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Report
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Long-term and short-term earthquake warnings based on seismic information in the SISZ

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The objective of this paper is to describe some experiences and procedures that can be useful to provide earthquake warnings on a long-term and short-term time scale, based on multiparameter seismic observations and evaluations.

1 Introduction

Our approach to earthquake warnings or predictions is to gradually assess the state of probability that a dangerous earthquake will occur, and to advise on precautions and to concentrate on scientific action to mitigate risks that the earthquake may cause.

In the PREPARED proposal we have defined warnings as following scenarios or stages (Stefánsson et al. 2002):

Years/month in advance. Concentration of various risk mitigating efforts; compiling background data, finding baseline, increasing research, increased monitoring and data analysis, strengthening of infrastructure.

Weeks/days in advance. Activation of civil protection and rescue groups; increased sensor based observations, raising in general the preparedness of people.

Hours/minutes in advance. Final civil preparations for an hazardous event that could occur. Such an alarm must have a specified time limit, i.e. if no event occurs within a specified period then this scenario ends.

The earthquake occurs. Actions aiming to mitigate impact on people and society; early information and warnings, now-casting, real-time damage assessment.

Post-quake information. Explaining the hazardous event; assessing and warning for further coupled hazards.

In Iceland we have to some extent tackled all these steps. We have had some good results. How have we done this and how can we get still further in providing better basis for warnings?

- By continuous relevant instrumental monitoring we approach such warnings through intensive watching.
- By automatic alert procedures we try to detect changes and patterns that may point to the place of an impending large earthquake on a long or short time scale.
- By concentrating our watching and research at places where we expect the next large earthquake to occur.
- By international multidisciplinary earthquake prediction research to understand the physical processes leading to large earthquakes.
- By developing monitoring and evaluation systems, based on these results, systems which are applied to make warnings and create data for further research and further enhancing our understanding.

- By developing information and warning systems which contain all the acquired knowledge and understanding, on-line observations and tools for fast evaluation and application of all the information we have. On the internet we set up a common table for scientists to cooperate and apply these tools for risk mitigation. This information and early warning system is the EWIS-system in build-up and operation at the Icelandic Meteorological Office.

The most significant basis for all this work are real-time observations in the earthquake areas. The basic system we apply for this in Iceland is the SIL-system for real-time acquisition and evaluation of information continuously carried by small earthquakes from deep down in the crust. But we also base our understanding to interpret these observations of older data and on modelling.

The aim of this report is to point out significant observations and results of research work, which is relevant for mitigating risks and for further studies. We will mainly concentrate on information and studies based on seismic research results and observations and on emerging models that can explain many features of the observations. We will point to some tools for immediate use in risk mitigation without claiming that these are the only tools. Many others are working in the same direction, even with the seismic data.

2 Earthquake release in large earthquakes in the SISZ since 1700

The South Iceland Seismic Zone (SISZ) is usually considered as a 70 km long and 10-20 km wide EW zone of left-lateral transverse plate motion. To the west it borders to the Hengill triple junction and to the east it borders to Hekla Volcano approaching the Eastern Volcanic Zone (Figure 1; Stefánsson and Halldórsson 1988; Einarsson 1991; Stefánsson et al. 1993). The SISZ is sometimes defined as a 10 km EW strip (based on microseismicity and surface earthquake faults), this is the SISZ in the narrow sense. Sometimes it is rather defined 20 km wide or even wider. The fault planes of the historical earthquakes as they have been interpreted here stretch outside the 10 km SISZ as seen in Figure 1.

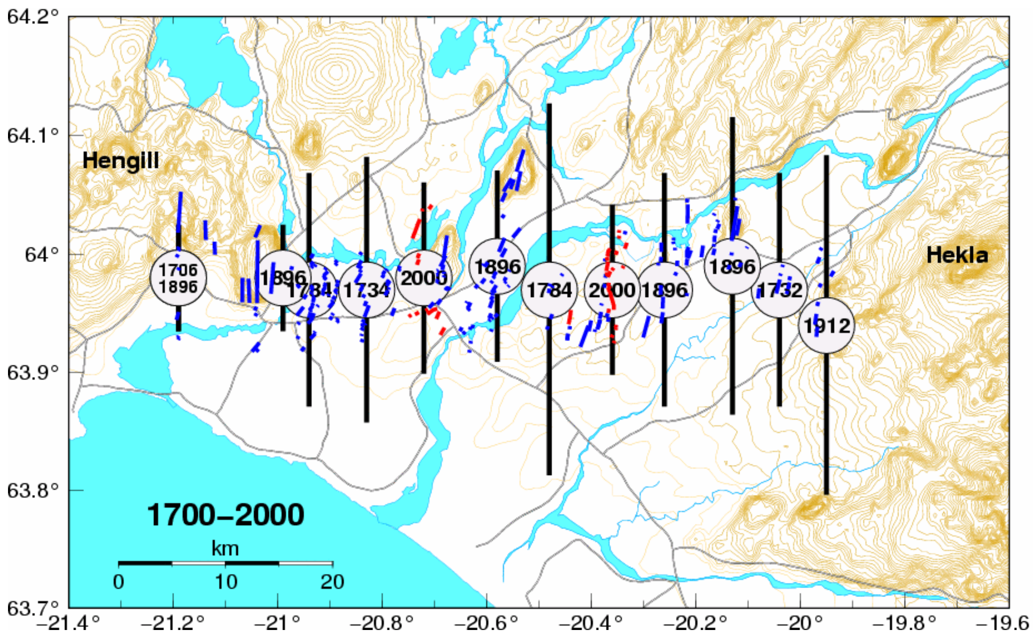

Figure 1. *Earthquakes in the SISZ since 1700. The earthquakes arrange side by side, each having right-lateral slip on a NS fault. The fault epicenters and magnitudes (Stefánsson et al. 1993; Einarsson et al. 2005) are estimated from historical data and the fault lengths are from Roth (2004).*

In Figure 2 we see a summing up of strain energy released in historical earthquakes along the SISZ (Stefánsson and Halldórsson 1988; Halldórsson 1987) On basis of such observations following conclusions were drawn from comparing the strain release in the zone due to plate motion and moment release in earthquakes in the area:

- More release of earthquake moment in the eastern part of the zone than in the western part. This was explained by thicker elastic/brittle crust in the eastern part (Stefánsson and Halldórsson 1988; Stefánsson et al. 1993).

- For explaining an apparent excess of strain energy in the zone as released in earthquakes, compared to that expected from plate motion only, it was suggested that energy from mantle fluids migrating upwards would probably contribute to the strain release energy in earthquakes in addition to the plate motion energy (Stefánsson and Halldórsson 1988).

- Apparent seismic gaps coinciding with high microseismicity was used 20 years ago to conclude that the most likely sites for the next earthquakes in SISZ were "at 20.3º-20.4°W and at 20.71°W", i.e. within a few kilometers from the sites of the two year 2000 earthquakes (Stefánsson et al. 1993; Stefánsson and Halldórsson 1988; Stefánsson et al. 2003). NS faults were assumed.

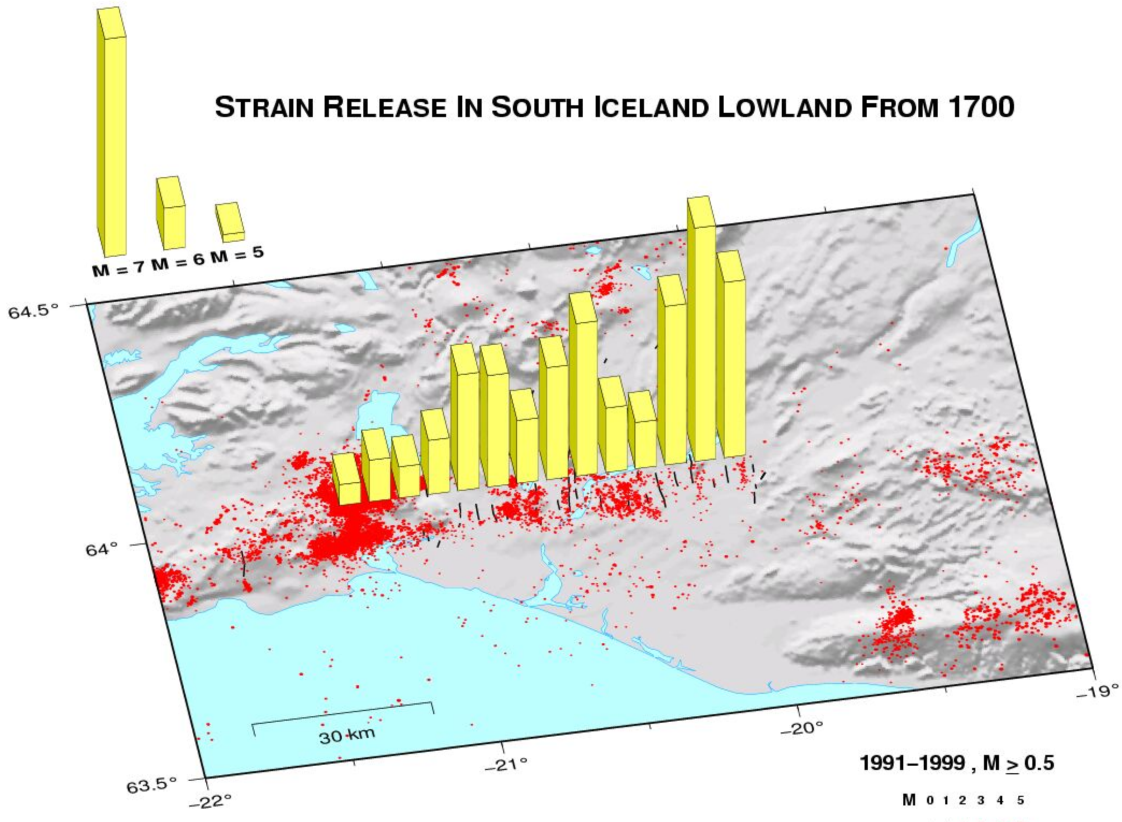

Figure 2. *The yellow pillars arranged along the SISZ in Southwest Iceland indicate release of strain energy in historical earthquakes from 1700 to 1999. Here Benioff strain (Benioff 1951) is calculated from the magnitudes of each of the historical earthquakes. The half of the strain of each earthquake is put at the most probable location, while ¼ is put at each side to try to allow for probable location errors. Red dots show microearthquakes. Black lines indicate the position of known earthquake faults. Earthquakes tend to be larger in the eastern part than in the western part, which has been explained by a thicker and stronger brittle/elastic crust there.*

Eruptions in the volcano Hekla, at the eastern end of the SISZ, probably play a significant role in the release of strain energy in it. Table 1 summarizes eruptions in Hekla since 1206 for comparison with earthquake activity.

Table 1. *Eruptions in and near the volcano Hekla. No comments means that the eruption was in Hekla. Most of the information come from Þórarinsson (1968). Other information come from Guðmundsson (2001).*

The eruption of the volcano Hekla since year 1200					
Year	Time	Lava km^2	Lava km^3	Tephra km^3	Comments
1206	December 4	>0.15		0.03(?)	
1222	(?)	-		0.01	
1300	July 11	>0.5		0.5	
1341	May 19	-		0.08(?)	
1389	Autumn	>0.2		0.08	Rauðöldur (7.5 km WSW of Hekla)
1440			?	?	N and S of Hekla
1510	July 25	>0.75		0.32	
1554	Spring		<0.1*	?	Vondubjallar (10 km SSW of Hekla)
1597	January 3	-		0.24(?)	
1636	May 8	-		0.08(?)	
1693	February 13	-		0.3	
1725	April 2		<0.1-0.2*	?	N and S of Hekla
1754					W of Hekla (Ferðabók Eggerts Ólafssonar og Bjarna Pálssonar and no other sources)
1766	April 5	1.3		0.4	
1845	September 2	0.63		0.28	
1878	February 28		0.1-0.2*	?	At Krakatindur (10 km ENE of Hekla)
1913	April 25		0.1?*	?	At Mundafell (6 km ESE of Hekla) and Lambafit (14 km NE of Hekla)
1947	March 29	40	0.8	0.21	
1970	May 5	18.5*	0.2	0.07	Both in Hekla and Skjólkvíar (5.5 km NNE of Hekla)
1980	August 17	22.5	0.12	0.06	
1981	April 9	6	0.03	-	

1991	January 17	23	0.15	0.02	
2000	February 26	18	0.11	~0.02	
* Estimated by Guðmundsson (2001).					

3 Earthquakes released in the SISZ 1896-1912

In the following we discuss the strain release in the zone since after the large 1896 earthquake sequence. After the breakthrough of the SISZ through a series of earth-quakes, 1732-1784, strain was rebuilt to be partly released in the five magnitude 6-6.9 earthquakes that swept over the central and western part of the zone in 1896.

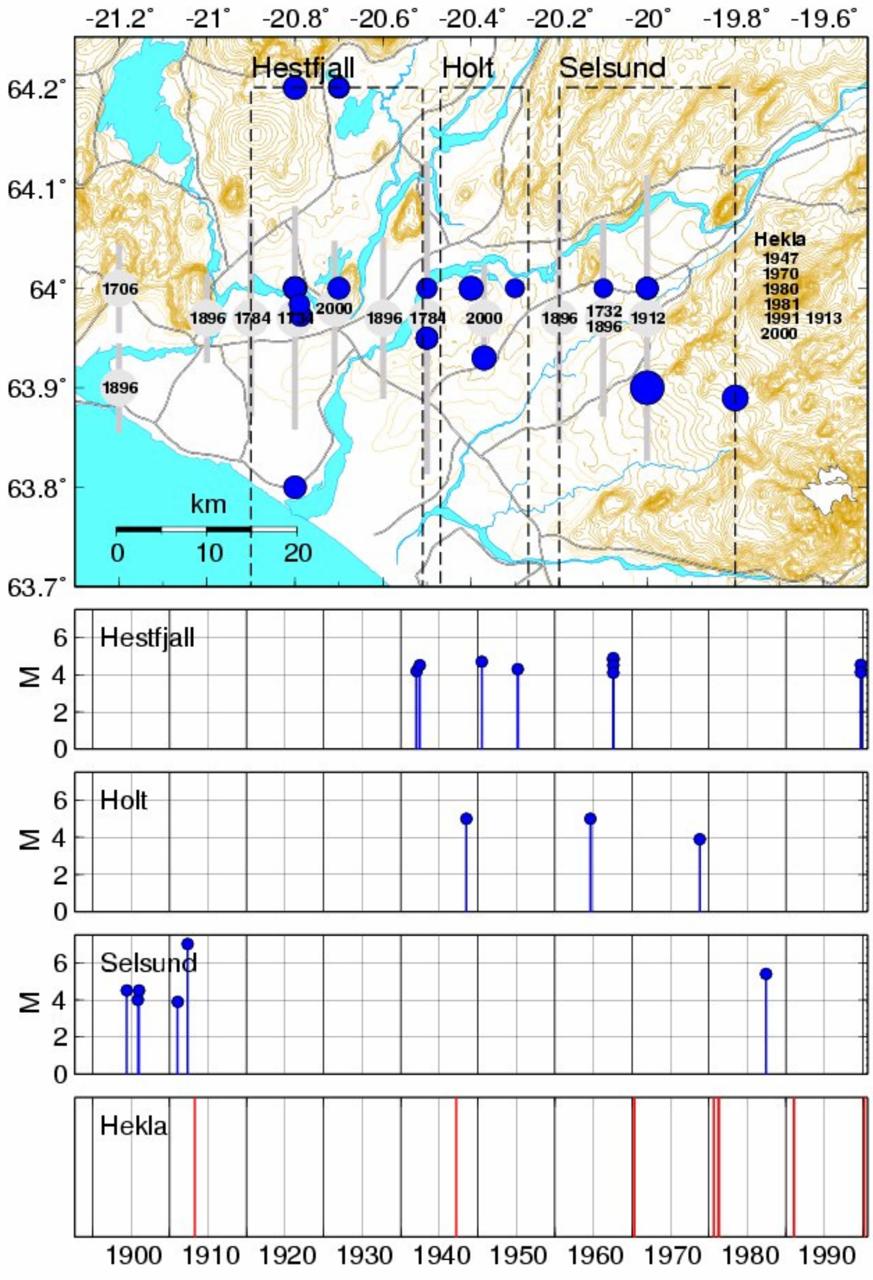

Figure 3. *Earthquakes estimated magnitude 4 or larger in the SISZ since after the sequence of 1896 and until the two large SISZ earthquakes in year 2000. The volcano Hekla with the years of its recent eruptions marked is at the eastern end of the zone. The Hengill volcanic complex is at and just off the left edge of this figure.*

This earthquake sequence, that started on August 26 east of the central part of the zone, ended September 6 in the western part (Figure 1). Smaller aftershocks were felt until end of May 1897. The strain in the easternmost 15 km of the zone, i.e. between longitudes 19.8º and 20.1ºW, was not released at that time. It has been pointed out that the earthquakes in the eastern part of the zone are potentially stronger than in the western part, explained by thicker elastic/brittle crust there than farther west (Stefánsson et al. 1993). This can also explain why the easternmost part did not break following the large earthquake sequence in the central and western part in 1896.

3.1 Silence and premonitory activity before the 1912 earthquake

SISZ was seismically relatively silent for 7-8 years after the 1896 sequence, i.e. until 1904. In 1904 to 1911 five earthquakes of magnitude 4-4.5 occurred with origin in the easternmost part, i.e. around 63.9°-64.0ºN and 20.0ºW. These earthquakes preceded the May 6, 1912, a magnitude 7 (Ms) earthquake which occurred in this easternmost part of the zone, i.e. the part of the zone which was not broken through in year 1896. The magnitude of the 1912 earthquake was instrumentally determined (Kárník 1969).

After 1912, i.e. after the complete breakthrough of the 1896 to 1912 series, there followed 30 almost silent years in the SISZ, i.e. until January 20, 1942, in the central western part of the area, the Hestfjall area (which was the location of the June 21 earthquake of the year 2000), and 36 silent years or until July 3, 1948, in the Holt area, the region of the first earthquake of year 2000 (Figure 3).

4 Earthquake release in the SISZ after the 1912 earthquake

The sensitivity or the limit of completeness of earthquake observations was of course variable during the 20^{th} century. Some work is left to unify magnitudes too. However, some patterns can be securely identified (Figures 3 and 4) on base of the data as they are now.

After 1912 the SISZ (as it is usually defined) was relatively silent until the magnitude 6.6 (Ms) earthquakes of year 2000. At both ends of the SISZ, i.e. in Hengill seismovolcanic area in the west and in Hekla volcanic area in the east (Table 1) there was high seismic activity at several occasions. Both these areas of the plate boundary show-up mixture of seismic and a volcanic character.

1926-1940: Earthquakes farthest east in the zone (Figure 4), i.e. in Hekla in 1932-1938, can be considered premonitory of the large Hekla eruption in 1947. A cluster in Hekla and Vatnafjöll area from 1940 to 1950 are forerunners to or related to the eruption.

1940-1950: See Figure 4. As seen also in Figure 3 relatively large earthquakes started to occur in the SISZ in this period, in the Holt area in 1948 and the Hestvatn area in 1942, i.e. where the two large 2000 earthquakes had their epicenters. The earthquake cluster farthest east in Figure 4 are in the Hekla Volcano, related to the 1947 eruption.

1950-1960: Increased activity is observed in the western part of the SISZ, but no activity in the eastern part. During this period, in 1952 to 1955, there was a large earthquake swarm period in the Hengill area at the western end of the SISZ, with magnitudes reaching 5.5.

1960-1970: Clusters of earthquakes occur in the Holt area 1964 and the Hestvatn area 1967 which are to a large extent aftershocks of magnitude 5 earthquakes at these locations. The epicenters of the 2000 earthquakes were at these locations (Figure 3). A third cluster is seen in the Vatnafjöll area south of Hekla, mostly in 1964 and 1967 at the location of the magnitude 5.8 earthquake of 1987.

1970-1980: On May 5, 1970 an eruption started in Hekla, and the earthquakes following this eruption make most of the cluster near the east end of the zone.

1980-1989: High microseismic activity especially in the central and western part of the zone. There was an eruption in Hekla in 1980. Very little seismic activity accompanied this eruption. In 1987 the largest earthquake since 1912 occurred in the SISZ, the Vatnafjöll earthquake of magnitude 5.8. The location of this earthquake is just SSW of Hekla and is probably tectonically closely related to the volcano.

1990-1991: The period is strongly characterized by the flurry of earthquakes that seem to have been triggered by and followed the Hekla eruption starting on January 17, 1991 (Figure 5).

July 1, 1991 to June 17, 2000: The characteristics during this period of detailed seismic observations indicate a general pattern as the older observations, however, much more detailed. Figure 5 only shows earthquakes larger than 1.5.

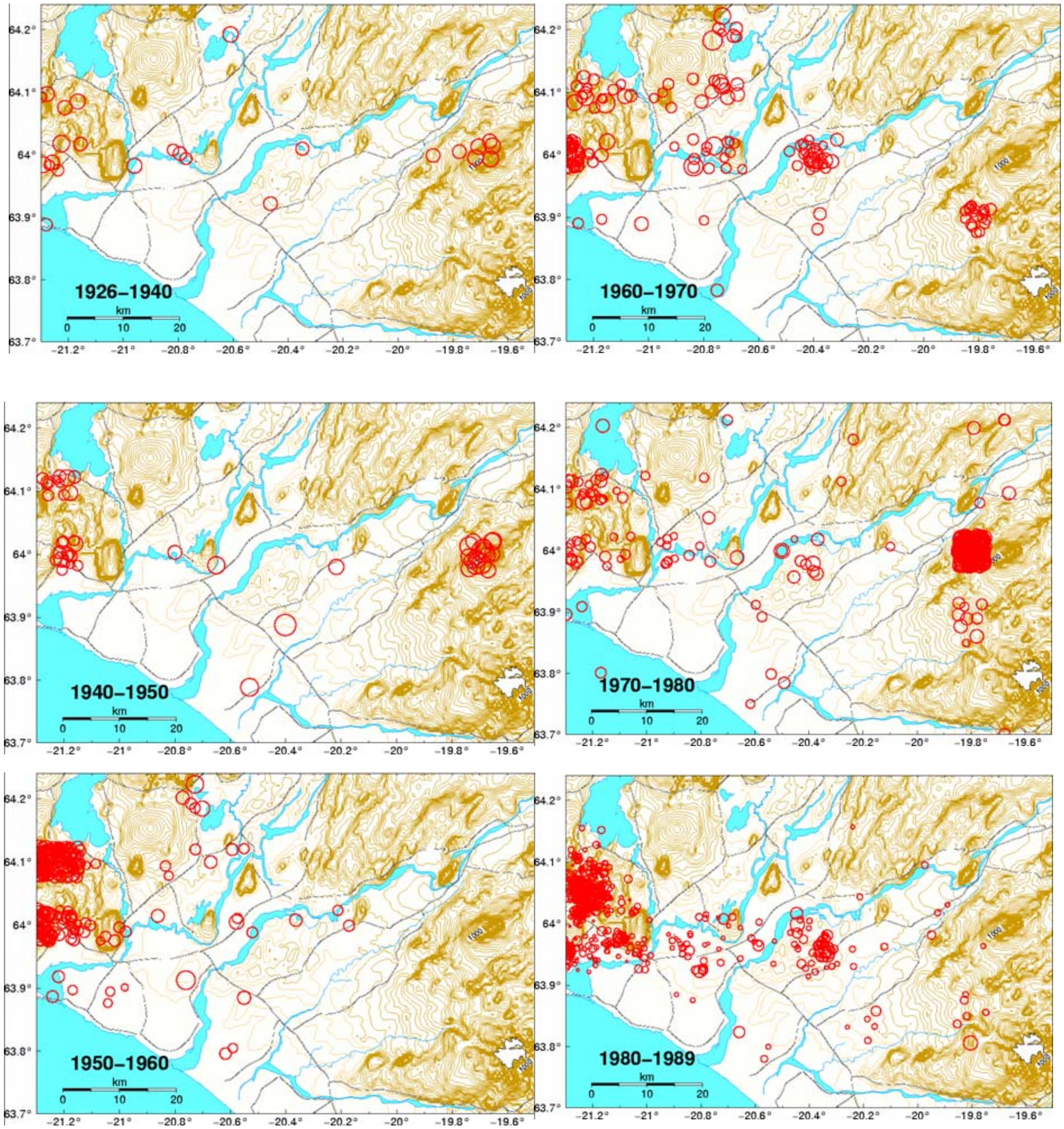

Figure 4. *Circles show epicenters of earthquakes registered by the Icelandic seismological stations from 1926 until 1989. The magnitudes (local scale, comparable to mb for the small earthquakes) are indicated by the size of the circles. In plotting in some cases we have applied random errors for earthquake swarm locations, which otherwise would have only been plotted in one point.*

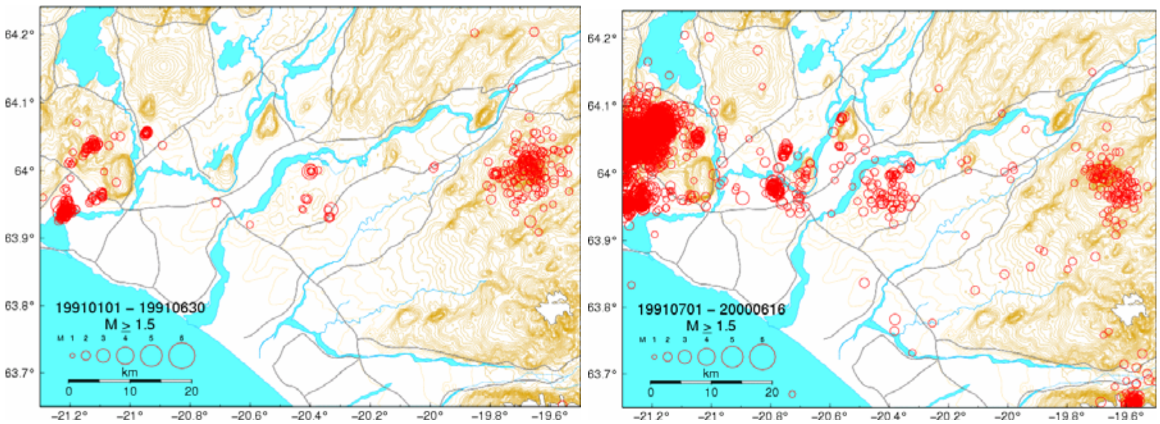

Figure 5. *Circles show epicenters of earthquakes registered by the Icelandic seismological stations during the first half of 1991 and for the period from July 1991 to June 17, 2000. The magnitudes larger than 1.5 (local scale, comparable to mb for the small earthquakes) are indicated by the sizes of the circles.*

4.1 Strain build-up and release in the SISZ, in relation to the east and west boundaries

The seismicity since the 1896-1912 earthquake sequence contains much information that can be used to understand and model the SISZ, strain build-up and coupling between events, or strain waves transmitted through the region. Much has been done as in the hazard assessment of the SISZ. However, the observations contain much information that have not so far been, with modelling, put into the general picture and the SISZ dynamics. Figures 4 and 5 demonstrate strong causative links between volcanotectonic events in the whole zone. It should be studied better with modelling how the frequent eruptions in Hekla Volcano since 1947 (Table 1) play a role for the strain build-up and strain release in the SISZ. The same is relevant for the volcanotectonic Hengill area at the western end. Strain build-up and strain release in the SISZ is strongly conditioned by its eastern and western ends, the Hekla Volcano and the Hengill area.

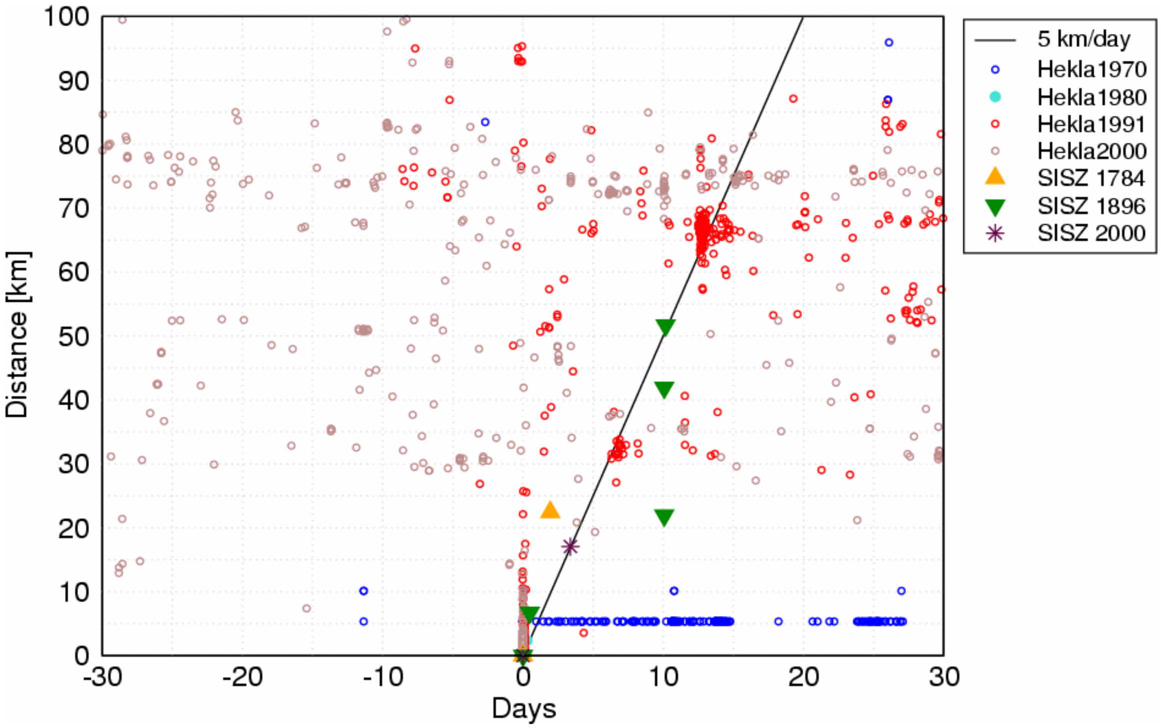

Figure 6. *Small earthquakes that followed four Hekla eruptions as described in the figure, distance from the volcano and time from the start of the eruptions. Also are shown times and distances of large earthquakes following a primary earthquake in SISZ set at point zero.*

The eruption of Hekla Volcano which started on January 17, 1991 triggered earthquake activity all along the SISZ. The time lag is being studied. It goes from direct dynamic triggering to several kilometres/day time lags (Figure 6). Such considerations were the basis for what among the seismologists at the Icelandic Meteorological Office was called a rule of thumb, i.e. that large earthquakes were expected to migrate along the SISZ with a velocity of 5 km/day. This was applied to make the time warning for the second of the two June 2000 earthquakes in the zone.

It is interesting for modelling of the dynamics of the SISZ and possibly for prediction that the 1991 Hekla eruption caused a flurry of earthquakes migrating westwards. This did not occur after the 1980 eruption. However, 7 years later we had the Vatnafjöll earthquake, SW of Hekla. And neither did it occur after the February 26, 2000 eruption, and 3-4 months later we had the June 2000 earthquakes in the SISZ. A question is if in both these cases an asperity hampered a strain wave to go westwards as it did in 1991.

The western end of the SISZ, i.e. the Hengill seismovolcanic boundary had a large episode of earthquakes and observed land deformation in 1994-1998. Earlier similar episodes have been observed in this area, for example in 1952 to 1955. The largest known such an episode occurred in 1789 and took to a large part of the Western Volcanic Zone, to NE and SW of the Hengill area. Such interactions are not discussed here, but are certainly of relevance for understanding the seismic activity in the SISZ.

5 Earthquake patterns in the SISZ from 1991 to 2000, the period from the start of the full operation of the SIL-system

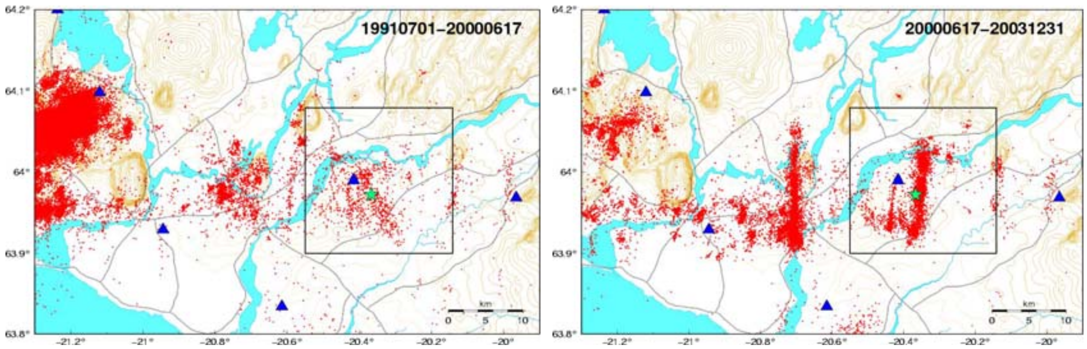

Figure 7. *Seismicity in the SISZ before (left) and after (right) the year 2000 earthquakes, shown by green stars. The area of dense seismicity at the western end of the SISZ is at the junction between the SISZ and the Western Volcanic Zone, displaying high seismic activity there during a volcanotectonic episode 1994-2000. At the eastern end of the zone we are close to the junction of SISZ with the Eastern Volcanic Zone.*

Figure 7 demonstrates the microseismicity (down to magnitude 0) during 10 years prior to the 2000 earthquakes (Stefánsson and Guðmundsson 2005). Two distinct clusters are seen in the epicentral area of the earthquakes. The main general features seen for the period 1926-1989 are there, however, since 1991 the SIL-system has been operated in the SISZ being almost complete in detecting earthquakes down to magnitude zero. Comparing the figure to the left, i.e. before the 2000 earthquakes, with the figure to the right we observe that the clusters before the 2000 earthquakes have a different configuration from the earthquakes after the large earthquakes. This has been discussed by Stefánsson and Guðmundsson (2005). The swarm volume that characterized the area prior to the 2000 earthquakes was called a dilavolume, i.e. the rock mass of the crust where lithostatic pore fluid pressures are in general elevated to shallow depths in the crust (mostly, however, below 3 km). High pore fluid pressures of this dilavolume which have volume distribution before the 2000 earthquakes were released in them and have changed to linear distribution on the earthquake faults after the earthquakes (Figure 7).

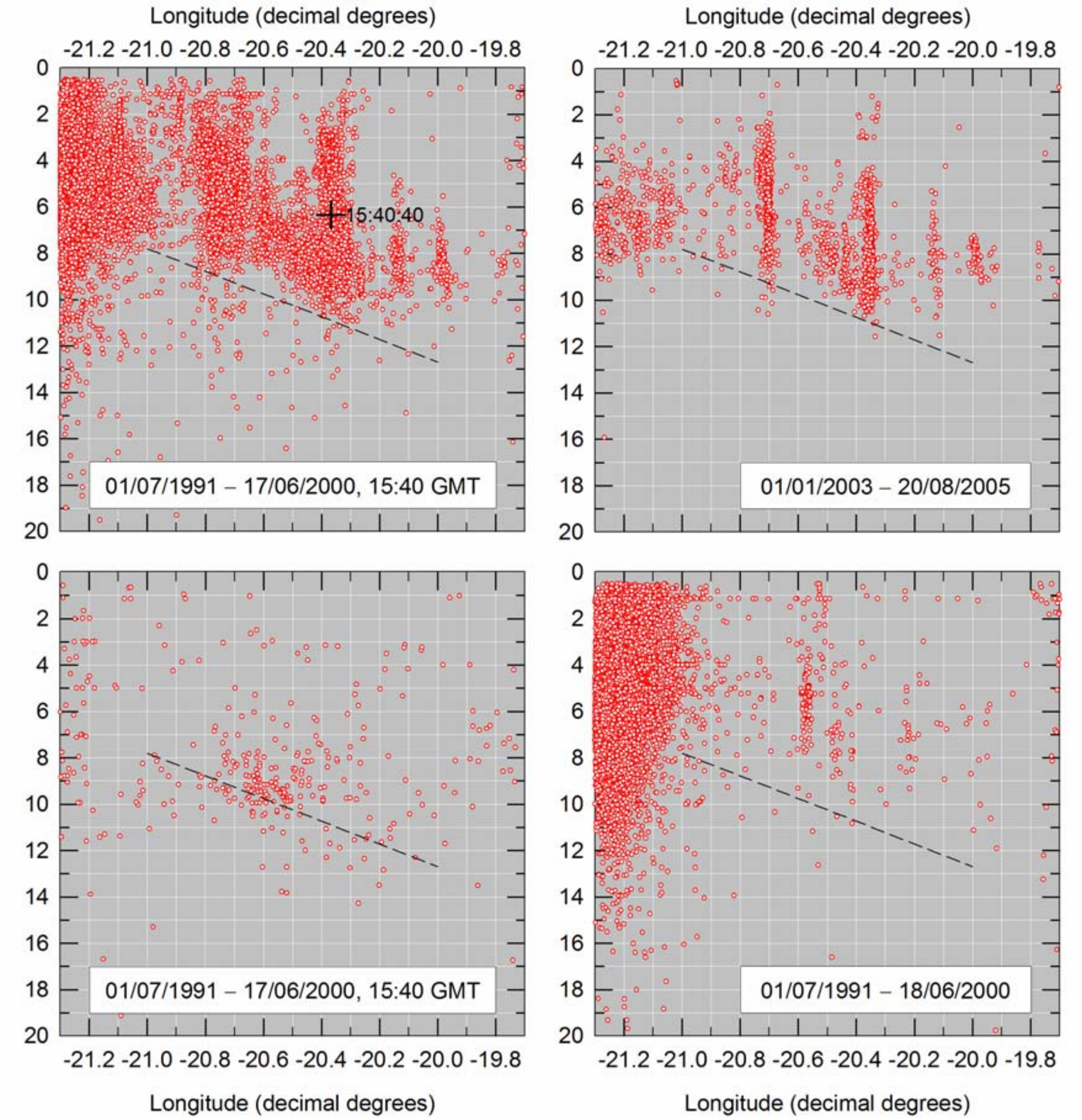

Figure 8. *The upper two figures show the depth of earthquakes in a 10 km slice in the SISZ for a period of years before the 2000 earthquakes. The most pronounced earthquake activity is further west, i.e. in the Hengill Volcano tectonic junction area releasing strong swarm activity especially in the period from 1994 to 1998, involving large tectonic strain release and volcanic intrusions at depth. The two lower figures are similar slices just south of and just to the north of the SISZ respectively.*

The limit of microearthquakes at depth have earlier been used to define the depth of the seismogenic crust at different points across the SISZ (Stefánsson et al. 1993) based on two years of recording in the zone. In Figure 8 a line has been drawn which indicates the depth of the elastic/brittle zone within a 10 km wide EW strip of the zone (between 63.92° and 64.02°N). The problem of drawing this line is that the depth limits of earthquakes depend on the stress, strain rate and variations in the pore fluid pressures. On one hand the microearthquakes tend to occur at higher levels at high shearing strains. On the other hand, during earthquake slip motion, strain rate is increased, so the ductile part of the crust acts as brittle, making the brittle zone look thicker. Nearby deformation and eruptions may deepen the microearthquake activity. In drawing this line we have to avoid such influences as possible.

It is assumed that the everyday activity of small earthquakes at 5 to 15 km depth is caused by high pore fluid pressures near a boundary separating the brittle and the ductile part of the crust. In the upper part of Figure 8 a line has been drawn to separate the apparently brittle crust from the ductile lower material within the zone. The sloping line of the figure is drawn at the bottom of the stable activity zone before the 2000 earthquakes, and it coincides with the depth of the aftershocks of the 2000 earthquakes, well after the slip motion has stabilized. Also are taken into account the changes of patterns seen in Figure 17, for different periods before year 2000 (see discussion in Chapter 5.4). It is assumed that the earthquakes occur preferably at the front of upwelling fluids in cracks penetrating into the brittle crust. On the other hand earthquakes occur in what usually is considered as a ductile crust during high strain rate. Thus intrusion pulses into the brittle crust cause high strain rates in the ductile part as microearthquakes. To draw this boundary as a sloping line is justified by the higher temperature gradient towards west coinciding with the aging of the crust towards east, i.e. away from the Western Rift Zone (Pálmason 1973; Stefánsson et al. 1993), which should be marked by a sloping boundary deepening towards east (see also Flóvenz and Sæmundsson 1993; Björnsson 2006; Ágústsson and Flóvenz 2005). The easternmost part of the central slice does not show earthquakes down at 12-13 km which the assumed ductile/brittle line indicates. It can also be argued that that short time since the last large earthquake on the fault at 20°W (released in 1912 earthquake) as well as the frequent Hekla eruptions in recent years have lowered the pore pressures near the brittle/ductile boundary, so therefore there are few microearthquakes there. The earthquake cluster at 8-9 km depth would then be remnants of high pore fluid pressures in the fault. It might also be explained as an expression of a hard core heterogeneity or asperity at 7-10 km depth in the fault, keeping the upper and the lower parts of the fault from moving and from creating small earthquakes.

The two bottom cross sections of Figure 8 represent the area south and north of the SISZ respectively (63.7°-63.85°N and 64.05°-64.2°N). A few earthquakes are seen down to the assumed brittle/ductile boundary even in the easternmost part. The data are much to scarce in these areas to draw definite conclusions, however, they indicate that the boundary could be of the order of 1 km deeper to the south of the SISZ than inside it.

It is of a great interest for modelling and for warnings to study better the brittle/ductile boundary, especially in this easternmost part of the zone, by introducing smaller earthquakes, applying denser network and relative location techniques and fault plane solutions. It is of interest in general to study more in detail the faint microearthquake activity in and around these faults to create a baseline for warnings.

5.1 Seismicity rate expressing stress changes

In Figure 8 we consider the time evolution of seismicity in the two SISZ clusters of Figure 2, approaching the large earthquakes, which occurred there in year 2000, by plotting the number of earthquakes larger than zero, which is near to the limit of completeness (Stefánsson and Guðmundsson 2005).

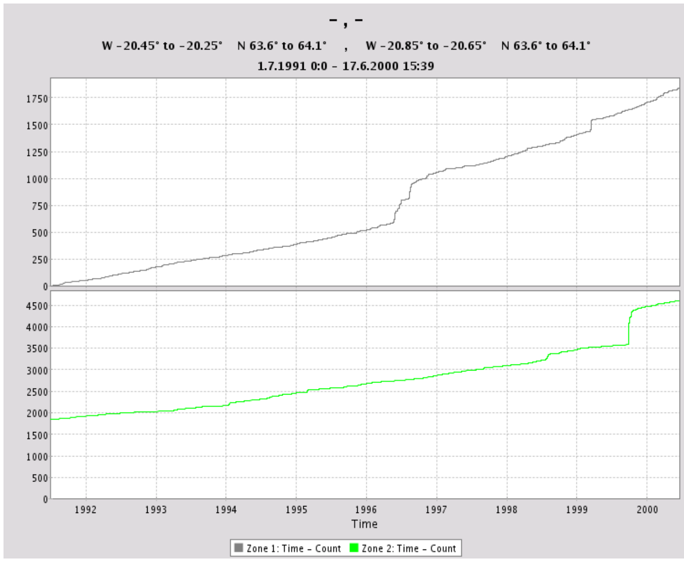

Figure 9. *The upper part of the figure shows cumulative number of earthquakes larger than 0 in the eastern cluster (cluster 1), i.e. where the first of the two earthquakes occurred., The lower part shows the evolution in the western cluster (cluster 2).*

Since after the activity swarm in 1996 there is a clear rate increase in cluster 1, i.e. where the first Ms=6.6 earthquake occurred (upper part of Figure 9). The even rate increase of the number (exclude the swarms) is expected to indicate stress increase in the area. However, we do not know how to use this for estimating, even very roughly, the time of onset of the probably impending event.

It is interesting that the area around the second M=6.6 earthquake (cluster 2) does not show such rate increase in stress from microearthquakes (lower part of Figure 9). Certainly it has more release of larger earthquakes near the end of the period. This indicates that the second earthquake was not preceded by a gradual build-up of stress in its surroundings as was for the first one. It was in the stress shadow of the hard core, the asperity of the first earthquake. The second earthquake was triggered by the first earthquake (Árnadóttir et al. 2003).

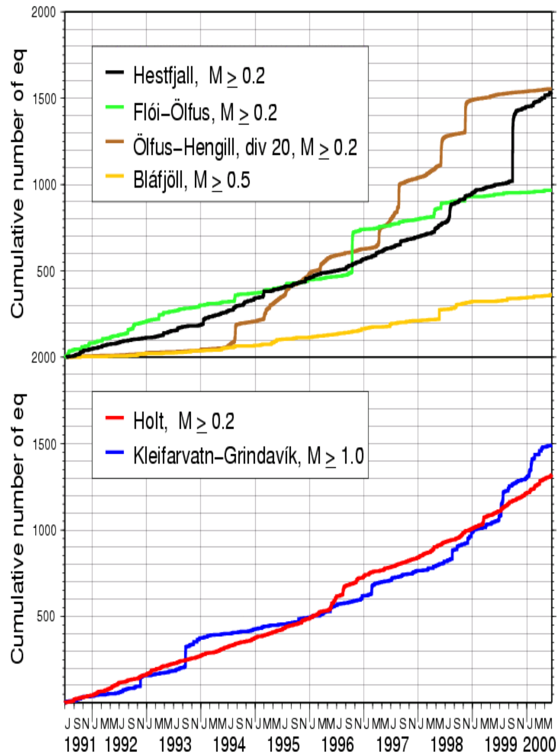

Figure 10. *The cumulative number of earthquakes above limit of completeness, different for each region of the SISZ and its prolongation in the Reykjanes Peninsula.*

It is difficult to draw conclusions about stress changes from seismic rate. It is usually considered that gentle changes may indicate stress increase or stress decrease, however, local swarms limit the possibility of using such methods. Also it would be more correct to say "increase in fracturing conditions", as the increase in number is also depending on changing pressures of lithostatic fluids penetrating high up into the crust. The two sites in the lower part of the Figure 10 indicate "stress" build-up. The Holt curve is in the focal area of the first earthquake while the Kleifarvatn curve is from the western end of the active zone where magnitude 5 earthquakes were triggered by the first earthquake, and thus were close to fracture criticality, i.e. had relatively high stress.

The seismicity indicates that the central part of the zone was in strain shadow from the asperity of the first earthquake and that the western end was also strained up close to the breaking limit when the June 17 earthquake occurred and triggered activity there.

5.2 Other patterns of seismicity indicating closeness to fracturing before the June 17 earthquake, i.e. the first main shock

Even if the general stress or stress build-up, expressed by the seismicity rate, before the occurrence of the first M=6.6 earthquake, is not indicating closeness to fracture, there was a clustering of microearthquake activity close to the becoming (expected) fault during the weeks before it, which may have predictive value about the rupture time. Some such patterns have been earlier discussed in Stefánsson and Guðmundsson (2005). Some more will be discussed in the following.

Figure 11 describes in map view the evolution of seismicity in the extended SISZ month by month from the beginning of January 2000. Looking only at the SISZ part a concentration is clear and elongated along the NS fault of the first 2000 earthquake especially during the last 2-3 weeks before the earthquake. Already this gives an indication of a rotation of the activity dominated by EW transform motion at depth towards perpendicular NS motion on the becoming fault of the first large earthquake.

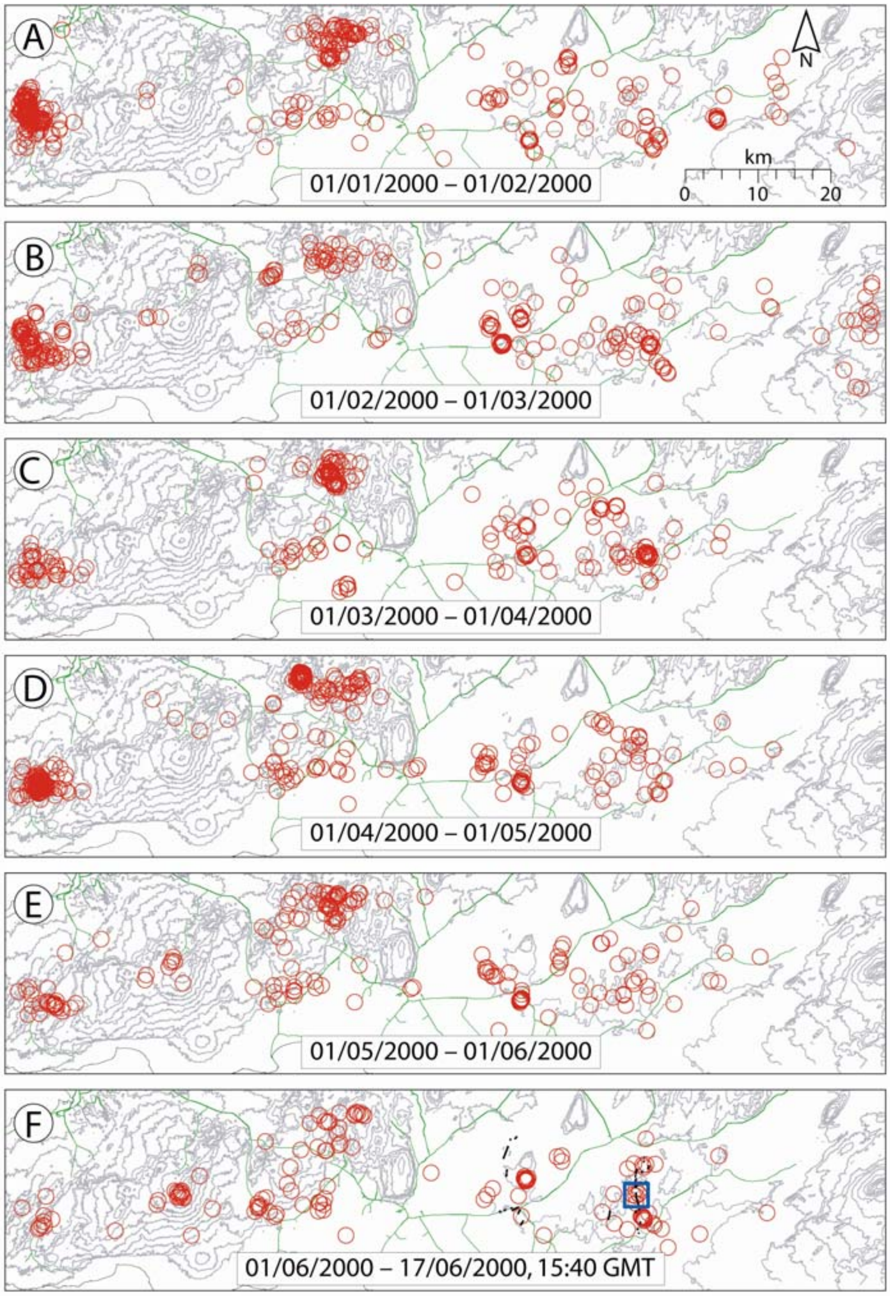

Figure 11. *The evolution of seismicity during 6 months before the June 17 earthquake. All earthquakes down to m=-1 are included. The SISZ is defined from longitude 19.8° to 21.2°W and latitude 63.9° to 64.0°N. We show here also the seismicity on the Reykjanes Peninsula further west which certainly is related to the activity in the SISZ. In between is the Hengill triple junction with very high swarm activity.*

The rectangular area of Figure 11 has a side length of 5 km, surrounding the June 17, 2000 asperity. In Figure 12 we see that microearthquake activity was evenly distributed in time at the depth of the asperity or just below it during the 2-3 weeks before the earthquake (see also Stefánsson and Guðmundsson 2005). Looking at the 5 km wide strip along the fault plane of the June 17 earthquake from April 1, 2000, we see that the depth distribution (Figure 13) starts to decluster with time around June 3. Looking at the latitude distribution along the fault we see the start of a similar declustering between the north and the south end of the fault a few days later (Figure 14).

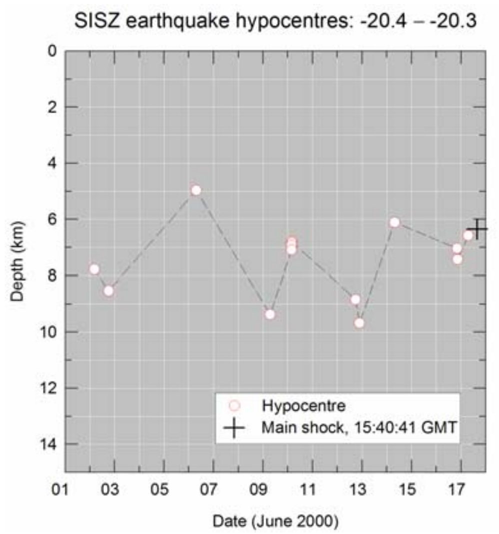

Figure 12. *Microearthquakes (magnitudes down to -1) at all observed depths in a rectangular 5 km area around the hypocenter of the 2000 earthquakes.*

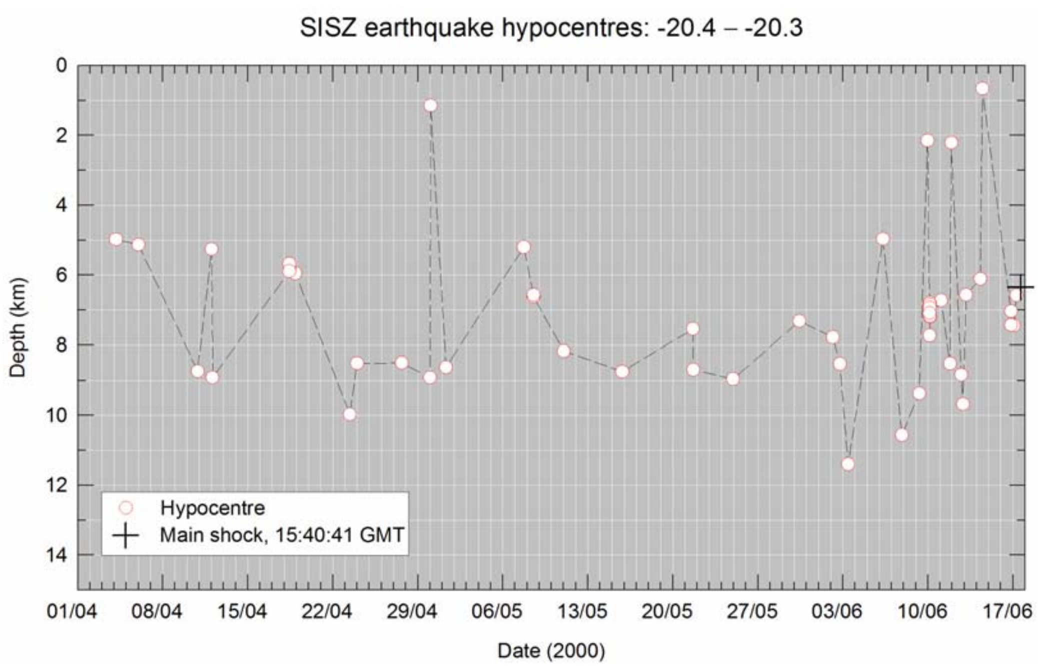

Figure 13. *Depths of microearthquakes (down to magnitude -1) with time from April 1, 2000 until the June 17 earthquake.*

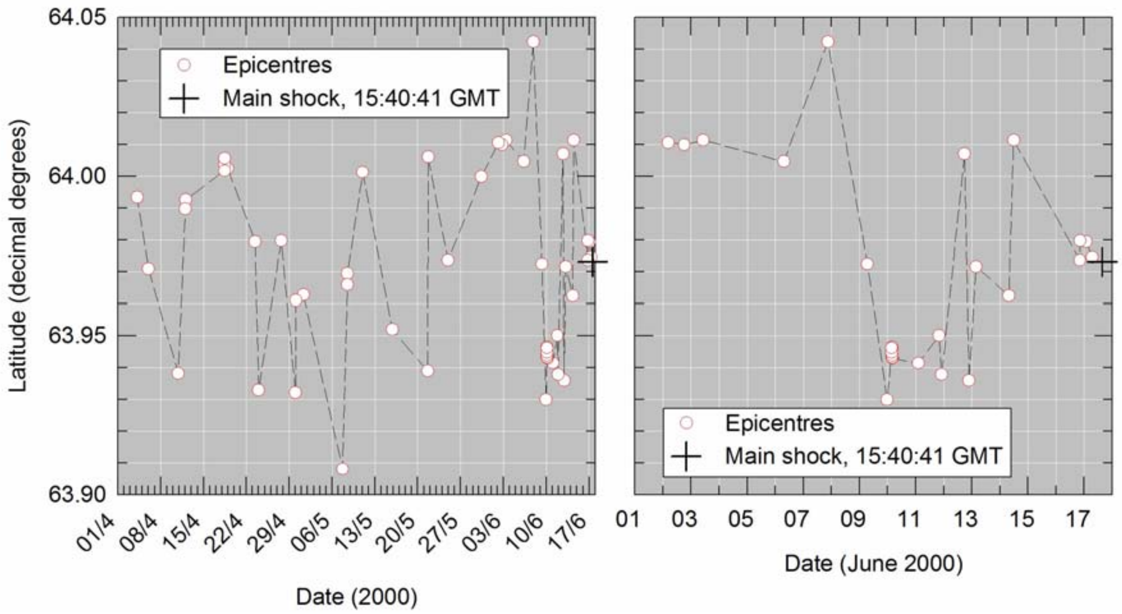

Figure 14. *Epicenters of microearthquakes along a horizontal strip along the NS elongated fault plane of the first 2000 earthquake during with time before the June 17 earthquake.*

The declustering of positions of the microearthquakes in time indicates that small motion started on a large part of the the fault 2-3 weeks before the June 17 earthquake. It is expected that if a motion starts on the fault as a whole, we would see the end effects at both ends of the fault, reflected in small earthquakes and the straining of the asperity. Based on the development of the microseismic activity the fault motion started at the northern end of the fault, declustering in depth. It moved to the southern end, and for a short time we recorded small earthquakes at both ends of the fault and at various depths. It started relatively deep as it seems below the brittle/ductile boundary. During the last 24 hours before the earthquake we see clustering in the asperity. Stable fault motion below the seismogenic zone started on a NS fault a few weeks before the earthquake of June 17. The earthquakes alternating between the N and S are fault-end effects caused by the motion across the fault. The asperity with a center at 6 km depth and diameter of approximately 3 km (Hjaltadóttir and Vogfjörð 2005) hampers the slip motion and is stressed by it until it breaks, nucleating the earthquake.

It is interesting that the only microearthquakes seen during the time from 1991 to 2000, below in the ductile crust, i.e. at 10-12 km depth, were directly below the asperity. Microearthquakes are exceptional in the ductile zone, and only observed where the strain rate is relatively high. One explanation to this is that there is a relatively hard core also deeper down and that the stable motion down there is locally hampered by this hard core, causing stress heterogeneity and thus microearthquakes.

The last phase in the earthquake nucleation process is seen as the breakthrough of the asperity of the June 17 earthquake in the first part of the P-wave of the seismic record. This phase of the earthquake is comparable to a magnitude 5-5.5. This nucleation phase is well recorded in the strainmeters in the neighbourhood of the earthquake epicenter. Figure 15 shows the nucleation signal recorded with a strainmeter at Skálholt, at 20 km distance from the epicenter. The first two seconds of the record show a signal before the real onset of the fault starts. The second and triggered earthquake of June 21 did not have such an asperity signal, i.e. it was not triggered by the breaking of such an asperity.

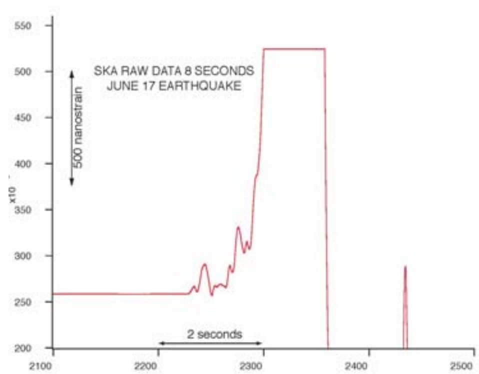

Figure 15. *Strainmeter record from SKA, 20 km to the west of the June 17 earthquake. It shows a nucleation phase 1.8 seconds before motions starts on the fault plane as a whole.*

5.3 An algorithm based on a microearthquake spreading along a becoming fault, for visualizing and alerting about a stable fault motion ahead of a large earthquake

The starting of the earthquake motion along the June 17 earthquake gives rise to a simple warning algorithm that takes into account that the seismicity is alternating along the whole fault during a short period of time before the real onset of the earthquake, i.e. declustering in space during a short time. We assume that the fault will be a 5 km wide NS strip at the right place, as predicted long time ahead of the earthquake. We select one week for a coincidence period for earthquakes, i.e. earthquake occurring within such period are defined coincident. The algorithm describes declustering along the fault by adding together the distances in consecutive microearthquakes within a predefined fault zone during 7 days ahead of any new earthquake, depths and distances along the fault plane. For simplification in this case of a NS fault it is enough to take the difference in latitudes.

In Figure 16 we see one warning curve for depths and another for horizontal motion. In fact this has a value for the visualization because the vertical declustering on the fault starts a few days ahead of the microearthquakes along the fault (Figure 16). Even on this figure of 7 days coincidence interval we can see the clustering just around the hypocenter/the 3 km asperity within 24 hours before the earthquake.

By taking into account clustering from a larger area towards the fault strip defined here, even a considerably stronger pre-earthquake signal is obtained. And of course assuming even a narrower fault area still increases the signal.

In reality in alerting and visualizing this initiation of the motion or this stable first part of the earthquake we multiply with an areal clustering. Take also into account the mechanism as these changes are described for example in Stefánsson and Guðmundsson (2005). For observing the earthquake clustering within 24 hours ahead of the earthquake it is of course better to apply a shorter coincidence time than a week which is applied in Figure 16.

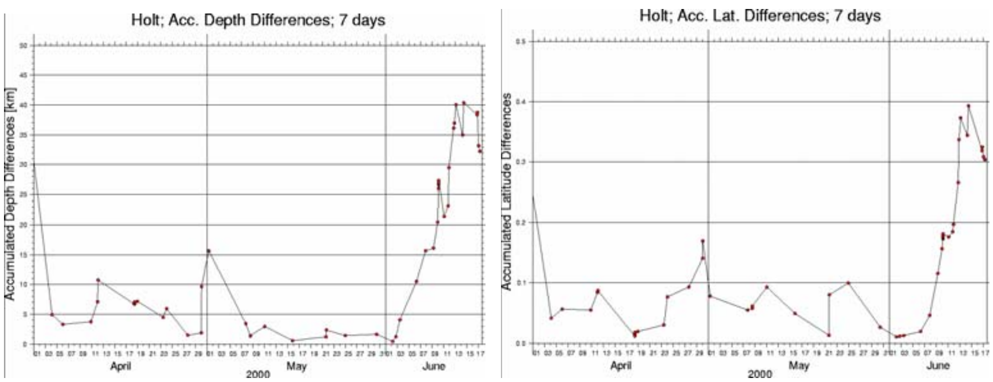

Figure 16. *Accumulated distances between consecutive microearthquakes with time. A week is selected as a coincidence period. The figure to the left shows accumulated depths, while the figure to the right shows distance changes along the fault, i.e. latitudes.*

This algorithm breaks down when the whole SIL period from 1991 is used. A possible reason for that is that this last period of time before the earthquake is an exceptionally low activity seismic period, and location errors may cause a lot of earthquake coincidences. Much work is left in testing this algorithm, especially for its significance during longer periods. It can, however, already at this moment be introduced into the visualizing procedures of the EWIS-system.

5.4 Depth variations indicating fluid pressure or stress variations

Fluids that migrate from lithostatic depths into the crust cause near lithostatic pore pressures at shallow depths in the crust, however, mostly below the lithostatic hydrostatic boundary at approximately 3 km depth, helping to release earthquakes by reducing the normal pressure on fault planes. Microearthquakes at shallow depths in the crust indicate high local fluid pressures and high local stresses. The difference to a nearby part in the fault zone which does not have a lot of microearthquakes at shallow depths does not indicate that it has lower stress, rather that it has lower fluid pressures.

The fluid pressure in the feeding volume at depth increases by time if it does not open for pressure release in some fracturing or faulting and related outflow of fluids. Deepening of earthquakes by time in an area may indicate lowering of fluid pressures by time and thus less probability of earthquakes, i.e. if the cause of the lowering is outflow outside the area (related to eruption or other intrusive activity or faulting). However, deepening can also indicate a pre-earthquake process in the area and inflow of fluids into the local earthquake fault and thus lowering fluid pressures locally.

In areas where we have lithostatic fluids penetrating into the crust the approaching to a large earthquake is not necessarily expressed in high stresses ahead of it. It would probably be seen in increased number of intrusive driven swarm episodes at shallow depth. To trigger a large earthquake we need high stress across a large fault and enough fluid in the surroundings to smear a large part of the fault area, or as it is sometimes claimed, to homogenize the stress around the fault so the fault motion does not stop before it gets to the fault area as a whole.

Depth of microearthquakes may be a good indicator about the faulting conditions in the zone. If the earthquakes are mostly deep it indicates relatively low stress and low fluid pressures. Getting shallower they indicate increase in either or in both. Coming to a certain high level of depths they indicate closeness to fracturing conditions, however, temporal migrations above that level do not so far give us any indication about closeness in time to a large earthquake. Rather if interpreted correctly a deepening may be an indication of a process ahead of an earthquake, comparable to stress relaxation.

It is clear from what here is said that we are still far from being able to use statistical mass evaluations of the number of earthquakes at shallow depths to predict the closeness to an earthquake. A study and modelling of the detailed features of the fluid migrations expressed in an earthquake is more adequate in approaching time prediction.

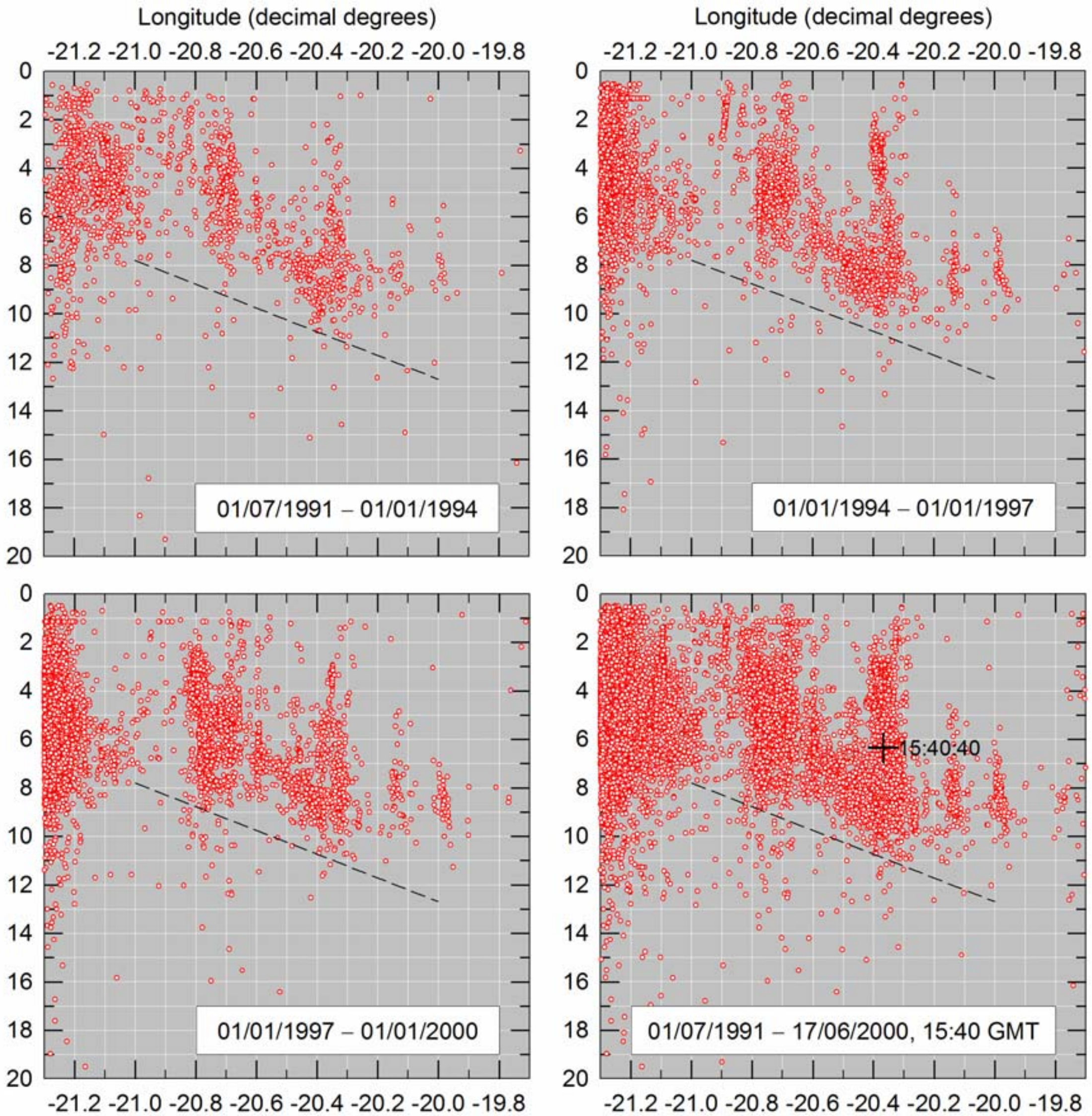

Figure 17. *Depth distribution of small earthquakes along the SISZ during 1991-1994, 1994-1997, 1997-2000 and 1991 - June 17, 2000, respectively. Earthquakes down to magnitude zero are used. The sloping lines are the depth lines of the brittle/ductile boundary as earlier discussed.*

Figure 17 contains patterns which indicate that apparent depth changes of the seismogenic zone at various places in the SISZ may express local and temporary stresses and fluid pressure changes. Assuming that the microearthquakes are triggered by the common action of tectonic shearing strain and the expanding pore pressures caused by the upflow of lithostatic fluids, the depth of the microearthquakes is inversely a measure of shearing stresses, i.e. if there is not a stress releasing fracturing on a larger scale active at the same time.

There is a distinct shallow line (indicating high stress) seen in the figure shallowing to west from 20.8°-21.1°W longitude of the first period, i.e. during 1991-1994. This line and some more activity there is not seen in the later periods, possibly because of stress or fluid pressure release related to the volcanotectonic episode in the Hengill area starting in 1994, releasing stresses in its surroundings and swallowing fluids from the fluid-feeding volume at depth.

Figure 18 gives an indication that depths may be used to study the build-up fracturing conditions for earthquake release. The left part of the figure shows depths of microearthquakes prior to a magnitude 4.8 earthquake in the SISZ on September 27, 1999. This was the biggest earthquake in the SISZ from the start of the SIL-system until the 2000 earthquakes.

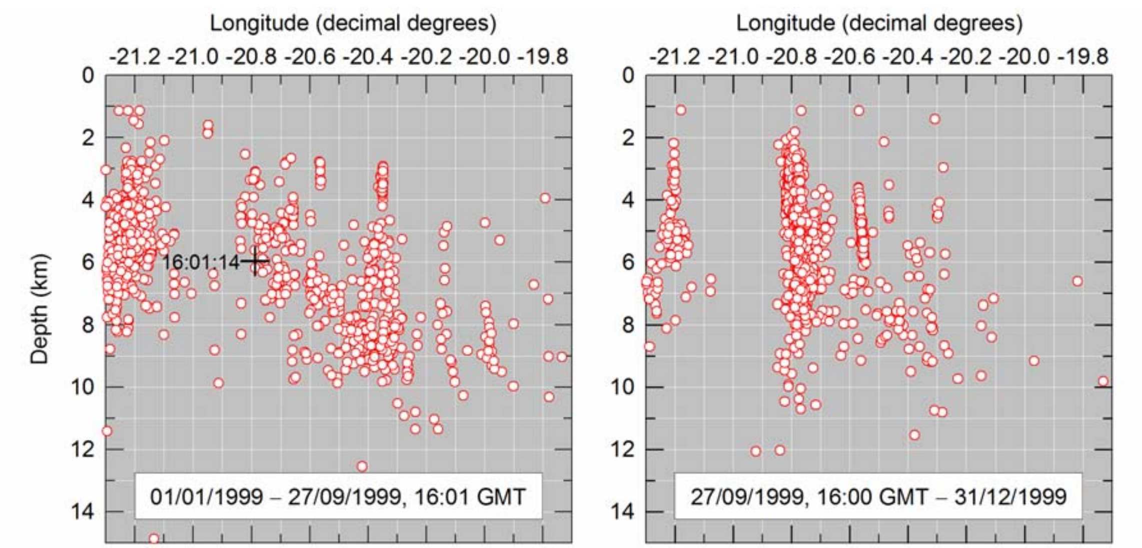

Figure 18. *Depths of earthquakes in SISZ prior to a magnitude 4.8 earthquake in the zone.*

The shallow microseismicity in the central and western part of the zone indicates high fracturing stress level there. The microearthquakes occur all the way up to the hydrostatic-lithostatic boundary, assumed close to 3 km. The period after the earthquake is shown in the right figure, i.e. aftershocks from 2-10 km depth. After the earthquake was released the microearthquake depth level is generally lowered (except of course in the source area). This is clearer one month later (see Figure 19).

Another feature that is interesting during the high „stress“ before the earthquake is that it is not seen in the easternmost part of the zone. This probably implies that tectonic strain has not developed to such conditions that lithostatic pore pressure build-up in the upper part of the crust has started. This waits for a partial break-up of the zone in medium-large earthquakes as was observed during 7 years prior to the 1912 earthquake in the very eastern part of SISZ. This may have happened again about 1950 in the central and western part of SISZ, half a century before the 2000 earthquakes. So it seems that we have to reach a certain strain level, starting to break up the seismic zone, before migrating fluids can penetrate into the zone and then possibly becoming an indicator for high fracturing conditions.

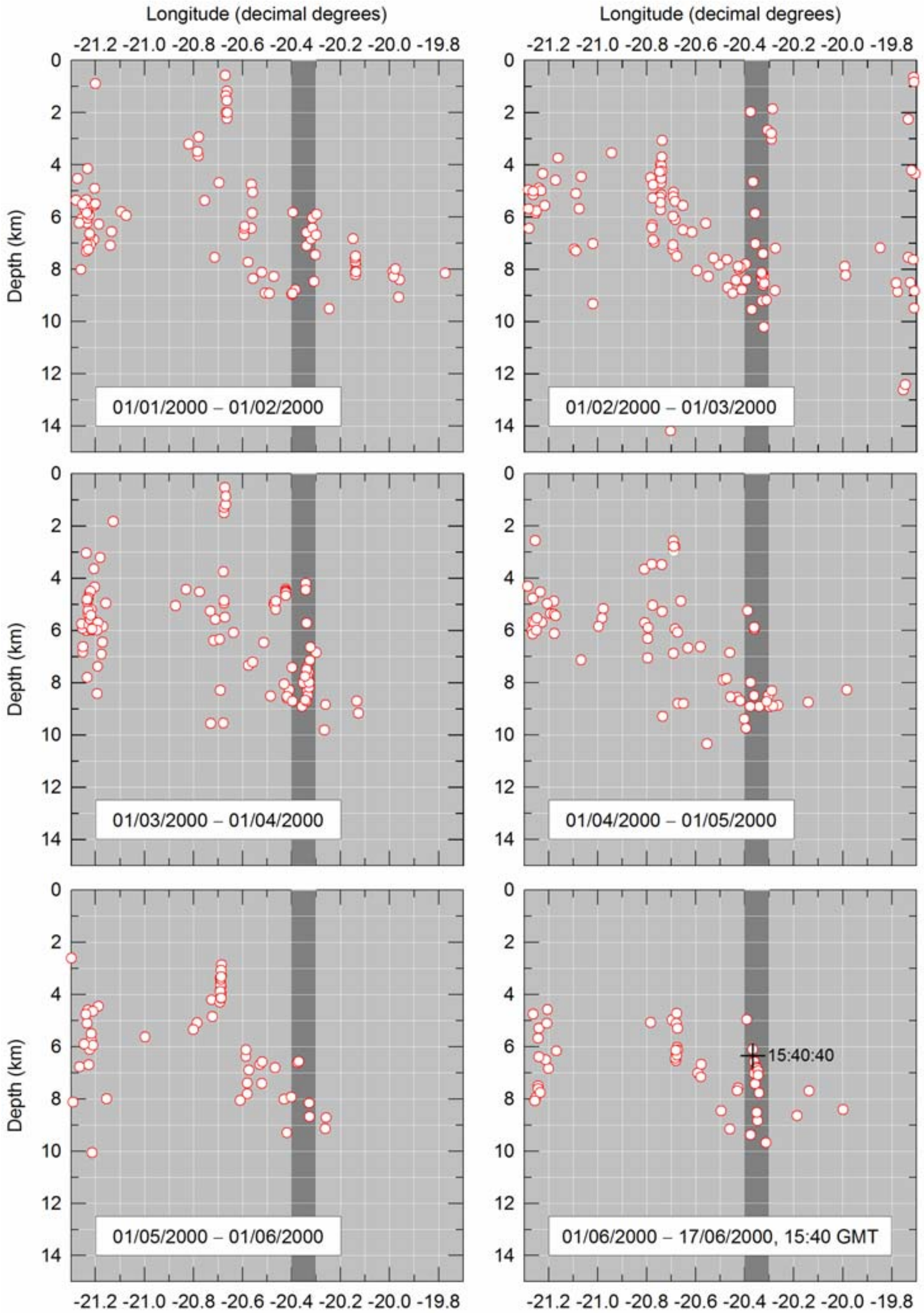

Figure 19. *Depths of seismicity in the SISZ during the months before the June 17, 2000 earthquake. The dark shaded field in the figures is a 5 km strip around the NS elongated fault. The relatively high activity in the easternmost part is mostly following the eruption of Hekla that started on February 26.*

Of course a question awakes if it is possible to use stress or fracturing level based on depths of microearthquakes to warn for large earthquakes. However, it is not so simple as that of just waiting for very shallow earthquakes. It is not so simple that stresses or fracturing level go up until the earthquake breaks out. In Figure 19 it seems indeed more like the opposite. The depth seem to go down during the year before the earthquake with an exception of shallowing in the second month, where we see general increase in seismicity at various depths, also shallow, possibly related to strain changes preceding and related to the Hekla eruption. This indicates that it may be useful to use the shallowness of earthquake sources as an indication of closeness to earthquakes, however, care must be taken by other observations and modelling what to relate the changes to. As will be discussed later the model we apply which involves much fluid mobility certainly involves stress fall during the homogenization of stress in the fault surrroundings, i.e. energy goes to this first part of the earthquake release process.

5.5 Some examples of the time evolution of depths

It has been shown that fluids with lithostatic pressure penetrate into the crust from below and increase pore or fluid pressures there (Zencher et al. 2006). This should provide a contribution to earthquake forecasting because elevated pore or fluid pressures in the fault area of a large earthquake will help to release it. Microearthquakes are, however, the only indicator for such stress increase and they occur intermittently (Figure 20 shows one such penetration of fluid from below). For many practical applications or for short-term warnings we are mostly talking of fluid pressure changes, rather than pore pressure changes. So number of fluid penetrations up through the crust gives more easily observable information about changes in fracturing conditions than the general shallowing of microearthquakes.

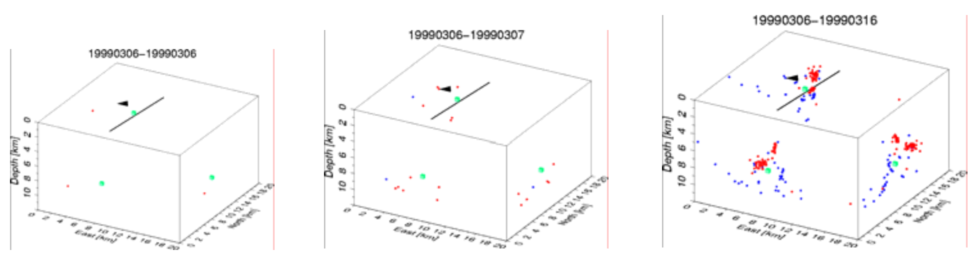

Figure 20. *Upward migration of fluid and thus fluid pressures from near the bottom of the brittle/ductile boundary during a period of 10 days expressed in microearthquake migration. The last day activity is in red dots, the older in blue. This episode occurred 15 months before the June 17, 2000 earthquake. Its hypocenter is marked by a green star.*

If the closeness to fracturing conditions were on a longer time scale fluid penetration would gradually build it up. Areas hit by this are weak areas and observations of elevated pore pressures are among the best methods to localize a becoming faulting.

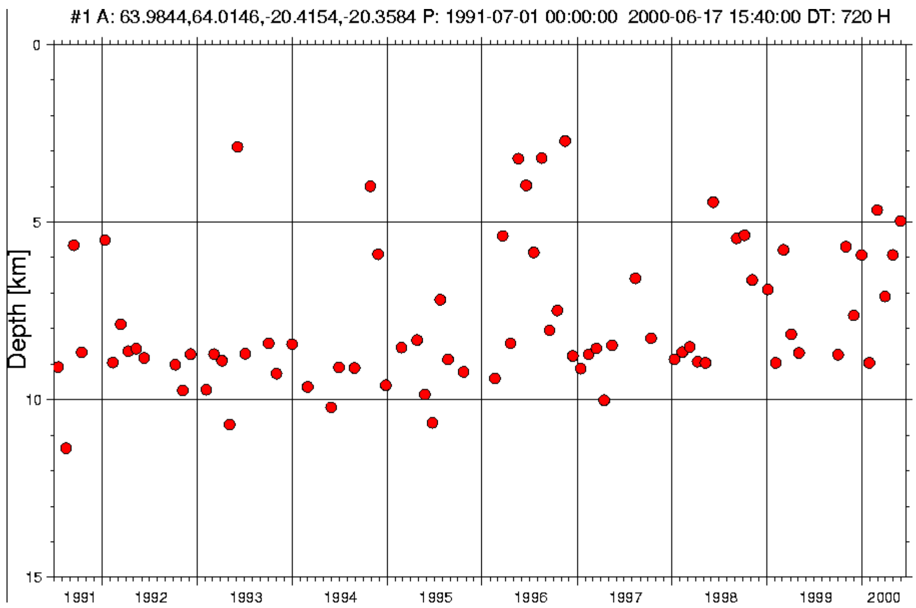

Figure 21. *Depth of earthquakes with time. A median value of 30 days in the most active area around the epicenter of the June 17 earthquake.*

To plot the time history of Figure 21 we have selected the most active area around the June 17 earthquake to try to monitor local apparent stress increase. The reason to select a limited area is that there seems to be an interplay between the various stress outlets of the area. The level in a nearby area may loose "stress" while the other is gaining. Expected long-term strain build-up in this area does not appear continuous. What we see in this figure is increasing number of fluid intrusions up in the crust. With probably gradually increasing strain around the area shallow depths become more frequent, especially after a stress pulse in 1996. There are relatively more shallow earthquakes on the later half of the graph, i.e. higher "stresses" were indicated. This is in line with stress increase before the June 17 earthquake as expressed in continuous microearthquake rate increase since 1996, as seen earlier in this report.

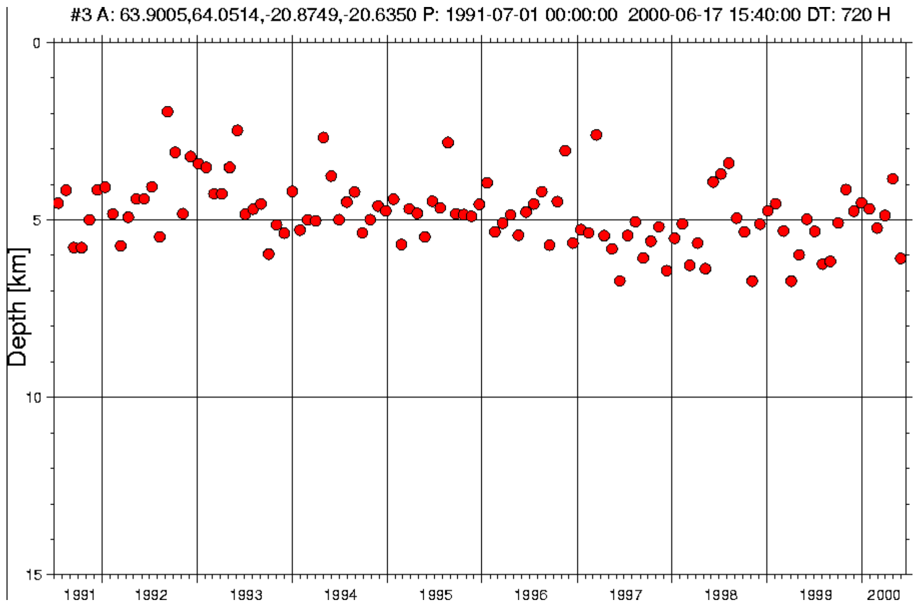

Figure 22. *Depth of microearthquakes with time. A median value of 30 days in the area of the second earthquake, June 21, 2000.*

In Figure 22 lowering of "stress" is rather indicated than build-up. This may be another indication that this earthquake stood in the shadow zone of the first earthquake, and its asperity. Lot of earthquakes occurred in this shadow, however, probably at low compressional stresses. This again is in general in agreement with the cumulative seismicity, which was not rising before this second earthquake except at local episodes. It looks as „stress" would rather be stable or lowering than building up, except possibly during the last year before the earthquakes.

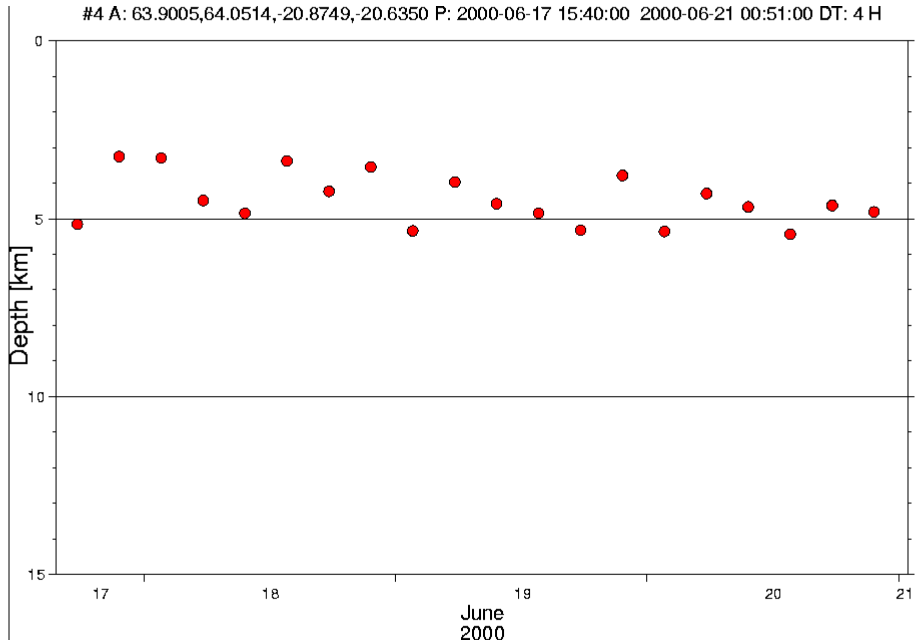

Figure 23. *The medians of earthquake depths after the first earthquake in the hypocentral area of the second earthquake, which was triggered by the first.*

We see again "stress" reducing before an earthquake, as indicated by the depths in Figure 23. In fact the seismic rate of microearthquakes also indicates stress decay with time between the times of the earthquakes. The second earthquake was triggered by the first (Árnadóttir et al. 2003) and we see large stress release first after the triggering slip started at depth at the south end of the fault and the fault motion which started there gradually migrated towards north to all parts of the fault. Fluid pressure that constituted a part of the initial stress or fracturing conditions on the fault flowed along the fault plane in response to the beginning motion had the role to homogenize the stress field along the fault and thus making movement along the whole fault possible. This homogenizing of the stress means also that the microearthquakes distribute along the fault instead of distributing high up in the zone. In other words to conserve energy, when spreading out laterally, the "stress" becomes lower.

6 Some significant characteristics of the earthquake processes in the SISZ since 1896 and their relevance for dynamic hazard assessment

In general the strain build-up and strain release in the South Iceland Seismic Zone, based on historical data and the general plate movements, has been described for example in Stefánsson and Halldórsson (1988), Einarsson et al. (1991) and Stefánsson and Guðmundsson (2005).

A striking feature is the silence after the large earthquake sequence of 1896-1912, pointing to a relatively complete stress drop in the whole SISZ.

It has been estimated that the released moment in the 1896 earthquakes occurring in the central and the western part of the SISZ was $5.2*10^{26}$ dyn cm. The first and starting earthquake of this sequence had the moment $2.8*10^{26}$ dyn cm based on Stefánsson and Halldórsson (1988). If we assume that all of the local deviatoric moment strain was released in this first earthquake and that the other shocks were triggered aftershocks as is a plausable assumption this is equivalent to the needed shearing potential or potential moment to cause earthquakes of this size in this part of the zone. Based on the same estimations the 1912 earthquake released $4*10^{26}$ dyn cm. If we assume that the moment in the whole of the zone was totally released in the two large 1784 earthquakes the moment build-up from 1784 to 1896 was $2.5*10^{26}$ dyn cm/year, which brings the potential moment in the easternmost area only to $3.2*10^{26}$ dyn cm, i.e. significantly lower than needed to release the 1912 earthquake. Increased local plate velocity may have increased somewhat after the 1896 earthquakes but probably of an order of magnitude lower than what is needed for so fast moment build-up. A remaing explanation would be that the moment at the easternmost part was not totally released around 1784 as assumed, $0.8*10^{26}$ dyn cm was left over from before, comparable to magnitude 6.6-6.7. The assumption that the whole zone was released in the 1784 earthquakes is not based on knowledge of a large earthquake in this area around 1784. The last known large earthquake in this part of the zone was in 1657 and another slightly further west (10 km) in 1630. The moment strain build-up from 1657 to 1912 was $6.4*10^{26}$ dyn cm calculated as above enough for a magnitude 7.1 earthquake.

What does the above indicate about the next probable large earthquake in the SISZ?

In the easternmost part of SISZ, i.e. near the 1912 earthquake origin, it may be assumed that a maximum of $2.4*10^{26}$ dyn cm was leftover moment in 1912. Assuming this the build-up time for a magnitude 7 earthquake in this area would be of the order of 160 years, i.e. reached in 2072, and of a magnitude 6.9 earthquake 112 years, i.e. reached in 2024. Assuming that moment release was total in 1912 would make these times much longer.

In the central part of SISZ around 20.4°-20.5°W longitude, earthquake of magnitude 7 occurred in 1784 (sometimes assumed to have been 7.1) releasing moment of at least

$4*10^{26}$ dyn cm. The first 2000 earthquake about 5 km from the 1784 earthquake, was magnitude 6.6 (Ms), i.e. released the moment 10^{26} dyn cm, which is only ¼ of the moment released in the nearby fault in 1784. Calculating build-up of strain moment since 1784 with same formula as above, subtracting the moment release in the 1896 earthquake at 20.5°W longitude and the 2000 earthquake at 20.4°W longitude potential moment for a magnitude 7 earthquake in this site will be reached in 2016. If we assume that the second of the 2000 earthquake moment release is also subtracted, i.e. 10 km away, we would have magnitude 7 earthquake potential here in the year 2058, and magnitude 6.9 in 2010.

6.1 The hope to be able to predict the time of the next earthquake in the SISZ relies on observing microearthquakes

These calculations above are not shown to be any kind of prediction of time and size of earthquakes. Firstly the calculations of potential moment at each place must be and probably can be better underbuilt. But the main problem comes when we try to estimate the size of the becoming earthquake. The size seems to be time-dependent in such a way that the longer the waiting time is the larger can the earthquake become at each place. However, the time of rupture relies also on the pore pressures or fluid abundance around the faults which are independently variable each place.

Only of the order of ¼ of the expected total moment release in the SISZ was released in the 2000 earthquakes according to the calculations above. If this is close to being right we can expect that the remaining will be released in two close to the order of magnitude 7 earthquakes in the area during the next few decades. It is most probable that this will happen where we have had such large earthquakes before, i.e. in the faults of the 1784 earthquake as in the fault of the 1912 event. An earthquake on the 1784 fault may be closer in time than the 1912 type earthquake.

Roth (2004) did calculate the shearing stresses near the 1912 earthquake based on stress contributions from earthquakes in the zone since 1700. According to him the 1912 earthquake occurred when the expected shearing stress was only a half of what was expected to be needed for a large earthquake there. This supports the suggestion that the build-up of moment in this part of the zone had been going on for much longer time than in the western parts of the zone.

Stefánsson and Guðmundsson (2005) have described earthquake clusters around the 2000 earthquakes before their occurrence having an areal distribution of 10*10 km, as seen by detailed SIL observations since 1991. As seen in Figure 4 these clusters are definitely observed since 1964 and possibly from 1926. The conclusion was that the observed clusters of small earthquakes were due to high pore fluid pressures caused by upwelling fluids with lithostatic pressure conditions in response to gradual plate motion strain build-up. These upwelling fluids gradually corroded or weakened the fault area to prepare for a large earthquake. After the earthquakes the aftershock activity was limited to a narrow zone around the earthquake faults as the areal high pore fluid pressures were released by the earthquake.

It is interesting to consider the preparation process of the 1912 earthquake of magnitude 7 in light of this. There was no observed seismicity in the 1912 epicentral area for the first 7 years after the 1896 earthquakes (Figure 3). A swarm activity period started to be observed there by magnitude 4-4.5 during 2003, occurring 8 years before the 1912 earthquake. The fluid driven activity which started at least 36 years before the 2000 earthquakes started only 8 years before the 1912 earthquake. It looks as if the 1896 magnitude 6.7-6.9 earthquakes at only around 10 km distance from the 1912 earthquake stirred up processes, which made fluids to be released from below the brittle/ductile boundary to gradually bring lithostatic pore pressures up into the brittle crust, triggering earthquake sequences in their way when reaching to shallower depth. It seems that the fluid pressure process was much faster before the 1912 earthquake than before the 2000 earthquakes, but of course it may have started before the 1896 earthquakes even if it did not reach same hights there as in the 1896 areas. The closeness to a large Hekla eruption of 1845 only of the order of 10 km to the east of the earthquake fault may have hampered the fluid pressure build-up there by releasing the pore pressures, so it was not ready to be released in 1896.

There are many unanswered questions here about the fluid pressure build-up, for example how much strain is needed so the brittle crust, below the lithostatic-hydrostatic boundary, becomes permeable towards fluids with lithostatic pressure. Also how much can nearby eruptions or earthquakes release of the pore pressures. These challenging questions should be dealt with.

A tentative model for the preparation and reactivation of the 17 June, 2000 fault is that of opening up by the magnitude 5 earthquakes in 1948 in the southern half of the fault and 1964 in the northern half of the fault, i.e. to the south and to the north of the 2000 earthquake asperity. The corrosion of the fault plane or damage zone started or reached a new level with these two earthquakes. The same is valid for the June 21 earthquake. The June 17 earthquake has an asperity in the middle, i.e. in a 3 km diameter hypocenter area while the second has a barrier at the southern end shortly to the south of the hypocenter. These conditions result in a somewhat different distribution of the seismic clusters before the two earthquakes. But what is common is the spread-out distribution of the clusters before the earthquakes as compared to the more linear distribution shortly before these and very clearly afterwards (see Stefánsson and Guðmundsson 2005).

6.2 The seismic activity at the eastern and western end of the SISZ

The east and the west ends of the SISZ are volcanoseismic areas. Episodes there can be defined as volcanoseismic.

The Vatnafjöll earthquake of 1987 (Bjarnason and Einarsson 1991) is only 10 km to the east of the 1912 earthquake, i.e. east of what usually is called the SISZ and is on a prolongation of SISZ. It occurred on a NS fault releasing moment which is only a fraction of the large SISZ earthquakes. The remaining strain moment can be expected to be released aseismically and related in recent years to the frequent Hekla eruptions, 1970-2000 (Table 1). Hekla eruptions often lead to a flurry of small earthquakes along the

SISZ. It was so in the 1991 eruption, but not in the 1980 and 2000 eruptions. It may be speculated that the strong asperity of the first June 2000 earthquake hampered strain build-up along the zone until it broke in the two large earthquakes. The same can be speculated for the silence after the 1980 eruption, i.e. that an asperity of the 1987 earthquake held against strain build-up towards west in the SISZ.

The western end of the zone, the Hengill region, had seismic swarm episodes culminating in magnitude 5-5.5 earthquakes in 1955 and 1998. It has been speculated if the 1998 episode in some way triggered the 2000 earthquakes. In this report it has been pointed out that the Hengill episode in 1998 seems to have reduced fracturing level in the area of SISZ closest to it, probably by reducing pore pressures there.

It has also been suggested that the Hekla eruptions may delay earthquakes in the nearby Hekla areas by reducing pore pressures there.

Such features and possible coupling must be studied much better than we have been able to do here, on basis of seismic information and on basis of deformation measurements. Understanding of this can be significant for earthquake prediction research. Taking into account the reduction of deep fluid pressures at depth in earthquakes and volcanic eruptions it can be expected that it will lower fracturing conditions in the areas near by, maybe of the order of 10 km away.

It has sometimes be speculated if the Hengill episode of 1994-1998 and the Hekla eruption in February 2000 loaded the SISZ to help gradually to make it break. It is just as well justified to say that the conditions for the Hengill and the Hekla episodes were that they were in a strain shadow of the “SISZ asperity” and were because of upflow of fluids, the first to break in a large event taking to the whole of the SISZ from Hekla in the east to Kleifarvatn in the west.

7 Earth-realistic models, the basis for predictions

Among very significant results of theoretical modelling within the PREPARED is that:

- There is expected a concentration of earthquakes in the proximity of the elastic brittle/ductile transition caused by increasing stress.

- High pore pressures can migrate from below the brittle/ductile transition to shallower depths. Thus episodes of fluid migrations can increase pore pressures up to lithostatic values and thus increase the instability of the faults (Zencher et al. 2006; Stefánsson et al. 2006).

7.1 Some baselines for modelling the SISZ as a whole

On basis of boundary type microearthquakes in the SISZ the depth to the brittle/ductile boundary is estimated 13 km near the eastern end of the zone (at 19.9ºW) and 7.5 km further west in the zone (at 21.0ºW) (see figures and discussion earlier in this report; Stefánsson 1993).

The bottom of the seismogenic zone is corrugated by cracks and fractures which at some places penetrate high up into the crust in response to tectonic straining of the zone. Former stress release/pore pressure release history and heterogeneities (asperities) accommodate for where this upstream of high pore pressures goes at different times.

Microearthquakes, caused by the interaction of fluids and the rocks in this tectonic environment, also define the SISZ in a narrow sense, i.e. in the sense of high effective pore pressures and fluid mobility. The SISZ in this sense is illuminated by the everyday seismicity and is only around 10 km wide (roughly between 63.9º-64.0ºN), running approximately from east to west.

A continuous transverse left-lateral motion along this zone is expected at depth, but expressed in the upper crust as right-lateral faulting in earthquakes with NS right-lateral fault planes. These earthquakes occur as an interaction between tectonic strain build-up in and near the zone and influx of fluids (assumed mostly water fluids).

The lithostatic fluids are slowly and continuously released from the upper mantle (at 50-100 km depth) to the ductile crust. Stable left-lateral semi-continuous slip on an EW fault (the SISZ direction) in the ductile crust releases fluids up to the bottom of the elastic/brittle crust where they are trapped until strain is enough in the brittle crust so it starts cracking in the presense of the fluids (Rivalta and Bonafede 2005). Relative high intrinsic permeability of the basaltic crust and the plate motion straining of the crust create conditions for fluids to flow up into the crust and weaken it by causing high pore pressures and earthquakes where other conditions are ready, first at the bottom of the brittle crust and gradually farther up where they are usually larger (Zencher et al. 2006). The interseismic pore pressures are released in large through going earthquakes, so pore pressures in the crust become lower around the particular fault than elsewhere in the zone. Thus it is likely that the next earthquake selects another site where pore pressures have not been released for a longer time, even close to the recently released one which is the most usual in the SISZ (Figure 1). So basically the upwelling fluid pressures decide which place is selected for the earthquake in response to the plate motion in the SISZ.

According to the modelling results pore pressures would be strong near the recent large earthquakes, illuminated in a narrow band of aftershocks slowly fading out with time, although released again during strong strain changes in the area, a new potential earthquake site, high pore fluid pressures would gradually help to release earthquakes higher and higher in the crust until they are released in large through-going earthquakes.

Figure 24 taken from Stefánsson (1999) illustrates schematically the pore pressure build-up of earthquake conditions at different parts of the zone at different times or sequences.

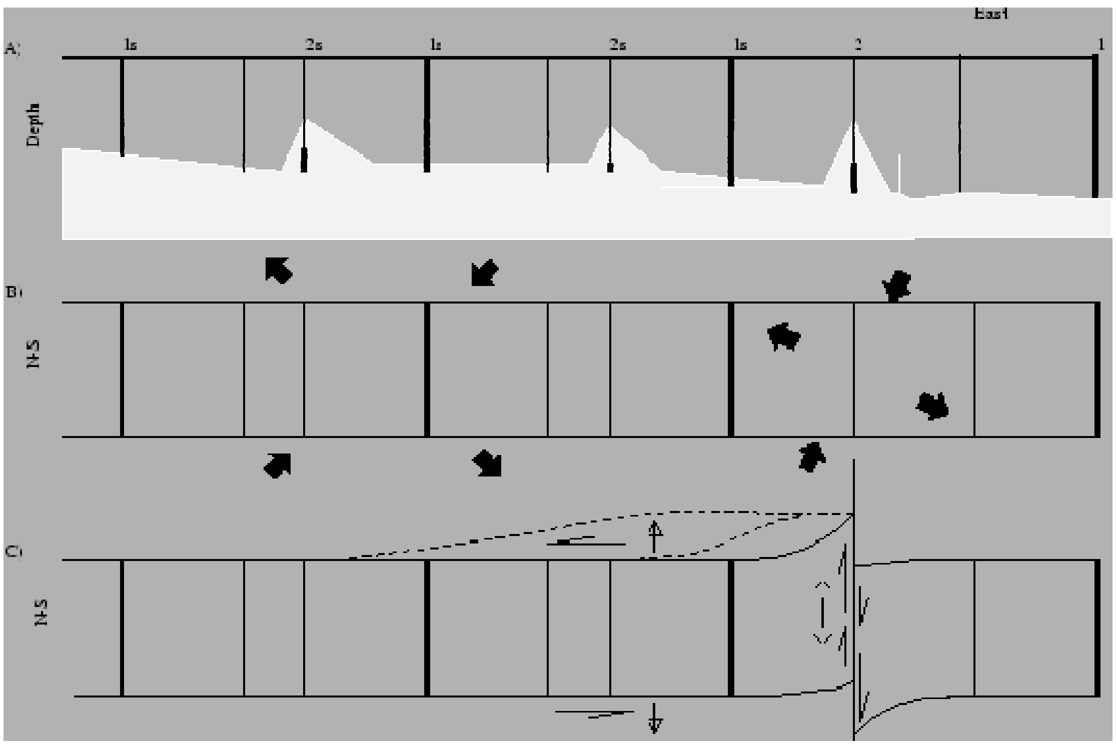

Figure 24. *A schematic picture illustrating the main features of the hypothesized model of earthquake processes in the SISZ. A) is vertical section of the earthquake zone. 1 indicates a large past earthquake more than 100 years ago, 1s are secondary earthquakes, 2 is an impending earthquake with a beginning of creation of a fault with fluid intrusion and high pore pressures near the bottom of the seismogenic zone. B) is a horizontal NS view of the same zone. The heavy arrows are the stresses and largest stress is turning northward near the approaching large earthquake source. C) indicates the movement of the NS fault in the large earthquake, right-lateral strike-slip and expansion which migrates along the zone.*

7.2 Towards modelling of individual large earthquakes

This process described here refers mostly to the June 17 earthquake, which triggered the June 21 earthquake. The heterogeneity of the crust especially the fluid mobility and its influence on increasing pore pressures gives a hope that there will be a considerable time delay between the start of the nucleation process and the onset of the real earthquake slip. Observations show that all earthquakes of magnitude 5 since the start of the SIL system have been preceded by foreshocks (Slunga 2003) support this idea.

These ideas here have been confirmed in studies of microearthquakes approaching the 2000 earthquakes (Figure 25) and tentative models with the same general ideas have already been the basis of warnings or time-dependent hazard assessments in the SISZ. They also explain the premonitory activity in the 1912 earthquake. The model described here as well as earlier experience is a basis for the procedures in Chapter 8 in making use of microearthquakes to provide time-dependent warnings or hazard assessments, as well as time-independent.

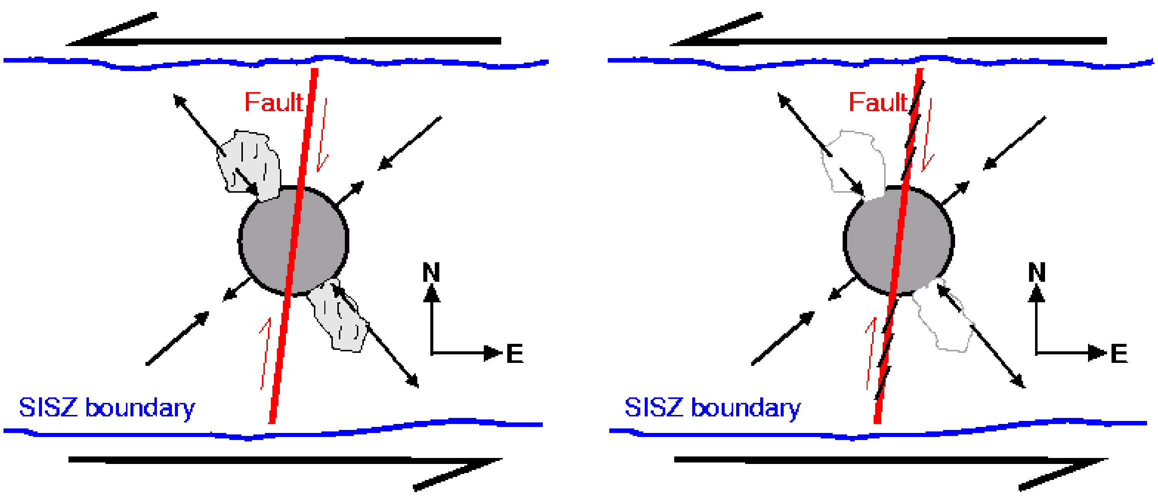

Figure 25. *Schematic picture of the conditions around the June 17 earthquake before its occurrence in the framework of the SISZ. The EW motion across the SISZ is shown by left-lateral arrows, and the boundaries of the 10 km SISZ are shown with blue lines. The June 17 fault (red) has a strike of 7ºNE. The 3 km diameter asperity is shown in dark grey. The regional horizontal stress axes are shown by arrows, and the maximum horizontal compression is here taken as 50ºNE. The local field heterogeneity caused by the left-lateral steady motion across the hard core is indicated by opposite short arrows. The areas of the premonitory swarm activity for decades, the "dilavolume" is in light grey (left figure). The line segments indicate frequent fault planes. The right figure describes the last 17 days before the earthquake. The microearthquakes now cluster near the fault plane and mostly below 6 km, with fault planes en echelon in accordance with the right-lateral motion which has started at depth (Stefánsson et al. 2006).*

The stages of the total earthquake process or the earthquake cycle in response to shearing caused by plate motion:

- Break up of faults not moving for a few hundreds of years in presence of subcrustal fluids in pores at the brittle/ductile boundary. Medium-size earthquakes only releasing parts of the damage area of the fault plane, in the early stage especially, close to the boundary of the weak SISZ zone.

- Inflow of fluids with lithostatic pressure from the bottom of the ductile crust in response to the fracturing and expanding fracturing.

- Long-term infiltering of the brittle crust by fluids from below, increased pore fluid pressures for a long time, at parts of the damage zone best open for such infiltering. The up-migrating places are often in compression shadows of hard cores/asperities.

- The opening and slow slip motion because of fluid pressure in conjunction with strain of the deepest part of the old fault, fluid transport along it leading to local stress modifications and homogenization of stress along the old NS-strike slip fault, preparing for the possibility of slip on the whole old earthquake fault.

- Release of motion shearing and pore pressure in a magnitude 6-7 earthquake.

The time spans of these stages seem to be very differently long.

8 Procedures for assessing place, size and the time of a large earthquake

8.1 To find the place of a large earthquake

The experience and new knowledge gained during the period approaching the 2000 earthquakes signifies the following methods:

- The microearthquake technology that has been developed within the SIL microearthquake monitoring technology provides a very powerful method for mapping of faults at depth that have been active for the last few hundred years and thus also to indicate probable magnitudes in future earthquakes on these.

- Paleoseismology is significant to fill in the gaps where the fault seismicity is too little for seismicity mapping.

- Mapping of surface faults from recent earthquakes is significant for this but also for predicting the expected destructive power.

- Modelling the accelerations or intensities and studying site effects at probable future earthquake sites to predict its effects.

- Historical seismology provides significant information to try to assess the place of the next earthquake.

All these methods have been applied within the PREPARED-project.

Mixing of historical seismicity knowledge with microearthquake information in fact led to an open prediction of both the large 2000 earthquakes, as the next large earthquakes in the SISZ (Stefánsson et al. 1993). In addition to this, in hindsight, several studies of the PREPARED-project pinpointed the earthquake location:

- Detailed studies of *b*-values in hindsight (*b*-value metdod) (Wyss and Stefánsson 2006) helped to localize the asperity of the next earthquake.

- The SRAM (Böðvarsson et al. 2005) in hindsight indicated the place of both the 2000 earthquakes.

8.2 To assess where the next large earthquake or the next earthquake sequence will occur, forecasting the place and size of the earthquake, long-term forecasting

Studying seismicity around the epicenters of the 2000 earthquakes for several tens of years indicates that the process leading to them could possibly have been observed 40 years before they occurred. Probable build-up process before the large 1912 earthquake can in hindsight be observed in medium-large earthquakes 8 years before.

Continuous monitoring around suspect faults will provide dynamic information which help to understand where we are in the earthquake cycle of the various faults. According to modelling results and experiences (Zencher et al. 2006; Slunga 2003; present report) closeness to large earthquakes will appear as high microseismic activity at depth in the crust, increasingly migrating to shallower depths, but also as medium-large earthquakes at shallower depths. Such activity creates the “hotspot” for a future earthquake. Both the historical activity and the observed microearthquake activity reveal that we can expect to see a gradual and stepwise corrosion of the fault plane or fault damage area until reorganization of stresses cause unstable conditions and homogenization of fault plane and make possible a slip motion on a large earthquake fault only short time before the earthquake. At present we do not see how we can estimate the time of the onset of an earthquake with any practical accuracy until possibly when the nucleation process starts.

Stress build-up can be watched by simply monitoring the frequency of earthquakes (FOE) above the Mc (magnitude of completeness), which should show a gentle increase, if we somehow allow for the disturbing effects of aftershocks of medium-large earthquakes. This will help in classifying the faults that we observe.

A more advanced method to do this is the SAG method which selects out singular events, or internally independent events. The number of these is a measure of stress changes (Lund and Slunga 1999; Lund and Böðvarsson 2002).

Shear-wave splitting (SWS) delay times observed above frequent microseismic activity at depth provides information on stress changes or closeness to fracture criticality in the medium around the ray paths (Crampin and Gao 2006).

Shallowing of microearthquake sources or maybe more frequent penetrations to shallow depths of high pore pressures may provide information about higher stresses or maybe more correct to say weakening of the source area.

Careful modelling on basis of the monitoring of the suspect fault to try to assess its state and compare it to the state of the fault in approaching the 2000 earthquakes, or other earthquake premonitory changes where we have some information like for example the 1912 earthquake and the Hengill earthquakes of 1998.

There are indications that different mechanism character of earthquakes in the region may provide indication about a proximity of an earthquake as well as anomalies in crustal velocities.

8.3 To predict the time of the next large earthquake

We do not have a method to do this in a definite and deterministic way on a long-term basis. On basis of history, of tectonics of the region and on the known energy release it is possible to provide a probabilistic time-dependent hazard assessment. This has been done in Iceland especially concerning the SISZ. According to this methodology the probability of an earthquake of magnitude 6 and larger to occur within the next 20 years was around 95% when the 2000 earthquakes occurred. Although this may sound good its reliability is limited, and it is merely useful for concentration of watching activities. Its reliability is of course most limited by the fact that it assumes that the seismic release in general behaves as it has done since 1700. However, we know from history that this activity is uneven in time.

The best method to approach the goal of seismic warning, short- or long-period, is a careful watch of the process that has started and can be observed on the suspect fault, some of the useful watching methods were described in a former chapter. The model that we have described earlier assumes that we will be able to see a preparatory process that we should expect to be observable for years and a nucleation process that takes from hours to weeks in a final stage before an earthquake. However, our understanding of the earthquake process in general is limited, and earthquakes do not repeat each other. Therefore it is significant to observe the starting of the premonitory signs of large earthquakes as early as possible and to monitor and model this ongoing process. Our hope is that during this process new understanding on the model of the impending earthquake will be revealed.

Modelling and the experience of the 2000 earthquakes show that the process starts up with a intrusion of fluids at depth, breaking up parts of the fault damage zone in an area which in extension is comparable with the fault length (of the order of 10-15 km). The fluids create general high fluid pressures in the fault area. However, the large throughgoing earthquake cannot be released until favourable conditions have been created for a slip motion which continues over a large fault plane. If we can follow and understand this nucleation process we may have the possibility of providing useful short-term warnings, i.e. of the order of hours to days in advance.

As was described in an earlier chapter the first earthquake of June 17 was preceded by the concentration of microearthquakes from being around in the damage zone towards a fault plane of an earthquake that probably occurred there more than 300 years ago. The place of this earthquake and that it would be large was forcast in 1993 (Stefánsson et al. 2003). In this report and in Stefánsson and Guðmundsson (2005) is described how the nucleation process was illuminated in microearthquakes. According to the interpretation of the microearthquake information the process started at depth, inside the ductile part of the crust with slow slip motion responding to shearing strain and to high pore fluid pressures.

By fluid mobility the stress around an old fault plane was homogenized and the whole fault plane started to move infinitesimally before the breaking of the asperity that for a long time had hampered motion of the fault. The first part was shown by declustering of microearthquakes on the fault plane for two weeks before the earthquake, and clustering in a 3 km diameter asperity for only 24 hours before the earthquake.

The large earthquake on June 21, 4 days after the first large earthquake was forecast as immediate, in real-time 24 hours before it occurred as most probable. It was done on basis of linearization of microearthquake seismicity on fault, that had been predicted as the place of the next large earthquake and the experience of short-term sequences in the zone, i.e. migration velocity of 5 km/day along the zone. The June 21 earthquake was not completely comparable with the first one, especially as it was in a stress shadow of the asperity of the first one, and does not seem to have an asperity as the first one. Rather the whole fault area had been corroded before it started with an inflow of fluids into the deep fault. This was seen by the deepening of the microearthquakes after the high stress caused by the first earthquake at its southern tip. This deepening was caused by inflow of fluids from the damage zone into the deep part of the earthquake fault. The fluid pressures in the fault reduced the normal stress across the fault as in the first earthquake to make the large earthquake release possible.

The magnitude 7 earthquake of 1912 at the eastern end of the SISZ may well fit into this modelling process, only with much shorter build-up time of observable changes.

What is said here does not indicate that we by careful watching can forecast all earthquakes. However, we have by study of the microearthquakes enhanced our methods of watching and monitoring the processes leading to earthquakes at least in Iceland.

We have here in first hand concentrated on the processes on and in the close neighbourhood of the becoming earthquake fault. The eruption of Hekla near the eastern end of the zone on February 26, 2000 (Table 1) can be considered to have had the effect in stress loading the SISZ to the west of it. It is interesting that while earlier eruptions in Hekla have caused a flurry of earthquakes in the whole SISZ during weeks and months afterwards, it was observed that it did not do so before the June 2000 earthquakes. No such high activity was in the zone until the large earthquakes occurred. We have not so far studied a possible coupling in detail. One possibility is that straining of the area as a whole was locked by the asperity core of the June 17 earthquake, so microearthquake activity did not occur until that asperity was broken. It is also probable that the large tectonovolcanic episode in the Hengill area, starting in 1994 and culminating and ending in 1998, had influence on the SISZ as a whole. However, much is left in studying the coupling and triggering effects of such events.

9 Conclusions

We have here described procedures to watch and to provide warning at various stages before large earthquakes in the SISZ. The work here is only based on monitoring of small earthquakes, down to magnitude 0. Even earthquakes down to magnitude -1 have rendered information that illuminate the nucleation process of a first earthquake. As had been stated in the first earthquake prediction research project, the SIL-project, starting 1988, monitoring earthquakes down to magnitude zero is essential for the possibility of using small earthquakes to monitor crustal processes leading to large earthquakes.

It is in fact probable that even smaller earthquakes can be essential for more secure results.

One of the significant results of this project is to observe and to model the significant role of lithostatic fluid pressures in the long-term preparatory process before the earthquakes. This implies that the detectability of small earthquakes should be greatly increased approaching the earthquake in suspect areas. This and the activity of high pressure and high temperature fluids in the process, implies that we should even go to higher frequencies by acoustic monitoring.

9.1 The next large earthquake in the SISZ, where, when and how big?

One of the questions to be answered in this project is how can we predict or warn for the next large earthquake.

Following paragraph is taken from Stefánsson et al. (2003) about the remaining moment stored in the SISZ after the 2000 earthquakes.

"As discussed earlier the moment released in the two large earthquakes of 2000 is estimated to be $1.2*10^{19}$ Nm, while the moment built-up and released during a 140 year earthquake cycle has been estimated to be $0.7\text{-}1*10^{20}$ Nm, where the higher value is based on the estimated size of historical earthquakes. Assuming that the lower value is more realistic, as the historical earthquake magnitudes may have been overestimated, and taking into account that only 100 years have apparently elapsed of the 140 year cycle (Stefánsson and Halldórsson 1988), the moment build-up before the earthquakes would have been $5*10^{19}$ Nm. This means that only a fourth of the stored moment would have been released in the two large earthquakes in 2000. The remaining moment is probably mostly stored in the easternmost part of the SISZ, where the largest earthquakes are to be expected as the elastic/brittle crust is thickest there. Judging from historical observations and the general understanding of tectonics outlined above, the build-up of strain from around 1900 to year 2000 has not been enough to produce a magnitude 7 earthquake in the easternmost part of the zone. We suggest, however, that further build-up of strain, in addition to what remains after the recent earthquakes will be enough to rupture the strong crust there within the next few decades. The above reasoning is based on a simple model of moment build-up, assuming steady plate motions with shearing deformation across a homogeneous SISZ, however, with increasing thickness and strength from west to east.

The total release of stress in such a simplified zone would have the tendency to delay until it starts at the easternmost, strongest part and trigger subsequent earthquakes further west during a relatively short time frame”.

So the prediction is that only a fourth of the stored energy in the SISZ was released in the 2000 earthquakes and most of this energy is stored near the eastern end of the SISZ and a possibility is for a magnitude 7 earthquake there during the next decades.

In the results of the PREPARED-project there is nothing which contradicts this prediction. One thing that has though not been taken into account in this long-term forcasting is the significance of frequent Hekla eruptions and a magnitude 5.8 earthquake at the eastern end of the SISZ. How much stored energy in the easternmost part of the zone was released in these volcanotectonic episodes? This has not been modelled so far. However, it is probable that we can speculate that if the Hekla eruptions (Table 1) with the continuation to south in the Vatnafjöll area would release stresses as a magnitude 6.5-7 earthquake, say in a 15 km diameter area and the same for the 17 June earthquake, we may have a 15 km diameter area left for an earthquake of similar fault length size as the June 17 earthquake near longitude 20.1°W, an earthquake having 20% higher moment relative to the thicker crust, i.e. around 6.7 (Ms).

It was discussed earlier in this paper that it is possible that the next earthquakes of magnitude 7 before the 1912 earthquake were the 1630 and 1657 ones. The pillar diagram shown in Figure 2 also does not lend much space for earthquakes at the eastern part.

Although there are not strong arguments for a magnitude 7 or larger earthquake near the eastern end of the SISZ during the next decades, we should prepare for such an earthquake in the area. By careful watching and research work it is possible that we can see such an earthquake long time before it occurs, for example from medium-size earthquakes, increasing pore fluid pressures high up in the crust and by microearthquakes. We have not seen this yet. The earthquakes that started 8 years prior to the 1912 earthquake give a hope for that we will see an recognizable activity period before such an eventual earthquake.

As to the question of earthquakes near the west end of the SISZ there is in many ways a similar situation. The large volcanotectonic event that culminated in 1998 may have released some stresses in its neigbourhood, i.e. within 10-20 km away from it. However, magnitude 6-6.5 earthquakes should be prepared for in this area, with monitoring and scientific watching.

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References

Ágústsson, K. & Ó.G. Flóvenz 2005. The thickness of the seismogenic crust in Iceland and its implications for geothermal systems. In: Proceedings of the World Geothermal Congress, Antalya, Turkey, April 24-29, 2005. Paper no. 0743.

Árnadóttir, Þ., S. Jónsson, R. Pedersen & G.B. Guðmundsson 2003. Coulomb stress changes in the South Iceland Seismic Zone due to two large earthquakes in June 2000. *Geophys. Res. Lett.* 30(5), 9(1-4).

Benioff, H. 1951. Earthquakes and rock creep, part 1. *Bull. Seismol. Soc. Am.* 41, 184-190.

Bjarnason I.Þ. & P. Einarsson 1991. Source mechanism of the 1987 Vatnafjöll earthquake in South Iceland. *J. Geophys. Res.* 96, 4313-4324.

Björnsson, A. 2006. Temperature of the Icelandic crust: Inferred from electrical conductivity, temperature surface gradient and maximum depth of earthquakes. *Tectonophysics*, accepted.

Böðvarsson R., B. Lund & A. Tryggvason 2005. Seismicity ratio with applied mechanism. In: Abstracts of the EGU General Assembly, Vienna, Austria, April 24-29, 2005.

Crampin, S. & Y. Gao 2006. A review of techniques for measuring shear-wave splitting above small earthquakes. *Geophys. J. Int*., submitted.

Einarsson, P. 1991. Earthquakes and present-day tectonism in Iceland. *Tectonophysics* 189, 261-279.

Einarsson, P., M. Khodayar, A. Clifton, B. Ófeigsson, S. Þorbjarnardóttir, B. Einarsson & Á.R. Hjartardóttir 2005. A map of Holocene fault structures in the South Iceland Seismic Zone. In: Abstracts of the EGU General Assembly, Vienna, Austria, April 24-29, 2005.

Flóvenz, Ó.G. & K. Sæmundsson 1993. Heat flow and geothermal processes in Iceland. *Tectonophysics* 225, 123-138.

Guðmundsson, A.T. 2001. *Íslenskar eldstöðvar*. Vaka-Helgafell, Reykjavík.

Halldórsson, P. 1987. Seismicity and seismic hazard in Iceland. In: D. Mayer-Rosa, J.M. van Gils & H. Stiller (editors), Activity Report 1984-1986 and Proceedings of the XX ESC General Assembly, Kiel, Germany, 1986. *Publication Series of the Swiss Seismological Service* 101. European Seismological Commission, 104-115.

Hjaltadóttir, S. & K.S. Vogfjörð 2005. Subsurface fault mapping in Southwest Iceland by relative location of aftershocks of the June 2000 earthquakes. *Rit Veðurstofu Íslands* 21. Icelandic Meteorological Office. Research Report, 18 pp.

Kárník, V. 1969. *Seismicity of the European Area, Part 1*. D. Reidel Publishing Company, Dortrecht, Holland, 364 pp.

Lund, B. & R. Slunga 1999. Stress tensor inversion using detailed microearthquake information and stability constraints: application to Ölfus in southwest Iceland. *J. Geophys. Res.* 104, 14947-14964.

Lund, B. & R. Böðvarsson 2002. Correlation of microearthquake body-wave spectral amplitudes. *Bull. Seism. Soc. Am.* 92, 2410-2433.

Pálmason, G. 1973. Kinematics and heat flow in a volcanic rift zone, with application to Iceland. *Geophys. J. R. Astrom. Soc*. 33, 451-481.

Rivalta, E. & M. Bonafede 2005. Analytical fault and fluid-filled fracture models in viscoelastic media. In: Abstracts of the EGU General Assembly, Vienna, Austria, April 24-29, 2005.

Roth, F. 2004. Stress changes modelled for the sequence of strong earthquakes in the South Iceland seismic zone since 1706. *Pageoph* 161(7), 1305-1327.

Slunga R. 2003. Microearthquake analysis at local seismic networks in Iceland and Sweden and earthquake precursors. In: *Lecture Notes in Earth Sciences* 98. Methods and Applications of Signal Processing in Seismic Network Operations. Springer Verlag, Germany.

Stefánsson R. 1999. A tentative model for the stress build-up and stress release in and around the SISZ. http://hraun.vedur.is/ja/prepared/reports/.

Stefánsson, R. & P. Halldórsson 1988. Strain release and strain build-up in the South Iceland seismic zone. *Tectonophysics* 159, 267-276.

Stefánsson, R., R. Böðvarsson, R. Slunga, P. Einarsson, S.S. Jakobsdóttir, H. Bungum, S. Gregersen, J. Havskov, J. Hjelme & H. Korhonen 1993. Earthquake prediction research in the South Iceland seismic zone and the SIL project. *Bull. Seism. Soc. Am*. 83(3), 696-716.

Stefánsson, R., F. Bergerat, M. Bonafede, R. Böðvarsson, S. Crampin, P. Einarsson, K.L. Feigl, C. Goltz, Á. Guðmundsson, F. Roth, R. Sigbjörnsson, F. Sigmundsson, P. Suhadolc & M. Wyss 2002. Application of practical experience gained from two recent large earthquakes in the South Iceland seismic zone in the context of earthquake prediction research to develop technology for improving preparedness and mitigating risk – PREPARED. An EC proposal. *Veðurstofa Íslands – Greinargerð* 02004. Icelandic Meteorological Office, Report. 119 pp.

Stefánsson, R., G.B. Guðmundsson & Páll Halldórsson 2003. The South Iceland earthquakes 2000 - a challenge for earthquake prediction research. *Veðurstofa Íslands – Greinargerð* 03017. Icelandic Meteorological Office, Report, 21 pp.

Stefánsson, R. & G.B. Guðmundsson 2005. About the state-of-the-art in providing earthquake warnings in Iceland. *Veðurstofa Íslands – Greinargerð* 05003. Icelandic Meteorological Office, Report, 26 pp.

Stefánsson, R., M. Bonafede, F. Roth, P. Einarsson, Þ. Árnadóttir & G.B. Guðmundsson 2006. Modelling and parameterizing the Southwest Iceland earthquake release and deformation process. *Veðurstofa Íslands – Greinargerð* 06005. Icelandic Meteorological Office, Report, 49 pp.

Wyss, M. & R. Stefánsson 2006. Nucleation points of recent main shocks in southern Iceland mapped by *b*-values. *Bull. Seism. Soc. Am.*, accepted.

Zencher, F., M. Bonafede & R. Stefánsson 2006. Near-lithostatic pore pressure at seismogenic depths: a thermo-poroelastic model. *Geophys. J. Int.,* submitted.

Þórarinsson, S. 1968. *Heklueldar.* Sögufélagið. Reykjavík.