers spreading across the plate boundary, and shallow and deep magma sources under Askja. We fit the 1993–1998 GPS data to this type of model. A deep Mogi source is required to explain contraction towards the caldera at far distances. This contraction is best observed north and south of the caldera, along the rift zone (Fig. 5a and 5b). East and west of Askja, this contraction towards Askja is counterbalanced by east-west plate movements associated with spreading, that increase with distance from the Askja centre, as suggested by Camitz et al. (1995). A model for the plate boundary spreading is needed to incorporate this effect. The best fitting parameters of our model were found with a grid search method, with some parameters constrained by prior assumptions. We demonstrate that the model fits the GPS and levelling observations much better than a model with a single Mogi source. Modelling approaches could be improved, but are unlikely to change fundamentally the conclusions presented here.

Divergent plate movements

There is only limited information on the regional spreading signal at Askja, and for our modelling we chose to use the simplest type of a model to account for the spreading. We apply a model with uniform horizontal strain within a plate boundary deformation zone (displacements increasing linearly with distance from a central axis of a plate boundary within the zone). Following Camitz et al. (1995), we apply this type of model with the central axis of the plate boundary crossing through 65.00°N and 16.75°W near the centre of Askja. Outside the plate boundary deformation zone, we adopted the NUVEL-1A model plate movement (DeMets et al. 1994), with a spreading direction of N106°E, and half-spreading rate of 9.5 mm/year. Local orientation of the plate boundary deformation zone is assumed to be perpendicular to the NUVEL-1A direction at Askja (N16°E). The width of the plate boundary deformation zone is the only parameter we attempted to constrain with our forward search for a best-fit model, considering a range of 30–150 km. Best fit was observed with a plate boundary deformation zone that was 100 km wide. This differs from the conclusions of Camitz et al. (1995), who inferred a width of ∼30 km from a more limited data set.

With a 100-km wide plate deformation zone, the precise location of the plate boundary is not crucial for our interpretation, as the strain gradient is 0.02 cm/year per km and the width of the area covered by GPS stations is about 20 km. Horizontal plate movements may amount to 0.4 cm/year (2.0 cm in 5 years) over the width of the area.

Fig. 5 a Horizontal GPS-displacements 1993–1998 relative to the reference station DYNG. **b** Horizontal displacements adjusted for inferred displacement of the DYNG reference station, according to the best-fit deformation model (Table 1). **c** Model displacements at all sites according to the model in Table 1. **d** Residual horizontal displacements for the model in Table 1. The largest residual is observed at station BATS which is influenced by local deformation. **e** Horizontal model displacements for one point source at 2.8 km depth presented by Sturkell and Sigmundsson (2000). **f** Residuals of the horizontal displacements from one point source
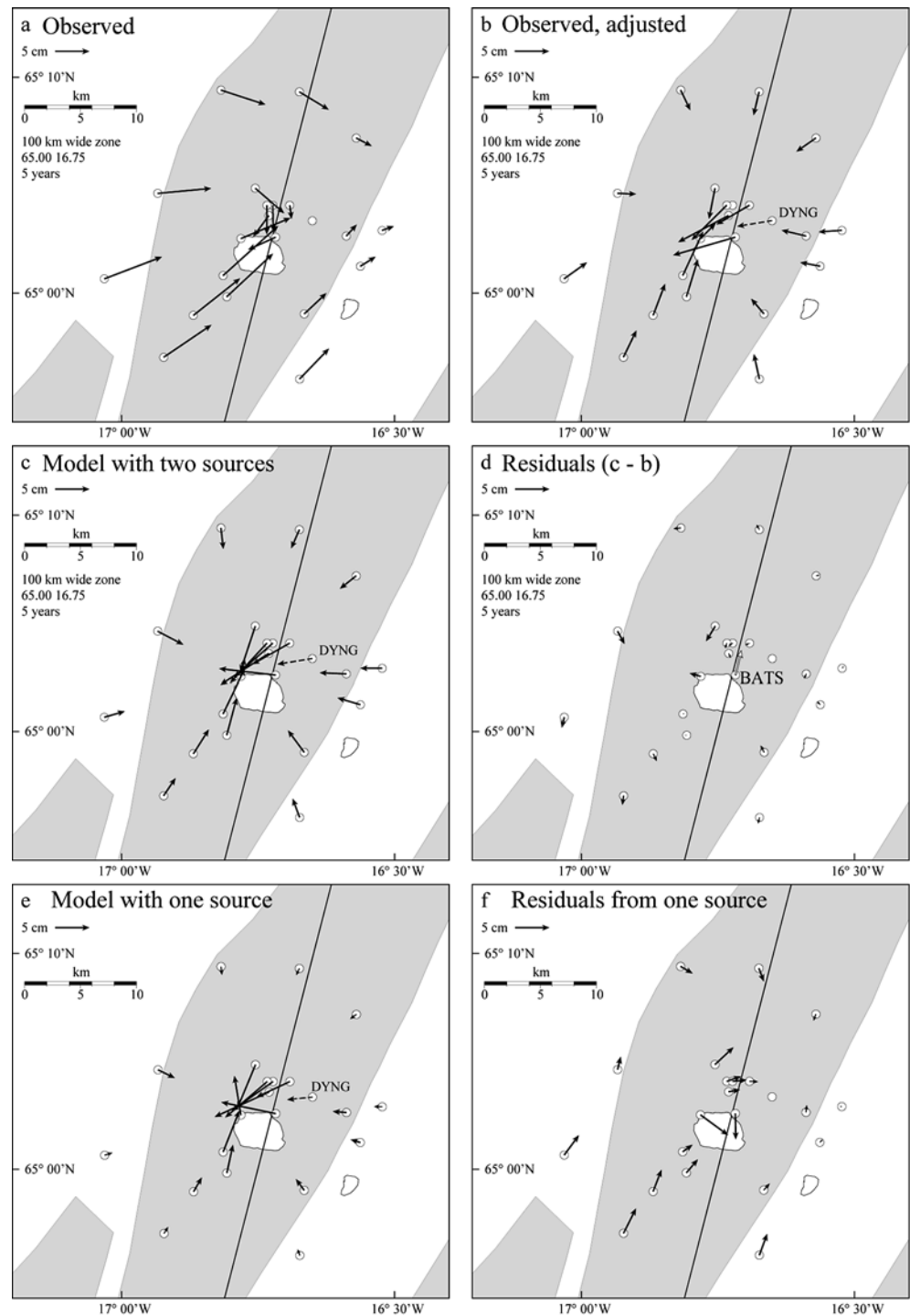
Model of the magma plumbing system

The 1993–1998 GPS-derived displacement field (Sturkell and Sigmundsson 2000), has been re-modelled and evaluated with the possible presence of a deeper source. Although one Mogi source has been used to fit much of the geodetic data, there are two main arguments for an additional deeper source. Firstly, a re-examination of the 1993–1998 GPS displacement field reveals a systematic misfit at distances far away from the caldera, mostly along the rift zone, that shows contraction towards Askja. Furthermore, petrological studies indicate also that a magma reservoir at 12–18 km depth is possible (Guðmundur E. Sigvaldason, personal communication, 2003). In an attempt to model an extensive plumbing system under Askja, we consider whether a simple approximation can explain the observations. A deeper-seated Mogi source is introduced to complement the shallow one.

Table 1 Deformation model for Askja, based on GPS-measurements 1993 and 1998. The strength is expressed as the vertical displacement above the centre of subsidence

	Latitude	Longitude	Depth (km)	Strength (m)
Mogi 1	N65.0456 (utm zone 28, 7214.9)	W16.7802 (utm zone 28, 416.2)	3.0	$h_s = -0.208$
Mogi 2	N65.0456 (utm zone 28, 7214.9)	W16.7802 (utm zone 28, 416.2)	16.2	$h_d = -0.052$

Plate boundary: Crossing point W16.75, N65.00, Direction 105.95°, Width 100 km, Half-spreading rate outside the zone 0.95 cm/year
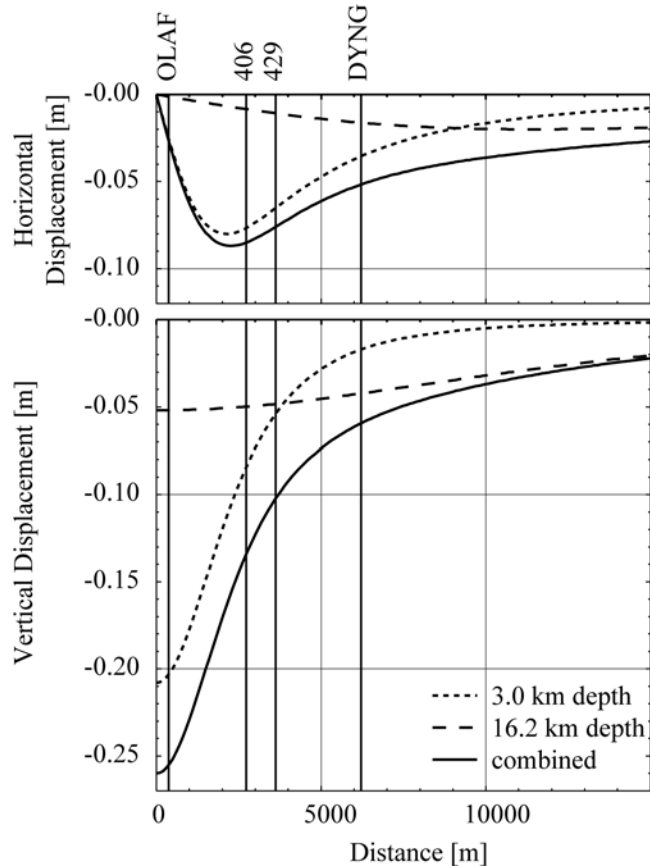

Fig. 6 Radial profiles for horizontal and vertical displacements for a shallow (3.0 km) and a deep (16.2 km) Mogi point sources, and for the two sources combined. The parameters for the models are from Table 1. For horizontal displacements the shallow source dominates the deformation out to 9 km from the centre, and for the vertical displacements out to 4 km. The distances of the GPS-points OLAF and DYNG from the centre are indicated, as well as benchmarks 406 and 429 on the levelling line
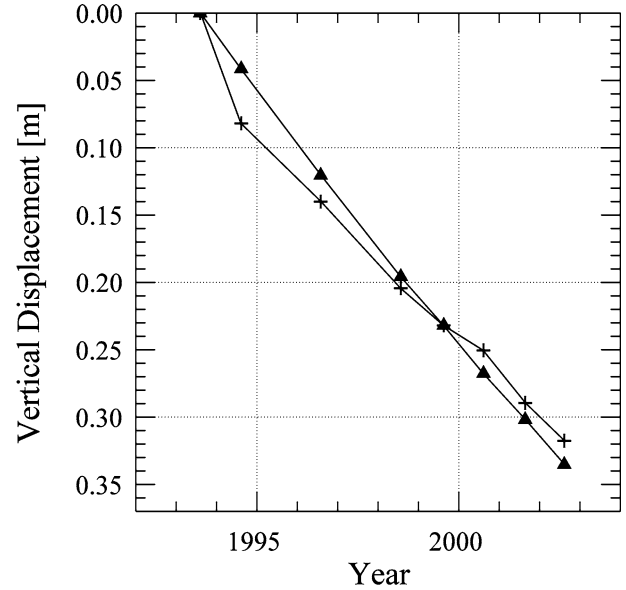

Fig. 7 Time series of vertical displacement at the Ólafsgígar (OLAF in Fig. 2), GPS station relative to the DYNG reference station (which is within the deformation field). Measurements are shown with a plus sign, triangles give predictions of the model presented in Table 1, function (2)

tical change, the shallow point source h_s contributes 80% and the deeper point source h_d contributes 20%. Then

$$h_0 = h_s + h_d = -2.8 + 2.8\, e^{(-(year-1983)/39.2)} \qquad (2)$$

The exponential decay curve equation (1) was implemented to express the observed time dependent decay of vertical displacement h_0 (Fig. 7).

The model presented above accounts well for the observed data. In the grid search we used a least squares criterion minimising the χ^2 merit function (Bevington 1969)

$$\chi^2 = \sum_{i=1}^{N} \left(\frac{res_i}{\sigma_i} \right)^2. \qquad (3)$$

where res_i is the residual between observations i and model prediction, and σ_i is the standard deviation of observation i. Firstly, the horizontal deformation field was evaluated, because a second component in the vertical deformation field is evident. This is a regional component, manifested as regional subsidence west of Askja relative to the area east of it (see Plate 2 in Sturkell and Sigmundsson 2000). In this case the number of observations, N, is 44 (two

We suggest that the location of the deeper source is directly under the shallow one. This is consistent with preliminary inspection of satellite radar interferograms of Askja that show just one concentric deformation pattern. The depth of this source, as well as its strength relative to the shallow one, were varied in order to best reproduce the surface observations. A Mogi source at 16.2 km depth, directly under the shallow one, accounting for 20% of the vertical signal directly above the sources, is our preferred solution (Table 1, Figs. 5c and 6). The deep source is important in accommodating the observed displacements occurring far away from the main caldera.

The total vertical displacements from the two point sources are expressed by the variable h_0. For the ver-

horizontal GPS displacement components at 22 stations). The estimated uncertainty of the horizontal components is 6 mm. The preferred model parameters (Table 1) result in χ^2_{min} equal to 101 (Fig. 5c and 5d). For comparison, a null model would give χ^2=2019, and a model with only the shallow Mogi source would give χ^2=235.

The vertical component of all the GPS points can be included with an uncertainty of 8 mm. In this case the number of observations, N, is 66. A forward search results in χ^2_{min} equal to 609. For comparison, a null model gives χ^2=3881. Distant stations west of Askja have a positive vertical residual, and stations to the east have a negative residual from the model. It is clear that there appears to be a regional gradient in vertical displacement, not accounted for by the model presented in Table 1. The regional vertical component is difficult to precisely evaluate because the GPS survey does not cover a large enough area. Both the horizontal and vertical combined model displacements for the two point sources are plotted in Fig. 6, showing the influence of the different source depth with distance from the centre.

The GPS-point at Ólafsgígar (OLAF), which has been measured frequently since 1993, is located within 500 m of the centre of subsidence, making its vertical record particularly sensitive to deflation of the caldera floor. Its vertical displacement is plotted in Fig. 7 together with the modelled displacements using the model from Table 1.

Discussion

Our deformation model for Askja consists of a local deformation field due to a magma plumbing system under the volcano and regional deformation due to plate spreading. The results from the local GPS-network at Askja are mostly sensitive to the local deformation field, but do nevertheless provide constraints on the regional strain field at Askja. The best-fitting model was found with a plate boundary deformation zone 100-km wide. This width is comparable to that observed further north from Askja along the plate boundary at the Krafla volcanic system, as well as new results from south Iceland (La Femina et al. 2005). GPS studies of plate spreading at Krafla, e.g. by Heki et al. (1993), show post-rifting stress relaxation after a rifting episode from 1975 to 1984. Temporary spreading velocities are in excess of the long-term average next to the plate boundary, with a maximum in spreading velocities at about 50-km distance from the central axis of the Krafla fissure swarm. This excess velocity appears to be gradually decaying. There is poor regional GPS coverage to detect spreading across the plate boundary at the latitude of Askja, but an earlier study by Camitz et al. (1995) suggested a plate deformation zone as narrow as 30-km wide. This conclusion is, however, open to a different interpretation with a wider deformation zone. Our revised estimate calls for a zone about 100-km wide, consistent with observations at from other parts of the plate boundary in Iceland.

The measured surface deformation has been modelled with two point sources while most studies of volcano de-

formation have been interpreted using a single source. One exception was the study by Tryggvason (1986) of deformation associated with the 1984 eruption at Krafla volcano. A model with shallow and deep sources was suggested, but the exact depth of the deep source remained unknown. In Askja the measured deformation can be explained by two Mogi sources located at depths of 3.0 and 16.2 km. The deeper source suggested by our geodetic results, is consistent with petrologic arguments. Analysis of augite from the Sveinagjá (Fig. 1) lava suggests crystallization at depths of 12–18 km (Guðmundur E. Sigvaldason, personal communication, 2003).

The deformation model and observations from the levelling profile can be used to evaluate volume changes of the inferred Mogi sources at Askja for the last decades. Extrapolation of vertical change at Askja into the 1972–1983 period of no observations suggests a turning point from uplift to subsidence in 1973. Total subsidence from 1973 to 2003 is 1.88 m directly above the point sources, assuming the inferred h_0 equation presented in Table 1 is valid. Furthermore, we assume that the relative contribution towards subsidence from the upper and deeper Mogi sources is the same for 1973–2003 as inferred for the 1993–1998 period (based on the GPS data), 80% and 20%, respectively. For the 1973–2003 period, subsidence due to the shallow source is estimated $h_s=h_0\times0.8=-1.50$ m, and due to the deep source, $h_d=h_0\times0.2=-0.38$ m. Volume change of a Mogi source cavity, ΔV_{ch} (chamber), can be inferred from the integrated ground-surface subsidence, ΔV_e (edifice), as $\Delta V_{ch}=2\Delta V_e/3$ (Delaney and McTigue 1994; Johnson et al. 2000). The subsidence since 1973 has generated an integrated surface deflation volume of $\Delta V_{es}=2\pi h_s d^2\approx0.083$ km^3, for the shallow magma chamber ($h_0=-1.50$ m and $d=3.0$ km) and $\Delta V_{ed}=2\pi h_d d^2\approx0.619$ km^3, for the deep reservoir ($h_0=-0.38$ m and $d=16.2$ km). The corresponding volume change for the shallow Mogi source is 0.055 km^3, and 0.412 km^3 for the deeper source. It must be emphasized that the volume-change value for the deeper source is very uncertain, as it is based on the assumption that 20% of the inferred subsidence at Askja throughout the 1973–2003 period is due to the deeper source. For comparison to the volume values, then the 1961 eruption at Askja generated ~0.09 km^3 of lava during 6 weeks. The volume change due to the shallow source 1973–2003 is only about half this value.

The volume estimates can also be compared to the 1875 Plinian eruption in Askja. It resulted in the creation of a 4.5 km diameter caldera with a volume of 1.77–2.5 km^3. The inferred volume of the eruptive products from 1874–1876 in the Askja system is 0.73 km^3 (Sigvaldason 1979). A dyke extending from Askja to Sveinagjá, which formed 1874–1876, was suggested by Thorarinsson (1968) to be 40 km long, 1 km high, and 4 m wide (0.16 km^3). Analogies with the Krafla 1975–1984 rifting episode suggests that cumulative dyke widening is likely to be much more as multiple intrusions occurred, and the dyke height may be more like 10 km, cutting throughout the brittle crust, rather than 1 km. Such an extensive dyke or dyke complex formed

in 1874–1876 may explain, at least partly the apparent volume discrepancy between volumes of the caldera and the eruptive products.

Some of the current subsidence in Askja probably relates to general pressure decrease at the plate boundary due to spreading. Some magma might flow out from the magmatic sources to accommodate the spreading. A micro-gravity study of the Askja volcano during the interval 1988–2002 suggests a sub-surface mass decrease of 1.6×10^{11} kg (de Zeeuw-van Dalfsen et al. 2004) at the caldera. This study suggests that the sub-surface mass decrease is best explained by magma drainage. No such events are known in Askja in the 1983–2003 period. Within the shallow crust such a process is likely to take place by dyking events, as observed during the Krafla rifting episode. Ductile processes might be important, especially for the deeper source that is located well within the ductile part of the lower crust. Volume discrepancies for 1874–1876 events at Askja have by some authors been contributed to lateral episodic drainage of magma; our observations would require such process to take place gradually during quiet periods.

We have suggested that there are two processes mostly responsible for the observed deflation of the Askja caldera, crystallisation of magma and pressure reduction along the plate boundary due the plate spreading process. If cooling and contraction by degassing and crystallisation of magma takes place it can lead to about 10–11% volume reduction. If the volume decrease associated with the shallow source would be solely due to this process, it requires the solidification and cooling of about 0.5 km³ of magma in the last 30 years. However, during the last 3000 years the Dyngjufjöll volcano complex (hosting Askja) has only produced lavas that are aphyric (Guðmundur E. Sigvaldason, personal communication, 2003) This may indicate that no large volumes of magma rests and crystallises in a shallow magma chamber prior to an eruption, resulting in aphyric lava (less than 5% phenocrysts). Although extrusives in the caldera are aphyric, some crystallisation may take place in the shallow magma chamber. Crystallization may also take place in the inferred deep magma reservoir, contributing its volume reduction. The Sveinagjá lava erupted in 1875–1876 and it contains 13% phenocrysts. Analyses of the phenocrysts suggest that they originated from a depth comparable to our deeper source (Guðmundur E. Sigvaldason, personal communication, 2003). Crystallisation and outflow from the magma sources are both considered contributing to the current deflation.

Levelling data provides the longest time series of deformation at Askja and show that the rate of deformation has varied significantly. A reversal from uplift to subsidence appears to have occurred in 1973, 2 years prior to the initiation of the Krafla rifting episode, 70-km further north along the plate boundary in Iceland. If draining of magma from depth under Krafla towards shallow levels initiated already in 1973, it could have influenced conditions along a major part of the plate boundary in North Iceland, reducing pressure under Askja and triggering the onset of subsidence. The observed gradual subsidence at Askja is unique in

magnitude for Icelandic volcanoes. However, similar process, but smaller in magnitude and shorter in duration has been observed at Krafla volcano after 1989 (de Zeeuw-van Dalfsen et al. 2006).

Conclusions

The Askja volcano has subsided for over 20 years. The work presented here indicates that the subsidence is gradually slowing down with an exponential decay constant of 39 years. The onset of the current subsidence is inferred to have been in 1973, 2 years prior to a rifting episode at the Krafla volcanic system in Northern Iceland. This correlation in time suggests links between the volcanoes, eventually by pressure connection through the lower ductile crust in Iceland. Re-interpretation of GPS results from Askja shows they can be better fitted with two Mogi sources, and a 100 km wide plate boundary deformation zone. The two sources model is taken as an indication of an extensive magma plumbing system under the Askja caldera. The improved deformation model presented in this paper fits better the observed contraction towards Askja over a wider area. The model has one Mogi point source at 3.0 km depth and another at 16.2 km depth.

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