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Rift-transform junction in North Iceland: rigid blocks and narrow accommodation zones revealed by GPS 1997–1999–2002

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SUMMARY

The current tectonics of North Iceland are characterized by rifting episodes along fissure swarms and by M_L 6–7 earthquakes in the transform zone between the northern rift zone and the Kolbeinsey Ridge north of Iceland. The last rifting period (1975–1984) was associated with an average opening of 5 m along the Krafla fissure swarm. Post-rifting deformation has been revealed by GPS investigations in Northern Iceland. A GPS network was occupied in 1997, 1999 and 2002 to quantify present-day displacements and their variation with time, on both sides of the on-land part of the Húsavík–Flatey fault and around the Krafla fissure swarm. The main deformational features observed are: (1) decrease in spreading velocities between the two timespans 1997–1999 and 1999–2002, (2) the existence, south of the transform zone, of a well-defined ∼20–30-km-wide plate boundary where extension dominates, (3) directly adjacent east and west rift shoulders where displacements are close to those observed on the Eurasian and North American plates and (4) a displacement gradient on Tjörnes that could correspond either to elastic deformation related to a currently locked dextral strike-slip fault or to an attenuated post-rifting effect northwest of Krafla.

Key words: deformation, GPS, Krafla, rifting, Tjörnes, transform zone.

INTRODUCTION

The current tectonics in Iceland are associated with the plate boundary that crosses the island from south to north. Along the rift segments, rifting events sometimes accompanied by volcanism and lasting from several months to several years are associated with opening reaching several metres (Tryggvason 1984). During these events, opening along the fissure swarms is equivalent to several hundred years of the average spreading of the North Atlantic at the latitude of Iceland (20 mm yr^{-1}) as proposed by NUVEL-1A (DeMets *et al.* 1994). Earthquakes along the rift zone, are related to volcanic activity or dyke injection and have magnitudes less than 4 (Gudmundsson 1999). On the other hand, current tectonics in transform zones such as the Tjörnes Fracture Zone in northern Iceland are characterized by destructive events with the highest magnitudes known in Iceland (M 6–7) (Halldorsson *et al.* 1996).

In this paper, for the periods 1997–1999 and 1999–2002, we specify how displacements on both sides of the Krafla fissure swarm varied with time, and we analyse the effect of these displacements on the transform movement. The aim of this study was to examine how rifting and transform movement are interacting 15 yr after the major rifting event of 1975–1984.

TECTONIC SETTINGS

The left-lateral shift between the North Iceland rift zone and the Kolbeinsey ridge implies a dextral component along the Tjörnes fracture zone (Rögnvaldsson *et al.* 1998). This zone is a succession of three parallel active lineaments, namely from SW to NE, the Dalvik seismic lineament, the Húsavík-Flatey fault and the Grimsey seismic lineament (Fig. 1) (Einarsson 1976, 1991). As these lineaments strike slightly oblique to the average spreading vector, an opening component can be expected in addition to the dextral strike-slip component. Present-day seismic activity is mainly located along the Húsavík–Flatey fault and around the Grimsey lineament (Fig. 1). The inland part of the Húsavík–Flatey fault is a clearly identified tectonic zone, exposed south of Tjörnes (Gudmundsson 1993; Gudmundsson *et al.* 1993; Palmason & Saemundsson 1974; Saemundsson 1974) but not linked to present day microseismicity or well-documented historical seismic events. The western part of the Húsavík–Flatey fault is offshore where it can be followed by dense seismic swarms (Fig. 1) (Rögnvaldsson *et al.* 1998) whereas the Grimsey lineament looks more like a N120°E alignment of *en echelon* N–S left-lateral strike-slip faults, in the upper 10 km of the crust, than a single right-lateral strike-slip discontinuity (Fig. 1). Geological analyses suggest that the main part of the transform movement between the Kolbeinsey ridge and the Northern Iceland Rift Zone has been absorbed in the past along the Húsavík–Flatey fault (Rögnvaldsson *et al.* 1998; Saemundsson 1974).

The last main event in the area was the rifting episode that occurred in the Krafla volcanic fissure swarm from 1975 to 1984 (Björnsson 1985). During this period, a large number of volcano-tectonic events produced unusually large displacements with both horizontal (up to several meters) and vertical (up to a few metres)

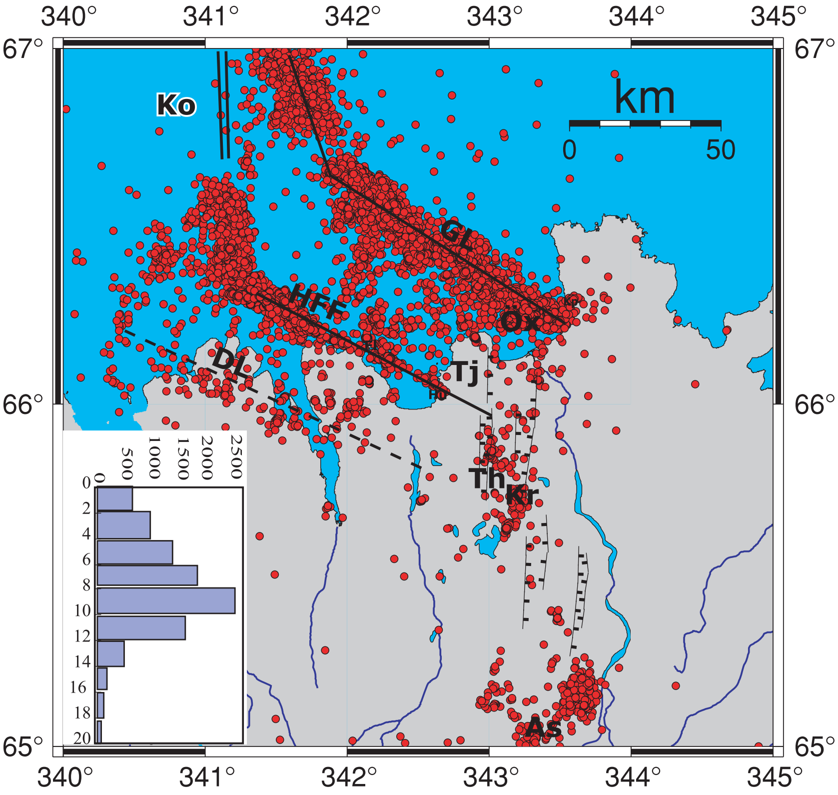

Figure 1. Microseismicity of northern Iceland during the period studied in 1997–2002. Each dot corresponds to an event with a magnitude larger than 1. The histogram shows the event distribution through depth during the same period of time but only for events along the Húsavík–Flatey fault. DL: Dalvik lineament, HFF: Húsavík-Flatey fault, GL: Grimsey lineament, Kr: Krafla centre volcano and fissure swarm, As: Askja centre volcano and fissure swarm, Th: Theistareykir fissure swarm, Tj: Tjörnes, Öx: Öxafjördur, Kp: Kopasker; Hu: Húsavík; Fl: Flatey. The seismic data were obtained from the data base of the Icelandic digital seismic network, SIL (Jakobsdottir *et al.* 2002; Thorbjarnardottir & Gudmundsson 2003; Thorbjarnardottir *et al.* 2003).

components (Tryggvason 1980, 1984, 1986). It must be underlined that the average horizontal displacement, 5 m, corresponds to 250 yr of opening of the medio-Atlantic ridge for northern Iceland (20 mm yr^{-1}). This period was followed by the disappearance of microseismicity along the eastern part of Húsavík–Flatey fault, the main structure of the transform zone (Rögnvaldsson *et al.* 1998). This could indicate that the section from Flatey eastwards of the Flatey–Husavik fault became locked during the Krafla rifting episode, and subsequently all rifting took place in the Krafla fissure swarm (from Myvatn into Öxafjördur). In recent years activity has picked up in the segment between Flatey and Husavik, although no activity is recorded from Husavik to the Theistareykir fissure swarm (R. Stefansson, personal communication, 1999).

The rifting episodes disturbed the stress and deformation fields as illustrated by the induced post-rifting displacements documented by GPS studies (Foulger *et al.* 1992; Hofton & Foulger 1996b), for at least the 10 yr after the end of rifting. These GPS measurements (Foulger *et al.* 1992; Hofton & Foulger 1996a) have locally revealed extension 2 or 3 times greater than the average relative plate velocity. Numerical simulations of present day (1987–1995) crustal displacements following the rifting event demonstrate that deformation is controlled by viscoelastic relaxation of the lower crust and upper mantle as far as 100 km from the spreading axis (Pollitz & Sacks 1996). New simulations (Berger 2004) taking into account the structure and the density of the lithosphere as documented by recent geophysical investigations and the new displacements proposed by (Völksen 2000) allow us to propose a new model with large viscosity of 6.10^{19} Pa s for the upper mantle and 6.10^{20} Pa s for the lower crust.

DATA ACQUISITION

The Tjörnes GPS Network (TGN) has been installed in Northern Iceland to study the displacements and their variations though time, on both sides of the Krafla fissure swarm and around the junction of the North Iceland rift system and the North Iceland transform fault zone during a period following the 1975–1984 rifting episodes along the Krafla fissure swarm. In this paper we present and analyse the results of three measurement campaigns made respectively made during the 1997, 1999 and 2002 summers. For each epoch, points were generally observed for two 24 hr sessions. Measurements in 1997, 1999 and 2002 were taken using Ashtech dual-frequency receivers with choke ring antennae in 1999 and 2002 and Ashtech Geodetic II and III type antennae in 1997.

Data were analysed with Bernese V4.2 software (Beutler *et al.* 2001) in the ITRF2000 reference frame using precise orbits (Altamimi *et al.* 2002), earth rotation parameters and data from IGS stations. The antenna phase centre offsets and the influence of elevation on phase centre variations were corrected using the standard antenna phase centre table. The troposphere-induced propagation delays were estimated from the observations, with troposphere parameters being estimated every two hours.

The 1997, 1999 and 2002 data sets were analysed with IGS station data (1997: HERS, KELY, KIRU, KOSG, ONSA, REYK, STJO,

VILL; 1999: HERS, HOFN, KELY, KIRU, NYA1, KOSG, ONSA, REYK, STJO, VILL; 2002: HOFN, KELY, KULU, NYA1, KOSG, ONSA, REYK, STJO, VILL). Six stations are common to all epochs (KELY, KOSG, ONSA, REYK, STJO, VILL). HERS and KIRU are additional stations for 1997 and 1999, HOFN and NYA1 are common to 1999 and 2002. Except for HOFN and REYK all these stations correspond to far-field points outside Iceland. In 2002 we also include in the processing data from the permanent network running at that time (Geirsson *et al.* 2006). The resolution of carrier-phase ambiguities used: (1) an ionosphere-free analysis without ambiguity resolution in order to assess residuals, (2) a 'Wübbena-Melbourne' ambiguity resolution (Melbourne 1985; Wübbena 1985) for the long baseline or, the local baseline, widelane ambiguity resolution with the use of local ionosphere model estimated from GPS observations and (3) a resolution of ambiguities using the ionosphere-free combination introducing the solved Wübbena-Melbourne or Wide-lane ambiguities and (4) a normal equation computation to allow solution combinations.

For each year's data, one normal equation and its weighting coefficient were obtained by combining the daily normal equations and the overall weekly solutions obtained from EUREF using a seven-parameter Helmert transformation. During the sinex–normal equation transformation, all *a priori* constraints present in the EUREF solution have been removed.

Velocities were then obtained by combining the weighted data from all three campaigns and all the weekly EUREF normal equations for the 1997–2002 periods, and expressed in the ITRF2000 reference frame using a 14-parameter Helmert transformation based on the best-determined EUREF points. Velocities were then expressed in the fixed Eurasia (Figs 2–4) and fixed North America reference

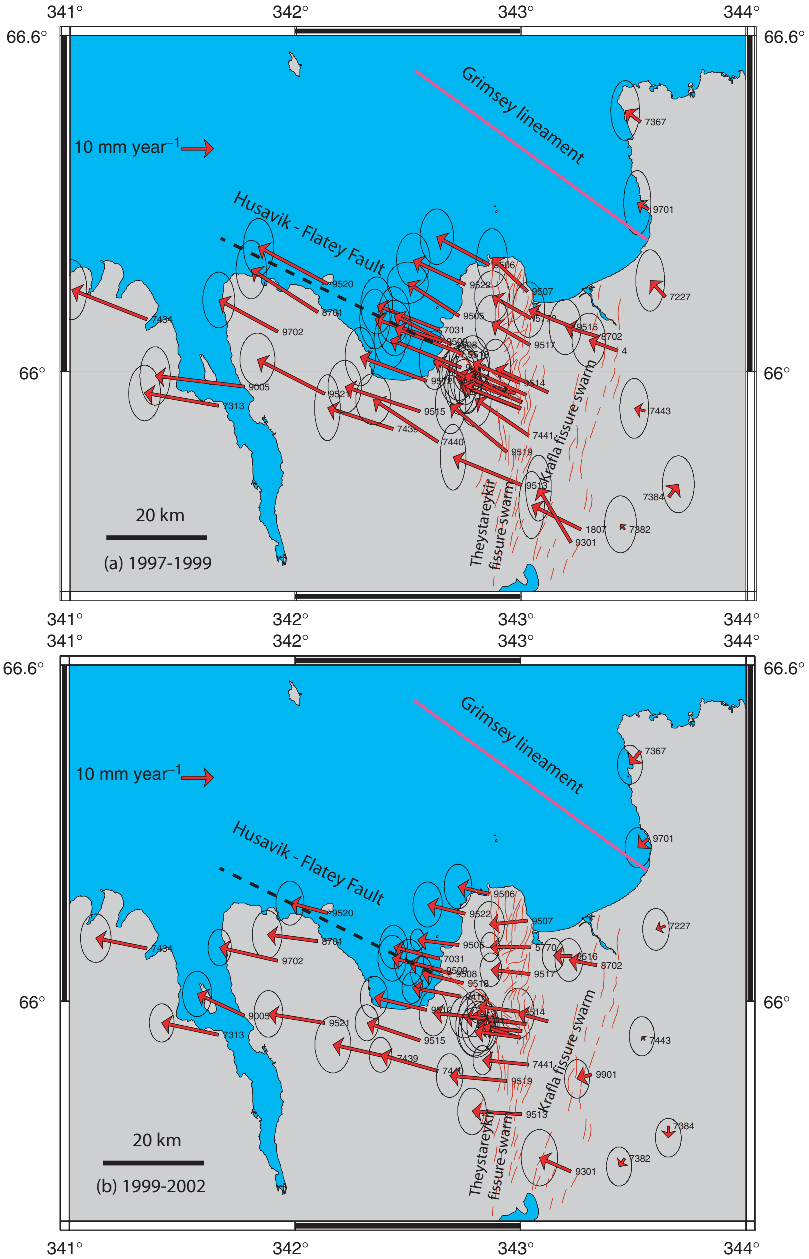

Figure 2. Velocity fields expressed in the Eurasia fixed reference frame. (a) for 1997/1999 timespan and (b) for 1999/2002. Ellipses correspond to a 2σ level (95 per cent confidence).

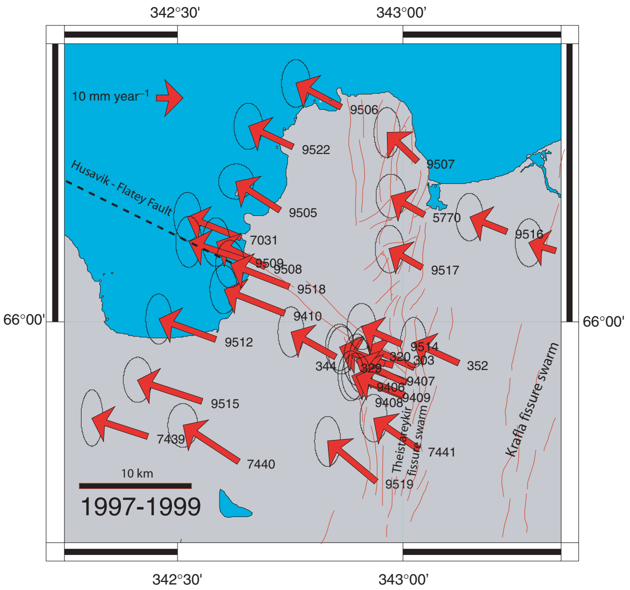

Figure 3. Velocity fields expressed in the Eurasia fixed reference frame for 1997/1999, detail on Tjörnes.

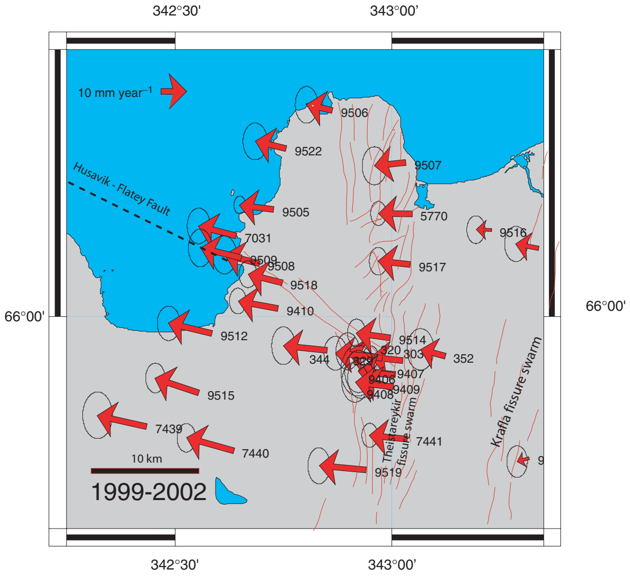

Figure 4. Velocity fields expressed in the Eurasia fixed reference frame for 1999/2002, detail on Tjörnes.

frames (Fig. 5) as defined by the rotation pole ITRF2000-Eurasia and by the Eurasia–North America rotation pole given by Calais *et al.* (2003).

RESULTS

Eurasia fixed reference frame

For both timespans (Fig. 2), 1997–1999 and 1999–2002, points east of the rift axis (7367, 9701, 7227, 7443, 7384 and 7382) and a few kilometres east of the Krafla fissure swarm present negligible displacements relative to the Eurasia fixed reference frame. Inside the Krafla and Theistareykir fissure swarms, a westward displacement gradient is observed (points 0004, 9516, 5770, 9517, 9507), whereas all points located in the SW part of the network present a westward displacement. A velocity gradient compatible with dextral N120°E strike-slip movements is visible on the 1999–2002 pattern (Fig. 2b). Displacements on both sides of the on-land part of the Húsavík-Flatey fault and around its junction with the Theistareykir fissure swarm do not show significant relative displacements excluding aseismic creep in this key area where the N120°E Húsavík fault meets the N–S Theistareykir faults.

North America fixed reference frame

Velocities expressed in the North America fixed reference frame (Fig. 5a), mainly for the 1999–2002 timespan (Fig. 5b), indicate that the SW part of the network (points 7434, 7313, 9005, 9702, 8521, 7439, 9515, 7440 and 9512) belongs to a rigid block moving like the North American plate. A velocity gradient perpendicular to the Húsavík Flatey fault (e.g. Tjörnes's points and 8701 and 9520) is particularly visible north of the Húsavík-Flatey fault. An E–W velocity gradient affects points East of Tjörnes, in the northern part of the Theistareykir fissure swarm (9515, 9517, 0352), whereas points located in the southern part of this fissure swarm are not affected by this gradient (7441, 9519, 9513), the gradient being located along the southern part of the Krafla fissure swarm (e.g. 9901/7441, 7382/9513). This displacement gradient along the rift axis (Fig. 5b) reveals evidence of a deformation area including the NE corner of Tjörnes, with a boundary striking N10°W, the fissure swarms being oriented N10°E. The obliquity of the area undergoing deformation to the rift axis has been also documented by INSAR investigations for the 1992–2001 timespan (Henriot & Villemin 2004).

Velocities changes between 1997–1999 and 1999–2002.

Major changes in the velocity field occur between the 1997–1999 and 1999–2002 timespans. In the Eurasia fixed reference frame (Fig. 2), points located in the North America block are affected by westward velocities reaching an average of 25 mm yr^{-1} during the first period of time whereas velocities decrease to an average of 18 mm yr^{-1} for the next three years. In the North America fixed reference frame, during the 1999–2002 timespan (Fig. 5b), points in the SW part of the network delineate a nearly undeformed area, whereas moderate deformation characterized by residual westward displacements occurred during the 1997–1999 span.

DISCUSSION

Post-Rifting relaxation

The time-dependent velocity field (see above) suggests that the deformation could be still affected by the post-rifting deformation following the 1975–1984 Krafla event (Heki *et al.* 1993). Since 1999, all deformation occurs in the rift zone as illustrated by the lack of relative displacements between points on either side of the rift zone east of the rift zone (Figs 2 and 5). For example, point 9901, a few hundred metres east of the Krafla fissure swarm, is not significantly displaced relative points on the 'European-fixed' part of the network. Opening seems to be restricted to a narrow band, only a few km wide, corresponding to the Krafla fissure swarm *sensu stricto*. Unfortunately, several different processes overlap in this area: post-rifting deformation inside the plate boundary, subsidence of the cooling magma chamber at Krafla, subsidence in the area around the geothermal power plant and widespread uplift due to magma accumulation 20 km down at the crust–mantle boundary, as demonstrated by (de Zeeuw-van Dalfsen *et al.* 2004). All those

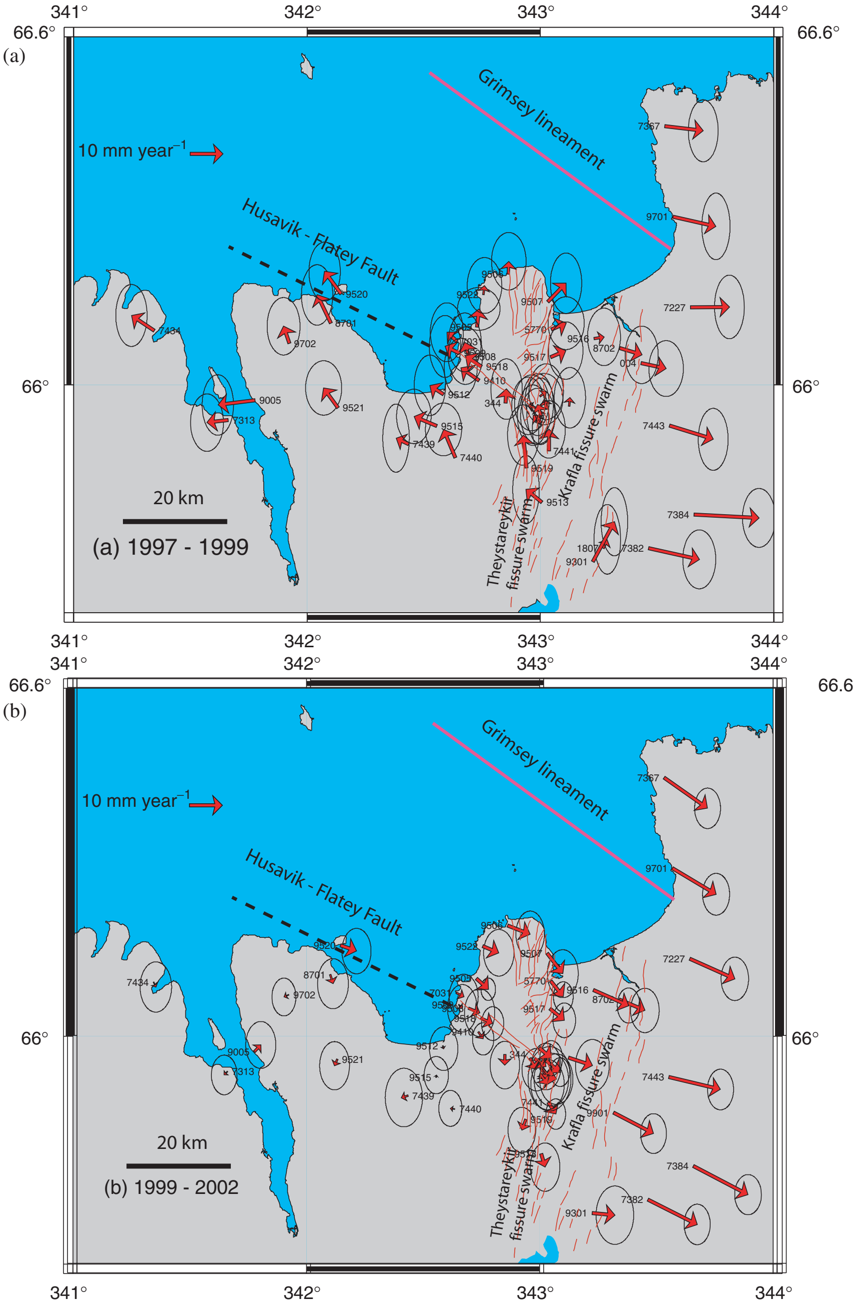

Figure 5. Velocity fields expressed in the North America fixed reference frame, (a) for 1997/1999 and (b) for 1999/2002. Ellipses correspond to a 2σ level (95 per cent confidence).

components necessarily produce a complex picture inside the rift zone itself where discrete GPS measurements are less appropriate than InSar deformation mapping.

To test the hypothesis that the velocity decrease between 1997–1999 and 1999–2002 is related to post-rifting relaxation, we have performed 3-D numerical simulations using ADELI3D software (Berger 2004) (Fig. 6). We have considered 3D models with crustal structures constrained by the geophysical investigations of ICEMELT (Darbyshire *et al.* 1998, 2000), FIRE (Brandsdottir *et al.* 1997; Darbyshire *et al.* 2000; Staples *et al.* 1997) and BP96 (Darbyshire *et al.* 2000; Menke *et al.* 1998). The Krafla rift zone has been modelled as a single dyke along which successive pressure increases were applied to simulate the different phases of rifting (timespan of rifting episodes, opening quantification and location of opening) as revealed by geodetic measurements (e.g. Tryggvason 1984). Velocities estimated for 1987–1992 and 1992–1995 (Völksen 2000) enable us to constrain the rheology parameters of the crust and the upper mantle. At the scale of our model, it has not been possible to distinguish the Husavik–Flatey fault from the Grimsey seismic lineament. As our modelling aims to study the deformation of the rift zone, we only have considered the Husavik–Flatey fault.

Simulations obtained by Berger (2004) reveal an elastoplastic upper crust and a viscoelastic lower crust overlying a viscoelastic upper mantle (Fig. 6). The average viscosities of the lower crust

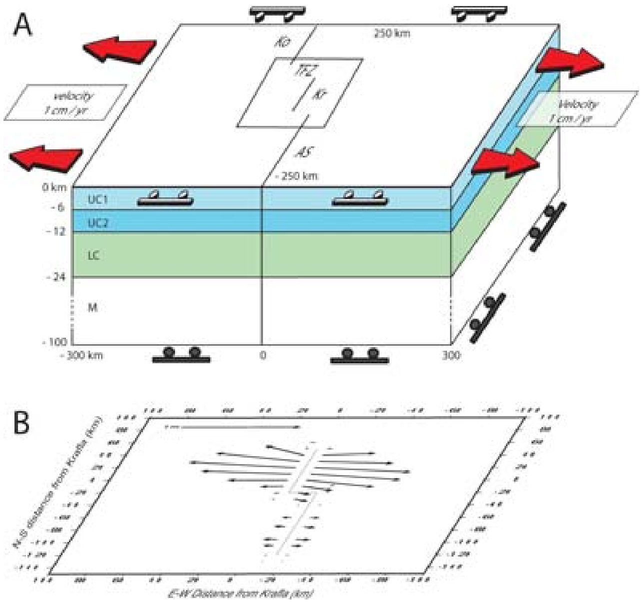

Figure 6. View of the 3-D model used for simulating the velocity fields presented in this paper (1997–1999–2002) and calibrated by (Berger 2004) on a large number of geophysical and geodetic investigations. (a) Rheological layering of the Icelandic lithosphere and boundary conditions. UC: upper crust divided in two layers UC1 and UC2; LC: lower crust; M: mantle. The rheological parameters are given in the text. (b) Displacements imposed in the model during the rifting crises period along Krafla and Askja fissure swarms.

and upper mantle have been estimated respectively to be 6.10^{19} Pa s and 6.10^{20} Pa s in average.

To simulate displacements measured in eastern Iceland, a pressure increase inducing one metre opening along the Askja fissure swarm need to be added to the opening of the Krafla fissure swarm (Berger 2004; Pollitz & Sacks 1996). Different simulations have been performed to test the influence of the Húsavík-Flatey fault: free sliding or high frictional strength on the upper part of the fault (0–12 km) to simulate respectively a slippery or a sticky comportment. These simulations did not reveal any significant influence of the rheological behaviour (millimetric differences between both solutions) of the Húsavík-Flatey fault in the simulation corresponding to the 1987–1992, and 1992–1995 displacements (Berger 2004).

We have applied the selected simulation of this study at our displacements fields 1997–1999 and 1999–2002 (Fig. 7): this model shows a relatively good agreement between observed and predicted displacements for the 1997–1999 and 1999–2002 velocities fields and allows us to take into account the velocity decreases between 1997–1999 and 1999–2002. As a conclusion, measured velocities fields reveal mainly a post-rifting relaxation due to the 1975–1984 Krafla rifting episodes.

The transform movements

The Húsavík Flatey fault

Three distinct cross sections enable us to analyse the behaviour of the Húsavík-Flatey fault system along its strike. The western one (points 9520, 8701, 9702, 9005 and 7313, Figs 2 and 5) unfortunately gives information about its south compartment only. The second section (points distributed along the line 9505 to 7439, Figs 3 and 4) is well documented and perpendicular to the fault main trace known on land. The third section (strip 9507 to 9519, Figs 3 and 4) is more complex to analyse because it strikes obliquely to the Húsavík fault and partly overlaps the rift zone.

At Flatey (Section 1, Fig. 5) there is no evidence for a complete lockage of the fault. Differences between velocities are smaller than uncertainties; at most the differences between 9520 and 8701 result from creep or elastic strain associated with huge microseismicity (Fig. 1).

On the other hand, a clear signal exists along the second profile where a difference of 8 mm yr^{-1} in average exists between points 9505 and 7439 (Figs 3 and 4). This difference is progressively accommodated from N to S, suggesting the lockage of a dextral strike-slip fault. The vectors are almost parallel and slightly oblique to the fault for both timespans. Unfortunately the profile is not long enough for estimating the thickness of the elastic crust and the locking depth of the fault. However, the accommodation suggests a minimum value slightly larger than the 10–12 km deduced from the microseismicity (profile on Fig. 1). It must also be underlined, as previously mentioned by Jouanne *et al.* (1999) that the point 9505 right in the middle of the fault-parallel motion is 10 km north of Húsavík instead of being on the fault itself.

The East-Tjörnes profile reveals a similar comportment, even if the N–S gradient in the fault-parallel velocities seems to be lower (Figs 3 and 4). Nevertheless, east of this profile, there is evidence neither of aseismic displacements nor strain accumulation as proved by the lack of velocity difference between the points 7382, 7384, 9901, 7443 and 7227 east of the Krafla fissure swarm (Fig. 5).

An alternative explanation can be put forward to explain the gradient observed in the fault-parallel component of the velocities. Indeed we know that the rifting episode resulted in about 8 m of extension at the latitude of Krafla. This extension decreases rapidly to the North at a rate of about 1 m each 5 km (Tryggvason 1984). As the initial tectonic events must have influenced the post-rifting effect, the result is a mimicry effect that could be confused with the lockage of a dextral strike-slip fault because post-rifting displacements decrease northwards. This alternative explanation should not change the seismic hazard for the area. According to both hypotheses, the section of the fault between Hùsavik and the þeistareykir is at present accumulating strain, the fault segment being locked. As the seismic activity has increased west of Húsavík, it seems likely that it is only a matter of time before the east section will rupture.

The Grimsey lineament

The network is not well configured to detect creep or accumulation of elastic deformation along the Grimsey lineament, this branch of the transform zone being entirely offshore. The modelling results show important misfits between observed and simulated vectors (points 7367, 9701 and 7227 on Fig. 7). The observed vectors are insignificant whereas the misfits suggest eastward motions that have not been properly simulated because we did not consider the Grimsey lineament in our modelling. Nevertheless some GPS points to the southeastern end of the microseismicity trace of the Grimsey lineament (e.g. 9701 and 7227). The displacements on these points do not differ significantly (Figs 2 and 5) although the points are on opposite sides of the on-land continuation of the lineament (Fig. 1). This supports an off-shore continuation in Öxafjördur Bay of the active rift that could connect with the Grimsey lineament somewhere north of Tjörnes rather than near Kopasker (Fig. 1). In that case the transform motion expected on the Grimsey lineament in the Kopasker should not be significant as the area is too far to the east of the rift transform junction.

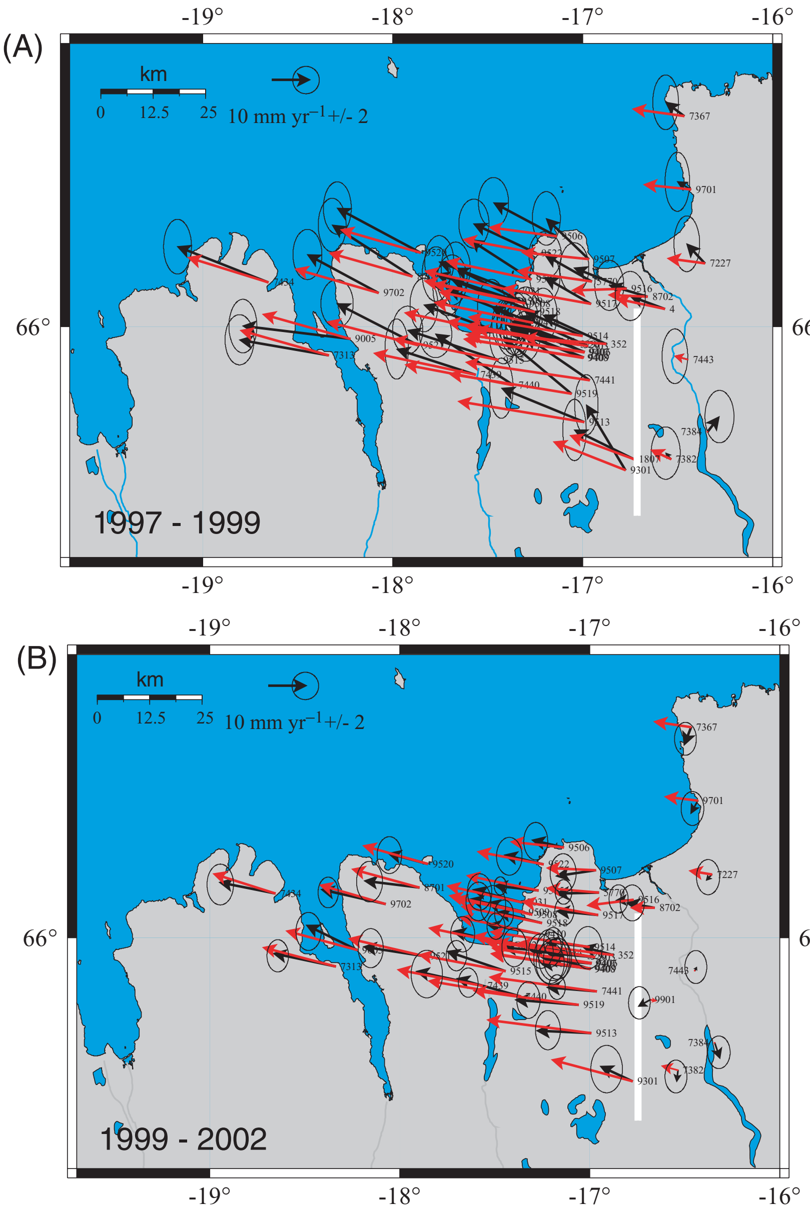

Figure 7. Result of the ADELI3D's simulation for 1997–1999 (a) and for 1999–2002 (b). Black arrows correspond to observed vectors, red ones to simulated velocities. All vectors are expressed in the fixed European reference frame. The white strip indicates the location of the dyke used in the simulation and active during the 1975–1984 events.

CONCLUSIONS

Present-day spreading velocities in northern Iceland are characterized by a decrease though time mainly induced by viscous deformation consecutive to the 1975–1984 Krafla rifting episode along the rift axis. Our modelling indicates that we are probably near the end of this viscoelastic effect and near the beginning of the steady state regime characterizing the inter-crisis periods. The velocity fields expressed both in Eurasia-fixed and North America-fixed reference frames allow us to outline two areas affected by very moderate deformation, relative to, respectively, the Eur and Nam reference frame, separated by a narrow extensional area with localized opening. This deformation area lies along the southern part of the Krafla fissure swarm and then widens westwards to the east of Tjörnes and north of the Þeistareykir fissure swarm. Extensional deformation seems to be more diffuse where the rift connects and overlaps the transform zone.

There is no geodetic evidence of creep along the eastern part of the Grimsey lineament and along the inland part of the Húsavík-Flatey fault and its connection with the Theistareykir fissure swarm. The western segment of the Húsavík-Flatey fault revealed displacements that could be due to either elastic strain accumulation or creep on this fault segment. This regime explains the intense microseismic activity presently occurring in this area.

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