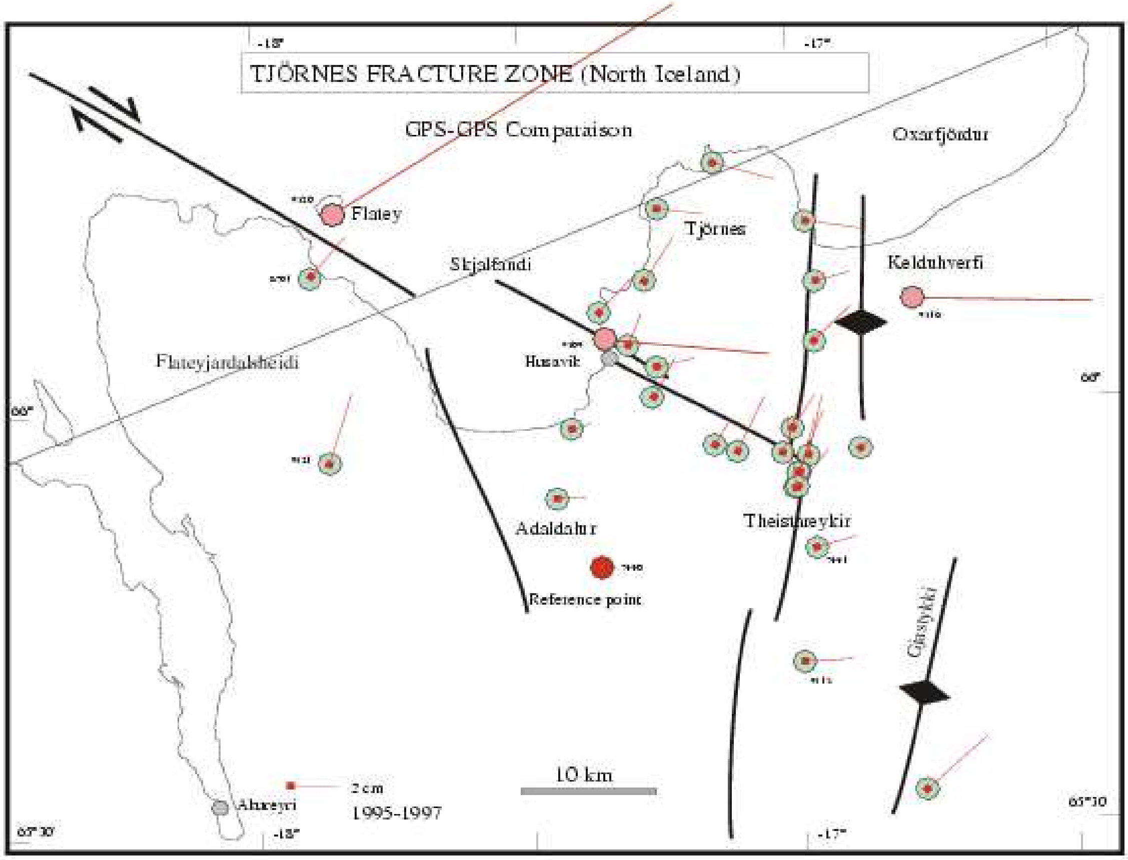

Figure 38. *1995-1997 horizontal displacements on the TGN.*

to the east have been computed for the points located in the fissure swarms. Points on Flateyjardalur move to the north. The point on the Flatey island reveals a large displacement to the NE that could be due to a local instability.

The assumption of an interseismic strain has been tested by using a simple dislocation model. This model assumes that a set of buried planar fault surfaces are locked above a given depth and are affected by uniform aseismic creep below this depth. In order to determine this brittle/ductile transition we used the microseismicity recorded by the SIL network from 1995 to 1997. We assume that most earthquakes are localized in the brittle crust. Thus gives us a limit at a 10 km depth in average. We founded a solution that minimizes the differences between simulated and observed velocities.

The model (Figure 39) assumes: (1) a dyke opening of 20 mm/year affecting all the brittle crust along the Kolbeinsey ridge; (2) two dyke openings of 30 mm/year and 20 mm/year respectively along the Krafla and Þeistareykir fissure swarms; (3) a dextral strike–slip fault striking N100°E between the two previous rift segments with a velocity of 50 mm/year below a depth of 10 km and completely locked above the brittle/ductile transition; (4) a 15 mm/year opening zone striking N140°E south of the HFF; (5) a fault along the Grímsey lineament with both a 15 mm/year opening and 20 mm/year dextral strike–slip movements.

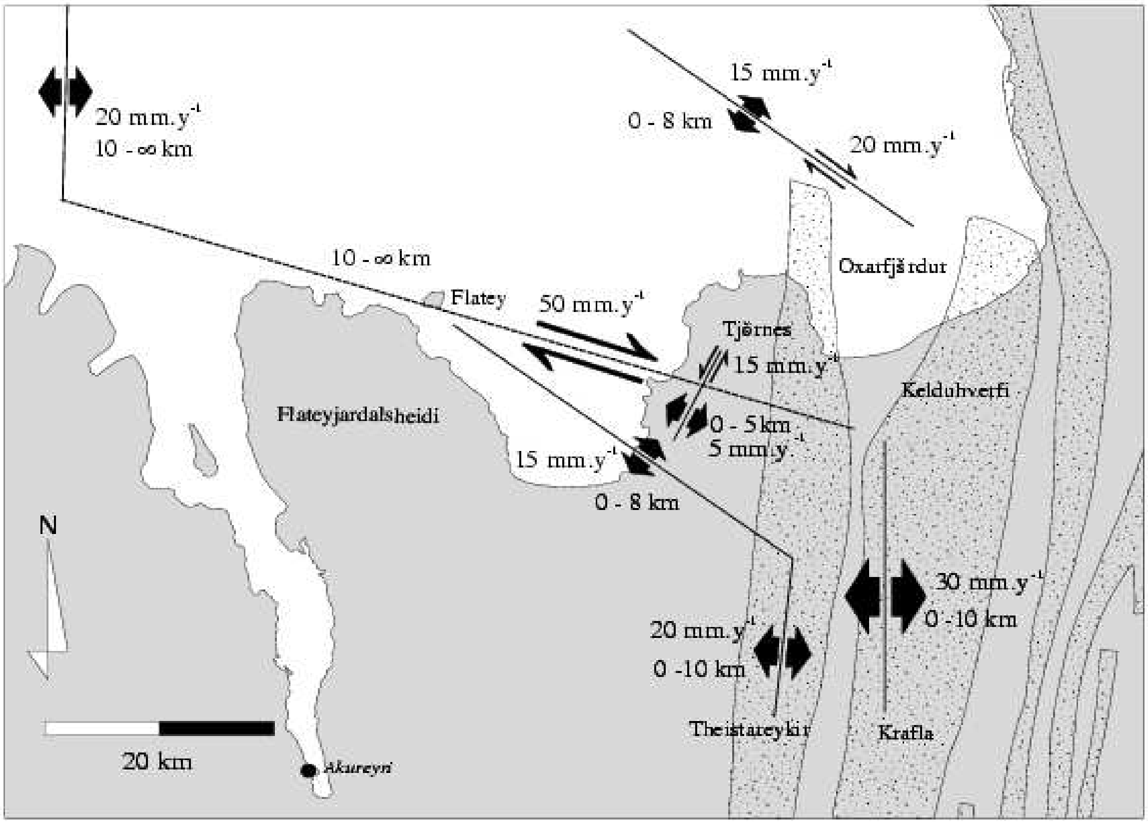

Figure 39. *Dislocations used in the simulation of the 1995-1997 displacement field (after Jouanne et al. 1999). Two main kinds of dislocation can be distinguished: superficial dislocations that affected the upper crust with mainly aseismic opening or strike-slip (Kolbeinsey ridge) and a major dislocation simulating a fault locked at 10 kilometers depth and affected by a constant slip below this limit simulating ductile slip at depth.*

In addition small wavelength tendencies has been adjusted by superficial faults.

This model based on our 1995-1997 TGN comparison revealed extension and strike–slip movements 3 to 4 times larger than the average velocity. The transform motion is locked on a large (150x10 km) fault surface and this represents the main risk for destructive earthquakes in the near future. From a mechanical point of view, the lockage could be due to the increase of normal stress on this surface following the double opening north and south of the fault zone.

1997-1999 velocity field

Comparison with the 1995–1997 velocity field

32 common points have been measured during the 3 campaigns (1995, 1997 and 1999). In comparison with to the first period and using the same reference point located in the southern part of the network, the 1997-1999 velocity field has the following main characteristics:

The eastward tendencies north of Tjörnes are always present but the average velocity has been divided by 2. We observed less than 1 cm/year of dextral displacement north of Tjörnes. The NNE displacements observed in 1995-1997 on both side of the Húsavík fault are always observed but their component are in the opposite direction of those observed during the first period. These points are now moving to the south also with smaller velocities. Similar conclusions are drawn with the 2 points on Flateyjardalur. The large displacement observed in 1995-1997 for Flatey is no more observed.

From this comparison we can argue that the displacements on Tjörnes have varied significantly in sense and size in less than 2 years. A model compatible with both period is presently being elaborated.

Velocities for the points added to the 1995 network

13 new points have been added to the 1995 network both to the east and to the west to the first study area. All eastern points shows very similar ESE azimuth of displacements. A difference of 1 cm/year has been observed between the points respectively inside and outside the Krafla fissure swarm. This demonstrates the activity of the eastern margin of this swarm. The points added to the west of the network show similar displacements to those located on Flateyjardalur, which mean a general sinistral displacement of the SW area, relatively to a point located in the southern middle part of the network.

3.6.3.2 Task 3.2: 1992-1998 deformation of the Krafla volcano

The Krafla fissure swarm, in the North Iceland rift zone, last underwent rifting between 1975 and 1984. Activity mostly concentrated at the Krafla volcano although it occurred all along the fissure swarm. We formed interferograms of ERS-SAR scenes (Figure 40) covering the area of Krafla with time span values of up to six years (1992–1998). This long period interferograms have been revealed steady rate of deformation at Krafla volcano, N–Iceland. The deformation rate at Krafla and within the fissure swarm, has values reaching +2.1 cm/year in the ground to satellite direction, at the volcano. The area affected by deformation extends 20 km both north and south of the volcano. The best fit dislocation model consists of sills, to the north and south of the volcano, and a magma chamber, located below the volcano, all of them undergoing contraction. The contraction of the magma chamber induced a subsidence that has been constant in rate during all of the studied period (Figure 41).

3.6.4 Task 4: Effects of fluid pressure on faulting

It is well known that seismogenic faulting is normally associated with zones of fluid overpressure (Guðmundsson 1999; Guðmundsson 2000a; Guðmundsson 2000b); in absence of fluids there would be hardly any tectonic earthquakes. The origin and magnitude of the fluid overpressure are, however, uncertain. For example, for the San Andreas fault, there are two models; one assuming the fluids to be of local and shallow origin, the other assuming the fluids to be of deep origin.

In order to throw light on the potential fluid overpressure in a major fault zone, as a part of a general study on the effects of fluid pressure on the probability of fault-slip, field measurements were made of 1717 mineral-filled veins in the damage zone in deeply eroded and

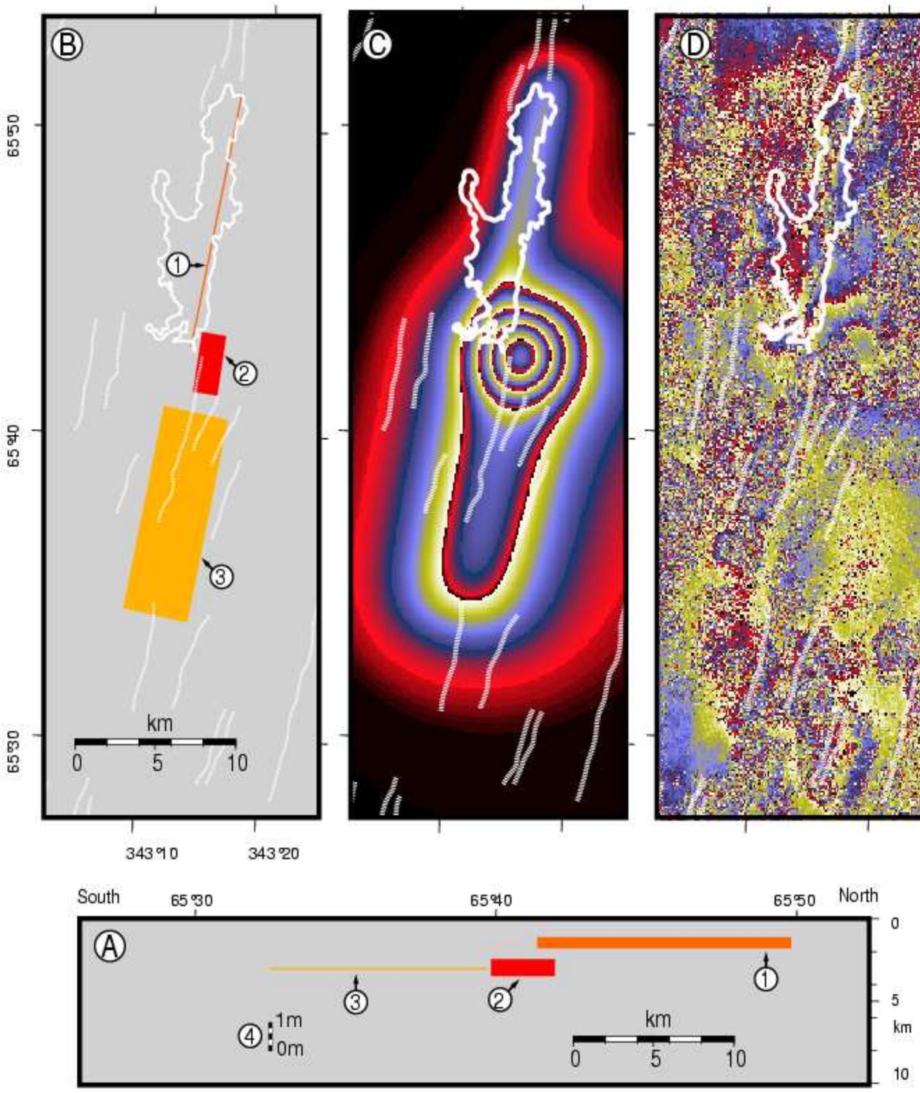

Figure 40. *Model geometry in cross section (A) and plan view (B), best fit model for the 6 year interferogram (C) and residual interferogram, i.e. model minus 6 year interferogram, (D) for the 6 year interferogram (after Henriot et al. 2000). Model elements represented in (A) and (B) are the northern sill (1), the magma chamber (2) and the southern sill (3). The model element dimensions represent their real geometries. The thickness of the lines in the cross-section (A) is proportional to the amount of contraction applied to the discontinuity in the model. The scale for the contraction in the 6 year interferogram model is given in (4).*

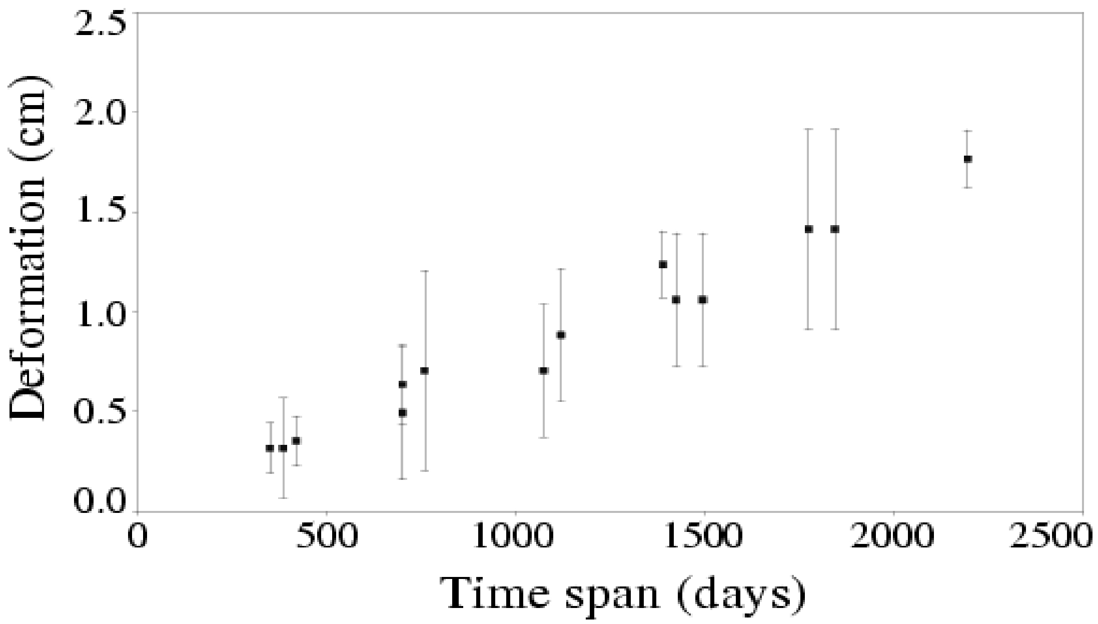

Figure 41. *Deformation at Leirhnjúkur, in the ground to satellite direction, with respect to the time span of the corresponding interferogram (after Henriot et al. 2000). The error is a function of the visual quality (Q) of the interferogram, 0 for no readability, 9 for perfect readability. It has been estimated as 1/Q of a fringe.*

well-exposed parts of the Tjörnes fracture zone, particularly on the Flateyjarskagi peninsula (Guðmundsson 1999; Guðmundsson 2000c; Guðmundsson et al. 2000). Most veins are composed of quartz, chalcedony and zeolites, strike roughly parallel or perpendicular to the fault zone, and are members of dense palaeo–fluid transporting networks. A common vein frequency in these networks is 10 veins per meter. Cross-cutting relationships indicate that 79% of the veins are extension (mode I) cracks; 21% are shear cracks. The apertures of most veins, measured as mineral-fill thicknesses, are from 0.1 mm to 85 mm, and the aperture frequency distribution is a power law. The outcrop trace lengths of 384 veins (of the 1717) could be measured accurately. These 384 veins are mostly small and range in length from 2.5 cm to 400 cm, in aperture from 0.01 cm to 0.9 cm, and have an average length/aperture ratio of about 400. Using simple fracture mechanics models, and the appropriate elastic properties of the host rock, this length/aperture ratio indicates and average fluid overpressure during vein formation of 20 MPa. If this fluid pressure acted on a potential fault plane, the effective normal stress across that plane would be zero or negative and, therefore, reduce the driving shear stress needed to trigger slip on that plane to only 4-6 MPa. This compares well with the most common stress drops, estimated at 3-6 MPa, for earthquakes worldwide.

Another potentially strong effect of fluid pressure on the probability of fault-slip in seismic zones is dyke emplacement in the nearby volcanic zones (Guðmundsson 2000d). In this model, dyke injection (and normal faulting) in the volcanic systems can lock or unlock the Húsavík-Flatey fault and the central parts of the South Iceland seismic zone. Dyke injection in the parts of the North and East volcanic zones between the Húsavík-Flatey fault and the South Iceland seismic zone tends to open (unlock) these zones and trigger seismogenic faulting. By contrast, dyke injection south of the South Iceland seismic zone and north of the Húsavík-Flatey fault tends to lock these faults and suppress their seismogenic faulting. Similarly, dyke injection in the north part of the West volcanic zone tends to lock, but dyke injection in its south part (including the Reykjanes peninsula) to unlock, the South Iceland seismic zone. Locking by dyke injection, however, is always temporary because plate pull gradually relaxes the compressive stresses generated by the dykes.

In terms of this model, the largest historical eruption in Iceland, Laki 1783, may have triggered the largest known earthquake sequence in S-Iceland, that of 1784. Conversely, the Húsavík-Flatey fault has recently experienced locking by dyke injection. There was considerable seismicity associated with the Húsavík-Flatey fault until early 1976. Then dyke injection and normal faulting in the northernmost part of the Krafla volcanic systems generated horizontal compressive stresses encouraging sinistral movement on the otherwise dextral Húsavík-Flatey fault, thereby locking the fault. Renewed seismicity on the Húsavík-Flatey fault, since February 1994, indicates that the Húsavík-Flatey fault is currently being unlocked by normal plate-pull movements which gradually relax the horizontal compressive stresses generated by the 1976 dyke. The unlocking began at the westernmost part of the Húsavík-Flatey fault, at the greatest distance from the 1976 dyke.

3.6.5 Task 5: Numerical models on faults and fault populations

Part of this work has focussed on quantitative field studies and modelling of the linking up of fractures into normal faults (Acocella et al. 2000) and strike-slip faults (Belardinelli et al. 2000). Part of the work, however, has focussed on the general evolution of the seismic zones, exploring the model of a stress-field homogenization being a necessary condition for the generation of large earthquakes (Guðmundsson and Homberg 1999). The work on the South Iceland seismic zone made in collaboration with Maria Elina Belardinelli and Maurizio Bonafede (Belardinelli et al. 2000) is described in Subproject 7.

Detailed, quantitative field and photogeological studies were made of the interaction and linkage of extension fractures and normal faults (Acocella et al. 2000), a fault type that is common in the South Iceland seismic zone and the Tjörnes fracture zone. 90 zones of interacting fracture segments in Holocene pahoehoe lava flows of the rift zone of Iceland were studied, each zone being located between a pair of extension fractures or a pair of normal faults, with lengths from tens of meters to several kilometers. Of all the zones, only 7% are underlapping, whereas 93% are overlapping and mostly with hook-shaped fracture pairs. The length/width ratios of most overlapping zones are from 2-6, with a mean value of 3.5. In the overlapping zones, most fracture pairs show moderate shear (strike-slip) components which are related to local variations in the extension (opening) directions. Vertical displacements on normal faults decrease as the overstep and length of overlapping zone increase; both, in turn, are proportional to the total lengths of the faults forming the pair. During their evolution, these zones develop from an underlapping stage, through an overlapping stage (the most common configuration) and, finally, to a linkage stage. The geometrical features of overlapping spreading centres at mid-ocean ridges show great similarities to those reported here. These similarities indicate that the architecture and evolution of overlapping zones are scale independent.

It is proposed that on entering crustal parts where the state of stress is unfavourable to any particular type of seismogenic faulting, the fault propagation becomes arrested (Guðmundsson and Homberg 1999). This model is supported by field and numerical studies on the propagation of fractures of various types (Guðmundsson 2000e). It follows that prior to the propagation of an earthquake fracture, the stress conditions in the zone along the whole potential rupture plane must be homogenized. The proposed homogenization of the stress field in a large rock volume as a precursor to large earthquakes implies that by monitoring the state of stress in a seismic zone, its large earthquakes may possibly be forecasted.

3.6.6 Task 6: Analyzing the fracture properties of Icelandic rocks in the laboratory

Our main overall objective has been to determine changes in the strength, deformability and physical properties of rocks subjected to elevated temperatures. In order to achieve this objective, we have designed and constructed an apparatus for the measurement of rock fracture and deformation properties at high temperatures and pressures, and performed a series of experiments to measure these properties.

3.6.6.1 Task 6.1: Apparatus development

We previously reported on the design and manufacture of an environmental cell for the measurement of fracture mechanics parameters under high-temperature/high-pressure conditions (PRENLAB–2 first annual report). We have now extended the range of capabilities of this apparatus so that it can perform four different types of experiment: (1) rock fracture mechanics experiments, (2) confined compression experiments, (3) confined extension experiments, and (4) direct tension experiments. The extended operating conditions are: (i) confining pressure up to 70MPa, (ii) test temperatures up to 500°C with water as the confining/pore fluid, (iii) test temperatures up to 900°C with dry nitrogen as the confining/pore fluid, and (iv) sample diameters from 25 mm to 60 mm. A schematic diagram of the apparatus and its operating modes is given in Figure 42.

3.6.6.2 Task 6.2: Experimental results and discussion

Starting material
The measurements reported here were made on samples of a macroscopically isotropic basalt collected from a roadstone quarry located southeast of Reykjavík, Iceland. Microscopically it has an aphyric texture, comprising euhedral laths of unaltered plagioclase averaging 0.2 mm in length, and anhedral augite microphenocrysts averaging 0.1 mm in diameter, with accessory anhedral oxides up to 0.1 mm in diameter. No free quartz was visible under either optical or SEM microscopy.

Summary of previously reported results
We previously reported results of measurements made at room temperature on samples that have previously been heat-treated to temperatures up to a maximum of 900°C in order to induce thermal crack damage. In a shallow crustal environment where the geothermal gradient is anomalously high, such as in Iceland, thermal stresses may well be large enough to induce such fracturing. Furthermore, where enough fractures propagate and link up to provide an interconnected network, they can provide permeable pathways for fluid flow which can in turn lead to embrittlement and weakening of the rock. The thermal cracking was monitored by measuring the compressional (P) and shear (S) wave velocities through the samples both prior to and following heat-treatment. Both P-wave and S-wave velocities remained essentially constant up to 400°C, with values of about 5.3 km/s and 3.0 km/s respectively. For higher temperatures the velocities decrease rapidly, so that by 800° they decreased to about 3.4 km/s and 2.2 km/s repectively.

We presented results from a series of fracture toughness measurements on heat-treated specimens of Icelandic basalt using the ISRM recommended methodology. Fracture toughness at ambient temperature was 2.71 MPa/m^2, and this remained essentially constant up to 400°C. There was then a very rapid decrease in fracture resistance between 400°C and 600°C, with relatively little change between 600°C and the highest heat-treatment temperature of 900°C. This pattern of behaviour was considered to be entirely consistent with the wave velocity data.

Finally, we presented the results of permeability measurements on heat-treated cores of basalt. The mean permeability of the basalt prior to heat-treatment was 9.4 nanodarcy (9.4 x $10^{-21}$$m^2$). Similar to the previous results, the permeability remained essentially constant after heat treatment to temperatures up to 300°C, and showed only a slight increase after treatment to 400°C. At higher temperatures, however, the normalized permeability changed dramatically, increasing by an order of magnitude at 700°C and by a factor of 40 by 800°C. We concluded that such a large increase in permeability was unlikely to result merely from the increase in the number or size of thermally-induced cracks, but from crack linkage processes

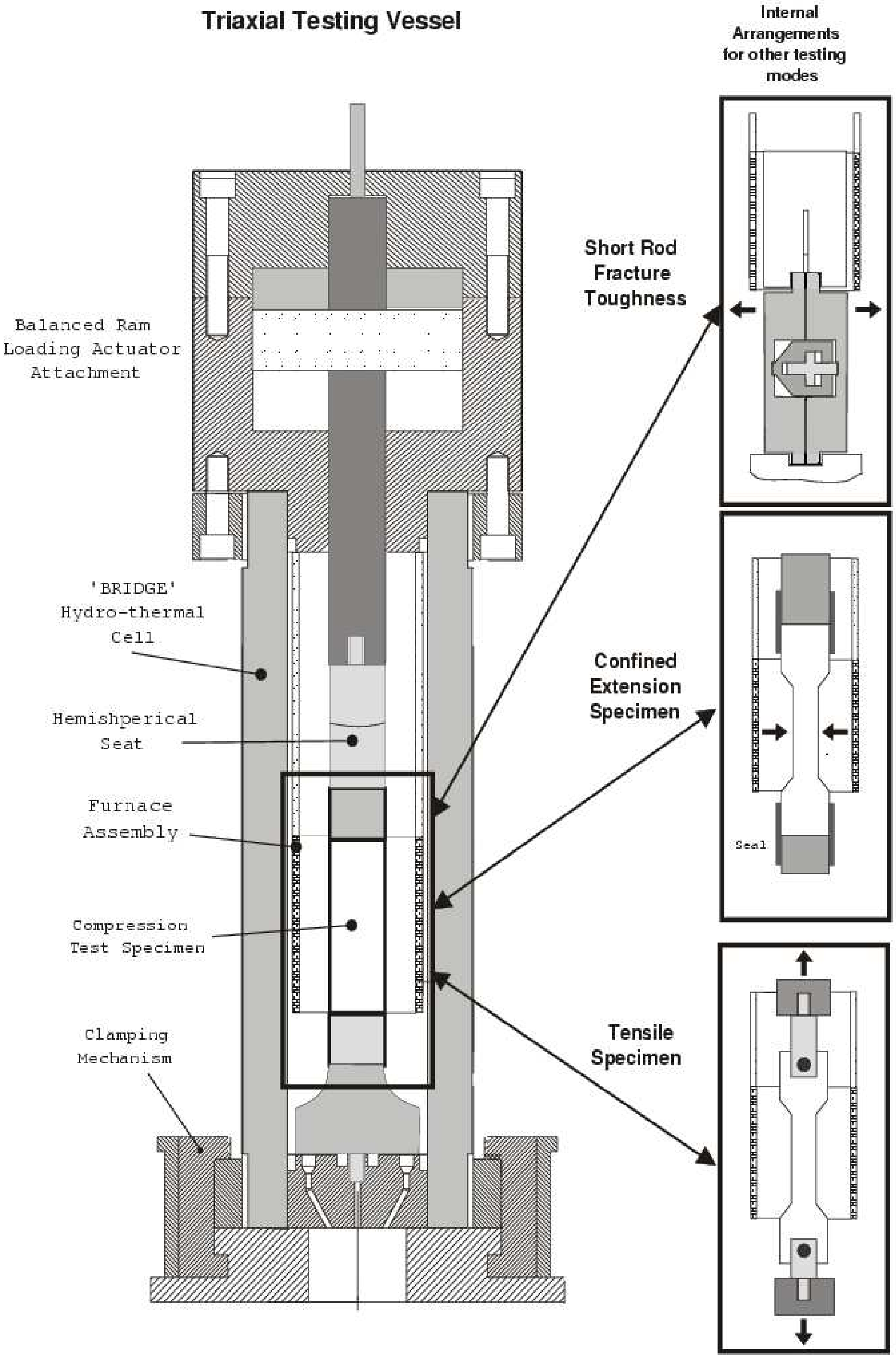

Figure 42. *Schematic diagram of the apparatus and its operating modes.*

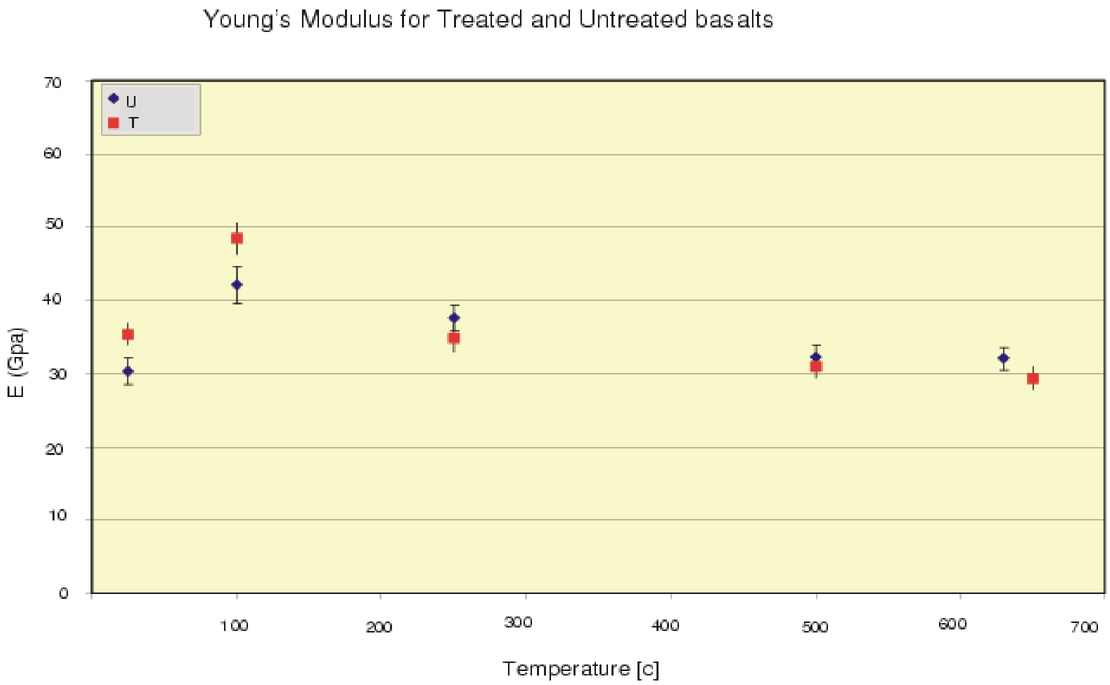

Figure 43. *Results of measurements of Young's modulus of elasticity for both PHT and NHT specimens up to 650°C.*

above some percolation threshold to form extensive sample-spanning permeable pathways for fluid flow.

New experimental results

In our most recent programme, we have performed suites of experiments on both pre-heat treated (PHT) and non-heat-treated (NHT) specimens of basalt. The reason for this is that when a measurement is made at elevated temperature, there are two potential effects on the measurement, which act simultaneously. First, the high temperature can lead to thermal cracking and hence in a change to the microstructure of the material being tested. Second, the temperature can influence the actual deformation mechanism. By performing these two suites of experiments, we hope to be able to discriminate betwen these two effects.

Figure 43 shows the results of measurements of Young's modulus of elasticity for both PHT and NHT specimens up to 650°C (all PHT specimens were pre-heated to 750°C). The two datasets show very similar trends and values, suggesting that the pre–heat treatment has relatively little effect on this parameter. As the temperature is raised from room temperature to 100°C, there is a significant increase in the modulus. This may appear counterintuitive, but is consistent with previously reported data on deformation properties of brittle rocks over this temperature interval (e.g. Meredith and Atkinson 1985). It is considered to be due to thermal expansion leading to microcrack closure and hence higher stiffness and crack resistance. At all higher temperatures, the modulus decreases monotonically with increasing

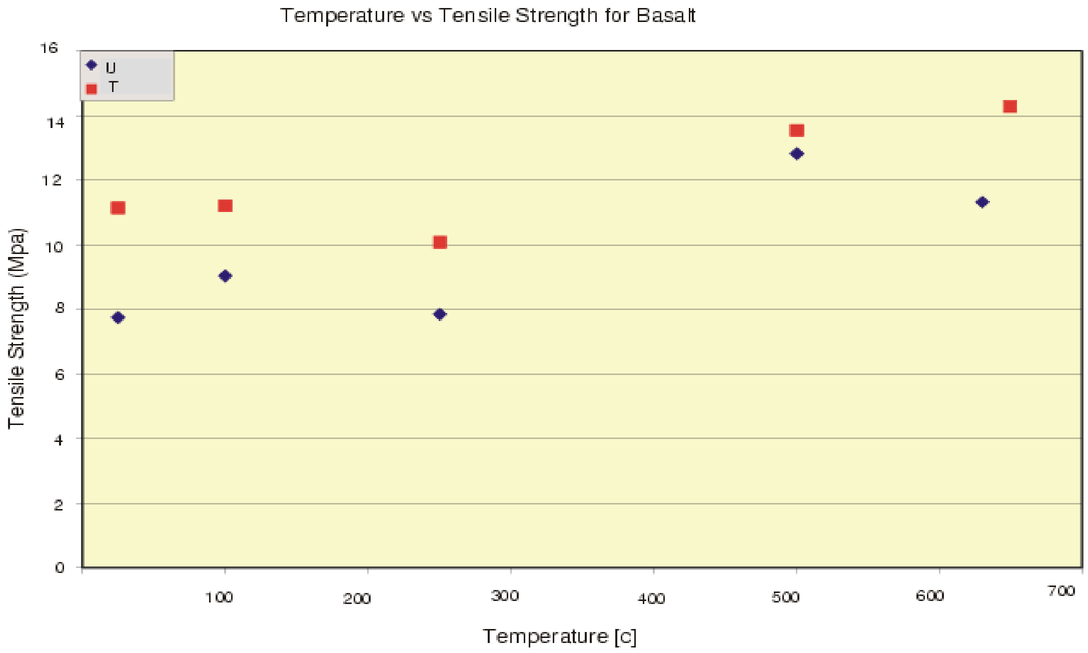

Figure 44. *Results of measurements of tensile strength up to 650°C for both PHT and NHT specimens.*

temperature.

Figure 44 shows the results of measurements of tensile strength up to 650°C, again for both PHT and NHT specimens. Here, the strength values for both datasets at first decreases for temperatures up to 250°C, and then actually increases for temperatures up to 650°C. Furthermore, tensile strength values for the PHT specimens are consistently higher than for the NHT specimens. The reasons for these variations are not fully understood at present, but we consider that it is likely to be due to local microplasticity at crack tips. Microplasticity is generally enhanced by high stresses, and hence is more likely to occur at crack tips than elsewhere in the rock because of the high stress concentrations at these sites. Microplasticity acts to blunt crack tips and therefore make crack propagation more difficult. If this is the case, then we could expect specimens to be less susceptible to tensile crack growth at higher temperatures. Furthermore, we might also expect the PHT specimens to exhibit higher strength because they have previously been subjected to a higher temperature (750°C) and hence potentially to more microplasticity. This explanation is consistent with the observation of little difference in modulus between PHT and NHT specimens. The modulus is controlled by the mere presence of cracks and not by their growth, and hence crack tip blunting is not likely to affect modulus values.

These are very new results, and we will be investigating this possible explanation microstructurally with a combination of scanning electron and optical microscopy.

3.6.7 Conclusions

Due to the high temperature gradient and low lithostatic stress, thermal cracking may be an important process in controlling fracture in Icelandic crust. In the absence of confining stress, such cracking starts in the temperature range 300° to 400°C in fresh basalt. Thermal cracking leads to increased fluid permeability above about 300°C, with the permeability increasing very non-linearly with temperature. Thermal cracking also appears to lead to significant decreases in mechanical strength and resistance to crack propagation of PHT specimens when tested after cooling at room temperature.

However, a somewhat different picture emerges when both PHT and NHT specimens are actually tested at elevated temperatures. There appears to be an increase in tensile strength with increasing temperature. We attribute this to enhanced microplasticity at highly stressed crack tips effectively blunting cracks.

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3.7 Subproject 7: Theoretical analysis of faulting and earthquake processes

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3.7.1 Subpart 7A: Ridge-fault interaction in Iceland employing crack models in heterogeneous media

3.7.1.1 Task 1: Magma upwelling as driving mechanism for the stress build-up in the elastic lithosphere

Tensile cracks are often employed to model magma migration in rift zones or within volcanic edifices, through lateral or feeding dykes. In a crack model, the overpressure of magma Δp with respect to the horizontal stress in the host rock, is assumed to be responsible for dyke opening and propagation. Most crack models of dykes have been developed so far in homogeneous media. The most simple heterogeneous medium has been considered, made up of two welded half-spaces, characterized by different elastic parameters. The analytical solutions available for the elementary dislocation problem in such a medium (Bonafede and Rivalta 1999) have been employed to set up an integral equation with generalized Cauchy kernel, representing the condition for static equilibrium. The unknown in such a problem is the dislocation density distribution, whose singular behaviour has been studied near the crack tips and near the intersection with the interface between the two media. When the crack is in half-space 1 but touches the interface, the order of singularity of the dislocation density distribution at the interface changes from the classical behaviour $\sim r^{-1/2}$ to $\sim r^{-b}$

(where r is the distance from the interface) and the order of infinity b is obtained solving a trascendental compatibility equation; some results are shown in Table 6.

$m = \mu_1/\mu_2$	∞	10	5	2	1	0.5	0.2	10^{-1}
b	0.255	0.312	0.352	0.430	0.500	0.576	0.678	0.752

Table 6. *Crack touching the interface.*

A crack crossing the interface $z = 0$ between the two half-spaces with rigidities μ_1 (in $z > 0$) and μ_2 (in $z < 0$) has been considered in detail. A system of generalized Cauchy equations is obtained, which is solved for the dislocation density distributions of each crack section. An internal singularity in the dislocation density distribution appears at the intersection between the crack plane and the interface. This singularity is again of the type r^{-b} on both sides of the interface and its order b depends only upon the elastic parameters of the media in welded contact and the ratio between the crack lengths in the two half-spaces (see Table 7). More specifically, the order of singularity b does not depend on the stress drop.

$m = \mu_2/\mu_1$	1	0.5	0.1	0.05	0.001
b	0	0.030	0.170	0.208	0.245

Table 7. *Crack crossing the interface.*

The horizontal stress component induced by crack opening is plotted in Figure 45, assuming 5 MPa overpressure within the crack. From a comparison with solutions pertinent to a homogeneous medium, it appears that layering can be responsible of stress changes, localized along the the interface, which may be considerably higher than the overpressure within the dyke. These results provide useful hints for the interpretation of induced seismicity in rift zones and in volcanic areas. The detailed results of this research are reported in Bonafede and Rivalta (1999).

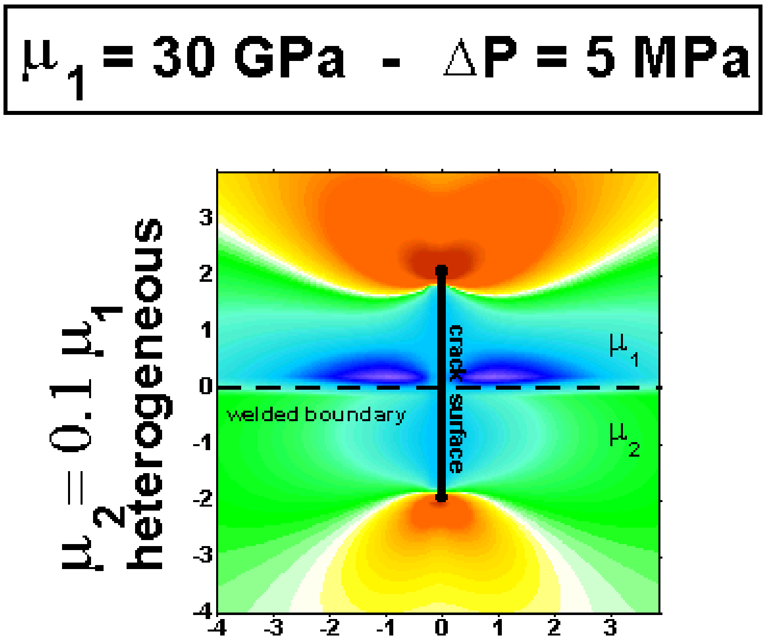
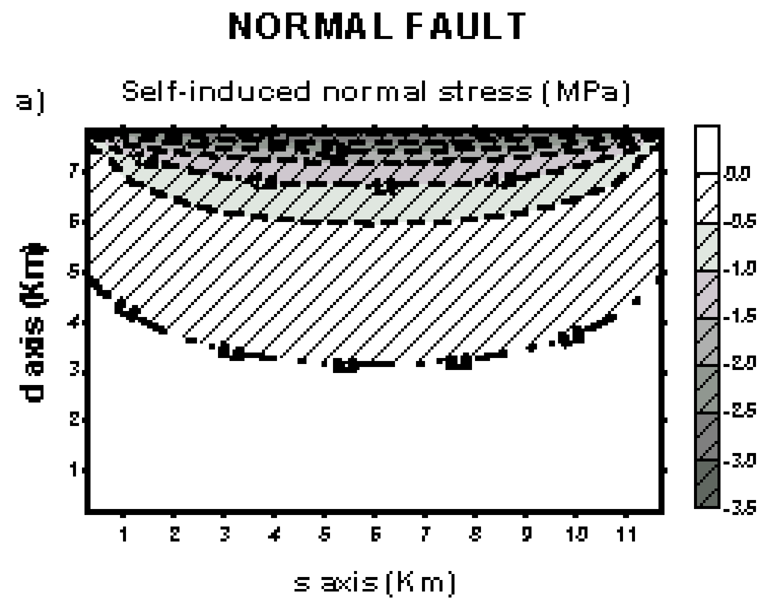


Figure 45. *Horizontal normal stress induced by rifting in proximity of a structural discontinuity (horizontal dashed line). The harder side of the interface is affected by strong compressive stresses. Other stress components (not shown) are also affected by layering, but to a lesser extent. Rifting is modelled as a tensile crack with overpressure* Δp.

3.7.1.2 Task 2: Space-time evolution of the stress field following earthquakes and episodes of magma upwelling

Mechanical effects left by an earthquake on its fault plane, in the post-seismic phase, were investigated employing the "displacement discontinuity method" and imposing the release of a constant, uni-directional shear traction. Due to unsymmetric interaction between the fault plane and the free surface, significant normal stress components are left over the shallow portion of the fault surface after the earthquake (Figure 46): these are compressive for normal faults, tensile for thrust faults, and are typically comparable to the stress drop. In Figure 46 the s-axis is along the strike of the fault, the d-axis is along the dip (positive upwards). Several observations can be explained from the present model: low-dip thrust faults and high-dip normal faults are found to be favoured, according to the Coulomb failure criterion, in repetitive earthquake cycles; the shape of dip-slip faults near the surface is predicted to be upward-concave; the shallow aftershock activity commonly observed in the hanging block of a thrust event is easily explained. The detailed results of this research are reported in Bonafede and Neri (2000).

The study of the effects induced by structural inhomogeneities on the stress and displacement fields around strike-slip faults has been recently completed. An elastic medium is considered, made up of an upper layer bounded by a free surface and welded to a lower



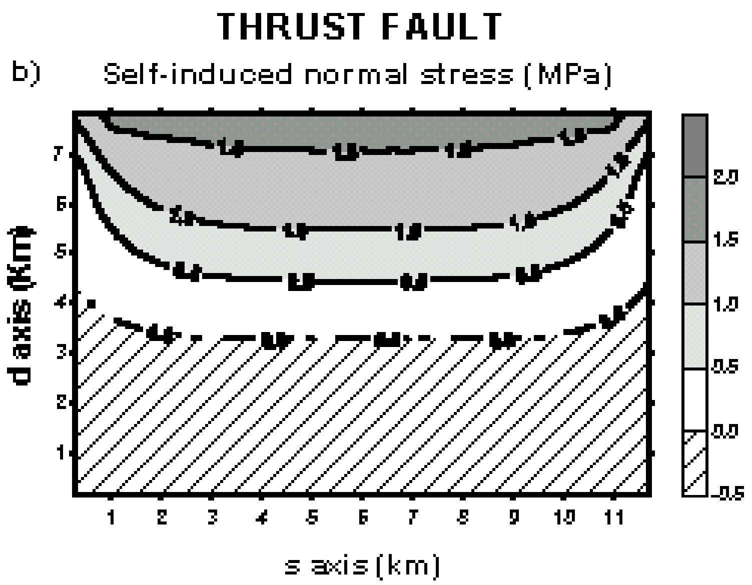


Figure 46. *Normal stress induced by uniform stress drop over a high-dip normal fault and a low-dip thrust fault. The near-surface part of a normal fault is affected by compressive stress, while tensile stresses act there for a thrust fault.*

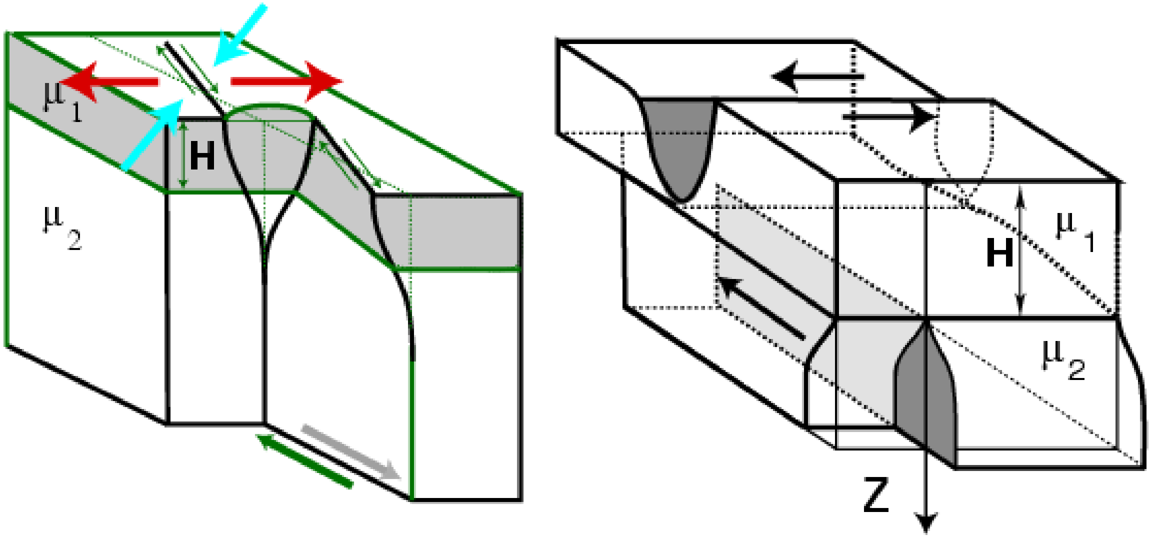


Figure 47. *According to the stress-drop-discontinuity condition, en-echelon fault segments, rotated in the direction of the tensile principal stress axis, must accompany brittle faulting at depth (left). On the contrary, anelastic deformation at depth is preferably accompanied by antithetic faulting in brittle surface layers (right).*

half-space characterized by different elastic parameters. Shear cracks with assigned stress drop are employed as mathematical models of strike-slip faults which are considered as vertical and planar. If the crack is entirely embedded within the lower medium (case A), a Cauchy-kernel integral equation is obtained, which is solved by employing an expansion of the dislocation density in Chebyshev polynomials. If the crack is within the lower medium but it terminates at the interface (case B), a generalized Cauchy singularity appears in the integral kernel. This singularity affects the singular behaviour of the dislocation density at the crack tip touching the interface. Finally, the case of a crack crossing the interface is considered (case C). The crack is split into two interacting sections, each placed in a homogeneous medium and both open at the interface. Two coupled generalized Cauchy equations are obtained and solved for the dislocation density distribution of each crack section. An asymptotic study near the intersection between the crack and the interface shows that the dislocation densities for each crack section are bounded at the interface, where a jump discontinuity appears. As a corollary, the stress drop must be discontinuous at the interface, with a jump proportional to the rigidity contrast between the adjoining media. This finding is shown to have important implications for the development of geometrical complexities within transform fault zones: planar strike-slip faults cutting across layer discontinuities with arbitrary stress drop values are shown to be admissible only if the interface between different layers becomes unwelded during the earthquake. Planar strike-slip faulting may also take place in mature transform zones, where a repetitive earthquake cycle has already developed. Otherwise, the fault cannot be planar: we infer that strike-slip faulting at depth is plausibly accompanied by en-echelon surface breaks (Figure 47) in a shallow sedimentary layer (where the stress drop is lower than prescribed by the discontinuity condition), while

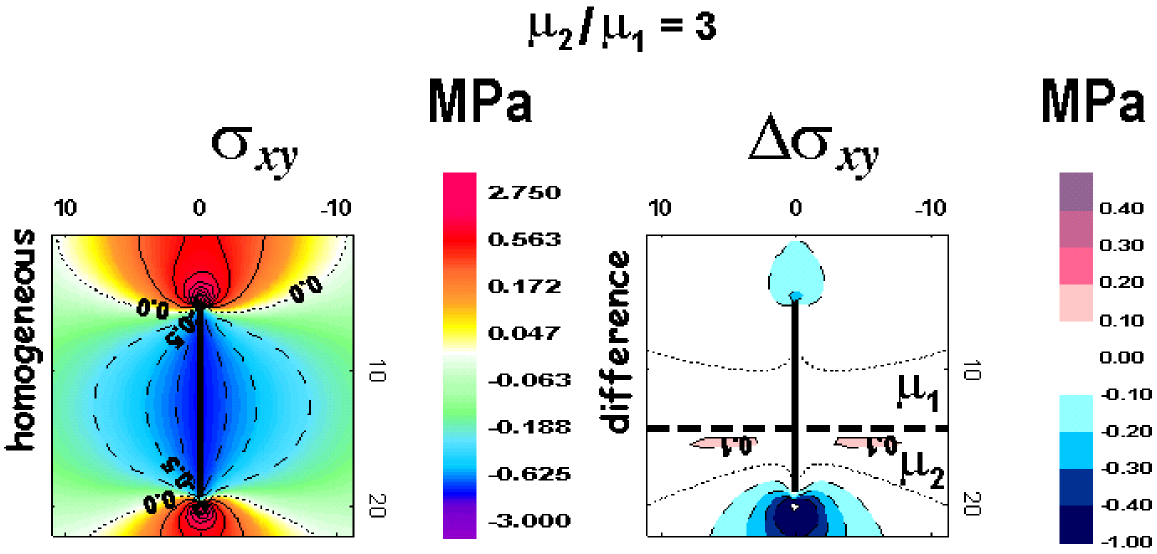


Figure 48. *Stress concentration induced by strike-slip faulting across layer interfaces. The panel on the right shows the incremental stress present in the layered model compared to the homogeneous half-space model (on the right). Accurate re-location of aftershocks in several areas of the world actually show a sharp concentration of events along bedding planes.*

ductile deformation (or steady sliding) at depth is preferably accommodated by antithetic faulting in the upper brittle layer (endowed with lower rigidity but higher stress), giving rise to bookshelf faulting (Figure 47). Results of this research were presented at several international conferences. A paper has been submitted for publication (Bonafede et al. 2000). The South Iceland seismic zone provides several instances of application of both types of complexities.

3.7.1.3 Task 3: Secondary earthquake fractures generated by a strike-slip fault in the South Iceland seismic zone

Most earthquakes in the South Iceland seismic zone occur on N-S trending dextral strike-slip faults. The resulting rupture zones display complex en-echelon patterns of secondary structures including NNE-trending arrays of (mostly) NE-trending open fractures (O.F.) and hillocks.

Three spatial scales characterize the surface faulting pattern: the length of the main fault (M.F. 104 m), the arrays here interpreted as surface evidence of secondary faults (102 m) and the individual O.F. (10 m). In order to improve our understanding of the genetic relationship between the O.F. and the M.F. we computed the stress field induced by slip on a buried M.F. using a dislocation model in a layered half-space: the fault surface is assumed to be embedded in the basement rock, topped by a softer near-surface layer. The O.F. were preliminarily considered as pure mode-I cracks opening in the near surface layer in the direction of the maximum (tensile) principal stress. Alternatively, secondary fractures were interpreted, as

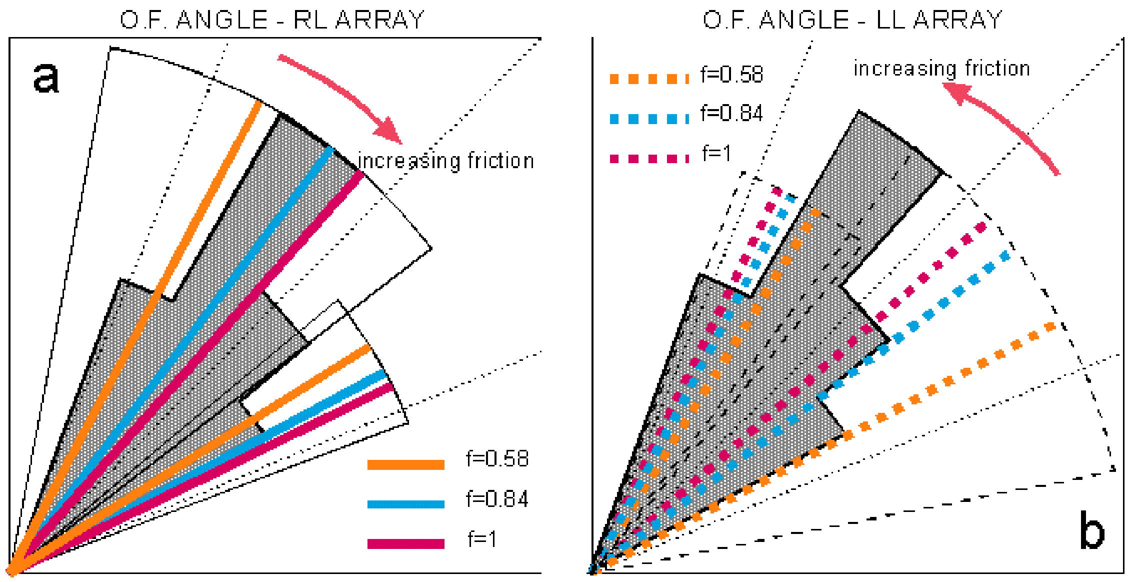


Figure 49. *Open fractures (O.F.) angles predicted from the combined effect of the main fault rupture and secondary fault (S.F.) rupture along the strike direction. The observed range and relative frequency of O.F. angles is shown shaded (from Bjarnason et al. 1993). If the friction coefficient f varies between 0.2 and 1.5 predicted O.F. angles vary within the circular sectors contoured in black. Solid lines in panel (a) represent the angle expected for O.F. belonging to dextral arrays. Dashed lines in panel (b) represents the angle expected for O.F. belonging to sinistral arrays. In both panels, longer lines indicate the predicted mixed-mode O.F. trends, while shorter lines indicate pure tensile trends. Mixed-mode O.F. are assumed to share the same style of faulting (sinistral or dextral) with the array to which they belong. The dotted lines indicate 22.5°, 45° and 67.5° directions (for reference). Coloured lines refer to particular friction values (indicated). Friction increases as indicated by the red arrow.*

mixed-mode cracks, slipping at depth as shear cracks and opening near the surface due to low confining pressure. The Coulomb failure function after the earthquake (obtained summing the M.F. stress change and the lithostatic stress) suggests that secondary faulting (S.F.) can be expected to occur in response to the main rupture below the upper soft layer down to few hundreds of meter depth. The total stress change induced by the M.F. and the S.F. (of smaller scale) is shown to yield quantitative explanations of the complex geometry observed in the fault region in terms of simple frictional laws and friction coefficients very close to those measured in the lab (Figure 48).

3.7.1.4 Meetings and conferences

XXIII EGS General Assembly, Nice, France, April 20-24, 1998.
LXXXIV Congresso Nazionale Societa Italiana di Fisica, Salerno, Italy, September 28 - October 2, 1998.
The third PRENLAB-2 workshop, Strasbourg, France, March 31, 1999.
XXIV EGS General Assembly, The Hague, The Netherlands, April 19-23, 1999.
XXV EGS General Assembly, Nice, France, April 25-29, 2000.
The fifth PRENLAB-2 workshop, Nice, France, April 27, 2000.
International School of Solid Earth Geophysics, Erice, Italy, June 17-23, 2000. 17^{th} course: Fault interaction by stress transfer: new horizons for understanding earthquake occurrence.

3.7.2 Subpart 7B: Modelling of the earthquake related space–time behaviour of the stress field in the fault system of southern Iceland

In the framework of the PRENLAB-2 project a model study which started during the PRENLAB project was continued to obtain forward models of the stress field and stress changes in the South Iceland seismic zone (SISZ).

This proposal has the target to model the space–time development of the stress field using data on strain and stress changes from the other experiments and from databases.

Two models were prepared during PRENLAB-1 and PRENLAB-2:

- a model of the South Iceland seismic zone and the adjacent part of the eastern volcanic zone.
- a scheme comprising the main ridge parts on Iceland and the North Atlantic ridge to the North and to the South of the island, including both the faults and the load due to Katla and Hekla volcanoes.

It was modelled:

- the changes in crustal strain and stress due to earthquakes and aseismic movement in the fault system of the South Iceland seismic zone.
- the interaction of faults.
- the mutual influence between volcanic and earthquake activity, e.g. magmatic upwelling and shearing at fault zones.

3.7.2.1 The model for the earthquake sequence at the SISZ

During PRENLAB-2, this model was further improved. The main features of this model are given again to ease comparison with the new results.

The method

The forward modelling of stress fields is done by applying static dislocation theory to geodetic data and data obtained through the seismic moments from seismograms. It allows to calculate displacements, strain and stresses due to double-couple and extensional sources in layered elastic and inelastic earth structures. Besides the change in displacement during the event, the changes caused by the movement of plates are included (for further details see e.g. Roth 1989).

Usually, for earthquake hazard estimation, the location, the magnitude and the statistically estimated recurrence period of former events is used. To improve this, here the rupture length and width as well as the tectonic setting and the crustal deformation rates are considered while calculating the space time development of the stress field.

The targets

In general, with these models, Subpart 7B aims:

- to achieve a better understanding of the distribution of seismicity in space and time, its clustering and migration in Iceland.

- to provide models for the joint interpretation of the data gathered in the whole research programme, of which this is one part.
- to compare models of stress fields at SISZ to those for stress fields in other regions, e.g. the North Anatolian fault zone.
- to make a contribution to the intermediate-term earthquake prediction in this populated and economically important region of Iceland.

The tectonic setting
The SISZ is situated between two sections of the mid-Atlantic ridge, the Reykjanes ridge (RR) and the eastern volcanic zone (EVZ). Even though the angle between the SISZ and the neighbouring ridges is far from 90°, it is considered as a transform fault. Following the transform fault hypothesis, left-lateral shear stress is expected along the E-W striking zone. This is equivalent to right-lateral shear stress on N-S oriented rupture planes. In fact, earthquakes seem to occur on N-S trending en-echelon faults (cf. Einarsson et al. 1981; Hackman et al. 1990), They are located side by side between the Hengill-Ölfus triple junction, where the RR meets the low activity western volcanic zone (WVZ) and Hekla volcano, in the EVZ (cf. Einarsson et al. 1981) (Figure 50). As we further know from Subprojects 4 and 5, the orientation of the larger horizontal principal stress is NE–SW, i.e. fits to an active N-S or E-W trending fault, which is (at least for the period of those investigations) not a weak fault like the San Andreas fault (cf. Zoback et al. 1987). Moreover, the stress orientation seems to have been constant since Pliocene time.

In detail, the questions to be solved are:

- Do these events, placed on parallel faults, release all the energy stored in the 3-D volume of the SISZ?
- Do the earthquakes always take place in areas of high stress?
- What is the critical stress level? How large is its variability?
- Where are the highest stresses nowadays?

The area investigated extends from 18° to 24°W and from 63° to 65°N. The origin is set to 24°W, 64°N (cf. Figure 51) it includes the SISZ, ±1° north and south of 64°N, the SW edge of the EVZ, and the northeasternmost part of the RR.

The initial stress field
The initial stress field is determined as follows: A tensional stress acting N103°E (nearly parallel to the SISZ; cf. DeMets et al. 1990) is assumed, due to the opening of the mid-Atlantic ridge in the region adjacent to the transform fault. While this rifting induces mainly shear stresses in the SISZ with a small opening component, the rift segments (RR and EVZ) are modelled with tensional stress and a small shear stress contribution. Tensional stresses at both ridges are modelled as constantly being released to end up with zero values at the rifts. This induces additional stress in the transform zone. The stress magnitude, which is unknown, is set to a value that produces left-lateral shear stresses in E-W direction as

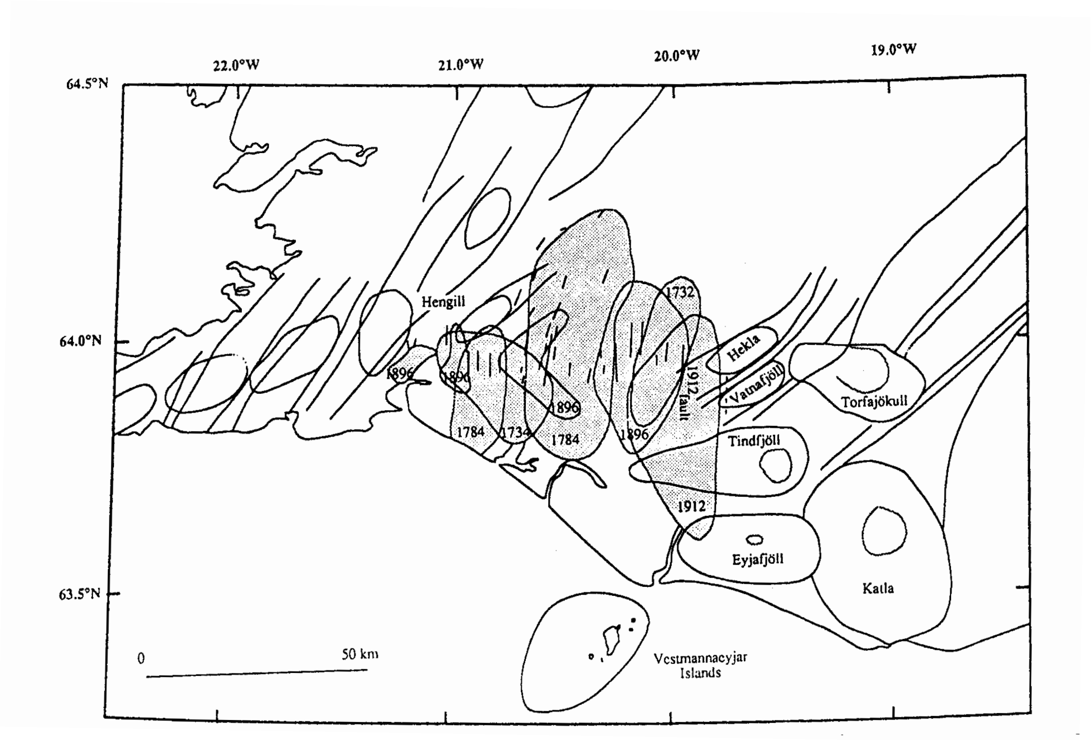


Figure 50. *The South Iceland seismic zone showing mapped surface breaks and regions in which over half of the buildings were destroyed in historic seismic events (after Einarsson et al. 1981). The north-south dashed line near Vatnafjöll indicates the estimated location of the fault on which the May 25, 1987, earthquake occurred (after Bjarnason and Einarsson 1991). The structural features and the coastline are after Einarsson and Sæmundsson (1987).*

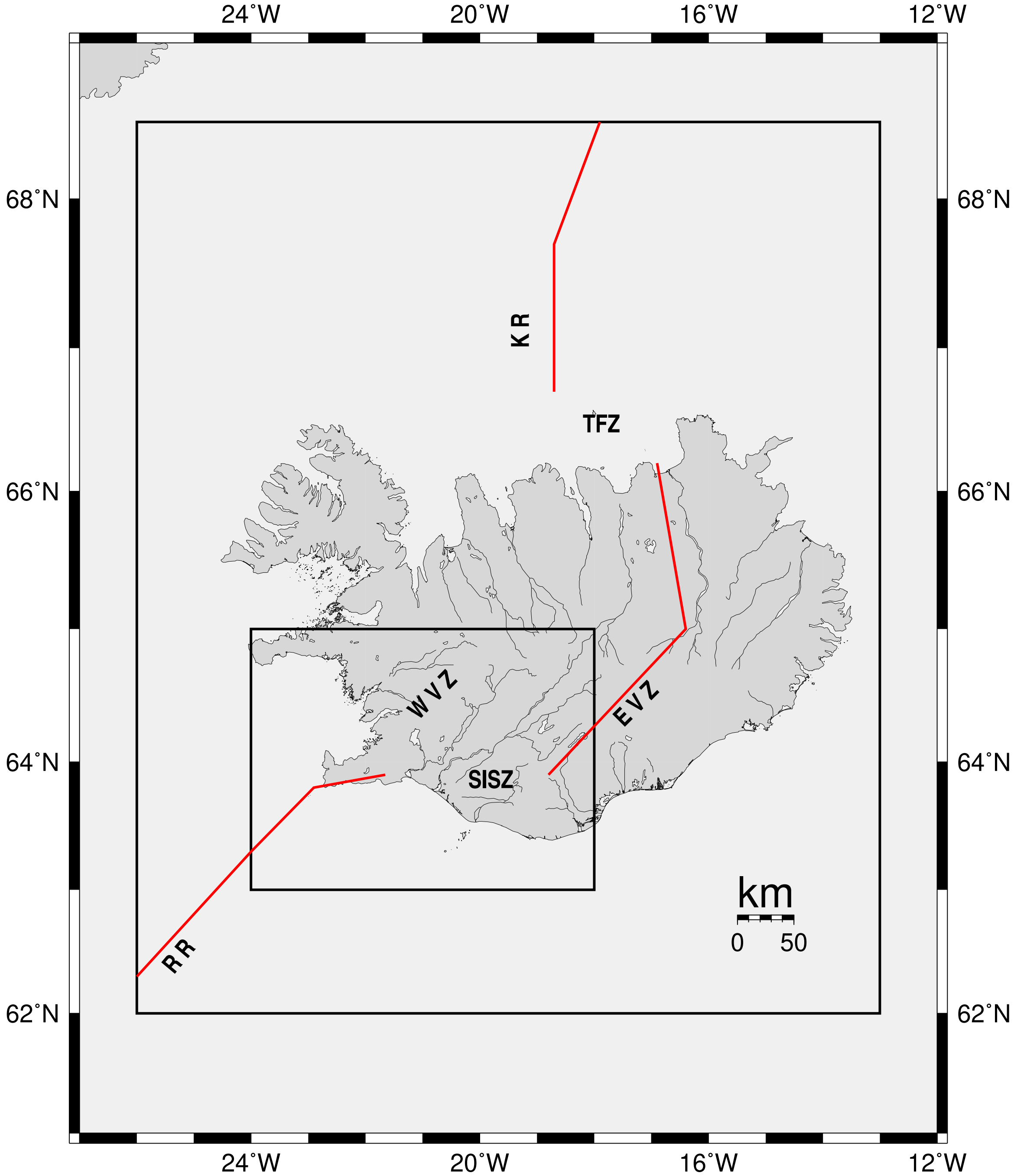


Figure 51. *Map of Iceland and surrounding area. Thick red lines indicate mid-Atlantic ridge segments, as used in the models. The smaller box shows the region of the model on the SISZ. The SISZ extends approximately between 21.4° W, 63.95° N to 18.8° W, 64° N. The large box gives the region for the Iceland rift model. RR: Reykjanes ridge, KR: Kolbeinsey ridge, WVZ/EVZ: western/eastern volcanic zone, SISZ: South Iceland seismic zone, TFZ: Tjörnes fracture zone.*

large as the stress drop determined for the largest event (M=7.1) in the studied earthquake sequence.

On this initial field, the stress changes due to earthquakes are iteratively superposed as well as the stress changes due to further spreading at the ridge segments. From global geodetic measurements an opening of 2 cm/year is found, e.g. in DeMets et al. (1990). This was used as a zero-th order approach but was reduced to only 1 cm/year, as discussed later. Further, as the simplest assumption, lacking other data, the spreading rate is taken to be constant during the modelled time period, even though this can be questioned as for instance the present debate on the stress increase in the New Madrid seismic zone shows (cf. Schweig and Gomberg 1999; Newman and Stein 1999) The stress field before every event is thus the sum of the initial field, the stress drop of all preceding events, and the plate tectonic stress build-up since the starting time of the model, which is set to 1706, when the first event in the series occurred.

Results were calculated for 280x220 test-points covering 280 km in E-W direction and 220 km in N-S direction. Stresses were computed for a homogeneous half-space, as a starting model. Although surface stress changes are calculated, these should be representative for crustal stresses using values for the moduli, that are typical for oceanic crust (see Dziewonski et al. 1975) and not for sedimentary layers at the surface. Moreover, as the faults are introduced vertically into the unlayered environment, the stresses do not vary much with depth, besides at the lower end of the fault.

Changes and improvements in PRENLAB-2

In the first phase of PRENLAB-2, the models developed in PRENLAB-1 were improved:

A The test-point density was increased from 56x44 (5 km distance) to 280x220 (1 km distance) to get more details of the stress field and to reduce interpolation errors.

B At the western end of the SISZ, segments with aseismic oblique slip (mainly normal faulting with a smaller component of left-lateral strike-slip) were introduced, to better fit the Reykjanes ridge (RR) between the southwest tip of the Reykjanes peninsula to Hengill-Ölfus triple junction (Figure 51).

C At the eastern and western tips of the SISZ two areas with steady stress release were introduced (see dashed lines in Figure 52). This could occur by a high rate of small events and maybe by creep. This is likely as these areas did never show strong events (M$\geq$6) in the seismic history of Iceland, with the exception of possible events in 1546 and 1632 at the western end (about 21.3°W) and one event in 1311 (about 18.9°W) (cf. Halldórsson 1991). See under Task 2 for a discussion of this.

D To investigate the model resolution a set of different models is produced: Besides the main model, several extreme cases are assumed and the variation of the main results under these assumptions is observed.

E The stress field was extrapolated to 1999 and was now updated to the situation after the June 21, 2000, event.

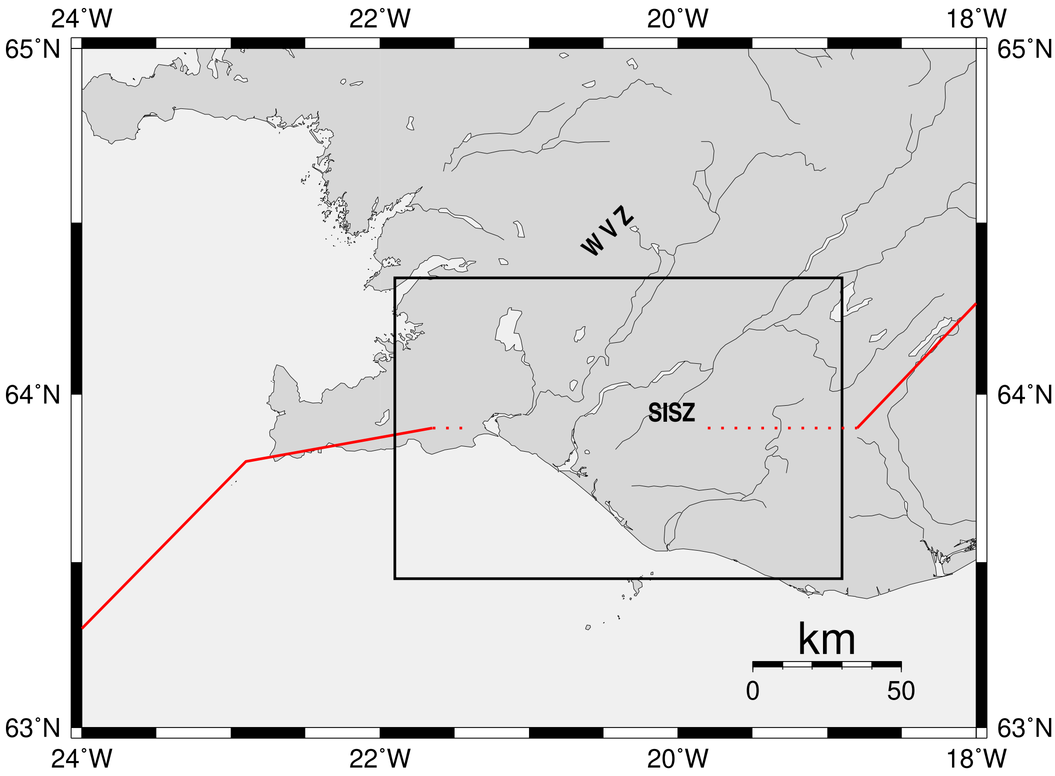


Figure 52. *The box gives the area in SW-Iceland used in the modelling as indicated with the small box in Figure 51.*

The earthquake data

All events with M $\geq$ 6 since 1706 were used (listed in Table 8., following Halldórsson et al. 1984; after Hackman et al. 1990; Stefánsson and Halldórsson 1988; and Stefánsson et al. 1993). The catalogue is supposed to be complete from 1706 for these earthquakes.

All ruptures were set to be oriented N-S, according to the isolines of damage intensity and surface ruptures shown in Figure 50. As only the events in 1912 and 2000 were instrumentally recorded, the source parameters are not very accurate – a problem to be discussed further below. The position of the epicenters and the rupture planes used in the models are given in Figure 53 and Figure 54, respectively.

Date[1]	Magnitude[1]	Epicenter[1]		South end of rupture[2]		Co-seismic slip[3]	Rupture length[4]
		Lat. °N	Long. °W	x [km]	y [km]	U_0 [m]	L [km]
1706	6.0	64.0	21.2	131	-5	0.30	10
1732	6.7	64.0	20.1	183	-11	0.77	22
1734	6.8	63.9	20.8	150	-23	0.96	25
14.08.1784	7.1	64.0	20.5	164	-18	1.9	35
16.08.1784	6.7	63.9	20.9	145	-22	0.77	22
26.08.1896	6.9	64.0	20.2	178	-14	1.2	28
27.08.1896	6.7	64.0	20.1	183	-11	0.77	22
05.09.1896	6.0	63.9	21.0	140	-16	0.30	10
05.09.1896	6.5	64.0	20.6	159	-9	0.48	18
06.09.1896	6.0	63.9	21.2	131	-16	0.30	10
06.05.1912	7.0	63.9	20.0	187	-27	1.5	32
17.06.2000	6.5	64.0	20.4	169	-12	0.9	16
21.06.2000	6.4	64.0	20.7	154	-13	1.1	18

Table 8. *Earthquakes M$\geq$6 since 1706 in the South Iceland seismic zone. 1) Data taken from Stefánsson et al. (1993); for the events in 1706, 1732, and 1734 no exact date is known. Data on the events of June 2000 are from Stefánsson, Guðmundsson and Halldórsson (pers. comm.), with magnitudes as moment magnitudes. 2) Position in the model coordinate system with origin at 64°N, 24°W. 3) Calculated via the magnitude moment relationship log M_0 [dyne cm] = $1.5M_S$+(11.8 - log(σ_a/μ)) with the apparent stress σ_a = 1.5 MPa and the shear modulus μ = 39 GPa (after Kanamori and Anderson 1975), followed by using the values of μ above, the rupture length as given in the table as well as a vertical fault width of 14 km east of 21° W and 7 km between 21° W and 21.2° W. Finally, the values for all events before those in 2000 were reduced by a factor of 2, following the discussion of Hackman et al. (1990). All this does not apply to the June 17 and 21, 2000, earthquakes, for which good instrumental data are available, and e.g. a maximum rupture depth of 9 and 7 km were given, respectively. 4) Besides for the June 2000 events, calculated using log L [km] = 0.5 M - 2 (after Qian 1986) which results in slightly lower values compared to e.g. Schick (1968).*

The results

The stress fields at 20 dates were calculated: the pre- and post-seismic situation for all 13 events. The time before 6 events was too short to accumulate appreciable plate tectonic stresses since the preceding event. In these cases, the post-seismic stress field of the preceding

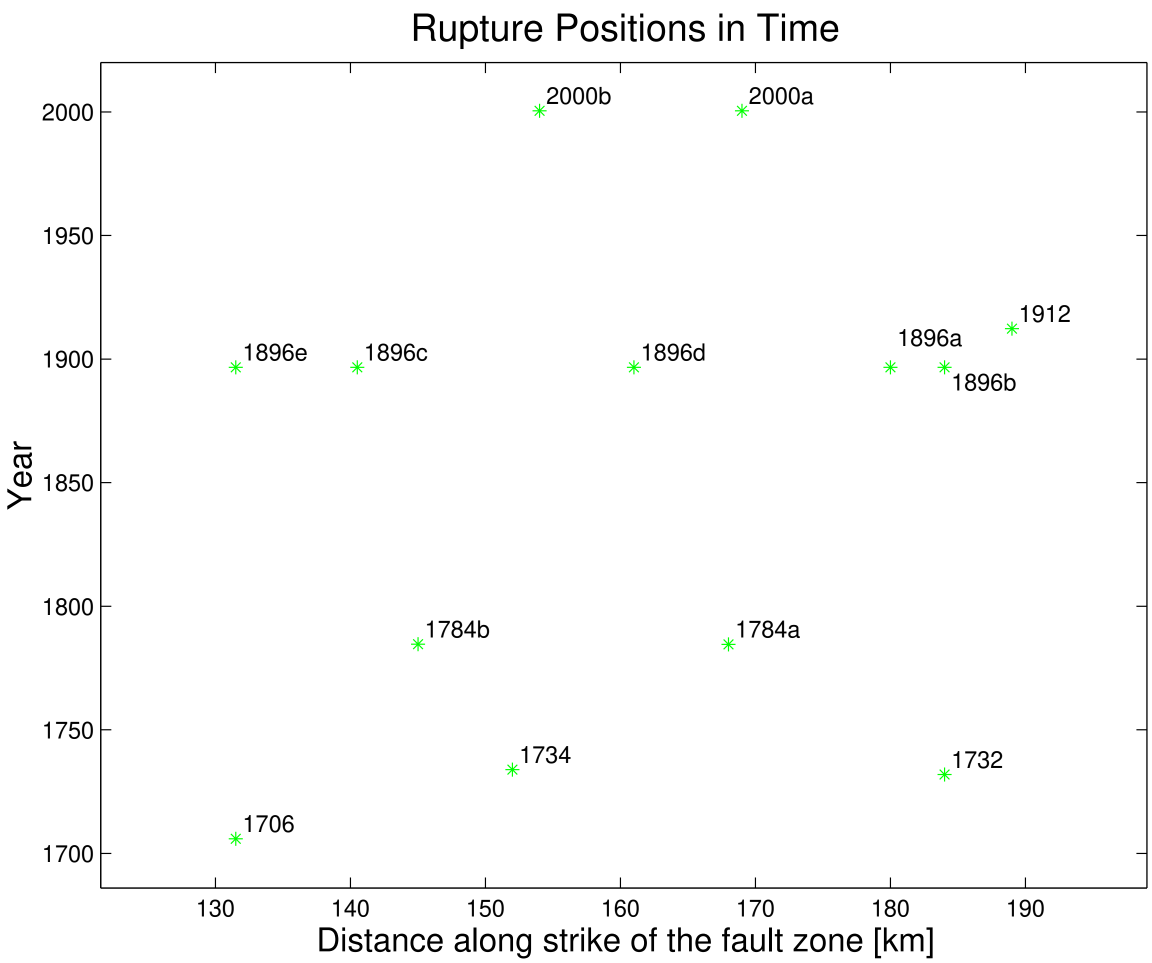


Figure 53. *Location of the earthquakes in space and time (cf. Table 8). As the events have been located on N-S trending faults in an E-W trending fault zone, their location is very accurately displayed in this graph.*

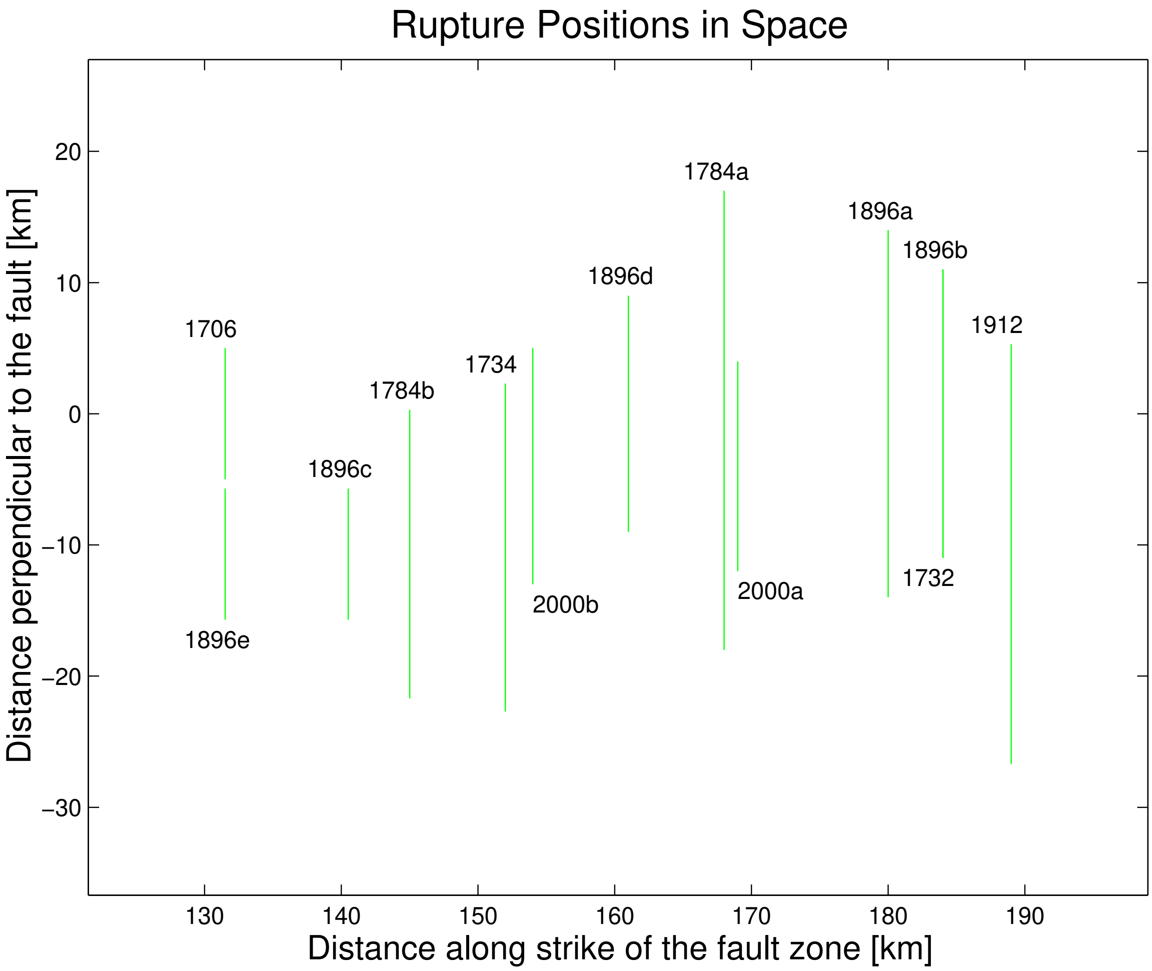


Figure 54. *The position of the rupture planes of the earthquake sequence as used in the models (cf. Table 8).*

event was set equal to the pre-seismic stress field of these earthquakes.

Originally, an extrapolation was done from the 1912 earthquake to spring 1999. After the two earthquakes happened this June, it was updated to June 17, 2000 (Figure 56) and the effect of both events were determined (see Figure 57 and Figure 58).

As a simple assumption, one might expect, that earthquakes in a certain fault zone usually occur at about the same critical shear stress level. We examine here if such an expectation matches the known facts about the earthquakes, given above, and the stress field in the SISZ from knowledge about plate motion and the modelling here. Figure 59 summarizes the mean shear stress level before each of the earthquakes at the area of the impending event. The stress level is near the average (1.8 MPa) or higher for most of the events. The highest value (for 1896e) is mainly influenced by the fact that the rupture area for the event in 1706 was located completely north of that in 1896 (cf. Figure 54). The second 1784 earthquake (two days after the first, 0.4 in magnitude smaller, 19 km away) might have been an aftershock and therefore situated in a lower stress area (1.6 MPa). The same might apply to the second of the 1896 events. The forth event in 1896 and the first shock in 2000 are both influenced by the largest event in the series, i.e. the first one in 1784. Thus, the accuracy of the source parameter of this 1784 earthquake strongly influences the whole model. Concerning the first event in June 2000, it has to be noted that it took place in a very inhomogeneous stress field, i.e. there are low stresses in the north of the rupture plane and high ones in the south (see Figure 56). Calculating an average pre-seismic stress level might be especially misleading for this event, when all the test-points around the rupture plane are considered equally.

Checking the performance of the model in a qualitative way, we examined if the earthquakes hit the high stress area and how large the high stress areas with no event were at the same time (the range in longitude with high stresses was summed when the N-S extension of the area was at least 5 km and the longitude range for the event that occurred was subtracted, usually 0.1 to 0.2 degrees in longitude). The results can be found in Table 9 and are quite satisfying with respect to the named question. Almost all events hit high stress areas and the size of high stress areas with no event was rather small after the earthquakes in the eighteenth century. Nevertheless, the problem remains, why some events did not occur earlier (at lower stress), just passing the "limit in pre-seismic stress" (i.e. here: the average pre-seismic stress).

As stated earlier, the plate velocity used in the model was set to 1 cm/year, only half of that what is measured. The reason can be seen in Figure 59 in comparison with Figure 60. The higher plate velocity leads to a steady and strong increase in the pre-seismic stress level, which is very unlikely, while the reduced rate entails an almost constant level. As the model cannot change the geodetic results, the reason for the discrepancy might be that only 50% of the stress build-up by plate tectonics is released seismically. For the rest, stable sliding or aseismic creep could be responsible.

Two problems were addressed next: (i) To see how sensitive the results depend on the model parameters, it was begun to check extreme cases and their outcome. (ii) The average stress level before some events is considerably lower than for others (cf. the discussion on Figure 59). This is, among other reasons, due to the fact that the rupture planes, used until now, extend rather far to the north and south of the SISZ. The damage areas from historical records are not gathered by scientists and are usually biased by uneven population density.

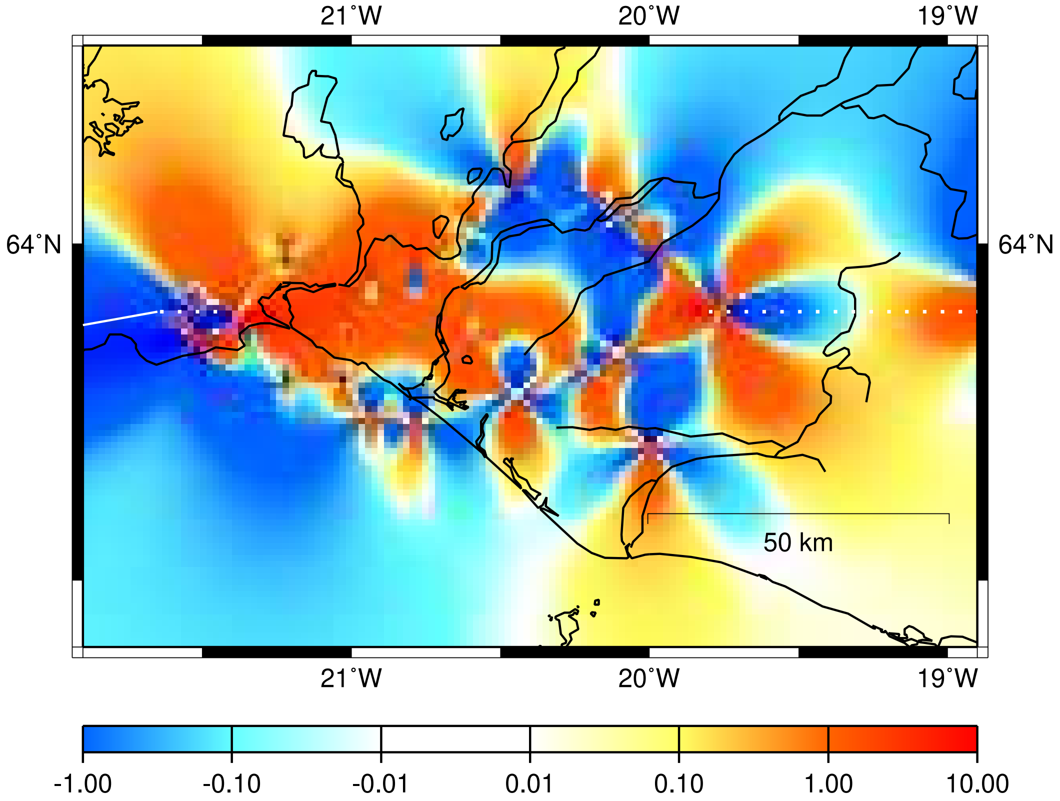


Figure 55. *The stress field after the May 6, 1912 earthquake.*

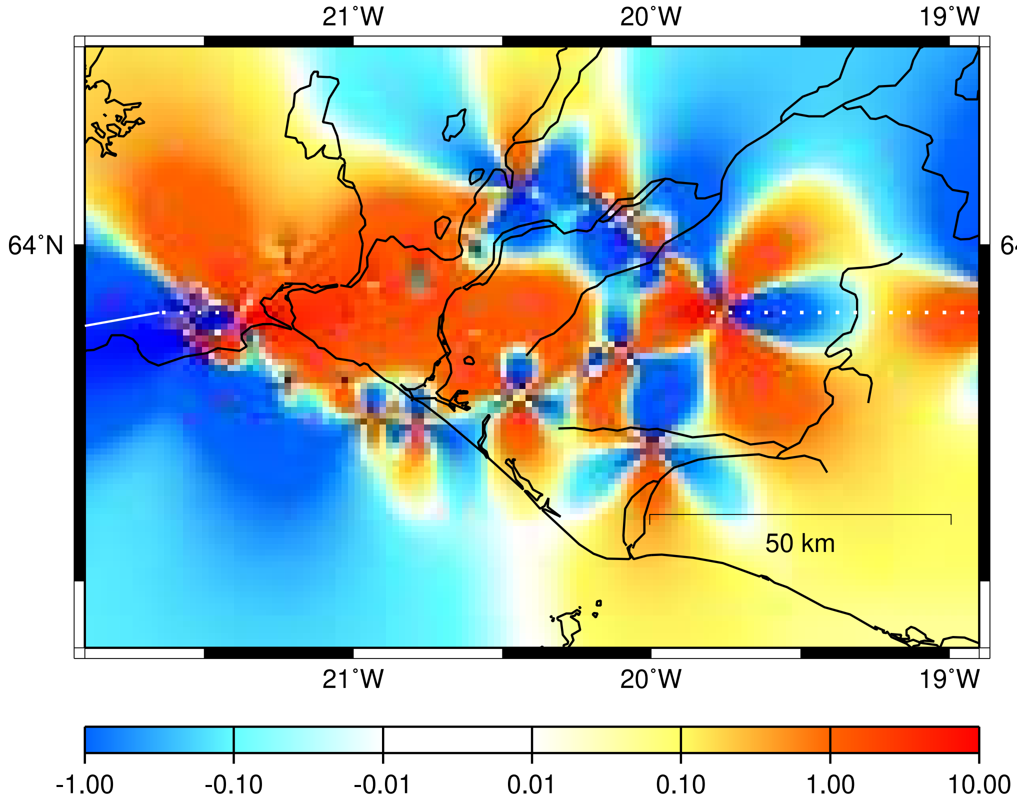


Figure 56. *The stress field before the June 17, 2000, M=6.5 earthquake occurring at 20.4° W, 64.0° N, i.e. (169, -12 — +4).*

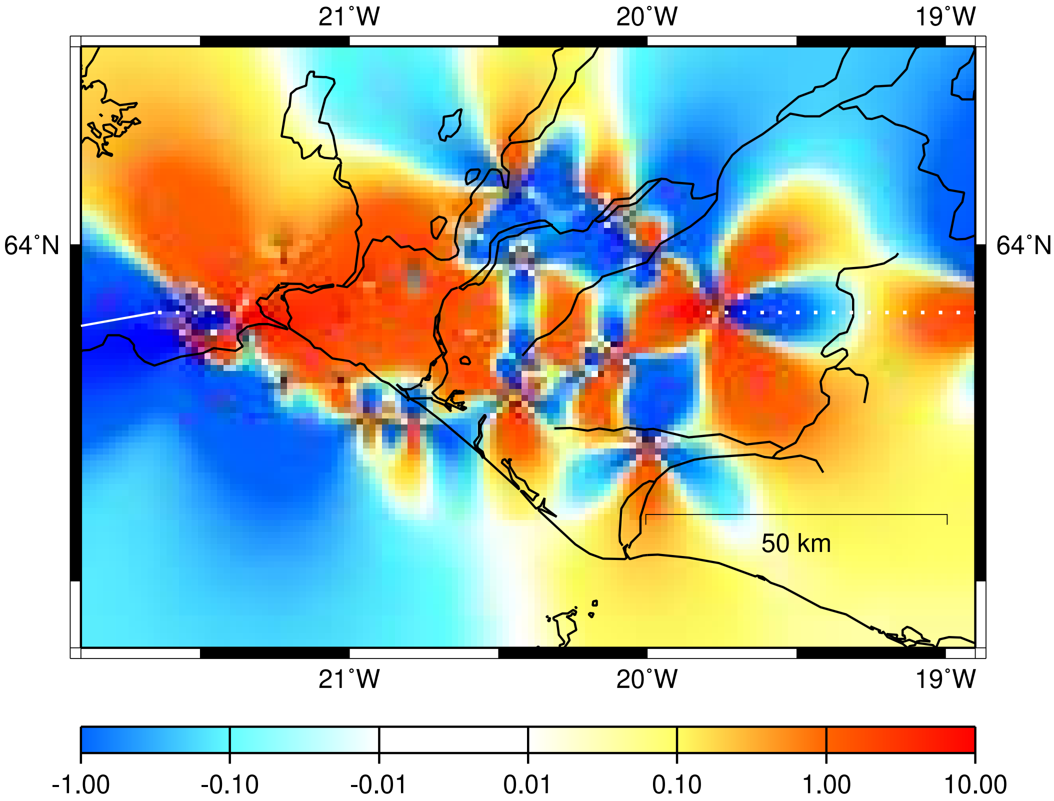


Figure 57. *The stress field after the June 17, 2000, M=6.5 earthquake occurring at 20.4° W, 64.0° N), i.e. (169, -12 — +4).*

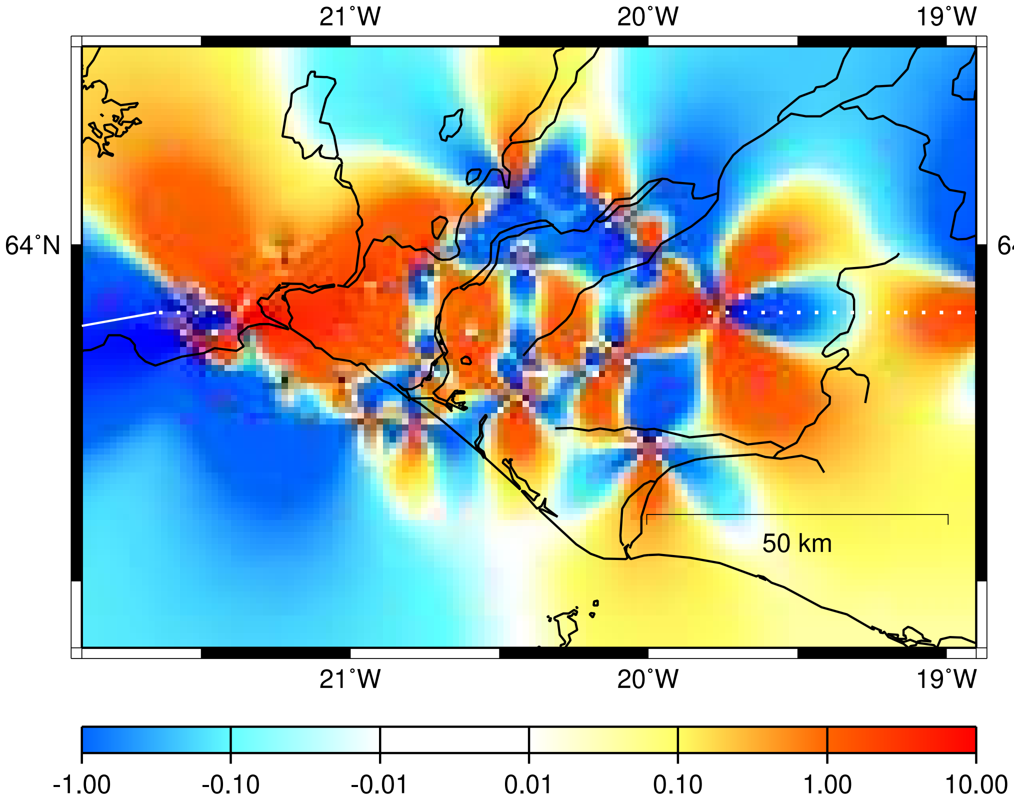


Figure 58. *The stress field after the June 21, 2000, M=6.4 earthquake occurring at 20.7° W, 64.0° N, i.e. (154, -13 — +5).*

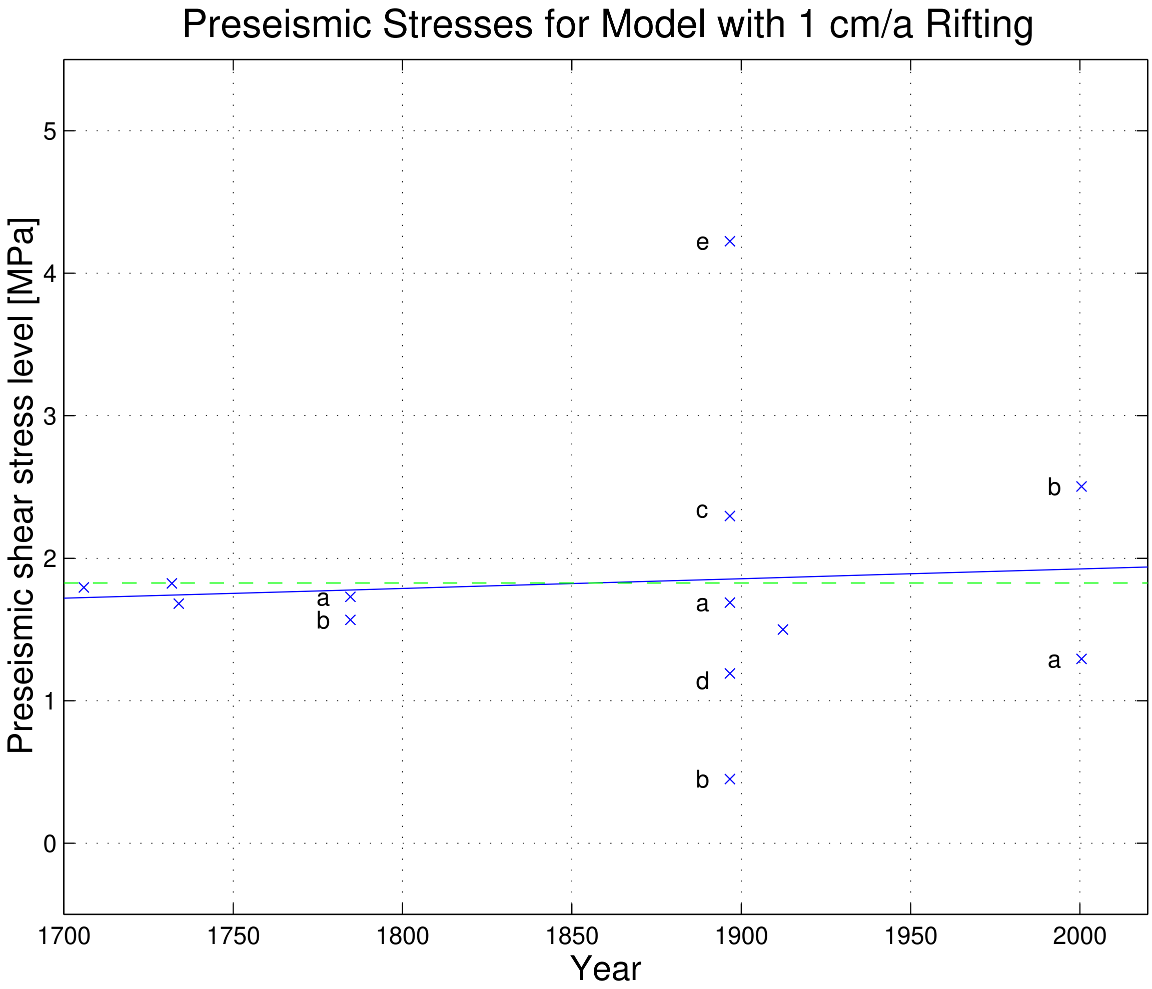


Figure 59. *Cross plot of the pre-seismic shear stress level at the site of the impending earthquakes vs. occurrence time. The stress values at 2 to 14 test-points at 0.5 km distance to the surface trace of the rupture plane were averaged. Instead of a plate velocity of 2 cm/year only 1 cm/year was used (see text). Letters "a" through "e" denote the events in one year in temporal sequence. The dashed line gives the average stress, the solid line represents the trend found by a linear least squares fit.*

Event time:	Pre-seismic stress high at °W:	Earth-quake at °W:	Hit high stress area?	Size of other areas with high stress:	Remarks:
1706	19.8 – 21.2	21.2	yes	large	tuning phase[1]
1732	19.8 – 21.1	20.1	yes	large	tuning phase
1734	19.8 – 20.0, 20.2 – 21.1	20.8	yes	large	tuning phase
1784a	19.8 – 20.0, 20.2 – 20.7, 20.8 – 21.1	20.5	yes	medium	tuning phase
1784b	19.8 – 20.0, 20.9 – 21.1	20.9	yes	small	tuning phase
1896a	19.8 – 20.3, 20.8 – 21.2	20.2	yes	medium	other high stress area hit 10 days later
1896b	19.8 – 20.0, ± 20.6, 20.8 – 21.2	20.1	no	medium	other high stress area hit 9 days later
1896c	19.8 – 20.0, ± 20.6, 20.8 – 21.2	21.0	yes	medium	other high stress area hit next day
1896d	19.8 – 20.0, ± 20.6, 20.8 – 21.2	20.6	yes	medium	other high stress area hit at the same day
1896e	19.8 – 20.0, 20.8 – 21.2	21.2	yes	small	
1912	19.8 – 20.0, 20.8 – 21.2	20.0	yes	small	
2000a	± 19.9, 20.5 – 21.2	20.4	no/ yes	small	hit short (N-S) high stress area[2]
2000b	± 19.9, 20.6 – 21.2	20.7	yes	small	

Table 9. *Qualitative evaluation of the model performance. 1) As the initial stress field is unknown, it was assumed to be homogeneous. So the first series of events that ruptures all across the fault zone is very strongly influenced by this assumption. 2) The stress field before the first earthquake in June 2000 is very inhomogeneous, i.e. there are low stresses in the north of the rupture plane and high ones in the south (see Figure 56). Looking at the development of the stress distribution, this situation is still influenced by the rupture position and magnitude of the first event in 1784 (cf. Figure 54). These historical data have some uncertainty (cf. Figure 50).*

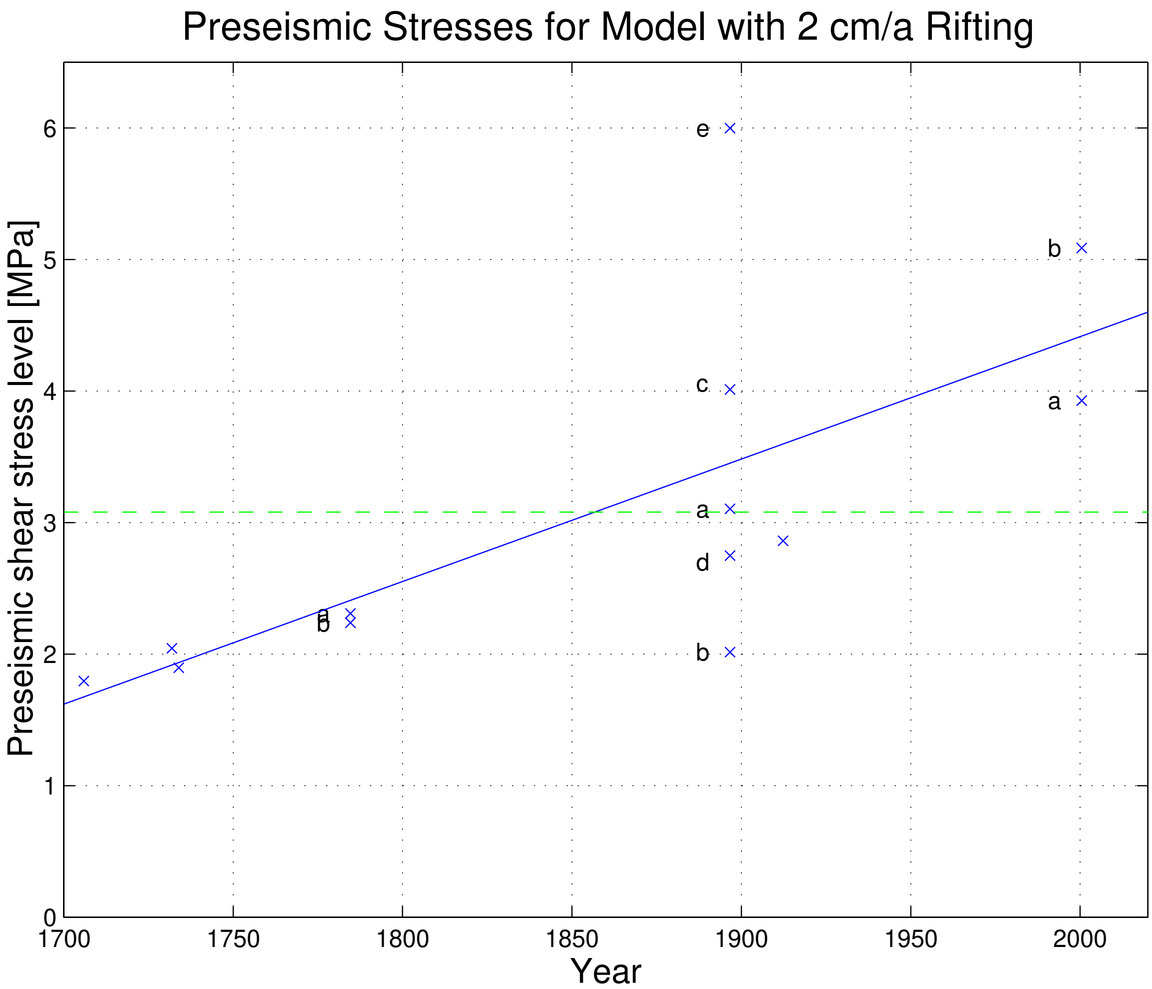


Figure 60. *Cross plot of the pre-seismic shear stress level at the site of the impending earthquakes vs. occurrence time. A plate velocity of 2 cm/year was used. For further explanations see Figure 59.*

So the magnitudes and locations are not very accurate, as stated earlier. And as mentioned in the footnotes of Table 9, there are doubts on the correct rupture size from global relations between magnitude and rupture length.

From both reasons given here, a model was calculated that uses the same seismic moment of the events, but cuts the fault length to 50% while doubling the co-seismic displacement. It will be termed "short rupture model". One side-effect of this change is an increase of the stress level, as the moment release is concentrated to a smaller area. The initial stress field amplitude was increased accordingly, because – as described above – this field is adjusted to the average stress change of the strongest event. It is important to note that the increase in stress level does not change the stress pattern of the initial stress field; as we are not looking for specific stress amplitudes but for stress concentrations, the change in level is not important. The resulting variation of the pre-seismic stress level is – as expected – smoother than before due to the concentration of stress release to high stress areas. However, this approach could only be used, once the historical events are re-evaluated with the result of shorter ruptures.

The processing of the recent earthquakes last June led to maximum rupture depth of less than ten kilometres. Even though there are smaller events located down to 13 km (cf. Stefánsson et al. 1993), the assumption that all ruptures extend to no more than a depth of 10 km seems to be reasonable. Such a model was calculated too, replacing the maximum depth of 14 km for most events (cf. Table 9) by 10 km. In this case too, the stress release by the events is higher, as it is concentrated to an area closer to the rupture plane. Moreover, the interaction of the events is lower due to this concentration in space. However, the stress level before the main events remains in a similar range as before (average now 2.1 MPa instead of 1.8 MPa), and the variation in the pre-seismic stress does not differ much (the standard deviation is 0.87 MPa instead of 0.88 MPa) from the model with deeper ruptures (cf. Figure 61).

In general, the variation of the model parameters shows that the models are stable, i.e. small changes in the parameters do not lead to totally different results. Therefore, they fulfill this important quality criterium.

3.7.2.2 Task 1: Extrapolation of the stress field for the next years

Originally, the stress field for the new models was extrapolated to spring 1999, with the additional stresses due to plate motion since 1912. This became obsolete when the earthquakes in June 2000 occurred. As presented above, the stress field before and after these events was determined. To examine if the stresses were high before the events was a good check for the model results. The present situation is essentially that after the second earthquake in June (see Figure 58). For more details see the discussion below.

3.7.2.3 Task 2: Pin-pointing of stress concentrations in space and time

At present, stresses are concentrated around 19.9°W and at 20.8–21.2°W. As there have not been earthquakes since 1912 and 1896, respectively, this is to be expected and was reproduced by the model.

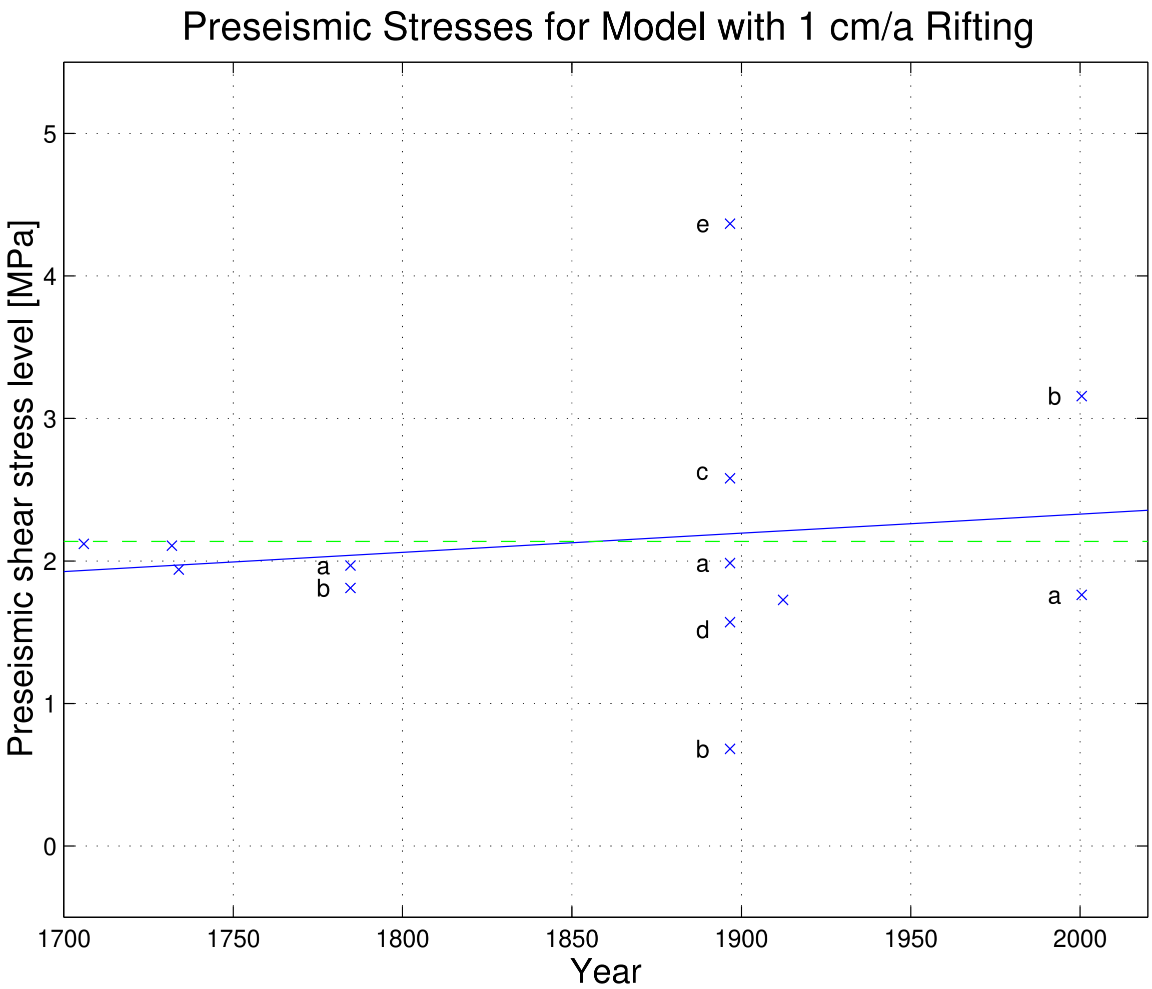


Figure 61. *Cross plot of the pre-seismic shear stress level at the site of the impending earthquakes vs. occurrence time. A maximum rupture depth of 10 km was used. For further explanations see Figure 59.*

Once more, we would like to point to the fact that, in the model, stresses are steadily released at those areas (marked with dashed lines) where the SISZ approaches the ridge segments, as noted under item "C" of the list of improvements above. It are sections of the SISZ between the high stress areas just mentioned and the rift tips.

Is this a weakness of the models or is it the real situation? Could these high stress areas be excluded from the model? The answer is essentially "no". The origin of the stress concentrations at the end of the SISZ, i.e. at the tips of the adjoining ridges, is the fact that the ridges do not extend to infinite depth, but are assumed to reach only 7 km depth and enter than an inelastic, hot region not capable of supporting stresses for time periods of years. Deeper penetration of the brittle layer there would homogenize the stress field between rift tips at some average value. Although this is not very likely, a similar effect would be produced by drag at the base of the adjacent plates. Nevertheless, such a redistribution of stress would not lead to low stresses at the ends of the SISZ.

Possibly, there might have been stronger events in 1546 and 1632 at the western end (about 21.3°W) and one event in 1311 (about 18.9°W) (cf. Halldórsson 1991). But all other evidence only shows small and medium sized events there. The main argument against high seismogenic stress release there is, that there was no large event ($M \geq 6$) since 1706. There are several indications that the stress release indeed mainly takes place in small and medium events at the ends of the SISZ. In 1987, there was a medium sized earthquake (M_S=5.8) at 63.91°N 19.78°W (198, -9) near Vatnafjöll (see Figure 50) at the east end of the SISZ. And in 1998, there were 2 similar events at the Hengill-Ölfus triple junction at the western end of the SISZ: June 4 (M=5.1) and November 13 (M=5) both accompanied by a lot of smaller events.

Therefore, we believe that the tips of the SISZ might only be the place where medium sized events take place and the stress concentrations given around 19.9°W and at 20.8–21.2°W represent those for future strong earthquakes.

As the stress build-up by plate motion is very low in these models (compare the stress field in 2000 in Figure 56 with that of 1912 in Figure 55), the uncertainty in time is very large in the occurrence time, see also the discussion below.

3.7.2.4 Task 3: Search for characteristic preseismic stress level

As a simple assumption, one might expect, that earthquakes in a certain fault zone usually occur at about the same critical shear stress level. Here, we tried to find out, if such an expectation matches the known facts about the earthquakes and the stress field in the SISZ.

In all models, the pre-seismic stress level for most main shocks is high and fairly stable. This is also true for the events at the end of the sequence, namely those of 1912, of June 20, 2000, and – with minor reservations – of June 17, 2000. It indicates that the rather simple model can already explain the main features of the behaviour of the SISZ. This is especially astonishing, when the fact is kept in mind, that most (all but one) events used are not instrumentally recorded. Before the June 2000 events the SISZ seems to have been prepared for rupturing at the specific locations. This fits very well to the findings from stress measurements in wells (see Subproject 4) that E-W left-lateral shear stress was acting on the fault zone. To improve the model further, the initial unknown stress field of 1706 could be reduced in the eastern part and the central part, where the first events did not occur before

1732 and 1734, respectively. Another improvement might be to include more basal drag as the source of plate tectonic stress increase as compared to ridge push.

Nevertheless, the problem remains, why some events, as those of September 6, 1896, or June 21, 2000, did not occur earlier (i.e. at lower stress), just passing the "threshold in pre-seismic stress" (here e.g. the average pre-seismic stress (cf. Figure 59). Only if this would be the case, a prediction of the occurrence time might be at reach.

Even though the earthquake rupture planes strike N-S, the stress changes calculated here affect the whole area of the SISZ.

A tendency with time towards higher values of pre-seismic stress was found. It is an indication that the stress increase due to rifting might have been assumed too high, i.e. not all of the stress increase due to the spreading rate of 2 cm/year (assumed to be constant between 1706 to 2000) was released by earthquakes. The assumption that only half of the accumulated stress is seismically released led to a rather constant pre-seismic stress level.

The variation in model parameters does not lead to totally different results, i.e. the model is rather stable.

In general, the models go beyond the standard earthquake moment release and hazard analysis as they include the spatial location and extension of the events, quantify the amplitudes of stress release and that of plate motion on the faults, as well as providing an extrapolation to the present stress situation. This report will be submitted for publication in an appropriate form.

3.7.2.5 Meetings and conferences

A meeting between Maurizio Bonafede and coworkers with Frank Roth took place in March 1999 in Strasbourg, France, parallel to the tenth biennal EUG meeting. Another meeting with Bonafede and Ragnar Stefánsson took place at the GeoForschungsZentrum Potsdam in June 1999.

3.7.2.6 Acknowledgements

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