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Earthquake activity in Sweden

Study in connection with a proposed nuclear waste repository in Forsmark or Oskarshamn

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This report concerns a study which was conducted for SKB. The conclusions and viewpoints presented in the report are those of the authors and do not necessarily coincide with those of the client.

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Summary

The aim of this report is to evaluate the risks for future earthquakes in the vicinity of the proposed nuclear waste repository sites at Forsmark and Oskarshamn. Time periods of 100 and 1,000 years will be considered, which implies that the focus of this study is on an evaluation of the current, general situation in the region. Major events on a longer time scale, such as an ice-age, will only be briefly considered.

Earthquakes are products of ongoing deformations within the Earth and this report will, therefore, concentrate on the current state of knowledge about deformations in the region. As earthquakes are our most important source of information about deformations at depth in the crust, we will focus on the available seismic data using the Nordic earthquake catalog maintained at the Institute of Seismology, Helsinki University, and the recent data from the new Swedish National Seismic Network. Direct measurements of surface deformation using the Global Positioning System will also be utilized in the analysis.

Sweden is a low seismicity area, with most earthquakes being observed in the south-west, around Lake Vänern, along the north-east coast and in Norrbotten. South-eastern Sweden is on the contrary relatively inactive. Seismicity is also, generally, episodic in time which together with the short period of instrumental observation, approximately 100 years, makes our knowledge about the activity far from complete. Although very large earthquakes (magnitude about 8) have occurred in Sweden, it is generally agreed that these were connected to the late stages of deglaciation at the end of the previous ice-age. At the time scales considered in this report, inferences from current seismicity is of more relevance. This data suggests that we should expect at least one magnitude 5 earthquake in our region every century and one magnitude 6 earthquake every one thousand years.

In order to illustrate the effects of static and dynamic deformation from a magnitude 5 earthquake, we use the Kaliningrad magnitude 5.0 event of September 2004 as a modeling example. The event occurred at 20 km depth but we vary the depth in order to see how the effects vary. At a reasonable depth of 12 km, as the 1976 Gulf of Finland earthquake, the static displacements at the Earth's surface do not exceed 0.2 mm and we would expect 0.05 g of acceleration, on crystalline bedrock.

Earthquake focal mechanisms reflect the state of stress which caused the earthquake. From the focal mechanisms we can therefore infer the stress state. Analysis of Swedish earthquakes show that in central and southern Sweden the crust below a few kilometers depth is in a state of strike-slip faulting, with the maximum horizontal stress directed WNW-ESE. This is confirmed by measurements in the two deep boreholes in Siljan. The stress state reflects the plate tectonic deformation caused by the opening of the Atlantic, which dominates the stress field in Sweden. Stresses due to postglacial rebound are much less significant today compared to the tectonic stresses.

Direct observations of large-scale surface deformation has been carried out in Sweden since 1993 in the BIFROST project, using permanent, continuous GPS-receivers. The project has produced high quality estimates of deformation rates which correlate very well with those obtained from glacial rebound modeling. In the residuals between model and observations there are indications of relative displacement between the stations. If these are fault movements, they indicate, although the relative displacements are less than 1 mm/year, deformation that is orders of magnitude larger than that observed in the seismic data, i.e.

aseismic deformation. Earthquake data from the 1980s have earlier been interpreted as indicating aseismic movement of the order 1 mm/year/100 km in southern Sweden. In order to determine whether or not aseismic deformation is present at this scale, a dense network of permanent GPS-stations would be required, preferably in an area of high seismicity. We recommend that such GPS-measurements are initiated, also in the regions considered for the nuclear waste repositories.

Sammanfattning

Syftet med den föreliggande studien är att utvärdera risken för framtida jordbävningar i de områden i närheten av Forsmark och Oskarshamn som nu studeras inför ett slutförvar för utbränt kärnbränsle. Studien omfattar två tidsperioder, 100 år och 1 000 år, vilket innebär att tyngdpunkten ligger på att utvärdera dagens situation i regionen. Storskaliga geologiska eller klimatologiska processer som försiggår under längre tider, som t ex en istid, kommer att endast kort beröras.

Jordskalv alstras av pågående deformationer i jordskorpan och denna rapport kommer därför att lägga stor vikt vid det aktuella kunskapsläget vad det gäller deformationer i Skandinavien. Eftersom jordskalv är vår huvudsakliga källa till information om deformationsprocessen djupare ner i jordskorpan kommer vi att fokusera på det tillgängliga jordskalvsdatat. Vi använder två jordskalvskataloger; den nordiska katalogen som förs vid det seismologiska institutet vid Helsingfors universitet och som omfattar jordskalv i Skandinavien från 1375 fram till idag, och det nya datat från det moderniserade och utbyggda svenska nationella seismiska nätet (SNSN). I analysen kommer vi också att lägga stor vikt vid det GPS-data (Global Positioning System) som analyserats inom BIFROST-projektet.

Sverige betraktas som ett område med låg seismisk aktivitet. Flest jordskalv har vi i sydväst, runt Vänern, längs Norrlandskusten och i Norrbotten. Sydöstra Sverige däremot, där Forsmark och Oskarshamn ligger, har relativt låg aktivitet. Seismiciteten är i allmänhet episodisk, vilket tillsammans med vår korta observationstid (ca 100 år med instrumentella observationer) gör att vår kunskap om seismiciteten är långt ifrån fullständig. Mycket stora jordbävningar, med magnitud omkring 8, har förekommit i Sverige. Man är dock överens om att dessa orsakades av isavsmältningen vid slutet av den senaste istiden, så på de tidsskalor som den här studien omfattar är det den nutida seismiciteten som är relevant. Det senaste århundradets data säger oss att vi kan förvänta oss minst ett magnitud 5 skalv per hundra år i regionen och därmed minst ett magnitud 6 skalv vart tusende år.

För att illustrera de statiska och dynamiska deformationerna från ett magnitud 5 skalv använder vi oss av Kaliningradskalvet, ett magnitud 5,0 skalv från september 2004, som vårt modellexempel. Skalvet skedde på ca 20 km djup men vi varierar djupet för att se hur det påverkar ytdeformationerna. På 12 km djup, samma djup som skalvet i Finska viken 1976, överskrider de statiska förskjutningarna på jordytan inte 0,2 mm och vi förväntar oss 0,05 g i acceleration på kristallint urberg.

Jordskalvens fokalmekanismer återspeglar de spänningar som orsakade skalven. Man kan därmed uppskatta spänningstillståndet från mekanismerna. Analysen av fokalmekanismer i Sverige visar att i jordskorpan, nedanför ett par kilometers djup, i mellersta och södra Sverige dominerar horisontella förskjutningar på vertikala förkastningar, dvs de största och minsta principalspänningarna är horisontella medan den mellersta är vertikal, ett s k ”strike-slip”-tillstånd. Den maximala horisontella spänningen pekar i riktningen VNV-ÖSÖ. Mätningar av spänningstillståndet i de två djupa borrhålen i Siljan bekräftar jordskalvanalysen. Detta spänningstillstånd stämmer väl överens med det man förväntar sig ifrån en plattektonisk analys, där Atlantens öppnande ger en VNV huvudspänningsriktning som dominerar spänningsfältet i Sverige idag. Spänningar orsakade av istiden är betydligt mindre än de plattektoniska idag.

Direkta observationer av storskaliga deformationer på jordytan har gjorts med kontinuerlig GPS-teknik i Sverige sedan 1993 inom det s k BIFROST projektet. BIFROST har redovisat mycket noggranna uppskattningar av deformationshastigheterna i Skandinavien som överensstämmer väl med modeller av landhöjningen. I de delar av observationerna som inte stämmer överens med landhöjningsmodellen finns indikationer på relativa rörelser mellan stationerna. Om dessa relativa rörelser utgörs av förkastningsrörelser skulle de, trots att de är mindre än 1 mm/år, innebära deformationer som inte motsvaras av seismisk aktivitet i samma storleksordning, dvs aseismiska rörelser. Jordskalvsdata från 1980-talet har tidigare tolkats som att aseismiska rörelser i storleksordningen 1 mm/år/100 km förekommer i södra Sverige. För att avgöra om aseismiska rörelser förekommer, och i så fall hur stora de är, krävs tätare GPS-mätningar med permanenta stationer under längre tid. Vi rekommenderar att sådana mätningar görs, helst i ett område med hög seismisk aktivitet men också i de regioner där slutförvar för utbränt kärnbränsle planeras.

Contents

1 Background

Earthquakes are not evenly distributed in the earth but are generally largely concentrated to limited seismogenic zones, which commonly correspond to the boundaries between the tectonic plates making up the earth's outer cold and rigid shell (the lithosphere). These plates are in constant slow motion relative to each other, and earthquakes occur as a consequence of these motions. Sweden does not lie close to a plate boundary, and activity here is low. The tectonic plates are relatively strong and rigid, and therefore stresses are maintained for great distances and tectonic deformation occurs even far from the plate boundaries. These processes generate earthquakes, and sometimes large earthquakes, even within plates. There are also a number of other phenomena which are significant in generating earthquakes, e.g. activity associated with mantle plumes, where warm material from the earth's mantle forces its way up towards the surface, or areas which have recently been loaded or unloaded by water or ice. Mining activity and dam construction can trigger earthquakes. The ongoing postglacial rebound in Sweden, where the earth's surface is still rising after the disappearance of the thick ice cover of the last ice-age, leads both to vertical and horizontal deformation. This process can lead to earthquakes, but the picture is in general dominated by the more normal plate tectonic forces.

The mechanical stability of the area is an important aspect in the process of planning a nuclear waste repository. It determines both where, in the regional as well as in the local geological context, the repository should be located and what characteristics the design of the repository must meet. This report will focus on mechanical stability in the context of deformations due to ongoing natural processes. Deformations within the earth's upper crust are mainly of two kinds, namely purely elastic deformations and deformations due to shear slip on fractures. Slow, stable slip and sudden, seismic slip both induce larger strains in an area than those caused by regional elastic deformation. Slip at seismic velocities, e.g. earthquakes, also release significant amounts of vibrational energy, which may lead to damage to constructions at considerable distances from the source. It is therefore important that an envisaged repository is so placed that it avoids predicted zones of significant local deformation.

Numerous studies of deformation and seismic risk, both general and specifically related to nuclear repositories, have been published. However, the issues are complex and not all aspects are completely understood. Analysis of seismic signals from earthquakes provides our most important source of information on deformation processes at depth within the earth, and such earthquakes are comparatively rare in low seismicity areas such as Sweden. Therefore the database for analysis is generally more limited here than in other more seismically active areas, and especially sophisticated and sensitive registration equipment is necessary in order to obtain a comprehensive data set for analysis. As seismic activity is also very unevenly distributed in space and is in any given area generally episodic in time, it is vital to have large enough areal coverage, even if it is the evolution of deformations in a small particular area that is of interest. Because of this, the Swedish Nuclear Fuel and Waste Managment Co (SKB) has supported the construction and operation of a modern, state-of-the-art, seismological network in Sweden /Bödvarsson et al. 1996, Bödvarsson 1999, Bödvarsson et al. 1999, Bödvarsson and Lund, 2003/.

This current report builds upon earlier studies of seismic risk in Sweden, complemented with the large amount of additional data which is now available through the new Swedish National Seismic Network /Bödvarsson, 2001–2006/, direct information about surface deformation from GPS (Global Positioning System) data, and new scientific insights into the relevant deformation processes.

The aim of this SKB financed project carried out by Uppsala University in collaboration with the Swedish Defence Research Agency (FOI) is, via the analysis of available data, to evaluate the risks for future seismic events in the vicinity of the proposed nuclear waste repository sites at Forsmark and Oskarshamn. Time periods of 100 and 1,000 years will be considered.

The project was mainly carried out in December 2005, with this final report being produced in February 2006. The defined time scale of interest means that the focus in this study is on an evolution of the current general situation in the area, and will not consider in detail e.g. the possibility of an ice-age.

2 Introduction

2.1 Earthquakes

An earthquake is a sudden movement on a fault (fracture plane). This sudden movement is normally the result of stresses which have built up over many years, due to ongoing deformation processes in the area. The part of a fault which moves in a given earthquake is limited in depth and width. How “large” an earthquake is depends on a number of parameters including the area of the fault which has moved, the size of the relative movement between the two sides of the fault (slip), and the elastic properties of the surrounding rocks. Earthquake size is commonly described in terms of its “magnitude”. Magnitude scales, including the well known Richter scale, are described below. Surface movement related to an earthquake is partly temporary (vibrations in the form of seismic waves) and party permanent (static deformation). Seismometers measure the temporary deformation directly, and indirectly provide information about static surface deformation. Other methods measure this directly (see below).

There is a limit to how large earthquakes can be because the rigid part of the earth has finite strength. The largest earthquakes observed on earth are of (Richter) magnitude about 9.5. The slip associated with such an earthquake may be some tens of meters over a fault area which is typically some hundreds of kilometers long and some tens of kilometers wide (deep). There is no simple relationship between fault area and slip, but in general larger fault area is associated with larger slip. Many decades of observation confirm that there is a relatively well defined relationship between the number of large and small earthquakes. Generally, the number of events observed decreases by a factor of ten for every unit increase in magnitude.

Sweden lies on a very old and stable part of the earth’s crust, and is a low-seismicity area. Earthquakes are observed every day in and around Sweden, but these are normally very small and thus not at all dangerous. Our current understanding of seismology and crustal deformation strongly suggests that a magnitude 9 earthquake in Sweden is extremely improbable on the time scale of relevance here (1,000 years). It is therefore of more relevance to discuss events of considerably smaller magnitude. While larger events are rather rare, existing data suggests that one earthquake of magnitude 5 or greater and within about 500 km of Oskarshamn or Forsmark can be expected about once per 100 years. A magnitude 5 earthquake in Sweden would normally be expected to correspond to a fault slip of the order of 50 cm over a fault area with a radius of about 500 m and at a depth of about 20 km, implying no fault fault slip at the surface. More shallow events can also occur. For example, the magnitude 3.1 Norrtälje event of 1979 had a hypocentral depth of only 800 m.

2.2 Seismic signals and seismological measurements

Large-scale bedrock movements are often seismic i.e. they produce seismic signals (ground vibrations) instrumentally detectable at the surface or in boreholes. However, even movements where a considerable proportion of the stored elastic energy is released in the form of seismic signals may not be instrumentally measurable, if the recording instrument (seismometer) is too far from the event, or if the signal is less than the background ground noise. The instruments used are extremely sensitive, measuring ground motion down to one nanometer per seconds (10^{-9} m/s). Ground movement may also be “aseismic” in that the movement is in the form of e.g. gradual deformation, not providing a fast fault slip of the type necessary for the generation of seismic waves. Seismic movements are detected using seismometers or some other similar instrument. The seismometers are generally distributed over the geographical region where the earthquakes occur in order to get observations of each earthquake from several different directions which allows accurate locations (epicenter and depth) and source mechanism for the event. The seismic waves can be of compressional or shear types which during the propagation may change type. Seismic signals are attenuated during propagation through the earth. The reduction of amplitude in the vibrations as they propagate away from the source is partly due to geometrical spreading i.e. that the energy is spread over a larger and larger area, and partly due to attenuation. This attenuation is partly due to the scattering by inhomogeneities within the earth. The seismic wave also loose energy by anelastic attenuation i.e. by internal friction. Ultimately, all the seismic energy is converted to heat distributed over the volume reached by the seismic waves. The amplitude of the signal depends upon the distance from the source and also upon the frequency of the signal. Higher frequencies attenuate more quickly. In studies of crustal deformation, where the events observed are often at many kilometers depth and many kilometers from the seismometers, normally no useful data at frequencies over a few tens of Hertz is recorded. The spacing between stations in Sweden is, for the relevant area, of the order of 100 km (about 50 km in the very vicinity of the proposed repositories). This allows detection of essentially all earthquakes of magnitude 0 and above. A magnitude 0 event has a movement of about 0.01 mm on a fault with a diameter of 100 m.

Aseismic motion can be measured using direct observations of the relative movements of geographically distributed points on the earth using e.g. Interferometric Synthetic Aperture Radar (InSAR), Global Positioning System (GPS), or more conventional surveying methods using e.g. lasers. In addition to producing seismic waves, larger earthquakes can also be associated with instantaneous surface deformations which can be analyzed using these techniques. However, the surface deformation caused by individual smaller earthquakes is so small that existing geodetic methods cannot normally observe the phenomena.

Prior to the advent of modern instrumentation, earthquakes were detected only by the direct observation of effects at the surface; e.g. ground movements felt by people and damage to buildings etc caused by the vibrations. The size of the effects observed and reported in e.g. newspapers of the time can be used to estimate the location, depth and magnitude of the event. Studies of such phenomena are called "macroseismic studies" and are of importance even today; not for estimating earthquake magnitude and depth, but rather to assess e.g. the damage caused to buildings of different types by bedrock vibrations of a particular amplitude and character. This information is also used to link the pre-instrumental observations to the instrumental ones. Prior to recorded history, the only specific information available about earthquake activity is provided by geological studies of surface faults, where it is in some cases possible to quantify what fault movement which has occurred, and when. Obviously, such "paleoseismological" observations, and even macroseismic studies, are far less reliable than modern instrumental studies, but they can be important because of the very limited time period for which we have instrumental data.

2.3 Magnitude

Magnitude is a measure of the size of an earthquake, and is related to the amplitude of observed ground vibrations at distance from the source due to the generated seismic waves. Magnitude scales have no upper or lower limits. The largest observed events have magnitude of about 9.5. The scale is logarithmic and is not directly proportional to energy released. A relatively shallow magnitude 6 earthquake generates elastic (seismic) waves containing about 32 times more energy than those from a magnitude 5 event, and about 1,000 (32×32) times that from a magnitude 4 event.

Many factors are considered in defining an earthquake's magnitude; focal depth, distance between the event and the recording stations, observed frequency content, and radiation pattern of the source. With a magnitude scale, we attempt to summarize the "size" of each event as a single number. In reality, earthquakes are complex phenomena and therefore we can choose to define "size" in slightly different ways. Different magnitude scales are used depending on the context of interest, the size and distance of the event, the frequency content of the data and the type of instruments registering the event. The most common magnitude scales are:

- Surface wave magnitude (M_S). This is used e.g. for events which have occurred at angular distance of between 20° and 160° from the recording station, and where the event is not more than 50 km deep. (on the spherical earth, one degree of arc corresponds to about 110 km). Surface waves (evanescent or boundary waves existing only close to the earth's surface) with periods between about 18 and 22 seconds are used for the calculation.
- Body wave magnitude (m_b). This is based on the amplitudes of body waves, i.e. waves which have traveled through the volume of the earth, at periods of between 0.1 and 3 seconds and for events at distances of more than about 5°.
- Moment magnitude (M_W) is calculated from the estimated integrated movement on the fault, the so called scalar seismic moment M_0, i.e. the size of the fault area which has moved multiplied by the fault slip (and further multiplied by the effective shear modulus of the medium). This is estimated by modeling of the observed seismic signals. The moment magnitude was designed to approximately agree with M_S but without saturation at large magnitudes and is defined as $M_W = 2/3 \log_{10} M_0 - 6.0$ for M_0 measured in Newton-meter. Note that the energy released from a given fault movement depends not only upon the moment, but also upon the difficulty of this movement which generally depends upon the normal stress across the fault.
- Local magnitude (M_L) is essentially the original Richter magnitude and is used for describing the size of "local" events i.e. relatively small events within about 600 km of the recording station and at depths less than about 70 km. Originally based on amplitude readings from a specific instrument in southern California, M_L is usually calibrated back to the original formulation. This is in spite of the fact that the swedish local magnitude is derived from the scalar seismic moment, using the relation $M_L = \log_{10} M_0 - 10.0$.

There are also a number of other scales used in earthquake contexts, for instance different intensity scales, one of which is called the "Modified Mercalli Scale". These do not describe the size of the earthquake per se but only the effect of the ground motion at a given point. The "intensity" of the ground motion is described by its macroseismic effect as damage to buildings etc. A small but close earthquake can give a much higher "intensity" of ground motion than a distant large earthquake.

As earthquakes are complex, the parallel use of different magnitude scales is often considered sensible, despite the obvious potential problem of (in)compatibility of the scales. From a scientific point of view, magnitudes based on the scalar seismic moment are the most attractive scales, because it explicitly describes a generalized physical but explicit characteristic of the fault movement. Or in other words, at least two parameters are needed in the description of the earthquake source to be able to compute all types of magnitudes for an earthquake, for instance the scalar seismic moment together with the static stress drop.

3 Regional stress fields

The "focal mechanism" of an earthquake is a summary of the movement on the fault plane associated with the event, and provides information about the fault area which has moved and the fault slip, as well as the orientation of the fracture and the shear slip in three dimensions. It is possible to deduce this information from recordings from a seismological network, because the vibrational energy being radiated from the slipping fault has a distinct radiation pattern i.e. different amounts of energy are emitted in different directions. Preexisting fractures or planes of weakness in the rock mass determines the orientation of the rupture plane but the orientation of movement is predominantly controlled by the stress field in the area. An elastic material, such as rock, has resistance both to compression and to shear deformation. The applied normal (perpendicular) and shear stresses in different spatial directions may be different, and in general this will be the case within the crust. There always exists one frame of reference (Cartesian coordinate system) within which the stress situation in a rock element can be described only by three perpendicular normal stresses, known as the major, minor and intermediate principal stresses.

Focal mechanisms for roughly 2,000 events in Fennoscandia and around the Baltic are now available. The size of these events varies from about M_L 0 to 5 (see above regarding magnitude scales). These events are generally at between 7 and 35 km depth in the crust. The observed mechanisms are dominated by horizontal movement on near-vertical fractures. Such "strike-slip" movement means that both the major and minor principal stresses are approximately horizontal. The major principal stress is oriented WNW-ESE and is in agreement with earlier results from microearthquake analysis in Sweden (N60° W ± 20°, /Slunga, 1991/). Measurements of the stress in the two deep boreholes in the Siljan area /Lund and Zoback, 1999/ show similar results: N72° W ± 7° and N53° W ± 9° in the Gravberg and Stenberg boreholes respectively. These directions are in accordance with that expected from ongoing plate tectonic movements around Europe, and are approximately perpendicular to the Scandinavian mountain chain.

An earthquake, or fault movement, means that the shear stresses on the activated fault have exceeded that which friction and any intrinsic material strength can stand, i.e. the Mohr-Coulomb failure envelope has been exceeded. Analysis of the data from the Siljan boreholes /Lund and Zoback, 1999/ shows that at 5 km depth the maximum and minimum principal stresses are horizontal, i.e. a strike-slip state of stress, and that the maximum shear stress is approximately 30 MPa. A Mohr-Coulomb analysis of the stresses also shows that the shear stress is approximately as high as the rock is expected to withstand, i.e. the shear stresses are maintained at failure equilibrium. The stress drop, i.e. decrease in shear stress, observed for larger earthquakes in the region, e.g. Skövde 1986 and Kaliningrad 2004, are around 50 MPa for events at depths 15–30 km. These stress drops are higher than the average of 12 MPa often quoted for intraplate earthquakes /Scholz, 2002/, however, high stress drops are not uncommon /e.g. Abercrombie 1995, Bouchon, 1997/. /Atkinson and Beresnev, 1997/ discuss some of the controversy regarding the stress-drop parameter.

The stress state discussed above pertain to crustal stress at seismogenic depths. It is well known from borehole and overcoring measurements that the state of stress close to the Earth's surface can vary considerably in orientation and magnitude. Much evidence points to a reverse state of stress in the uppermost crust in large parts of Sweden /e.g. Stephansson et al. 1991/.

It is well established that significant movement (up to about 1 cm/year vertically) is ongoing in Scandinavia due to postglacial rebound after the last ice-age. Very tectonically active areas of the earth e.g. subduction zones, can show (horizontal) movements of about 10 cm per year. Thus a movement of 1 cm/year can be considered to be very significant. It might be deduced from these large movements that the stress situation in the area is dominated by the deglaciation effect, but this is not the case. Due to the lack of external resistance to upwards movement of the surface of the earth in our context a vertical movement is much less significant than horizontal relative motion. In fact, the stress regime is dominated by the plate tectonic regime, which explains the homogeneous stress picture which emerges from the analysis of focal mechanisms over a large geographical area.

4 Earthquakes in Sweden

4.1 The geographical and temporal distribution of earthquakes

Earthquakes which affect constructions are comparatively rare in Sweden. However, smaller earthquakes occur continuously and on a longer time scale it is not improbable that a larger earthquake, of a size which may lead to significant risk for constructions, may occur. The most recent such event was in 1904 when an earthquake of magnitude M_S 5.4 occurred near the Koster islands. It follows that despite the relatively low seismic activity in and around Sweden, it is motivated to investigate the possibilities of earthquake damage to constructions. On a geological time scale, earthquakes have been observed instrumentally only for a very short period of time (since 1904 in Sweden). Especially in low-seismicity areas it takes a considerable time before the seismological data set provides sufficient coverage for the underlying patterns to be thoroughly observed. In considering seismic risk, it is not, however, feasible to wait for the considerable time (perhaps several hundred years) for a near complete data set to be obtained before attempting interpretation. Thus any analysis of deformation and seismicity in Sweden requires sensible extrapolation from the existing data set. By basing such extrapolation on additional data, such as geodetic measurements, and on solid physical principles, the risk picture can be analyzed and assessed with a reasonable degree of reliability.

Seismic activity in Sweden is far from evenly distributed, see e.g. Figure 4-1. Areas of relatively high activity include the north-east coast and the region west of Lake Vänern. In the north of the country a number of neotectonic faults (i.e. faults which have moved since the last ice-age) show significant seismic activity. In contrast, the south east of the country is relatively inactive.

Because of the relatively low activity level in Sweden and the limited time period for which reliable data is available, the picture of seismic activity is far from complete. With Uppsala University's new seismological network, the Swedish National Seismic Network (SNSN), an average of almost two new events per day are recorded and analyzed. With improved coverage in the south-west of the country (where station spacing is today relatively large), the number of detected events will increase further. Despite the limited data coverage, it is already quite clear that the activity is very episodic in character; behavior which is in complete agreement with observations from other parts of the world where the activity level is much higher. Especially in such a low activity area as Sweden, this episodicity means that a very low historical activity rate in an area can not necessarily be interpreted to imply that the area will continue to show a low rate of activity.

Seismological research has traditionally been primarily directed towards analyzing the larger observed events. There are a number of reasons for this. One is that manual collation and analysis of data is time consuming, and with manual analysis only a limited number of events can be treated. Another reason is that it is the largest events which are destructive. A third is that despite the very large number of small events relative to larger ones, the aggregated total deformation from the smaller events is small in relation to that from the few very large events. This means that the few larger events dominate even in terms of the ongoing tectonic deformation processes. For example, a magnitude 2 event represents a movement of perhaps 1 mm of a fault area of 100×100 m^2 (r=56 m), and a magnitude 6 gives 1 m movement on a 10,000×10,000 m^2 (r=5.6 km) fault. This traditional picture has, however, been dramatically altered by developments over the last decade or so. Today, it is

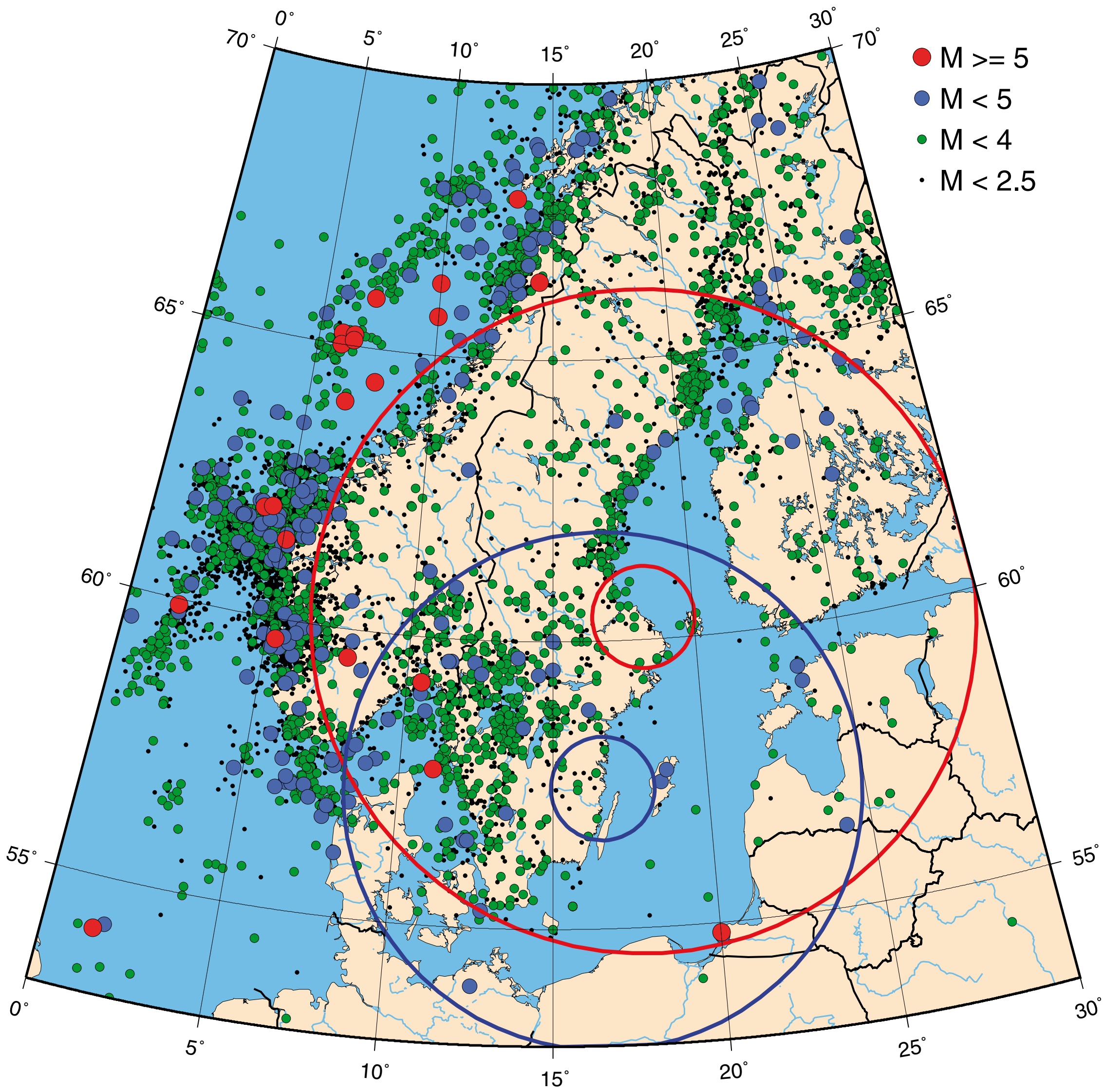

__Figure 4-1.__ Known earthquakes in the Nordic region from 1375 to 2005. The large red circle has a 650 km radius from Forsmark and the large blue circle has a 500 km radius from Simpevarp. Small circles have radius 100 km.

possible to acquire and analyze data with a high degree of automation, allowing the routine analysis even of small events. It has also been appreciated that while the small events only represent a minor part of the ongoing deformation, each event carries important information about the stress regime in the crust. By interpreting the information from the many small events, a much more complete picture can be obtained than if the analysis is restricted to a few larger events. The new SNSN, see Figure 4-2, is now one of the world's most modern and advanced seismological networks, and routinely analyzes events with magnitudes well under zero.

4.2 Seismology in Sweden

Seismological instrumental observations commenced in Sweden in 1904 with the installation of the 1,000 kg Wiechert seismometer in Uppsala. When this had been running for only 19 days the largest earthquake in Sweden's vicinity in historical time occurred.The

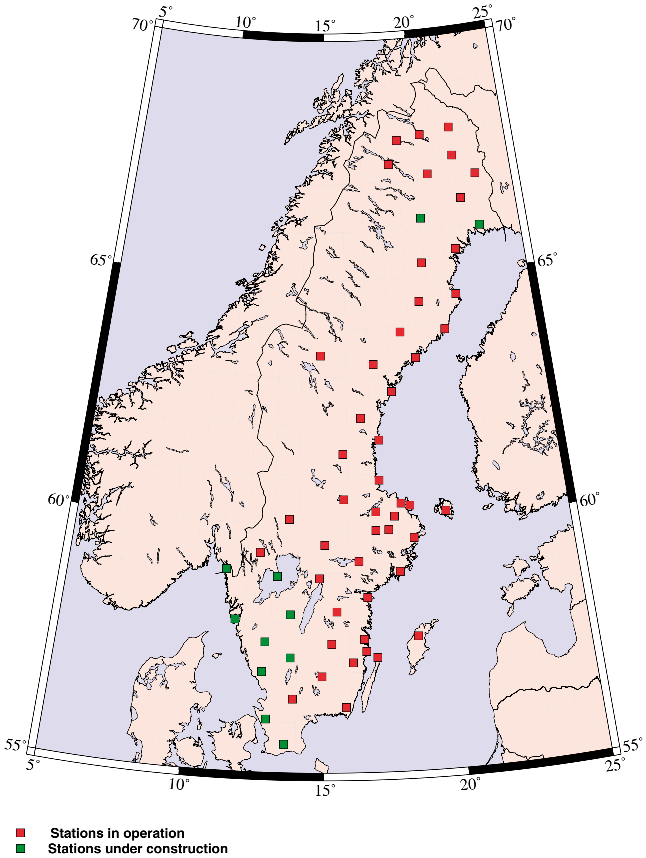

Figure 4-2. *As of this writing, in February 2006, the Swedish National Seismic Network (SNSN) has 51 stations in operation and 9 sites awaiting instrument installation.*

event occurred offshore, by the Koster islands, and had magnitude M_S 5.4. Various buildings (church towers, chimneys etc) close to the epicenter were damaged, and the event was felt over large parts of Sweden, Norway and Denmark, even as far away as in St. Petersburg.

Based on paleoseismological studies it is now generally accepted that approximately 10,000 years ago several very large earthquakes occurred in the north of Sweden. The largest of these may well have exceeded magnitude M_W 8, on a par with the size of the largest events observed on earth. It seems highly likely that the occurrence of these events was related to unloading due to deglaciation at the end of the ice-age. While post-glacial rebound continues, the rate of this has decreased greatly since the deglaciation. Therefore activity in Sweden is today generally in line with what is expected from a stable shield area far from a plate boundary (the closest of which is the mid-Atlantic ridge, over 1,000 km away).

As can be seen from Figures 4-1 and 4-3, a considerable number of earthquakes do occur in Sweden, unevenly distributed over the country. One concentration of activity is along the north-east coast, possibly partly related to changing relative sea level due to the post-glacial rebound. Another area of higher activity to the south-west is related to a major geological suture zone in this area. In the far north of the country, several of the glacially induced faults discussed above show considerable activity. As these faults have been significantly active at some point during the last few thousand years, it is not surprising that they still show small but very significant levels of activity today. It is unclear if this activity reflects an exponentially decreasing relaxation process after the earlier larger motions, or if there is still a significant risk of large movements on these faults.

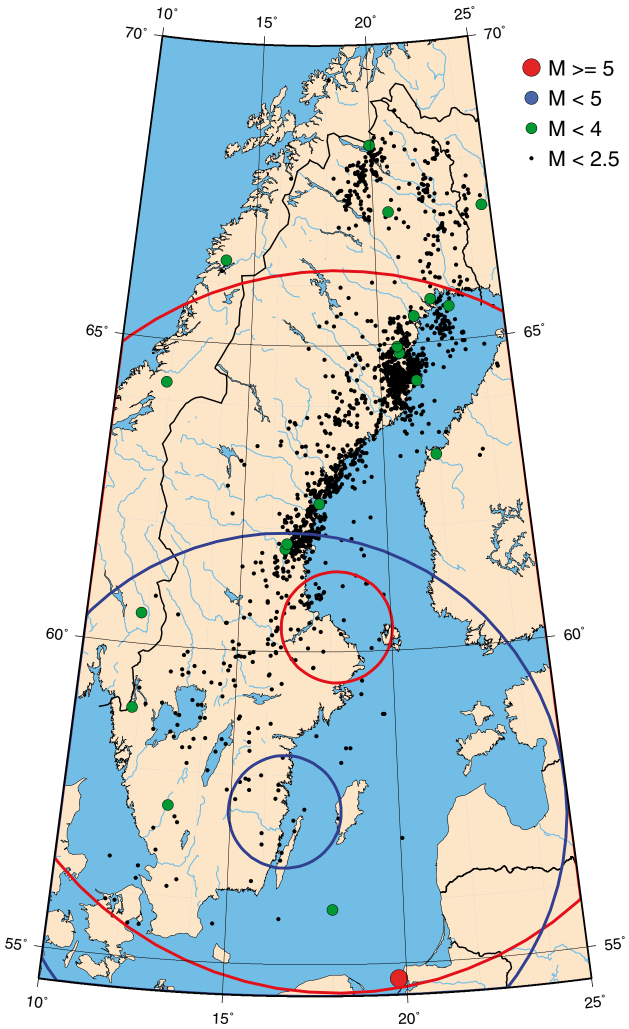

Figure 4-3. *Earthquakes registered by the Swedish National Seismic Network between August 2000 and December 2005. Large red circle has a 650 km radius from Forsmark and the large blue circle has a 500 km radius from Simpevarp. Small circles have radius 100 km.*

The data in Figure 4-1 extends back to 1375, based on historical documents. Obviously, the reliability of some of the data, especially the older data, must be regarded as questionable. However, the existence and approximate position and magnitude of the larger events can probably be regarded as reliable. Event detection improved with the installation of the Wiechert seismometer in 1904, but the very poor geographical distribution of instruments meant that epicenter and magnitude estimates were very uncertain. In various periods during the middle of the last century, various temporary stations were deployed in different parts of the country, leading to geographical and temporal variations in the detection and correct localization of events. In the 1960s more modern instruments were deployed, and some of these (about 6 stations) were run continuously. Such data were complemented by various more dense temporary networks. Starting in 1998 the earlier permanent stations were replaced by modern digital instruments, and since then the number of stations has been expanded to about 60, see Figure 4-2. The data recorded by the modern network is shown in Figure 4-3. Note the difference in pattern in Figures 4-1 and 4-3 for the south-west of the country. This is a simple consequence of the poor station coverage in this part of the country for the modern instrumental data. As indicated in Figure 4-2, the situation will dramatically improve during 2006. Clearly, when interpreting Figures 4-1 and 4-3 it is important to be aware of the uncertainties and limitations in the data sets.

As of this writing, the new SNSN has been operational for five and half years, with a yearly increase in the number of stations. To date, roughly 1,900 earthquakes have been recorded, located and had focal mechanisms calculated. This can be compared to the data base of historical (1375–2005) Nordic earthquakes used for Figure 4-1, which comprises approximately 1,800 earthquakes which can be classified as Swedish (with today's borders). Very few of these events have focal mechanisms. The SNSN currently records approximately two earthquakes per day and when the newly constructed stations in southwestern Sweden become operational, this rate is expected to increase by approximately one earthquake per day. In addition to the new stations, an agreement has been made between the SNSN and the Finnish Institute of Seismology in Helsinki about cooperation on detection and analysis of earthquakes in the Bay of Bothnia, which today falls between the two national networks. This is expected to increase our knowledge about the processes that produces earthquakes in the area.

4.3 Observed earthquake occurrance rates

The frequency-magnitude, or Gutenberg-Richter, relation describes the number of events equal to or larger than a certain magnitude in a specific data set. It is usually expressed as $\log N = a - bM$ where M is the magnitude, N the number of events equal to or larger than magnitude M, a is the intercept (which depends on the number of events in the time and region sampled) and b is approximately 1. b is referred to as the b-value. The Gutenberg-Richter relation uses the 10-based logarithm, which we will denote log below. In Figure 4-4 we show frequency-magnitude relationships for the Forsmark and Oskarshamn areas. The figures have been created using earthquakes within 500 km radius from Oskarshamn and 650 km radius from Forsmark. The reason for choosing these large areas is to include enough events in the analysis, and also to allow for changes in the seismicity pattern from the current paradigm of low seismicity in south-eastern Sweden and higher seismicity in the south-west. For Forsmark we use a 650 km radius in order to include the Kaliningrad event of 2004. Larger radius than 650 km would include offshore Norway seismicity in the analysis, which is undesirable. The magnitudes in the two catalogs have been homogenized using the relationships of /Slunga et al. 1984/. Only data from 1904 until today is used from the Nordic catalog. We see clearly in Figure 4-4 how much more sensitive the new SNSN is with respect to small earthquakes. The step-like behavior of the SNSN line at larger magnitudes is due to the short period of observation, we have recorded very few events above magnitude 3 and the curve is therefore very unreliable above approximately magnitude 3.

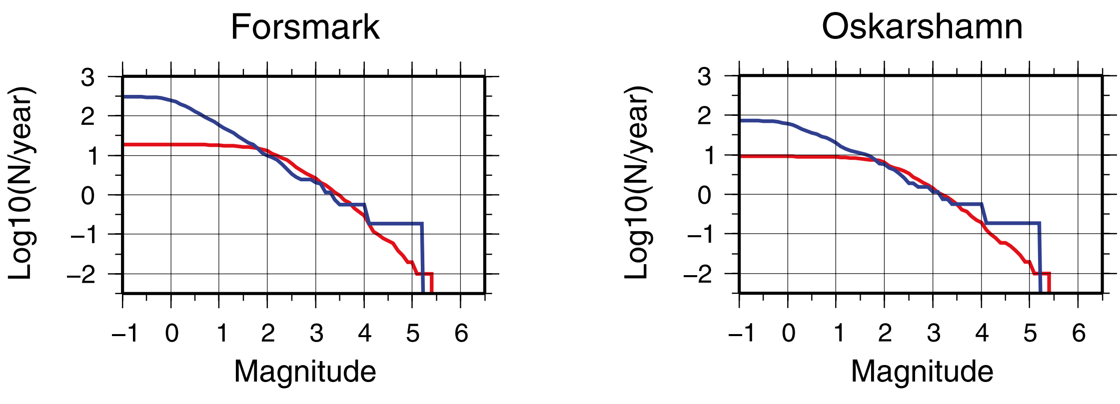

Figure 4-4. *Frequency-magnitude relationships using number of events per year. SNSN data, blue lines, Nordic historical (1904–2005) data, red lines. Left) Forsmark, events within a 650 km radius. The SNSN data has b = 0.75 and the Nordic data has b = 0.97. Right) Oskarshamn, events within a 500 km radius. The SNSN data has b = 0.60 and the Nordic data has b = 0.87.*

A least-squares fit to the slope in the frequency-magnitude plot gives the b-values. For Forsmark, the SNSN data gives b = 0.75 and the Nordic data b = 0.97. For Oskarshamn, the SNSN data gives b = 0.60 and the Nordic data b = 0.87. The smaller the b-value is, the higher is the occurrence rate of large earthquakes compared to small earthquakes. Although seemingly different, we cannot confirm that the b-values of the two data sets are statistically different. This is because the SNSN does not cover the entire circles in Figure 4-1 and because there so few observations of larger events. Also, the station coverage for the data reported in the Nordic catalog has varied substantially since 1904. Our b-values are generally lower than those reported by /Kijko et al. 1993/, who for Sweden south of latitude 60° have b = 1.04 ± 0.05 and for north of latitude 60° b = 1.35 ± 0.06. /Slunga et al. 1984/ report a b-value of 0.75 for southernmost Sweden, latitude south of 56.5° and longitudes between 12° E and 14° E. The frequency-magnitude relationships in Figure 4-4 are used to quantify the discussion on the frequency and probability of larger earthquakes in Sweden below. Before that, however, we will use the observation of two events with magnitude larger than five in the region for a discussion on expected rates of earthquake occurrances. With approximately one magnitude 5 earthquake every 100 years (the Koster islands 1904 and Kaliningrad 2004), we expect approximately one magnitude 6 earthquake every 1,000 years and one magnitude 7 earthquake every 10,000 years in the region, assuming that the external conditions do not vary significantly over these time periods. We note in passing that the very small number of larger events makes a statistical treatment uncertain. If we include both magnitude 5 events the average rate is one magnitude 5 every 50 years, although we know they occurred with a 100 year interval.

As south-eastern Sweden has had relatively low seismic activity during the last 100 years, it is interesting to compare the seismicity in the vicinity of the proposed repository sites with that in the larger areas studied above. We use circles with radius 100 km (31,416 km^2) around the sites as our smaller areas, see Figures 4-1 and 4-3, and use only data from 1 January 1904 until today from the Nordic catalog, i.e. two magnitude 5 evens in 102 years. Note that Figure 4-1 includes all the Nordic data, from 1375 onward. Below we will refer to the number of magnitude 5 earthquakes but this is in the Gutenberg-Richter sense of number of magnitude 5 and larger events. Normalizing the number of magnitude 5 events to the 100 km radius circles (we will refer to these areas as R100), we obtain 0.078 magnitude 5 earthquakes/100 years/R100 for Oskarshamn's 500 km radius circle and 0.046 events/100 years/R100 for the Forsmark 650 km radius circle, the difference being due to the size of the areas. Within the R100 radius of Oskarshamn we have had 7.84 events/100 year/R100 with magnitude greater than or equal to 2.0, compared to 25.76 events/100 year/R100 in the 500 km radius circle. This higher number reflects the higher activity to the north and west. Extrapolating to magnitude 5 events using the Gutenberg-Richter relationship of the 650 km radius area, we obtain 0.024 events/100 years/R100 instead of the 0.078 calculated for the larger region, i.e. a factor 3.3 lower occurrence rate. Repeating the exercise for Forsmark gives 13.73 magnitude 2.0 or larger events/100 year/R100 in the 100 km radius circle, compared to 31.07 events/year/R100 in the 650 km radius circle. This corresponds to 0.021 magnitude 5 events/100 years/R100 in the 100 km radius circle instead of the 0.046 events/100 years/R100 in the 650 km radius circle. The occurrence rate of magnitude 5 earthquakes is thus a factor 2.3 lower in the smaller area around Forsmark than would be expected from the data in the larger area.

The result of these calculations, that the vicinity of Oskarshamn and Forsmark has a factor of approximately 2–3 lower rate of magnitude 5 earthquakes than southern Sweden in general in the last 100 years, has to be interpreted with great caution. First, there are very few events in the smaller areas, 8 above or equal to magnitude 2.0 in Oskarshamn and 14 in Forsmark, which makes the calculations statistically unstable. Second, episodicity of the seismicity on time scales of a hundred years or more have not be taken into consideration. Third, since no magnitude 5 events have been observed within a 100 km radius circle of either Oskarshamn or Forsmark, we have used the frequency-magnitude relationships of the larger areas to calculate the occurrence rates. This may be incorrect. In conclusion, we do not have enough data or time of observation to be able to state that the probability of a magnitude 5 event is lower in south-eastern Sweden than in the south-west. The south-east certainly does have a lower earthquake rate currently. In this context, however, the Kaliningrad magnitude 5 earthquake of 2004 should be kept in mind as it occurred in an area of very little seismicity for the last 100 years.

4.4 Probabilities of future earthquakes

This section is rather similar to the preceding section in that we present estimates of future earthquake occurrances. However, instead of using the point observation of two magnitude 5 events in 102 years, we will use a regression line through the frequency-magnitude distribution and a model for earthquake occurrence to estimate actual probabilities. The simplest model used to describe earthquake occurrance is the Poisson distribution. Using the Poisson distribution, the probability, *P*, that an earthquake of magnitude *M* or larger will occur within an exposure time t, given an average recurrence time *T* (M), is:

$$P(\%) = 100(1 - e^{-t/T(M)}) \qquad (1)$$

In order to obtain average recurrence times that reflect all available data, we splice frequency-magnitude data from the Nordic and SNSN networks, using the 650 km circle around Forsmark, see Figure 4-5. We use the Forsmark circle instead of the smaller Oskarshamn circle since the SNSN network has significantly more observations in the larger circle. The results of this and the following sections would be slightly different if we used the Oskarshamn circle instead. The data is, however, not plentiful enough to statistically show a difference between the two sites. The spliced data set enables us to use data from magnitude 0 to 5, giving a Gutenberg-Richter relation with b-value 0.81 and a-value 2.63. Note that we assume a constant b-value over the entire magnitude range, which according to Figure 4-5 may or may not be correct. As seen in Figure 4-5, the regression line slightly over-estimates larger events. The recurrence time for magnitude 5 events, for example, is 25 years using this model.

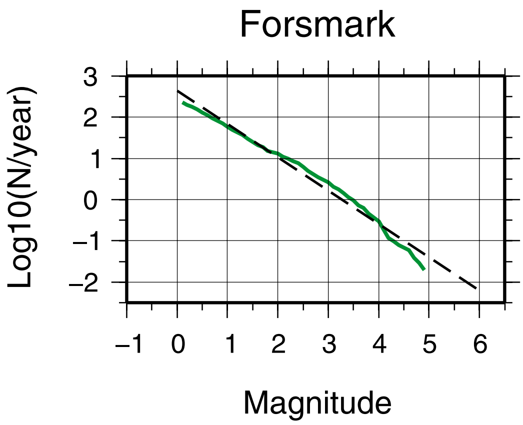

Figure 4-5. *Frequency-magnitude relationships using number of events per year in the 650 km radius around Forsmark. Spliced Nordic and SNSN data, green line. Maximum likelihood estimate of best fitting Gutenberg-Richter relationship, dashed black line. b = 0.81 and a = 2.63.*

The recurrence times from the Gutenberg-Richter relation are used with Equation 1 to calculate the probability of a certain magnitude event within a certain time period, see Table 4-1. We see in Table 1 that the low b-value produces a return time of 1,000 years for a magnitude 7 event instead of the 10,000 years of our rough estimate based on one magnitude 5 every 100 years and a b-value of 1. This indicates how important the b-value is for the long time estimates.

Repeating the calculations for the 100 km radius area around Forsmark we obtain the results in Table 4-2, which can be compared with the R100 discussion above. We see that the recurrence time for a magnitude 5 event is 1,061 years according to Table 4-2, which gives 0.094 events/100 years/R100 instead of the 0.046 that the analysis above yielded. The approximate factor of two difference is due to the b-value we use. Table 4-2 also shows that there is an appreciable probability, 61%, of a magnitude 5 event within 100 km of Forsmark in 1,000 years and a non-trivial probability, 14%, of a magnitude 6 event in the same time span.

Going one step further, we restrict the analysis to a circle of radius 10 km around Forsmark. A 10 km radius allows ample room for the full rupture length of a magnitude 6 earthquake, which is usually on the order of 10 km. Table 4-3 shows that the recurrance times increase by a factor of 100, as expected. The probability of a magnitude 5 earthquake drops to 0.9% in a 1,000 years and for a magnitude 6 the probability is now 0.15%. Note that when using an area as small as this, geological considerations become very important. A large earthquake requires a large fault to occur on, which then has to be present in the considered area.

Table 4-1. Probabilites, in percent, of earthquakes within the Forsmark 650 km radius area. M is the magnitude, T the recurrence time in year and t the exposure time, also in year.

M	T	t=10	t=20	t=30	t=50	t=100	t=1,000
3	1	100	100	100	100	100	100
4	4	91.9	99.3	99.9	100	100	100
5	25	32.8	54.9	69.7	86.3	98.1	100
6	158	6.1	11.9	17.2	27.1	46.8	99.8
7	1,000	1.0	2.0	3.0	4.9	9.5	63.2
8	6,310	0.2	0.3	0.5	0.8	1.6	14.7

Table 4-2. Probabilites, in percent, of earthquakes within the Forsmark 100 km radius area. M is the magnitude, T the recurrence time in year and t the exposure time, also in year.

M	T	t=10	t=20	t=30	t=50	t=100	t=1,000
3	27	31.3	52.8	67.5	84.7	97.7	100
4	168	5.77	11.2	16.3	25.7	44.8	99.7
5	1,061	0.94	1.87	2.79	4.60	8.99	61.0
6	6,696	0.15	0.30	0.45	0.74	1.48	13.9
7	42,250	0.02	0.05	0.07	0.12	0.24	2.34
8	266,579	.004	0.01	0.01	0.02	0.04	0.37

Table 4-3. Probabilites, in percent, of earthquakes within the Forsmark 10 km radius area. M is the magnitude, T the recurrence time in year and t the exposure time, also in year.

M	T	t=10	t=20	t=30	t=50	t=100	t=1,000
3	2,666	0.37	0.75	1.12	1.86	3.68	31.3
4	16,820	0.06	0.12	0.18	0.30	0.59	5.77
5	106,127	0.01	0.02	0.03	0.05	0.09	0.94
6	669,617	.002	.003	.005	.008	0.01	0.15

4.5 Expected distances to large earthquakes

The Guthenberg-Richter relation can also be used to assess within which distance from a certain position we expect to have an earthquake of a certain magnitude within a certain time. We normalize the a-value from above to number of events per year per square kilometer, i.e. $N/(S \times T)$ where N is the number of events, S the area in square kilometers and T the time in years. That gives us a = –3.52. The b-value does not change. We can then write the non-normalized form of the Gutenberg-Richter relation as /Jacob 1997/:

$$\log N = a - bM + \log S + \log T \tag{2}$$

We define a specific source area $S' = \pi r^2$ which is needed to produce one event, i.e. $N = 1$, of magnitude M or larger, in a period of T (years). In order to find the distance to an event of magnitude M or larger from a point in a region characterized by the a- and b-values, we express S' in terms of a median-probability (50-percentile) distance d. $d = r/\sqrt{2}$ which gives $S' = 2\pi d^2$. Inserted into Equation 2 above we have

$$\log 1 = 0 = a - bM + \log S' + \log T = a - bM + 2 \log d + \log 2\pi + \log T \tag{3}$$

or

$$\log d = \frac{bM - a - \log 2\pi - \log T}{2} \tag{4}$$

Inserting our a = –3.52 and b = 0.8 and a number of different recurrence periods into Equation 4, we obtain the results in Table 4-4.

The median distance can be seen to increase with magnitude and decrease with average recurrence time. We see, for example, that a magnitude 5 event is expected within 230 km of Forsmark with a recurrence period of 100 years and a magnitude 6 event within 183 km with a recurrence period of 1,000 years.

Table 4-4. Median distances *d* (km) expected for earthquakes with magnitudes M and recurrence periods T (years) for an unconfined seismic source with uniform cumulative rate log *N* (per year per km^2) = –3.52–0.8*M*.

M	T=50	T=100	T=250	T=500	T=1,000
3	52	37	23	16	12
4	130	92	58	41	29
5	326	230	146	103	73
6	818	579	366	259	183
7	2,056	1,453	919	650	460
8	5,163	3,651	2,309	1,633	1,155

5 Maximum movements related to individual earthquakes

5.1 Modeling of static deformation

The general pattern of permanent (static) deformation due to a single earthquake can be simply modeled by regarding the earth's crust as a homogeneous elastic half-space. We apply the methodology of /Wang et al. 2003/ which is based on the theoretical expressions of /Okada 1992/. We stress that this modeling refers only to the static deformation effects, not the dynamic effects which occur temporarily due to the radiation of seismic energy from the source. These dynamic effects are discussed in section 5.2 below.

For our modeling experiment, we choose a scenario based on the parameters of the Kaliningrad earthquake of 2004. This event had Richter magnitude M_L 5.0 with approximately 50 cm maximum slip on a fault area of about 450 m radius. The event was at about 20 km depth, but in our examples we compare the surface static deformation from a Kaliningrad type event at three different depths. The first case is based on the magnitude 3 Norrtälje event of 1979, which was at only 800 m depth. The second case shown is for a slightly deeper event (3 km), and the third case is for 12 km depth, corresponding to the (well determined) depth of the 1976 magnitude M_L 4.5 event in the Gulf of Finland /Slunga 1979/. The first two scenarios can be regarded as significantly less probable than the third, because the low normal stress at shallow depths implies that there is less likelihood of a large event there.

The modeling presented assumes a near-vertical fault striking east-west and pure strike-slip motion, which is a very common type of focal mechanism in central- and southern Sweden /Slunga, 1991/. Using seismological terminology, the model fault is described by strike 90°, dip 90° and rake 180°. This is also in agreement of mapped east-west structures in Oskarshamn and Forsmark.

Modeling the reference event at 800 m depth produces fairly large surface deformation where the maximum displacements are of similar magnitude both horizontally and vertically, and do not exceed 1 cm. It follows that the maximum relative motion at the surface across the fault is about 2 cm in this scenario. Similarly, the maximum relative vertical motion is about 2 cm. The distance between lobes of maximum displacement is on the order of 2 km.

The second example, shown in Figure 5-1, is a scenario with higher probability where the hypocentral depth is 3 km. The maximum displacements have decrease to 1.5 mm, with relative displacements being less than 3 mm and these loci are approximately 5–6 km apart. The final example is for 12 km depth. The maximum static deformation at the surface is now very small with 0.2–0.3 mm of maximum relative displacement.

We can conclude that the surface static deformation from our reference event is very small. Of course, larger events produce larger static deformations, but the probability of a larger event decreases logarithmically, according to the Gutenberg-Richter law. However, we should recall that it is not uncommon that large earthquakes in seismically active areas can produce rupture movements at the surface of meters or even tens of meters, and that the post-glacial neotectonic faults in Sweden appear to show throws (relative movements between the two sides of the fault) of up to 10 m.

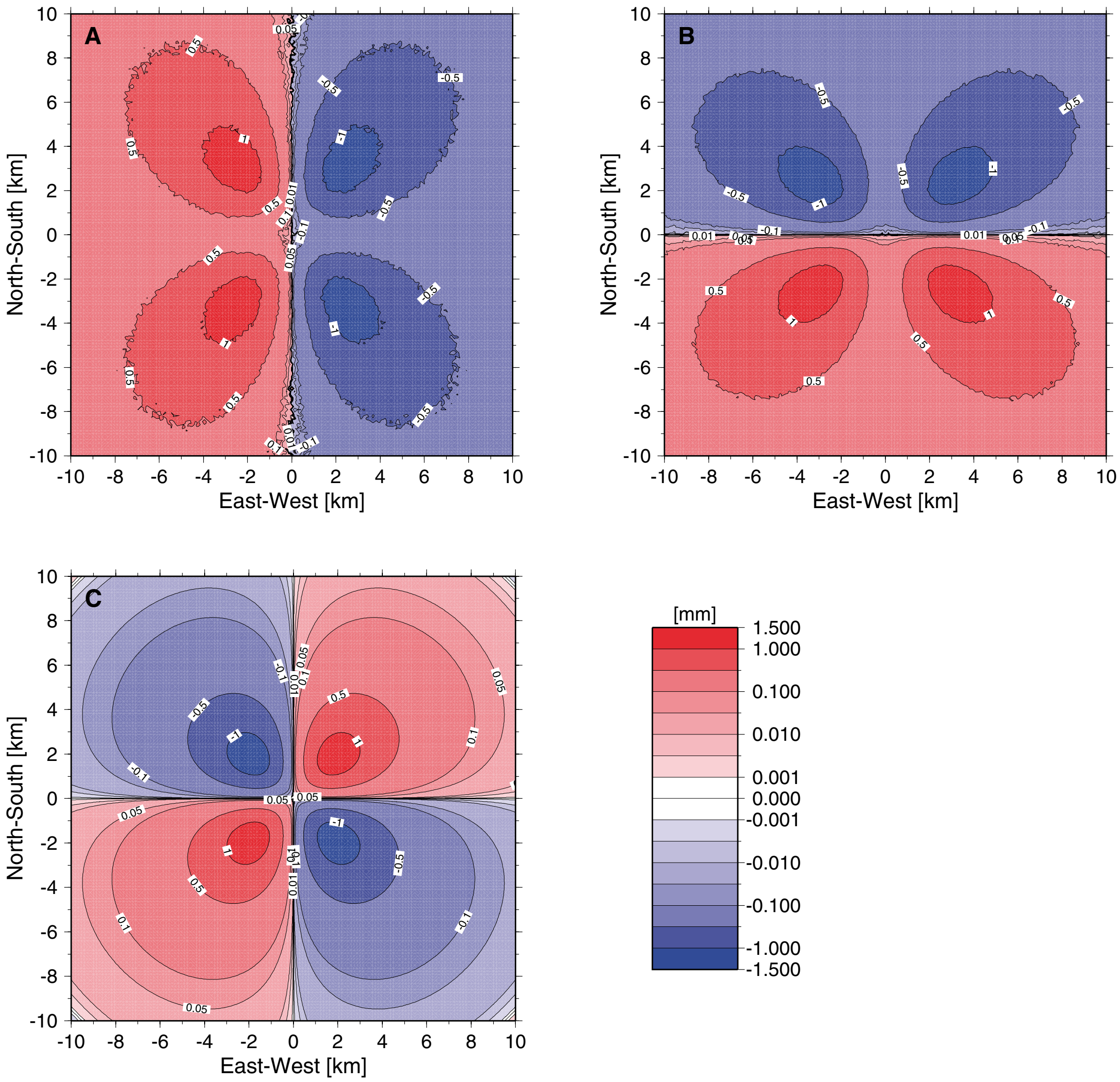

Figure 5-1. *Static deformation at the surface from a M_W 5.0 earthquake at 3 km depth. The epicenter is at (0.0) and the fault is vertical, striking east-west with pure right-lateral strike-slip motion. Contours in millimeter. A) North-south horizontal motion, south positive. B) East-west horizontal motion, west positive. C) Vertical motion, up positive.*

5.2 Dynamic loading

By far the most common way of describing the damage potential of large earthquakes is peak ground acceleration, i.e. the maximum acceleration of the ground motion at a particular place on the ground surface. This "peak ground acceleration" (PGA) is usually expressed as a proportion or percent of the acceleration due to gravity (i.e. 9.8 m/s^2). The observed horizontal and vertical PGAs are often different. Because of the ubiquitous force of gravity, which buildings are always constructed to withstand, with de facto a significant margin of safely, the most destructive movements are generally those which are horizontal rather than vertical, largely irrespective of the relative vertical and horizontal amplitude of the movements. In general, there is however a close correlation between the two, and it is common to quote PGA without a detailed consideration of the difference between surface motion in the different directions. The popularity of PGA as a standard expression of earthquake risk is partially because it is specific, simple (a single number) and unambiguous. In many countries, building codes referring to earthquake safety are expressed explicitly in terms of PGA.

There are also other measures of the potential destructive capacity of large-amplitude seismic waves. Experience from e.g. the USA demonstrate that a useful complement to PGA, which in many cases may be a better measure of risk, is "cumulative absolute velocity" (CAV). CAV is an integrated measure of the velocity of ground movement during a period when this is over some predefined threshold value. Clearly, several cycles of motion of a given amplitude can cause significantly more damage than one cycle of the same maximum amplitude. Furthermore, the damage potential of ground vibrations depends upon their frequency content. Because of the properties of buildings, some frequencies (e.g. those which can cause a construction to resonate) can be far more destructive than others. For many constructions, the resonant frequency is such that lower frequency seismic signals are potentially far more destructive than higher frequency signals of the same maximum amplitude. CAV is being increasingly used today by e.g. the nuclear power in- dustry in the USA. The geology of Sweden and many of the most risk-prone areas of the USA is significantly different. Preliminary studies /Zemichael et al. 2004/ suggest that in Swedish conditions, the CAV concept may be more generally applicable than in many parts of the USA. However, as risk in Sweden is both relatively low and difficult to assess (because of the low seismicity), given the uncertainties the practical difference between PGA and CAV may not be so significant here.

Existing ground vibration models for Swedish bedrock (we base our analysis on /Slunga et al. 1984/) together with parameters corresponding to the magnitude of the Kaliningrad event of 2004 (M_L 5.0) and the depth of the "Gulf of Finland event" of 1976 (12 km) lead to maximum surface accelerations of 0.05 g on crystalline rock. The corresponding maximum ground velocity is estimated to be 16 mm/s, and the differential dynamic movements over a length of 500 m to be 0.4 mm. The expected static differential movement over this length is, of course, much less.

If a "Kaliningrad" event should occur at a depth of 3 km, the corresponding values are 0.3 g, 74 mm/s and differential movement over 500 m about 2 mm. The static differential movements here would be less than 1 mm. This is, however, a relatively improbably scenario because of the lower normal stress at shallower depths. Compensating for this effect modifies the expected values of maximum acceleration to 0.16 g, velocity to 42 mm/s and differential dynamic movement to about 1 mm over 500 m.

6 Seismic and aseismic deformation

As we have described above, earthquakes are produced by rapid (dynamically unstable) movements along fracture surfaces in the earth's crust. The size of these movements is strongly correlated to the events (seismic) magnitude. Smaller events with M_L less than 1 typically correspond to lateral movements of 0.001–1 mm, whereas larger events with magnitude around 5 have slips of 50–500 mm. These rapid (seismic) movements occur on a limited area on the fault plane, for example a roughly circular area with a radius between 30 and 900 m.

Slower fault motions (relative velocity across the fault plane of less than perhaps 0.1 mm/s) do not in general generate seismic signals of sufficient amplitude to be detected by normal seismometers. Such movements are commonly considered "aseismic", or referred to as fault creep. In accordance with standard Mohr-Coulomb friction theory, movement is initiated on a fault when the friction and cohesion (intrinsic strength of the material) becomes less than the applied shear stress. Whether or not an initiated movement is dynamically unstable and therefore accelerates depends upon the properties of the fracture and the surrounding medium, and the characteristics of the local stress field. In situations where rapid acceleration does not occur, the movement ceases when the shear stress has been reduced to a value corresponding to less than the dynamic friction of the fracture. In general, only in cases of seismic motion (10–1,000 mm/s) does the inertia of the rock mass become significant. Aseismic fault creep is most commonly observed at plate boundaries, such as the creeping section of the San Andreas fault or "slow earthquakes" in subduction zones.

In addition to dislocation movement on fractures, the earth's crust can deform elastically when submitted to an external change in stress. Such deformations are distributed throughout the entire volume of the stressed area in accordance with Hookes law. The stress changes can be positive or negative and the deformation varies accordingly. In general, ongoing tectonic deformation, where stresses and stress changes are transmitted from plate boundaries into plate interiors, is elastic in the upper crust until sufficient shear stress has been built up on a fault. Once the strength of the fault is overcome the fault moves and we have either an earthquake or aseismic creep on the fault. The motion on the fault also produces an elastic deformation field, as shown in section 5.1, which is added to the regional field. Other sources of, short lived, elastic deformation are e.g. earth tides and atmospheric pressure changes. Elastic deformations have to be taken into account when relating observed geodetic movements to motion deduced from seismic activity, also because the elastic deformation processes may not be directionally monotonic.

Movement on a fault, whether seismic or aseismic, causes the total elastic energy of the source volume to decrease. The greatest reduction is in the central part of the active fault, where the motion is greatest. In some areas around the fault the deviatoric stress can increase, leading to aftershocks. The mechanisms leading to the time delay between main shock and aftershocks is relevant to a discussion about elastic versus non-elastic processes. An adjustment to pore pressure changes due to volume strain produced by the main event may play an important role /Lindman et al. 2006/.

The surface deformation associated with larger earthquakes is sometimes directly observed using geodetic methods (InSAR, GPS etc). However, most events in the shield area are at considerable depth (7–35 km) compared to the deduced dimensions of the slip area. If an event has a maximum slip of 10 mm (e.g. $M_L = 3$) and a slip radius of 300 m, at a distance of only 300 m from the event the deformation has decreased to a value of less than 0.1 mm. The static deformations at the surface due to such movements at depth cannot be geodetically observed using currently available technology, cf section 5.1.

Clearly, it is of great significance in understanding deformation processes whether or not significant aseismic movements occur in Fennoscandia. If this is the case, then movement on faults may be very much greater than that deduced by summing the motions (moments) of observed earthquakes on the faults. There is some empirical evidence which can be used to support a hypothesis of significant aseismic movements, both at regional and local levels.

6.1 Regional deformation

Studying the temporal and spatial behavior of microearthquakes from southern and northern Sweden, /Slunga 1991/ hypothesized that there is significant stable (aseismic) sliding along whole faults. This sliding accumulates stress at asperities along the faults which, when they break, cause microearthquakes. For southern Sweden, from 1980 to 1984, /Slunga 1991/ estimated that there is aseismic horizontal deformation across the region of about 1 mm/year/100 km.

In 1993 a network of permanent GPS-stations was established in Sweden (SWEPOS) and in the following years permanent GPS-stations were established also in Finland (FinnRef). Together with some Norwegian GPS-stations, these networks were incorporated into the BIFROST project /e.g. Johansson et al. 2002/ in order to provide data for glacial isostatic adjustment (GIA) analysis. With almost ten years of data, the thorough BIFROST analysis shows a maximum vertical rebound velocity of approximately 11 mm/year and horizontal movements which exceed 2 mm/year at several sites. Formal uncertainties are as low as 0.1 mm/year for the horizontal rates and 0.2 mm/year for the vertical rate at most stations /Johansson et al. 2002/. The velocity estimates are constant rates fitted to the time series data of each GPS-station, in agreement with GIA which should not have rate changes over a ten year period. The deformation rates observed by the BIFROST project agree very well with predicted deformation rates from models of the GIA process, both vertically and horizontally /Milne et al. 2004/. The Earth model used by /Milne et al. 2004/ has a uniformly thick, elastic lithosphere overlaying two mantle layers of varying viscosity. No lateral thickness variations or variations in elastic/non-elastic parameters is included. There is, thus, no fault slip in the model. The horizontal strain rates have sizes up to 2 mm/year/500 km, which corresponds to an elastic stress change of 0.0004 MPa/year.

Removing the modeled GIA deformations from the observed deformation field, the BIFROST group presents residual movements of up to 2 mm/year in Kuusamo, Finland, and in southern Sweden the residuals are 0.5–1 mm/year /Milne et al. 2004/. In terms of differential movements of the residual field in SW Sweden, the stations Jönköping and Borås shows approximately 0.8 mm/year/70 km /Milne et al. 2004/, which corresponds to an elastic stress change of 0.0011 MPa/year. This is more than twice the stress change caused by the deglaciation according to the BIFROST analysis. In a model of the whole of western Europe, the GPS observations show that the land uplift and plate tectonic forces works together in the deformation processes, /Marotta et al. 2004/.

The largest residual found in the GPS analysis is in the Kuusamo area in Finland, approximately 2 mm/year. It is interesting to note that this area is the seismically most active area in Finland. As was shown in the discussion above in section 5.1, the static deformations due to seismic slip (the earthquake source) cannot be observed at the surface for most of the Fennoscandian seismicity. Again, the hypothesis that the earthquakes are related to extensive fault slip would explain the coincidence in space and time of high earthquake activity and GPS anomalies. Note also that the earthquake activity is in general episodic in character which means that if it is significant one should expect a large formal error estimate in the estimation of the stationary movements, as is observed.

6.2 Local deformation

Campaign GPS measurements have frequently inferred local deformations rate in Sweden on the order of 1–4 mm/year /Pan and Sjöberg 1999, Pan et al. 1999, Sjöberg et al. 2004/ but are hampered by high uncertainties. As is evident from the discussion above, it is very difficult to use the regional, high resolution SWEPOS network to confirm these measurements. SWEPOS is not dense enough, considering that a 1 mm/year movement on a 20 km long fault zone produces less than 0.2 mm/year of deformation at distances beyond approximately 5 km.

The area of Äspö north of Oskarshamn has been studied by GPS campaign measurements during 2000–2004 /Sjöberg et al. 2004/. The GPS network was located north and south of an E-W fault. The results were only partly significant but were interpreted as possible slip along the fault at a rate of about 1 mm/year, with the northern side moving eastwards. This is in agreement with the general stress field in southern Sweden having principal compressive horizontal stress in the WNW-ESE direction. No seismicity was observed in the area. If the geodetic data does indeed correspond to lateral movements of the surface, and if these movements do reflect motion also at greater depths, then significant aseismic movement has occurred. We can illustrate this by considering the seismic moments. The geodetic data is suggested to correspond to movements of 1 mm/year over a fault length of at least 3 km. Assuming that fault movement must extend to a depth roughly comparable to the horizontal fault length that moved, say 1 km depth, the moment (integrated slip) is approximately 10^{14} Nm/year. If this moment was released seismically by magnitude 0 events it would require approximately 10,000 events per year.

The earthquake closest to the site that has been thoroughly analyzed is the earthquake of September 8, 1988, with $M_L = 1$, focal depth 16 km, peak slip 0.4 mm, south of Oskarshamn /Slunga and Nordgren 1990/. It supports the likelihood of slip on east-west faults as its fault plane solution indicated strike-slip movements on either a E-W or N-S vertical fault. The direction of motion was in agreement with the results of /Sjöberg et al. 2004/.

The GPS results from SWEPOS show a westward movement of approximately 0.7 mm/year at the Oskarshamn station, about 40 km south of Äspö. The two stations north of Äspö, Jönköping to the west and Visby to the east, both show only about 0.2 mm/year westward motion /Milne et al. 2004/.

These three observations pertain to totally different crustal deformations but the directions of motion are consistent, right lateral on E-W faults. Even if one assumes that the estimated movement of 1 mm/year is correct (as indicated by the significant part of the GPS campaign) it does not mean that the movement will go on. Fault movements are often episodic.

6.3 Discussion

The inferences on aseismic deformation presented above are in most cases barely statistically significant. They do, however, seem to come to similar estimates in terms of the magnitude of deformation, approximately 1 mm/year. If, as the above data appears to suggest, aseismic movement is much larger than seismically observed motions, then there may be much more fault movement than otherwise expected. However, as aseismic movements are generally less destructive than seismic movements, in the context of a repository the aseismic movements may only be of direct relevance if the movements occurs on a fault in the repository itself. If significant and large scale aseismic deformation is ongoing in Fennoscandia, this will affect stress propagation through the region which may be significant when assessing seismic risk in the context of stress accumulation. When assessing the probability of large earthquakes in an area based on observed seismicity, the aseismic deformation does not directly affect the assessment.

Making geodetic measurements of campaign type with the necessary degree of accuracy is difficult. Despite the significant movements suggested by the data and associated error estimates, it must be considered more than possible that the errors are underestimated, and the deduced movement does not exist. We also point out that /Johansson et al. 2002/ explicitly conclude that they see no evidence of the relative horizontal motions inferred by /Pan and Sjöberg 1999/.

We conclude that, while (the sparse) data supports the existence of significant aseismic movement on faults, this data is limited in volume and often associated with large uncertainties, implying that it must be interpreted with caution. Further studies along the lines of /Slunga 1991/, complemented with suitable surface (GPS) data appear to be motivated. In order to gain more definite information on the deformations at the local sites it is recommended that continuous GPS measurements are establish for several years using a dense network in the area.

7 Climate related issues

The time spans considered in this study, 100 years and 1,000 years, are short relative to major climatic changes such as ice ages. The influence of the climate on earthquake activity during these times is expected to be minor and we will just briefly discuss two climate related issues here.

7.1 Global warming

The Intergovernmental Panel on Climate Change (IPPC) expects the current global increase in temperature to continue into the future, although the duration and magnitude of the increase is very uncertain /IPPC 2001/. In the context of earthquake activity, an increase in sea-level is probably the only effect of global warming that needs to be considered. Thermal expansion of water and melting of continental ice sheets both contribute to sea-level rise. Studies of the decimation and/or accumulation of ice in Greenland and Antarctica are, as of yet, not conclusive. For the next century, the IPPC projects an increase in mean sea-level by 0.09 to 0.88 m between 1990 and 2100 /IPPC 2001/. For the next millennium, the Greenland ice sheet is likely to disappear if the warming trend continues whereas the fate of the Antarctic ice sheet is less certain. If both these ice sheets disappear, sea-level is estimated to rise more than 70 m.

Increasing sea-levels will slowly add both a load and an increasing pore pressure to the rock where today's coast-lines lie. It is well known that reservoir impoundment often induces earthquakes below the reservoir /e.g. Simpson et al. 1988/, an activity which usually decays away in a few years. On the other hand, studies of how ocean tides affect earthquake activity have not reached consensus, although a recent study indicates that there is a correlation between the highest Earth tides and shallow thrust fault events /Cochran et al. 2004/. The sea-level rise discussed here is slow both compared to the filling of a dam and to tides. It also extends over a large region, thereby not inducing large strains. Although a pore pressure increase will tend to decrease fault stability, we estimate that the increase in seismicity due to sea-level rise will be very small.

7.2 A future ice-age

Climate models and geological evidence indicates that ice-ages in northern Europe initiate in similar patterns /Näslund 2006, Lambeck 2005/. We therefore assume here that a future glaciation will initiate similarly to the latest, Weichselian, glaciation. Models of the Weichselian glacial history /Näslund 2006, Lambeck 2005/ indicate that the build up of an ice sheet is a slow process. Both these models show that it takes tens of thousands of years before an appreciable ice sheet has advanced down to the latitudes of Forsmark and Oskarshamn. An advancing ice sheet disturbs the stress field, first by decreasing the horizontal stresses (in the fore-bulge) before the ice sheet arrives, and then by increasing the horizontal stresses when the ice covers the area of interest. How these stress perturbations affect fault stability depends on the pre-existing stress field. /Lund 2005/ showed that if the pre-existing stress field is a strike-slip state, as at depth in most of southern Sweden, fault stability can be decreased by the fore-bulge. When the ice is in place, fault stability increases. In a reverse faulting stress environment, as in the upper kilometer in Forsmark, fault stability is increased both by the advancing ice sheet and under the emplaced ice sheet. For both time periods considered in this report, it is unlikely that glacially induced stresses will have any appreciable influence on fault stability.

8 Final remarks

Roughly 1,900 microearthquake fault mechanism are now available from the new Swedish National Seismic Network (SNSN). With the additional 9 stations that will be instrumented within the next few months we expect to record approximately 1,000 earthquakes per year by the whole network. This should be compared with the approximately 1,800 earthquake locations in Sweden available in the data base of Nordic earthquakes from 1375 to 2005. It is essential that microearthquake monitoring is performed simultaneously (in time and space) with the GPS measurements in order to increase the possibilities to interpret the geodetic observations.

There is a lack of exact knowledge and understanding regarding fault motions and the relation between seismic and aseismic deformation. With the present observational networks SWEPOS and SNSN when these have been in operation simultaneously for long enough time, there are possibilities in the future to get a better understanding of the deformation. The interpretation should also include possible episodic signals that could be identified from the SWEPOS network.

As is obvious from the discussions in the report it is not possible to give numbers on the size of expected permanent fault movement due to aseismic slip for time periods of 100 or 1,000 years based on the available data. Instead permanent dense GPS networks are recommended to be installed and operated for several years (probably 10) in the areas.

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