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Review of paleo-, historical and current seismicity in Sweden and surrounding areas with implications for the seismic analysis underlying SKI report 92:3

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SSM perspective

Abstract

The report contains a review of the current state of knowledge on Swedish paleoseismic events, as well as information on the Swedish earthquake catalogue updated to early autumn 2016 and the revised Nordic earthquake catalogue (Fencat). It includes a section on the possibilities to assess spatial and temporal variations in seismicity, given the sparse early catalogues and the low earthquake activity rate in Fennoscandia. The updated earthquake data is compared to the SKI Technical Report 92:3 (SKI 92:3) data and it comments on how this may affect the ground motion spectra and how this work could continue. SKI 92:3 contains envelope ground response spectra especially developed for Sweden in a joint research project between the Swedish nuclear power inspectorate (SKI) and the Swedish licensees. It was not within the scope of this study to perform a quantitative assessment of how the existing spectra are affected by the updated data set.

Background

The number and quality of instrumentally recorded earthquakes in Fennoscandia have significantly increased since the publication of SKI 92:3, mainly depending on the expansion and modernization of the Nordic seismic networks. The Fencat catalogue accordingly contains many more events than when data were extracted for SKI 92:3. In addition, the historic part of the catalogue has been updated with new estimates of location and magnitude of the larger events, and cleared of a number of frost and blast related events. The large number of smaller earthquakes recorded in the last two decades give a good indication of current spatial variability in the earthquake rate and it is also making it possible to outline fault systems in Sweden that are currently seismically active, although with a low rate.

Therefore, the Swedish Radiation Safety Authority (SSM) has commissioned Seismology group at Uppsala University to carry out the present study with the objective as it set out below. SSM especially emphasized the need to include information on paleoseismic events, which are not included in the SKI 92:3 data.

Objectives of the project

The objective was to review the current status of paleoseismology and historical earthquakes in Fennoscandia and to update and evaluate the earthquake data used in SKI 92:3, as well as to qualitatively assess how the new results would affect the envelope ground response spectra in SKI 92:3.

Results

There are twelve confirmed post- or endglacial fault scarps in Sweden where glacial landforms have been displaced. For many of these there is, however, lack of stratigraphic information which would help constrain both the timing of the events and whether they formed as a result of a single rupture or not. It will also provide information critical to magnitude calculations. There are, however, no indications of observations of multiple large ruptures on the endglacial faults.

Swedish paleoseismic events associated with a fault scarp, as well as confirmed events from Norway and Finland are included in the final inventory. The report presents an updated table of events with location, timing and magnitude, as needed by a seismic hazard assessment.

Based on comparisons of different time slices of Fencat, it could be find that on the large scale the spatial variation in seismicity in Sweden has been relatively stable during the last century.

The results in SKI 92:3 are not expected to be affected by the paleoseismic data, if analysed with the SKI 92:3 methodology.

Occurrence rates estimated with only the recent Swedish earthquake data or the Fencat data indicate that the SKI 92:3 estimate may be conservative for earthquakes of magnitude 5 and smaller but that it may underestimate the rate of large events.

Project information
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Summary

The Swedish Radiation Safety Authority (SSM) initiated a project with the Seismology group at Uppsala University in order to update and assess the earthquake data which forms the basis of the currently used envelope ground response spectra for Swedish nuclear power plants in SKI Technical Report 92:3 (SKI, 1992). SSM especially emphasized the need to include information on paleoseismic events, which are not included in the SKI (1992) data. In this report we will perform an initial comparison of the SKI (1992) data and the updated earthquake catalog and comment on potential qualitative effects on the earthquake occurrence probabilities. A quantitative assessment of how the existing spectra are affected by the updated data set is not within the scope of this project.

The literature on post- or endglacial faulting in Sweden has been reviewed in Appendix 1 of this report. We conclude that there are 12 confirmed endglacial fault scarps in Sweden where glacial landforms have been displaced. For many of these there is, however, lack of stratigraphic information which would help constrain both the timing of the events and whether they formed as a result of a single rupture or not. There are no indications in the literature of observations of multiple large ruptures on the endglacial faults. We discuss claims of paleoseismicity in the absence of surface ruptures and find that there is great uncertainty in many of them, in terms of occurrence, location, magnitude and timing. Only paleoseismic events associated with a fault scarp are included in the final inventory. We add to the inventory of Swedish paleoseismic events the confirmed events from Norway and Finland and present an updated table of events with location, timing and magnitude, as needed by a seismic hazard assessment. It should be noted that research on postglacial faults in Fennoscandia is currently developing very rapidly, bringing new information about the known faults and possibly identifying new ones.

Since the publication of SKI (1992) there has been a tremendous increase in the number and quality of instrumentally recorded earthquakes in Fennoscandia, thanks to the expansion and modernization of the Nordic seismic networks. The joint Nordic earthquake catalog Fencat (Ahos & Uski, 1992; FENCAT, 2016) therefore contains many more events than what it did when data were extracted for SKI (1992). In addition, the historic part of the catalog has been updated with new estimates of location and magnitude of the larger events, and cleaned of a number of frost and blast related events. The large number of smaller earthquakes recorded in the last two decades not only give a good indication of current spatial variability in the earthquake rate, it is also slowly making it possible to outline fault systems in Sweden that are currently seismically active, although with a low rate. Comparing different time slices of Fencat we find that on the large scale the spatial variation in seismicity in Sweden has been relatively stable during the last century. There are significant temporal variations in the seismicity rates, both in the country as a whole and in different subareas. Much of this variability is, however, related to varying degrees of observational efforts and, later, instrumental deployments, making it difficult to draw firm conclusions about variations in crustal processes.

The seismic data underlying the analysis in SKI (1992) is not explicitly described, neither the area used nor the magnitude range included. It is therefore difficult to make comparisons of the updated data to the SKI (1992) values for seismicity rate per area and year and the variation of that rate with size of the events, as expressed by the seismic moment. We find for the paleoseismic data that these are unlikely to affect the results of SKI (1992) if analyzed with the SKI (1992) methodology. However, in a modern

probabilistic seismic hazard analysis they are likely to increase the estimated hazard. Occurrence rates estimated with only the recent Swedish earthquake data (2000-2016) or the joint Fennoscandian earthquake data (Fencat) for a region including Sweden, Finland and the Baltic and two different time periods (1970-2012, magnitude 2 or larger and 1875-2012, magnitude 3 or larger) indicate that the SKI (1992) estimate may be conservative for earthquakes of magnitude 5 and smaller but that it may underestimate the rate of large events. These conclusions do, however, depend on which data were included in the SKI (1992) estimates and that we do not know.

Advances in data acquisition, hazard methodology and international recommendations has made SKI (1992) outdated. We suggest that an update of SKI (1992) is considered, and emphasize the need for expertise in intraplate seismicity if such a project is initiated. We see a number of important issues related to the Fennoscandian intraplate seismicity that needs to be addressed in future analysis: the level of paleoseismicity in Sweden, the stationarity of earthquake occurrence, the maximum possible magnitude and uncertainties in ground motion prediction.

Sammanfattning

Tillsammans med seismologigruppen vid Uppsala universitet initierade Strålsäkerhetsmyndigheten (SSM) ett projekt för att uppdatera och utvärdera de jordbävningsdata som ligger till grund för de markrörelsespektra, dokumenterade i SKI Technical Report 92:3 (SKI, 1992), som styr hur jordskalvsrisker behandlas vid svenska kärnkraftverk. SSM pekade särskilt på behovet av att inkludera paleoseismiska data, något som inte alls berörs i SKI (1992). I detta projekt har vi sammanställt och uppdaterat seismiskt data och kvalitativt jämfört nya jordskalvsfrekvenser med de i SKI (1992). Det har inte ingått i projektet att kvantitativt utvärdera hur de existerande spektra skulle förändras med ändringar i den bakomliggande jordskalvskatalogen.

Vi har sammanställt en litteraturstudie över postglaciala jordskalv i Sverige (i Appendix 1 i denna rapport) och finner att det finns 12 bekräftade postglaciala förkastningar i Sverige som förskjutit sediment från den senaste istiden. Många av dessa saknar dock stratigrafisk information vilken skulle kunna minska felmarginalerna på uppskattningar av både när skalven inträffade och huruvida de inträffade som enstaka stora händelser eller som flera mindre. Det finns i sammanställningen inga indikationer på att dessa förkastningar skulle ha rört sig upprepade gånger i stora skalv. Vi diskuterar också de paleoseismiska händelser som definierats utifrån spår i lösa sediment, där det inte finns en förkastning på jordytan. Tolkningen av dessa data är som regel är behäftade med stora osäkerheter; om det verkligen är ett skalv som orsakat sedimentstörningarna, när skalvet inträffade, var det hade sitt epicentrum och hur stor magnituden i så fall var. Endast postglaciala skalv med en förkastning synlig på ytan är medtagna i den slutgiltiga listan över svenska postglaciala jordskalv i denna rapport. Till den listan har vi lagt bekräftade postglaciala skalv från Norge och Finland och vi presenterar en uppdaterad tabell över skalvens lokalisering, magnitud och när de inträffade, parametrar som alla behövs för en seismisk riskstudie. Vi noterar att forskningen om postglaciala jordskalv i Fennoskandien utvecklas mycket fort just nu med ny information om kända förkastningar samt indikationer på skalv på nya platser.

Sedan SKI (1992) publicerades har det skett en enorm ökning i antalet jordskalv som registrerats instrumentellt i Fennoskandien, och i kvalitén i bestämningen av skalven. Detta beror framförallt på utbyggnaden och moderniseringen av de nordiska seismiska nätverken. Den samnordiska jordskalvkatalogen Fencat innehåller därför betydligt fler skalv idag än den gjorde vid tiden för datauttaget till SKI (1992). Den historiska delen av Fencat har dessutom uppdaterats med nya bestämningar av lokaliseringar och magnituder för många av de större skalven och rensats på händelser som visat sig vara relaterade till frostknäppar eller sprängningar. Det stora antalet småskalv som registrerats under senare år ger inte bara en god bild av skalvens rumsliga fördelning, det har under senare tid visat sig att de sakta men säkert börjar ge en bild av förkastningssystem i Sverige som är seismiskt aktiva. Genom att jämföra den rumsliga fördelningen av jordskalv i olika tidsperioder i Fencat ser vi att den storskaliga skalvfördelningen i Sverige är relativt stabil över det senaste århundradet. Det finns signifikanta tidsvariationer i datat, både i Sverige som helhet och i olika mindre områden. En stor del av den variationen måste tillskrivas variationer i hur noggrant man samlat in data från allmänheten, och senare i hur instrumenteringen sett ut, vilket gör det svårt att dra slutsatser om geologiskt signifikanta variationer.

De seismiska data som ligger till grund för SKI (1992) är tyvärr inte beskrivna i detalj. Inte vilket område som inkluderats och inte heller vilka magnituder som tagits med i

beräkningarna. Det är därför svårt att göra jämförelser mellan det nu uppdaterade datat och siffrorna i SKI (1992) på skalvaktiviteten per år och ytenhet, och hur aktiviteten varierar för olika stora skalv. För de postglaciala skalven finner vi att dessa inte påverkar resultaten i SKI (1992) om analysmetoden i SKI (1992) används. En modern seismisk riskanalys (s.k. PSHA) skulle dock inte vara komplett utan de postglaciala skalven, och de skulle öka risken. Skalvaktivitet beräknad utifrån det moderna svenska datat (2000-2016) eller från Fencat för ett område som innefattar Sverige, Finland och Östersjön under två tidsperioder (1970-2012, magnitud större än 2, 1875-2012, magnitud större än 3) tyder på att siffrorna i SKI (1992) är konservativt uppskattade för skalv med magnitud 5 eller lägre, men att de skulle kunna underskatta risken för större skalv. Dessa slutsatser vilar dock på ofullständig information om hur SKI (1992) tagit ut sitt seismiska data.

Förbättringar i datainsamling, riskanalys och internationella rekommendationer har gjort att SKI (1992) nu är föråldrad. Vi föreslår man överväger att uppdatera SKI (1992), och betonar att om ett sådant projekt initieras är det viktigt med expertis inom så kallad ”intraplate” seismologi, d.v.s. jordskalv i geologiskt gamla områden långt ifrån plattgränserna. Vi ser att ett flertal viktiga frågor om den Fennoskandiska seismiciteten behöver belysas ytterligare, som t.ex. hur stor var den paleoseismiska aktiviteten, hur stabil är jordskalvsaktiviteten i tid och rum, vilken är den största magnitud vi kan ha i Sverige och hur kan osäkerheter i markrörelseberäkningar kvantifieras på ett rimligt sätt.

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1. Introduction

The Swedish Radiation Safety Authority (SSM) initiated a project with the Seismology group at Uppsala University in 2015 in order to assess and update the earthquake data which form the basis of the currently used envelope ground response spectra for Swedish nuclear power plants in SKI Technical Report 92:3 (SKI, 1992). SSM especially emphasized the need to include information on paleoseismic events, which are not included in the SKI (1992) data. It is not within the scope of this project to quantitatively assess how the existing spectra are affected by the updated earthquake catalog, the project will only perform an initial comparison of the data sets and comment on potential qualitative effects on the earthquake occurrence probabilities.

During the last decade a number of probabilistic seismic hazard analysis (PSHA) projects have been performed for planned or existing nuclear power plants. These projects include for example the PEGASOS Refinement SSHAC Level 4 project in Switzerland (Swissnuclear, 2013), the Thyspunt SSHAC Level 3 project in South Africa (Bommer et al., 2015) and two projects in Finland, one for the planned Fennovoima power plant at Hanhikivi (Korja & Kosonen, 2015) and one for the existing power plants in Lovisa and Olkiluoto (TVO/Fortum project, no public reports). The projects have developed methodology for PSHA and also analyzed problems, and application of procedures, related to various tectonic environments. The Fennovoima and TVO/Fortum projects were collaborative efforts between seismological and geological institutes in Finland, Sweden, and for TVO Estonia, designed to take into account expertise from the countries included in the areas of influence for each site. These two projects also initiated a study to revise the joint Fennoscandian earthquake catalog, Fencat, maintained at the University of Helsinki, in order to remove events that were not earthquakes and homogenize magnitudes. The final results of the Finnish projects are not yet in the public domain, but may be acquired from the companies and could be used as a starting point if SKI (1992) was to be updated.

This report contains a review of the current state of knowledge on Swedish paleoseismic events, information on the revised Fencat catalog and the Swedish earthquake catalog updated to 30 September 2016. We include a section on the possibilities to assess spatial and temporal variations in seismicity, given the sparse early catalogs and the low earth-quake activity rate in Fennoscandia. The updated earthquake data are compared to the SKI (1992) data and we comment on how this may affect the ground motion spectra and how this work could be continued.

2. Brief review of the relevant sections of SKI (1992)

The earthquake data which form the basis for the occurrence rate relations used in SKI (1992) are discussed in Report No. 1 of the six report compilation that is SKI (1992). The fundamental parameters used in the relations are the scalar seismic moment and the hypocentral distance. Report No 1 contains five Appendices, of which Appendices 2-5 discuss the earthquake data underlying the conclusions in Report No 1. As we will refer frequently to these Appendices we denote them SKI (1992) A2 to SKI (1992) A5. Earthquake data for the spectral characteristics and ground motion attenuation with distance are mostly discussed in Report No. 2 of SKI (1992).

2.1. Earthquake occurrence data

The earthquake data in SKI (1992) come mainly from the joint Fennoscandian earthquake catalog Fencat (Ahjos & Uski, 1992; FENCAT, 2016), maintained at the University of Helsinki but contributed to by seismic networks in Sweden, Finland, Norway, Denmark, Estonia and NW Russia. The SKI (1992) analysis includes Fencat data until 1987. As the depth estimates for the Fencat data up until 1987 usually had high uncertainties, SKI (1992) used the earthquake depth distribution from Slunga et al. (1984), based on data from the FOA southern Sweden seismic network of 1980-1984. In addition, SKI (1992) A2 & A3 support the analyses by use of the western Scandinavia data from Ambraseys (1985), analyses by Slunga and collaborators (Norrman & Slunga, 1984; Slunga et al., 1984; Slunga, 1986) and various European and international earthquake catalogs (e.g. Leydecker, 1986; Karnik, 1969,1971; Bulletin of the International Seismological Centre, U.K, and the Preliminary Determination of Epicenters from the U.S. National Earthquake Information Center). When discussing seismicity rates in the appendices it is worth noting that the authors had instructions to focus on the Ringhals and Barsebäck nuclear power plants, so the seismicity analysis did not include earthquakes north of latitude 61°.

In SKI (1992) A2, Slunga argues that the temporal variability in seismicity and the difficulty in predicting where larger earthquakes occur implies that the seismic zones defined around the nuclear power plants should reflect a larger, regional seismicity rate. For Barsebäck he proposed the area in Sweden south of the Tornqvist zone (a major geological boundary east-west through northern Skåne), including Zealand in Denmark. For Ringhals, Slunga proposed to use the seismicity in all of southwestern Sweden west of the Protogine zone (another major geological boundary that runs north-south from Värmland, through Vättern to northern Skåne), including the Kattegatt area north of the Tornqvist zone. There is no analysis of the frequency – magnitude distribution around the nuclear power plants in SKI (1992) A2. However, Slunga presents frequency - magnitude relations for various circular areas around the 1904 Oslofiord magnitude 5.4 earthquake. He notes that the radius has to be at least 500 km for the 1904 event to fall within the (extrapolation) of the frequency-magnitude distribution of the events in the area. There are no results in terms of the Gutenberg-Richter a- and b-values.

Arvidsson et al. in SKI (1992) A3, proposed a different zonation scheme based both on the seismicity and a then recent map of tectonic lineaments and faults in southwestern Sweden (Kornfält & Larsson, 1987). They propose 6-7 seismic source areas in southwestern and southern Sweden, and Denmark. Ringhals would be in a zone along the southwest coast, north of northwestern Skåne. For Barsebäck, it is unclear from SKI (1992) A3 which zone it would belong to, either the zone along the Tornqvist zone, or the zone comprising Denmark. There is no analysis of the frequency – magnitude distribution in SKI (1992) A3.

In SKI (1992) A4, Skordas & Wahlström analyze frequency – magnitude distributions for Barsebäck and Ringhals, using circular areas centered on the power plants with radii 50, 100, 150 and 200 km. The data used come from the merged Arvidsson et al. catalog in SKI (1992) A3 (it is not explicitly stated but probably refers to Appendices I and II in Arvidsson et al.) and comprises events from 1497 to 1985. It is unclear if the study uses epicentral or hypocentral distances. Skordas & Wahlström use both a linearized least-squares method and the Aki (1965) maximum likelihood method to estimate the a- and b-values of the Gutenberg-Richter distribution. The numbers stabilize for radii of 100 km or more, when there are enough events in the analyses. The maximum likelihood method gives b-values of approximately 0.9 – 1.1 for Barsebäck and for Ringhals 1.0 – 1.1.

As is evident from the above, SKI (1992) A2-A4 give three different suggestions as to how zonation around the power plants should be done. In the main text of Report No. 1, none of these schemes were eventually used. Instead the main report compares the various frequency-magnitude estimates from the appendices and other studies (Norrman & Slunga, 1984; Ambraseys, 1985) and adopts a more conservative "large-scale Fennoscandian average" frequency-magnitude (actually a frequency – scalar seismic moment) relation, referred to as the "seismicity function". It is not clear from SKI (1992) which data or region were actually used for the calculation of the epicentral density used in the seismicity function. The report states that averaging over hundreds of thousands to millions of square kilometers can produce a reliable assessment of the epicentral density for magnitudes up to approximately 5. The report also states that the seismicity function "is basically determined on the basis of earthquake observation data extracted from Fencat (1987) […] comprising 733 events in Scandinavia". This number is the same as that quoted in Arvidsson et al. in SKI (1992) A3 for their Fencat data for only Sweden, south of latitude 61°. It seems, however, unlikely that this very restricted data set is the basis for the seismicity function as the number of large earthquakes is very small (five events with local magnitude above 4.5), it does not include Norwegian nor Danish seismicity, not even the large 1904 event, and the frequency-moment comparisons in Figure 7 of Report No. 1 indicate that a much larger data set was used to produce the seismicity function.

The seismicity function has a fixed point ("is anchored" in the language of SKI (1992)) at seismic moment $M_0 = 10^{15}$ Nm. The number of earthquakes per square kilometer and year with seismic moments which exceed $M_0 = 10^{15}$ Nm is referred to as the epicentral occurrence density N_A and is anchored at $N_A(M_0 = 10^{15}) = 1.5 \cdot 10^{-7}$. As discussed above, it is unclear how this number was obtained and which area was used. Since SKI (1992) uses the scalar seismic moment instead of a magnitude, magnitudes from Fencat has been converted to scalar seismic moments. This has been done using the relationships between macroseismic magnitudes, local magnitudes and scalar seismic moment outlined in Slunga et al. (1984). Unfortunately, there is no analytical expression in Slunga et al. (1984) for the relationships used between macroseismic magnitude, local magnitude (Wahlström, 1979) and seismic moment for seismic moments larger than 10^{13} Nm. We have only the two graphs of the relationships provided in Appendix 2 of Slunga et al. (1984), making it difficult to repeat the calculations in SKI (1992). Reading off the graphs, the SKI (1992) anchoring moment corresponds approximately to local magnitude 4.7 and macroseismic magnitude 5.0. We note that on the moment magnitude scale (Hanks & Kanamori, 1979), used nowadays for larger magnitude events, the anchoring moment corresponds to M_W 3.9.

The frequency – moment function adopted in SKI (1992) is: $\log N_A = a - b' \cdot \log M_0$ where the b'-value is different from the usual frequency – magnitude b-value as it has to incorporate the relation between the magnitude scale and $\log M_0$. SKI (1992) uses a b' of 0.87. How this value was obtained is not discussed. Figure 7 of Report No 1 compares various frequency-moment distributions from Appendices 2 and 4 and from other studies. The b' of 0.87 may just be an adaption to these results. SKI (1992) does not incorporate events with moments larger than $3 \cdot 10^{18}$ Nm, corresponding to a local magnitude of approximately 6.5 and a moment magnitude of 6.3. Such a "cut-off" is referred to as the maximum possible magnitude in the area of interest, M_{max}, in current earthquake hazard analyses. SKI (1992) notes that the cut-off does not affect the statistical result "very much", as the contribution from events with moment larger than 10^{18} Nm is "fairly small".

In order to produce estimates of the occurrence rates of different magnitude earthquakes at different distances from the power plants, SKI (1992) uses the hypocentral distance and integrates over crustal volumes at various distance intervals. Although not explicitly stated in the text, Figure 1 in Report No. 1 indicates that the seismogenic crustal thickness is assumed to be 35 km. This figure also defines the hypocentral depth distribution and distance intervals used in the calculations of the tabulated values in SKI (1992) A1.

2.2. Ground motion data

As data on ground motion characteristics from larger earthquakes in Fennoscandia were very scarce in the late 1980s, and still are, SKI (1992) turned to the so called Standard Response Spectra for rock sites in Japan (Hisada et al., 1978; Ohsaki, 1979; Watabe & Tohdo, 1979; Katayama, 1982; Ohta et al., 1983). These are about 300 acceleration spectra with mathematical models for the description of the spectra in terms of e.g. Peak Ground Velocity (PGV) or spectral velocity (Sv). As the spectra come from Japanese earthquakes in a plate boundary setting where both the earthquake source functions and the wave propagation effects can be markedly different from those in the Fennoscandian intraplate environment, SKI (1992) identified three main factors which cause significant differences between Japan and Fennoscandia; (i) the earthquake stress drop, (ii) the fault area and associated strong motion duration, (iii) the anelastic attenuation along the wave path. All these three effects were analyzed and the spectra modified accordingly.

The Japanese magnitudes, M, were converted to scalar seismic moments using the relationship: $\log M_0 = 9.1 + 1.5 \cdot M$, in accordance with Slunga (SKI (1992) A5).

The modeled response spectra for Swedish hard rock sites were compared to data from three earthquakes in the Eastern USA and Canada. The comparisons indicate that the Swedish response spectra agree fairly well with the Eastern North America data at frequencies below approximately 10 Hz, but that they underestimate the response at higher frequencies.

3. Update on Fennoscandian seismicity

The 24 years that have passed since the publication of SKI (1992) has seen a significant increase in data and knowledge on Fennoscandian seismicity. The evolution of seismic instrumentation, data acquisition and processing power has made possible a large expansion of the Nordic seismic networks, especially in Sweden where the Swedish National Seismic Network (SNSN) grew from five analogue and one digital seismic station in 1997 to 65 modern digital broadband stations in 2012. The denser seismic networks now detect much smaller events and thus the number of analyzed earthquakes has grown rapidly. From the year 1497 to August 2000, when the SNSN entered automatic network processing, there are approximately 1,400 earthquakes in Fencat within the Swedish territory. From August 2000 to September 2016, the SNSN has recorded almost 7,900 earthquakes within Sweden, increasing the database by more than a factor 5. The implications of the new instrumental data are discussed in Section 3.3 below.

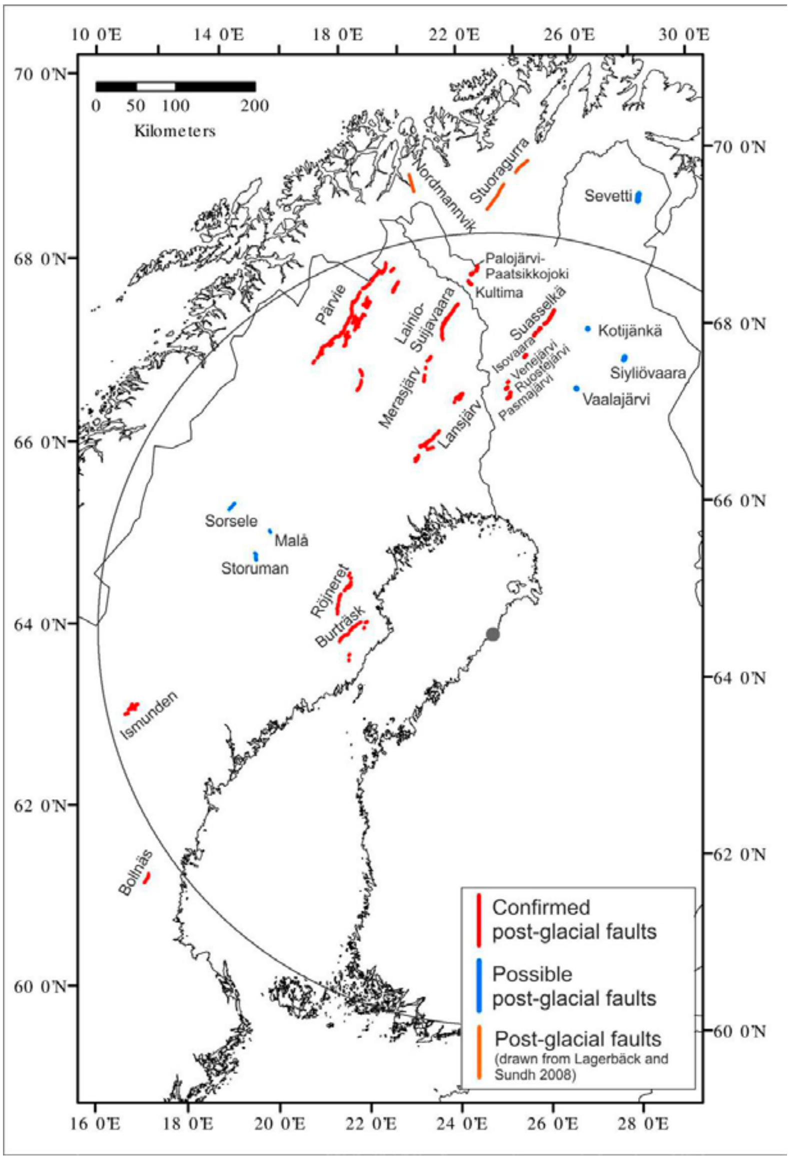
A second strong trend in the last decade or so has been a renewed interest in postglacial, or endglacial, faulting. As will be reviewed in Section 3.1, new Lidar (Light detection and ranging) data have revolutionized geomorphological studies and made possible a new way to identify and analyze postglacial faults. In addition, reflection seismic studies, microearthquake studies, InSAR and a host of geophysical studies have been performed over some of the faults, see below.

3.1. Paleoseismicity

Quoting Bolton (2015), "The aim of paleoseismology is to generate a record of past earthquakes (i.e. magnitude, recurrence interval, timing, etc.) from a range of geological observables preserved within a landscape. This is achieved by identifying features associated with a single paleoearthquake as opposed to the long-term deformation along a fault or within a basin." As the written historical earthquake record in Fennoscandia, which goes back approximately 500 years, contains very few large earthquakes, paleoseismology could potentially be a powerful tool to complement the data on occurrences of large events. However, in contrast to more seismically active areas, faults suitable for trenching that have been seismically active in the last few thousand years are difficult to identify in Fennoscandia, due to the low seismicity rate. There is one notable exception to this, the large post- or end-glacial fault scarps mostly found in northern Fennoscandia, see Figure 1.

In Appendix 1 of this report we review the current knowledge on post-glacial paleoseismic earthquakes in Sweden. The conclusions of the review are:

"According to the literature, there are twelve scarps in Sweden that appear to cross cut glacial sediments. These 12 features often include multiple segments and complex geometries. Most of these structures are in the northern part of the country, but examination of LiDAR-derived imagery has revealed previously unknown scarps in central Sweden. For all of these features, fault rupture is interpreted to have occurred around the time of deglaciation generally between 10,500 and 9,500 years before present. Magnitude estimates for the seismic events associated with fault ruptures range from as low as Mw=6.1 to as high as Mw=8.2 [range for the different faults]. Review of the



literature has also revealed the complete lack of stratigraphic information relating to most

Figure 1. Copy of Figure 3.2.3.1 from Korja & Kosonen (2015). Inventory of post-glacial fault scarps in Fennoscandia. Note that since the publication of this Figure, the Nordmannvikdalen fault has been reclassified and is no longer considered to be tectonic (Redfield & Hermanns, 2016).

of the scarps. Stratigraphic information would not only help constrain the timing of fault rupture, but also provide information critical to magnitude calculations. The assumption that each scarp formed as a result of a single event remains untested on six of the ten scarps.

Despite the vast body of literature related to proposed paleoseismicity in Sweden in the absence of surface rupture, great uncertainty surrounds many of these claims. Sediments may be disturbed in a number of different ways in a glacial environment, and they do not necessarily indicate paleoseismicity. The assigning of ages and magnitudes to proposed paleoseismic events defined by disturbed sediments is often ad hoc or attained through misuse of published empirical data. Without the presence of a scarp, significant uncertainties exist regarding the location (ie occurrence), timing, and magnitude of proposed paleoseismic events. Thus, they are excluded from the current inventory."

There is little evidence for paleoearthquakes in the time period between the apparent burst of seismicity occurring as the ice sheet disappeared, 11,000 – 9,500 years before present (BP), and the start of written historical records. Mörner (2004) reports five events, based on soft sediment disturbances, with associated magnitudes between the time of the Pärvie rupture (~9,500 years BP) and 5,000 radiocarbon years BP. In addition, Mörner (2009) reports an additional eleven events between 4,800 radiocarbon years BP and 900 radiocarbon years BP, stating however that only two of these are "recorded by multiple factors and firmly dated." One of the two was later assessed by Gregersen and Voss (2014) as an unlikely earthquake. As the uncertainties in occurrence, location, timing and magnitude of these soft sediment based events may be significant, as pointed out in Appendix 1, we will not include them in the paleoseismic catalog here until they are more thoroughly investigated.

For the purpose of this report, i.e. assessing how paleoseismicity may affect the seismic hazard analysis in SKI (1992), the events need to have an associated magnitude. Of the twelve paleoearthquakes identified in Appendix 1, only eight have reported magnitudes. These are listed in Table 1, taken from Table 1 in Appendix 1. Of the four paleoearthquakes without magnitude estimates in Appendix 1, two have fault scarp lengths of approximately 40 km and two have shorter scarps (11 and 17 km), see Table 1 in Appendix 1. Depending on the depth of the rupturing fault and the amount of slip in the event, the magnitudes of these events are probably in the 6-7 span. The apparent mismatch between the magnitude of the Pärvie event in Table 1, and the magnitude range given in the conclusions quoted above comes from an earlier magnitude estimate for Pärvie, discussed in Appendix 1. As pointed out in Appendix 1, there are significant uncertainties associated with the magnitude estimates for the endglacial events. The estimates are usually based either on statistical relationships between surface rupture or fault offset and magnitude (e.g. Leonard, 2010) or the moment magnitude definition from the scalar seismic moment, which is estimated from surface rupture length, width of the fault plane from the seismogenic thickness of the crust and an average slip often taken as the observed surface offset. In addition, the faults are always assumed to have ruptured the full extent at one instance in time. There are no indications in the literature of observed multiple large ruptures of the same endglacial fault. Many of the assumptions needed for a magnitude estimate are not as strongly based in the data as would be desired, and therefore the magnitude estimates are often worst case scenarios. With the review in Appendix 1, we conclude that during a time span of up to 1,500 years around the disappearance of the ice sheet, Sweden experienced at least a dozen earthquakes of magnitudes 6 – 8.

Large endglacial faulting also occurred in Norway and Finland, see recent reviews in Lund (2015) and Korja & Kosonen (2015). In Norway, the NEONOR project (Olesen et al., 2013) investigated a large number of claims of neotectonism and concluded that only two endglacial faults could be identified with certainty, the Stuoragurra fault in Finnmark and the Nordmanvikdalen fault in Troms. Recently, the Nordmanvikdalen fault has been reclassified as non-tectonic (Redfield & Hermanns, 2016). In Finland, recent Lidar investigations have revealed a number of new endglacial faults and extended the length and complexity of some of the previously known faults (e.g. Sutinen et al., 2014; Korja &

Kosonen, 2015). In Table 1 we have included those faults which have been confirmed as postglacial and have a magnitude estimate documented in the literature.

A few claims of end- or postglacial faulting have been made in Denmark and northern Germany. In Denmark, Sandersen & Jörgensen (2015) used Lidar, borehole and airborne electromagnetic data to analyze irregularities on the Tinglev outwash plain in southwestern Denmark. They interpret these as Holocene strike-slip movements along graben faults, but do not identify the causative faults and can therefore not provide location or magnitude of the proposed events. In Germany, Brandes et al. (2012) identified meter-scale faults in a Pleistocene age alluvial-aeolian sand complex and inferred that movement on these was due to one or more earthquakes on the nearby Osning Thrust fault system, some 16,000 – 13,000 years before present. The causative earthquake locations were, however, not identified and therefore no magnitudes could be estimated.

Name	Country	Central latitude	Central longitude	Timing [years before present]	Moment magnitude	Reference
Pärvie	Sweden	67.93	19.28	9,500	8.0	Appendix 1
Lainio	Sweden	67.98	22.32	11,000 – 10,000	7.1	Appendix 1
Merasjärvi	Sweden	67.53	22.00	11,000 – 10,000	6.3	Appendix 1
Lansjärv	Sweden	66.59	22.11	10,500 – 10,390	7.8	Appendix 1
Röjnoret	Sweden	64.78	20.12	11,000 – 10,000	7.1	Appendix 1
Burträsk	Sweden	64.42	20.53	11,000 – 10,000	7.1	Appendix 1
Lillsjöhögen & Ismunden	Sweden	63.18	15.16	After deglaciation < 10,000	7.0	Appendix 1
Bollnäs	Sweden	61.33	16.35	10,670 – 10,200	6.1	Appendix 1
Stuoragurra	Norway	69.54	23.91	postglacial	7.3	Olesen et al. (2013)
Suasselkä	Finland	67.97	25.30	postglacial	7.0	Kujansuu (1964), Olesen et al. (2013)
Pasmajärvi/ Ruokojärvi/ Venejärvi	Finland	67.26	24.16	Postglacial	6.5	Kujansuu (1964), Olesen et al. (2013)

Table 1. Large paleoseismic earthquakes in Sweden, Norway and Finland with associated location, timing and magnitude. Coordinates give the approximate midpoints of the faults, see Figure 1.

As alluded to above, research on endglacial faulting has been significantly revived in the last decade. New faults have been found (e.g. Appendix 1) and some old faults have been reclassified as probably not end- or postglacial (e.g. Olesen et al., 2013; Redfield & Hermanns, 2016). Doubts on the estimated magnitudes come from e.g. Lidar data which indicate that perhaps the Pärvie fault did not rupture all at once (Appendix 1) and new microearthquake data from Burträsk which indicate that the fault may not have ruptured through the entire seismogenic crust (Lund et al., 2015). More geophysical techniques such as reflection seismics (e.g. Ahmadi et al., 2015), magnetics and gravity (e.g. Malehmir et al., 2016), lake bottom mapping (Vogel et al., 2013) and electromagnetics (Kamm et al., 2016) are being applied to the faults to determine their extent and orientation, and satellite techniques such as InSAR have been used to try to infer current surface motion (Mantovani et al., 2013). A deep drilling project on the Pärvie fault (DAFNE) is in the planning stages in order to probe the fault at depth (Kukkonen at al., 2011). In addition, the last decade has seen renewed modeling initiatives in order to better understand the mechanics of the process of endglacial faulting (e.g. Lund et al., 2009; Steffen et al., 2014). It is likely that our knowledge of endglacial faulting will continue to expand in the near future, and that this will increase our understanding of the potential hazards that such faulting, and the current activity on the faults, pose to our society.

3.2. Development of Fencat

The current version of the Fencat joint Nordic earthquake catalog contains events from 1375 up until the end of 2012 and has a total of 21,385 events, 4,582 of which occurred prior to 1 January 1988, see Figure 2. As SKI (1992) does not explicitly describe the data set used we do not know how many of the latter events were part of the 1987 Fencat edition used for the seismicity function. SKI (1992) probably extracted data in some specific region, with some specific quality and magnitude thresholds.

The earthquake data in Fencat come from a wide variety of sources. The early data are assembled from publications by natural scientists, often on other topics, and from church records in Sweden and Finland which from 1749 were required to note information also on odd and unusual events. The establishment of the Royal Swedish Academy of Sciences in 1739 created a forum for notices and discussions that included earthquake phenomena but it was not until the end of the 19th century that systematic collection of earthquake information in Sweden was initiated by the Swedish Geological Society. This function passed to the Geological Survey of Sweden after the large earthquake of 1904. The other Nordic countries saw similar developments.

Instrumental recording of earthquakes in the Nordic countries started with the installation of a Wiechert long-period seismograph in Uppsala in 1904. This was followed by seismographs in Bergen in 1905 and Helsinki in 1924. These first seismographs were not very sensitive to local or regional earthquakes, and from the records of felt earthquakes it seems that even earthquakes with magnitudes up to 4 in the Nordic countries were sometimes not recorded by these instruments. This is, however, a point for further research. Seismic networks that were sensitive enough to pick up most of the felt events were not in place until the late 1960s, early 1970s for all of Fennoscandia. Note that the earthquake maps in Figure 2 only contain Fencat data, implying that the lack of events in northern Germany and Poland is incorrect. Incorporating data from a European wide catalog is necessary to assess seismicity there. It is unclear how this was addressed in SKI (1992).

The long time span implies that a number of different magnitude scales have been used over the years. Definitions of macroseismic scales, based on reports of felt ground motion and intensity estimates (also on different scales) have varied over the years. The instrumental period has not alleviated the problem, there are a number of different instrumental local magnitude scales in Fencat. Seismic hazard studies require a uniform magnitude definition for the events used, which makes it necessary to homogenize the magnitudes in the catalog. As discussed above, in SKI (1992) this was done by conversion to a seismic moment scale.

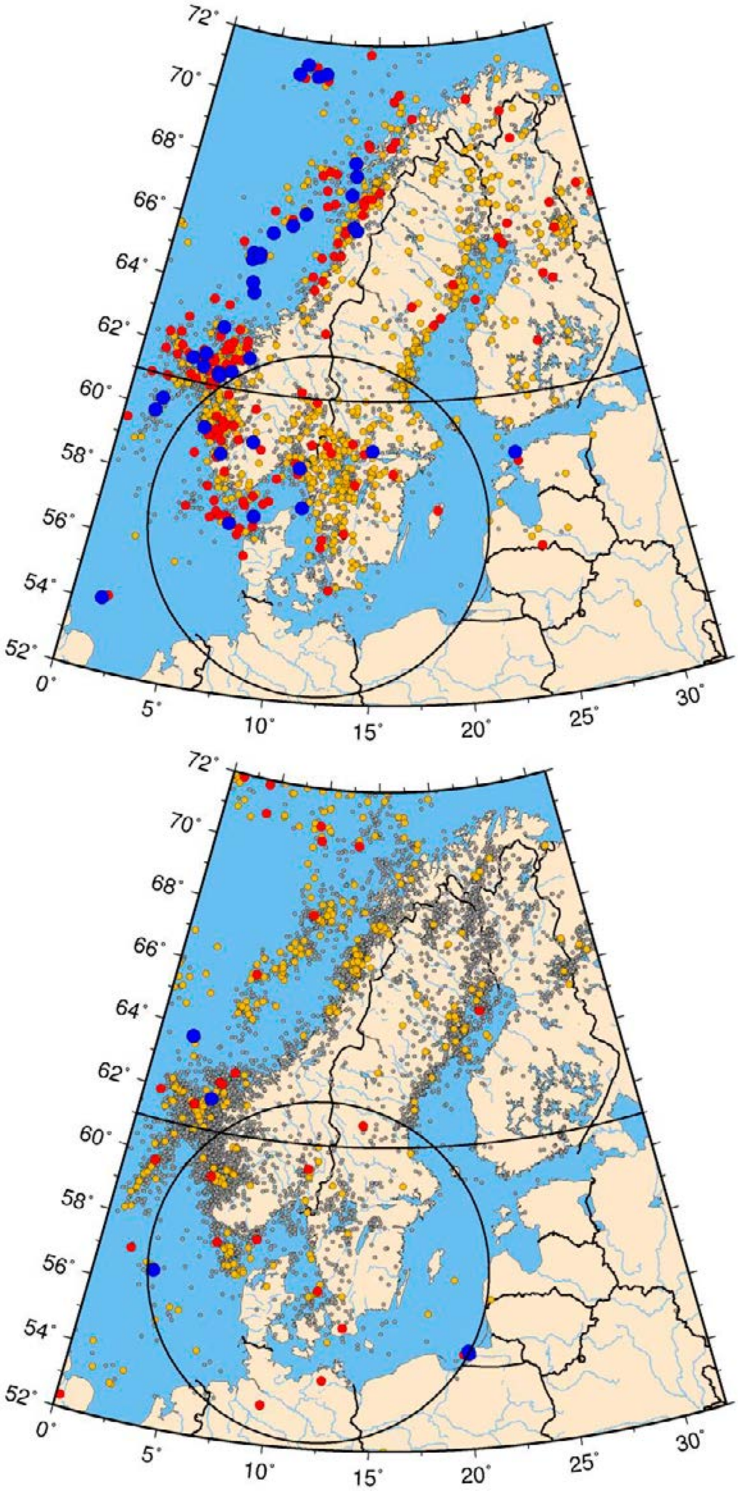


Figure 2. Map of earthquakes from Fencat. Small grey circles, magnitude M < 3, yellow circles, 3 <= M < 4, red circles 4 <= M < 4.6(5.0) and large blue circles correspond to magnitudes with equivalent seismic moment M_0 >= 10^{15} Nm, see text for details. The line

at N61° indicates the cut-off used in SKI (1992) and the circle shows a one million square kilometer circular area around Ringhals. Top) Events prior to 1/1 - 1988. Bottom) Events between 1/1 – 1988 and 31/12-2012, plus the 2014-09-15 Sveg (M4.1) and 2016-03-19 Bottenviken (M4.1) events.

The recent seismic hazard projects in Finland (e.g. Korja & Kosonen, 2015) initiated a review of Fencat for the relevant areas (500 km and 300 km radius circles centered on the locations of the future and current nuclear power plants, respectively) and a new effort of homogenization of the magnitudes. This process is now complete for earthquakes in Finland but not yet for all of Sweden. For Norway and Denmark the process has only begun. For Sweden and Finland we have returned to the references where the early data were compiled during the late 19^{th} and early 20^{th} centuries in an attempt to identify events that have erroneously been classified as earthquakes. We have also reviewed particular issues such as proximity to known blast sites (mines, quarries, military etc) and temperature. One problem in the data prior to about 1920 is a strong seasonal signal, see Figure 3A. The prime suspect for that is frostquakes, cracking of the frozen ground with accompanying loud noise and sometimes also shaking. We acquired weather data from the Swedish and Finnish Meteorological Institutes for a number of locations, going back to approximately 1845. We also received the Uppsala temperature series, which starts in 1722. For some time periods we also had snow cover data, which aids the analysis since thick snow cover insulates the ground and prevents frost cracking. For Sweden in the time period 1904 – 1965 we have only investigated events with estimated magnitude over 3.5. We studied temperature fluctuations around the dates of the earthquakes and could identify a number of events which correlated with a rapid decrease in temperature from around zero to below -10. In some instances these “events” were also accompanied by cracking of the ground, as noted in the original descriptions. The events remain in Fencat, but are marked as doubtful. Most of the frost related events are small, but our analysis of the 2-4 January 1894 events indicates that the mainshock, assigned macroseismic magnitude 5.1, is actually widespread frostquakes in south-central Sweden and southern Finland. It is extremely unlikely that a magnitude 5.1 earthquake in Sweden would cause cracks in the ground in Finland. After removal of the frost and blast related events, the seasonal distribution can be seen in Figure 3B. We see that there is still an obvious seasonal signal, indicating that it is likely that there are more frost quakes in the data, but that this signal is now significantly reduced as compared to Figure 3A. In Figure 3C we show instrumentally recorded data from Fencat for 1965 – 2012, and we see that the seasonal signal is not present in that data.

During the investigation we also corrected the dates for the early events where the conversion from the Swedish version of the Julian calendar to the standard Gregorian calendar had not been properly performed. We also marked as doubtful some events which are likely to be mining induced, or rock bursts, and some events which are likely to be blasts from Navy operations.

The magnitudes of small historical earthquakes are difficult to assess and often rely on only a few observations in a small area, leading to considerable uncertainty. Not even for the very largest earthquakes is it always straightforward to assign a magnitude, as it very much depends on the reliability and interpretation of reports of shaking. Recent studies of the 1759 Kattegatt and the 1819 Lurøy earthquakes, probably the largest in Fennoscandia, have caused a significant debate on the interpretation of both distant reports of shaking and nearby reports of non-shaking (e.g. Huseby & Kebeasy, 2004, 2005; Wahlström, 2004; Bungum & Olesen, 2005). The magnitude estimate for e.g. the Lurøy event varies from M_S 5.1 to M_S 5.8. Since the time of data collection for SKI (1992) in 1987 there has been a number of studies reevaluating older magnitudes, such as Muir Wood & Wu

(1987) for Norwegian events, Bungum et al. (2009) for the 1904 Oslofiord event and the studies referred to above.

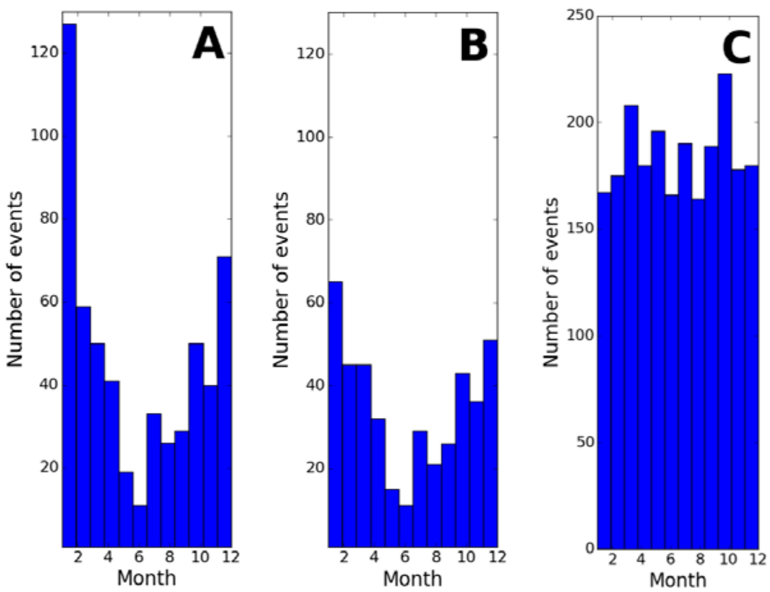


Figure 3. Seasonality in the earthquake data in Fencat. Number of events versus month. A) All events in Fencat 1375 - 1920. B) Events considered to be frost, weather and blast related removed from 1375-1920. C) Instrumental earthquake recordings 1965-2012 (note the different scale).

Summarizing, a revision of the Fencat catalog has been performed for Finland and Sweden from the first entry to 1903. For Finland, the catalog has been cleaned up to the most current entry, while for Sweden only events with magnitude larger than 3.5 have been checked between 1904 and the present. The smaller events will be investigated in the near future. A number of events with magnitude greater than 4 have been considered doubtful during this process, most importantly the 1894 event with macroseismic magnitude 5.1 (corresponding to M_0 just above 10^{15} Nm, using the graphs in Appendix 2 of Slunga et al. (1984)), which probably has direct implications for the SKI (1992) anchoring point considering the low rate of earthquakes of that size. The revisions imply that Fencat is soon ready for a modern study of earthquake hazard in the Baltic Shield. However, since the Norwegian and Danish catalogs are still under revision, a study of seismic hazard to Swedish nuclear power plants, using a standard circular area with 500 km radius, will require further scrutiny of the relevant Norwegian and Danish entries in Fencat.

We see clearly in Figure 2 how seismic activity varies spatially in Fennoscandia and how most of the larger magnitude events are associated with the Norwegian west coast region. We may not have been able to reproduce the exact same magnitude/moment homogenization scheme as SKI (1992), the blue circles in Figure 2 show events with our estimated moment larger than 10^{15} Nm. However, it is unlikely that SKI (1992) would have a significantly different number, or distribution, of these events. Figure 2 then shows how important it is to properly define which area is included in the calculation of the "seismicity function". We also note that there are very few larger magnitude events in the

Baltic Shield (i.e. the region from the Swedish-Norwegian Caledonian mountain range to western Russia), four events prior to 1988 and two events after. Looking more in detail at the post-1987 map in Figure 2, we see the two, surprising, earthquakes with moment magnitudes around 5 in Kaliningrad in 2004. These occurred unexpectedly in an area which has seen very little seismicity in the last centuries (Gregersen et al., 2007). The blue circle in the North Sea west of Jutland is an event in the Ekofisk oil field in 2001 induced by water injection (Ottemöller et al., 2005) and it would therefore be excluded as man-made in a seismic hazard analysis. The remaining two larger events belong to the Norwegian west coast seismicity and occurred in 1988 and 1989.

The Fennovoima (Korja & Kosonen, 2015) and TVO/Fortum seismic hazard projects used the Fencat earthquake catalog for the PSHA. Just as in SKI (1992) the magnitudes had to be homogenized but this was done not by conversion to scalar seismic moments but instead to moment magnitudes. The homogenization process took advantage of the seismic moments routinely calculated by the SNSN and then added published information on moments for Fennoscandian events, usually for single larger events. The homogenization process is based on the local Helsinki magnitude scale, and the result is a piece-wise linear function between seismic moment and magnitude, with one linear relationship for moments below $\log(M_0) = 13.5$ and another linear relationship for larger moments. The details of this is not yet in the public domain. As the homogenization process was performed primarily for events within the zones of interest for these two studies, it is not possible to directly apply to all event of interest for a similar Swedish study, as the Norwegian events would have to be investigated in more detail first.

3.3. Earthquakes in Sweden 2000 – 2016

The modernization and expansion of the Swedish National Seismic Network (SNSN) from 1998 to 2012 (Bödvarsson & Lund, 2003; Bödvarsson et al., 2006) has provided a significantly more detailed picture of earthquake activity in Sweden than was previously available. Figure 4A shows the current station network and Figure 4B shows the seismicity. Comparing Figures 2A and 4B, we note that the large scale features are the same, the northeast coast and the Vänern areas have the highest seismicity whereas the mountain range and the southeast have very low seismicity. Perhaps the most striking difference between the two maps is the clearly mapped seismicity along the endglacial fault scarps in northern Sweden. Pärvie, Lainio, Merasjärvi, Lansjärv and the Burträsk faults are all seismically active and the events occur along and southeast of the fault scarps, as expected from the inferred reverse mechanisms, with faults dipping to the southeast. Although the large ruptures are inferred to have occurred some 10,000 years ago, the faults are still active. Other pronounced areas of earthquake clustering are the events in a north-south extension in the Bay of Bothnia, and a northeast-southwest lineament of seismicity from north of Hudiksvall to Arbrå, which has no obvious geologic structure associated with it. Since the SNSN is able to record very small earthquakes, down to magnitude -1, we start to see more activity on some structures where earthquakes have been previously absent, or very rare. One such location is on a branch of the Tornqvist zone, on land from the Bjäre peninsula eastward along Hallandsåsen, which potentially links the area of magnitude 4+ events in Kattegatt with an onland extension. Other locations are the weak line of earthquakes coincident with the Lillsjöhögen/Ismunden endglacial fault and the line south of the Burträsk fault.

During the time period of the modern SNSN four events with magnitude larger than 4 have occurred in, or in close vicinity to, Sweden: the 2008 M4.3 Lund, the 2012 M4.1 Kattegatt, the 2014 M4.1 Sveg and the 2016 M4.1 Bottenviken events. This is an interesting temporal clustering of large events, as no earthquake with magnitude above 4 occurred between 1986 and 2008 in Sweden. Incidentally, a similar cluster of four M4+ events occurred between 1983 and 1986. Temporal clustering thus seems to be a recurring phenomenon and these temporal clusters and the long hiatus in between strongly suggests that the occurrence of larger events is not stationary in Sweden. In SKI (1992) A2, Slunga points out that two of the largest events in Fennoscandia, the 1819 Lurøy and 1904 Oslofiord events, occurred in areas which did not have a high prior seismic activity. Although significantly smaller, the recent M4+ events follow a similar trend. The Lund and Kattegatt events may be associated with the geologically very significant Tornqvist zone, but there is not much prior seismicity there, see Figures 2 and 4. The Bottenviken event occurred in a more seismically active region, whereas the Sveg event occurred in an area almost void of earthquake activity, Figures 2 and 4.

Since the modernization of the SNSN in 1998 local magnitudes are calculated using a seismic moment based scale derived from the work of Slunga et al. (1984). In Figure 5 we show the frequency-moment curve for earthquakes occurring within the SNSN network (and azimuthal gap less than 180 degrees) between 2000 – 2016. In addition, all four M4+ events are in Figure 5, in-spite of the fact that not all of them have an azimuthal gap of less than 180 degrees. The number of events has been normalized to the area covered by the outermost stations of the SNSN, 465,983 km^2 and the 16 years of operation of the modern SNSN. It should be noted that the SNSN expansion took place over many years, starting in 2000 and finishing in 2012, which distorts the areal normalization. This has not been taken into account here. The b'-value (using the nomenclature of SKI (1992)) for the SNSN data is 0.77, estimated after Aki (1956) and Marzocchi & Sandri (2003) and with a formal uncertainty of 0.02. In Figure 5 we have also added the seismicity function of SKI (1992). We see that the frequency-moment distribution lacks events with moment larger than 10^{15} Nm, and that there is a deficiency of events with moment higher than about 10^{13} Nm. There is also a bending of the curve starting just below 10^{12} Nm, which is similar to the bend in Slunga's curves in Figure 7 in SKI (1992) Report No. 1. The SNSN curve cuts 10^{15} Nm at $8.8 \cdot 10^{-8}$ events per square kilometer and year, below the SKI (1992) seismicity function at $1.5 \cdot 10^{-7}$. However, due to the lower b'-value the SNSN function predicts a larger number of events with moment above 10^{17} Nm. The formal uncertainty of the b'-value is very small, due to the large number of events in the calculation. As is evident from Figure 5, that uncertainty is not appropriate to describe the uncertainty associated with the SNSN frequency-moment distribution for extrapolation beyond the data, and not even for the larger events in the data, as the distribution is not linear. The non-linearity could be due to the short period of observation, methodological problems in assessing the moment for large and small events consistently or actual physical differences in the faulting mechanisms.

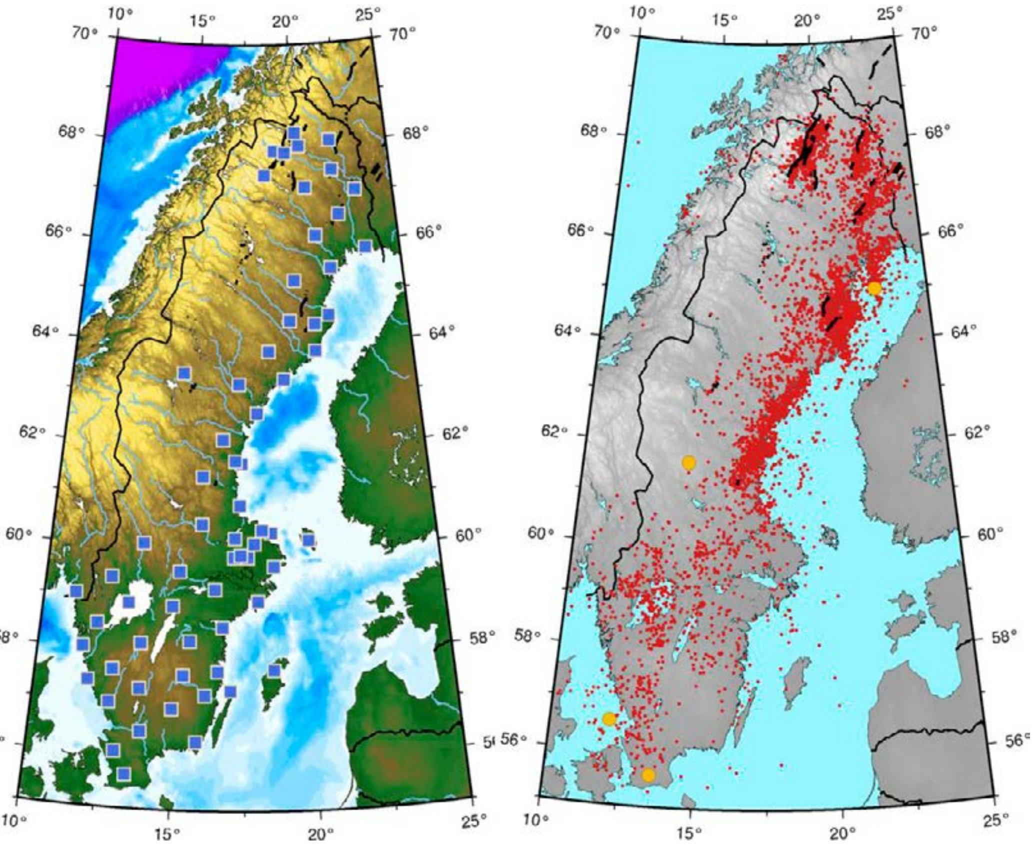


Figure 4. Left) Stations in the Swedish National Seismic Network (SNSN). Right) Seismicity recorded by the SNSN between 2000 and 2016 (red circles), events with magnitude 4 or larger (orange circles). Black lines show endglacial fault scarps.

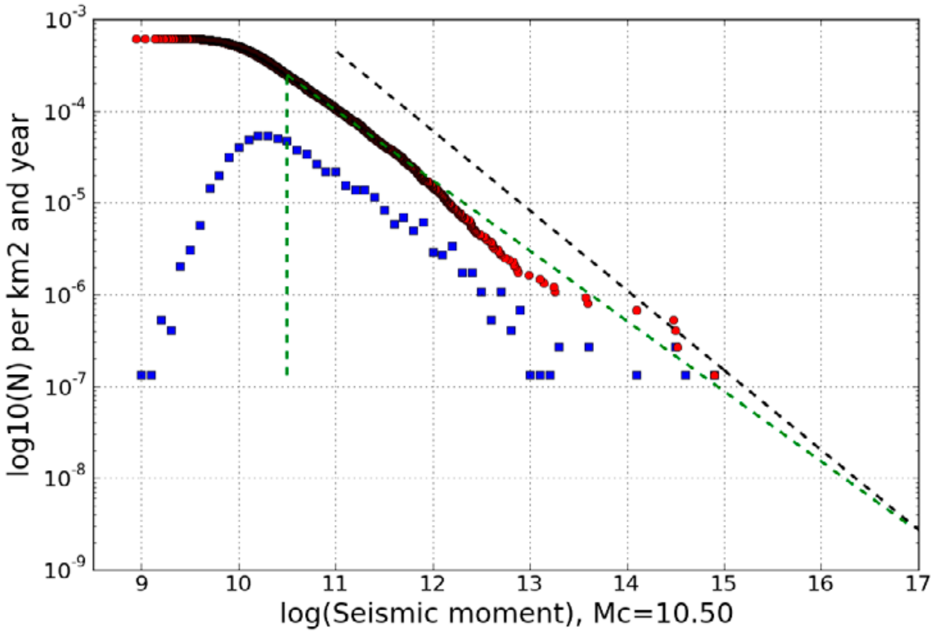


Figure 5. Frequency – seismic moment distribution of the events recorded by SNSN between 2000 – 2016 which have an azimuthal gap of less than 180°. Cumulative numbers (red circles), histogram (blue squares), the green dashed lines shows the "moment of completeness", Mc = log(M_0) = 10.5, and the b' = 0.77 ± 0.02 line. The black dashed line shows the seismicity function from SKI (1992), b' = 0.87.

4. Temporal and spatial variations in seismicity

At plate boundaries, large earthquakes tend to occur in relatively well defined zones in the vicinity of the boundaries. Just as important is that large earthquakes tend to recur in the same zones. This is in contrast to intraplate seismicity where there is still considerable uncertainty as to where large earthquakes occur and whether or not they tend to recur in similar locations. Much interest in intraplate seismicity has been fueled by the New Madrid earthquakes in central USA (e.g. Stein et al., 2009), and the efforts to estimate seismic hazard in the region. With even the most recent GPS networks being unable to detect strain accumulation in the New Madrid region, Stein et al. (2009) propose that intraplate seismogenic faults interact in a complex system which cannot be understood by analyzing an individual fault, and that hazard assessments focusing on recent seismicity therefore may overestimate the risk in one region and underestimate it in another. The model has been proposed also for the migrating system of large events in northern China (Liu et al., 2011), where in the last 2000 years no large earthquake has occurred on the same fault twice. Calais et al. (2016) go one step further and propose that the concept of recurrence for large intraplate earthquakes may even be incorrect. Instead they argue that earthquakes in stable continental regions (SCR) are better explained by transient perturbations of local stress or fault strength that release elastic energy from a prestressed lithosphere. As a result, they propose that SCR earthquakes can occur in regions with no previous seismicity and no surface evidence for strain accumulation.

In Fennoscandia, the crustal deformation field is dominated by glacial isostatic adjustment (GIA) both in the vertical and horizontal directions (e.g. Lidberg et al., 2010; Kierulf et al., 2014). Any estimate of a tectonic signal in the strain rate field must therefore attempt to remove the GIA affect, which is very difficult within the uncertainty limits of the GPS signal (e.g. Scherneck et al., 2010; Keiding et al., 2015). It has not been possible to estimate strain accumulation on individual fault systems in Fennoscandia, which makes it very difficult for quantitative models of temporal and/or spatial variations in seismicity. Stress measurements and focal mechanisms in Sweden and Finland generally show strike-slip to reverse faulting conditions (e.g. Slunga, 1991; Lund & Zoback, 1999; Uski et al., 2003, 2006; Heidbach et al., 2008) with the maximum horizontal stress directed approximately NW-SE. This has led a number of authors to conclude that the Fennoscandian stress field is dominated by ridge-push (e.g. Slunga, 1991; Bungum et al., 2010). Redfield & Osmundsen (2015) on the contrary propose that Fennoscandian seismicity is principally the product of locally derived stress fields and that far field stress from the oceanic domain is unlikely to penetrate deeply into a hyperextended continental margin.

In the subsections below we investigate events in the Fencat catalog located east of a line along the Swedish-Norwegian border down to Zealand in Denmark, for which magnitudes have been homogenized. The homogenization scheme is the one developed in the Finnish nuclear power plant projects discussed in Section 1 and is a magnitude scale based on scalar seismic moments and scaled to be consistent with the Helsinki local magnitude scale. The details of this are not yet in the public domain but will be released. Homogenized magnitudes are important in the comparisons below as a difference in magnitude of 0.3 translates to a factor of 2 difference in occurrence rate, if the b-value is close to 1.

4.1. Temporal variations

Already the descriptions of the Fencat data above make it clear that there will be temporal variations, such as varying numbers of reported earthquakes per year or the number of events larger than a certain magnitude per year. The largest sources of temporal variation are the variation in observation techniques, and in the pre-instrumental period varying interest in earthquakes as a phenomenon and to some extent also variations in population density. Extracting information on real, physical, temporal variability from the data is therefore not easy in our low seismicity area.

In an attempt to study temporal variations in the data we use the homogenized catalog for Sweden, Finland and the Baltic described above. In Figure 6 we plot the number of events in three-year central sliding windows for three different magnitude intervals, magnitudes larger than 1, 2 and 3. The upper panel in Figure 6 clearly shows the impact made by the modernization of the Swedish and Finnish seismic networks in the early 2000s, when the rate soars. The peak in the 1980s is due to the FOA network in Sweden, a temporary deployment of seismic stations that was significantly more dense, and more modern, than the then existing permanent network. The middle panel provides a better picture of the earlier seismicity. This shows the improvement in detection and location that came about through the modernization and expansion of the seismic networks in the mid to late 1960s and this figure seems to indicate that after approximately 1970 we record most of the magnitude 2 and larger events in Sweden and Finland. We still see a peak associated with the FOA deployment, and peaks in 1994 and 2006 which require further investigation before they can safely be deemed to have a natural cause. The middle panel also shows a concentration of events in the early 1900s, with the 1904 earthquake and its aftershocks and perhaps an elevated general awareness of earthquakes after the 1904 event and the 1908 Messina disaster. We see a significant decline in seismicity during the two world wars, and in their aftermath. This is probably to some extent an artifact of a lower public focus on earthquakes. Finally, the middle panel shows a marked change in awareness and interest after the events of the 1750s. First the Swedish north coast event of November 1751, then the Lisbon earthquake of 1755 and finally the large 1759 Kattegatt earthquake. In the lower panel of Figure 6 we show the variation in magnitude 3 or larger events. We now see the effect of the start of systematic gathering of earthquake related data in the late 1870s. Again there are dips in the rates during the two world wars and well into the postwar 1950s. The cause of this absence of M3+ events is not clear to us. There seems to have been a decline in M3+ events since 2000, when the seismic networks grew. It should be more closely investigated if this is related to methodological differences in the pre- and post-2000 periods. Although the simple qualitative analysis above cannot provide proof of either stationarity or non-stationarity, the occurrence of the very largest events, with magnitudes of 4 or larger, strongly indicate non-stationary behavior, as discussed in Section 3.3 above.

It should be pointed out that Figures such as Figures 6-8 can become a little complicated when aftershocks are taken into account. There are clearly aftershocks to the 1904 event in Figure 6 in the upper and middle panels, which broadens the peaks. Even smaller events, such as a magnitude 3 event could have aftershocks of magnitude 2. Aftershock sequences in intraplate regions may also last for a very long time (e.g. Stein & Liu, 2009). In addition, there seems to be a general increase in earthquake awareness after larger events, as clearly seen in the historical data. These figures must therefore not be interpreted as temporal variations of independent events.

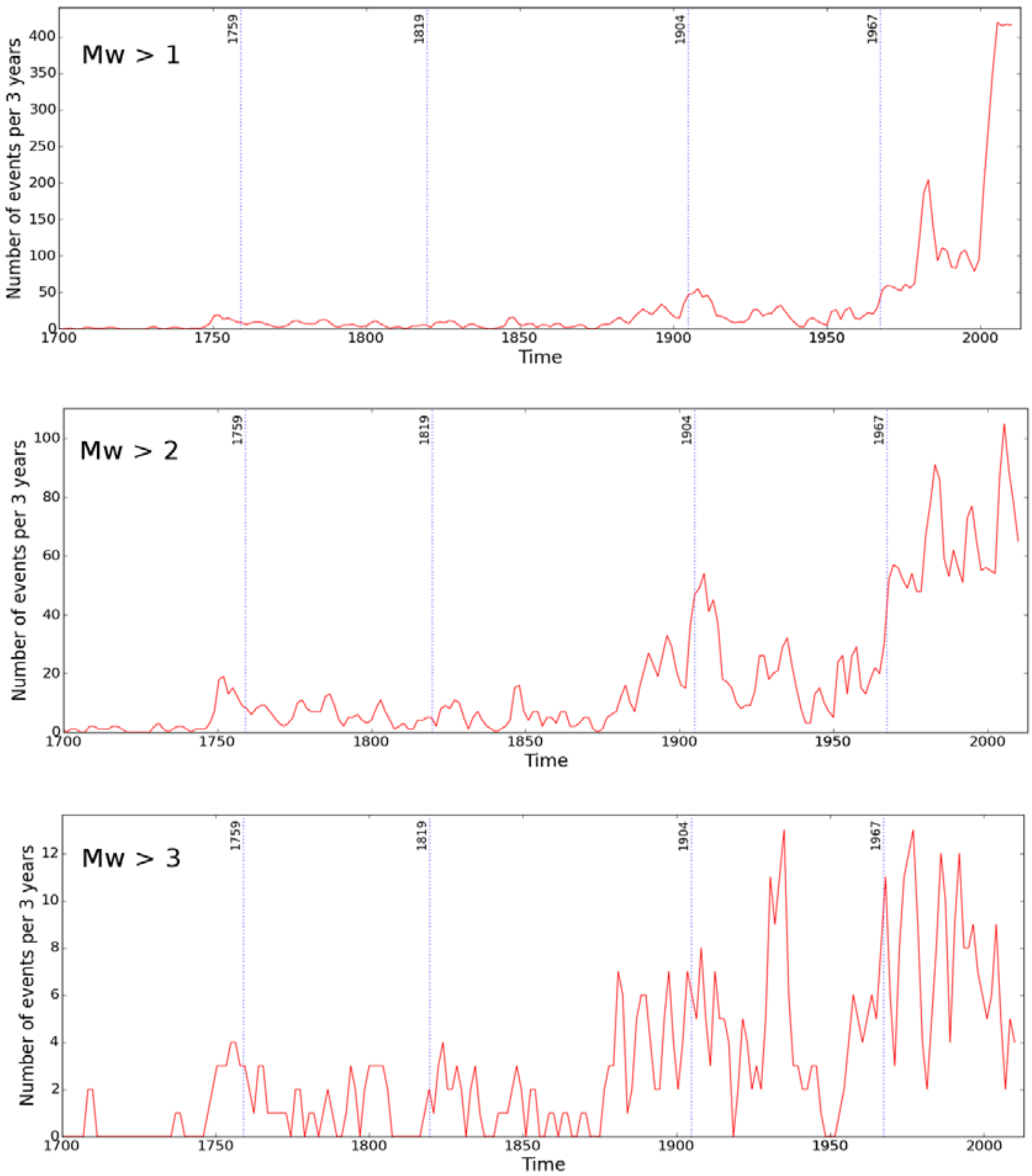


Figure 6. Temporal variation in the number of earthquakes in a sliding three year period in Sweden, Finland and the Baltic from 1700 to 2012. Homogenized magnitudes. All events above magnitude 1 (upper), 2 (middle) and 3 (lower). Large Fennoscandian events are indicated by blue dotted lines.

In general, however, we observe that the recent large events in Sweden have had very few, less than 10, aftershocks with generally very low magnitudes, less than 1.5.

Real physical temporal variations seem difficult to conclude from the data set used in Figure 6. It may be easier if a more confined area is considered, one with a higher seismicity rate. In Figures 7 and 8 we repeat the exercise by extracting data from southwest Sweden, Figure 7 and the Burträsk area, Figure 8.

The data in Figure 7 come from an area approximately from Skagerack to Vättern east-west (longitude 11.2 to 14.7) and from Göteborg to Sunne (latitude 57.8 – 59.8) north-south. The panels mirror some of the features we observed for the whole data set, such as the significant change in observations from about 1750, the peak after 1904 (this area includes the 1904 earthquake and aftershocks), the low activity in the 1940-1950s and the increase after 1970 due to better seismic networks. A little surprising is the hiatus after about 1860 and the lack of the clear increase associated with better reporting in the 1870s.

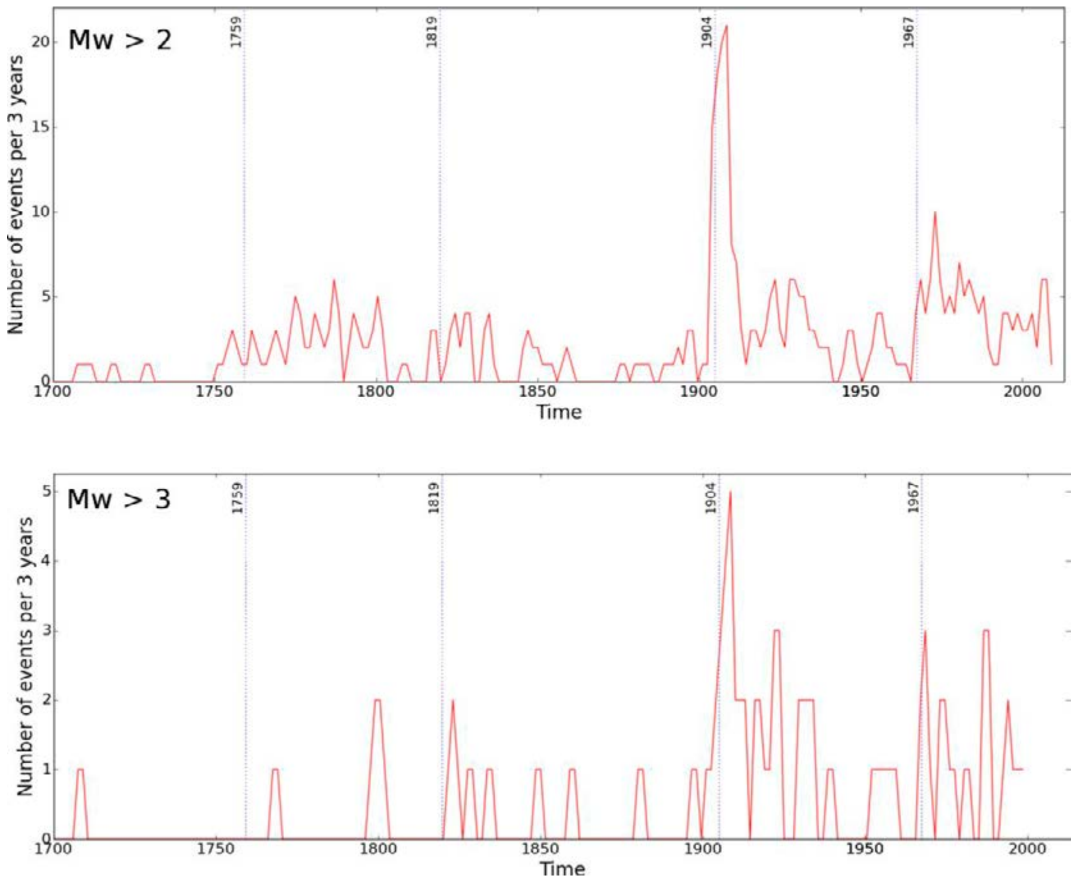


Figure 7. Temporal variation in the number of earthquakes in a sliding three year period in southwestern Sweden from 1700 to 2012. Homogenized magnitudes. All events above magnitude 2 (upper) and 3 (lower). Large Fennoscandian events are indicated by blue dotted lines.

The Burträsk seismicity is displayed in Figure 8 and although we see in the data from about 1970 onward that it is just as seismically active as southwestern Sweden (the Burträsk area here is only 30% the size of the southwestern Sweden area above), the data from the early times are quite different. The events in 1750 have clearly left their mark in Figure 8, and this is due in part to Mr. Gissler, a teacher in Härnösand who enthusiastically studied events in northern Sweden between about 1710 to 1760. There is then a long time period of almost no events until the resumption of earthquake studies in the 1870s. Interestingly, the Burträsk region has some seismicity during the wars, but virtually no larger activity between 1920 and 1939. This is an interesting point which should be further looked into.

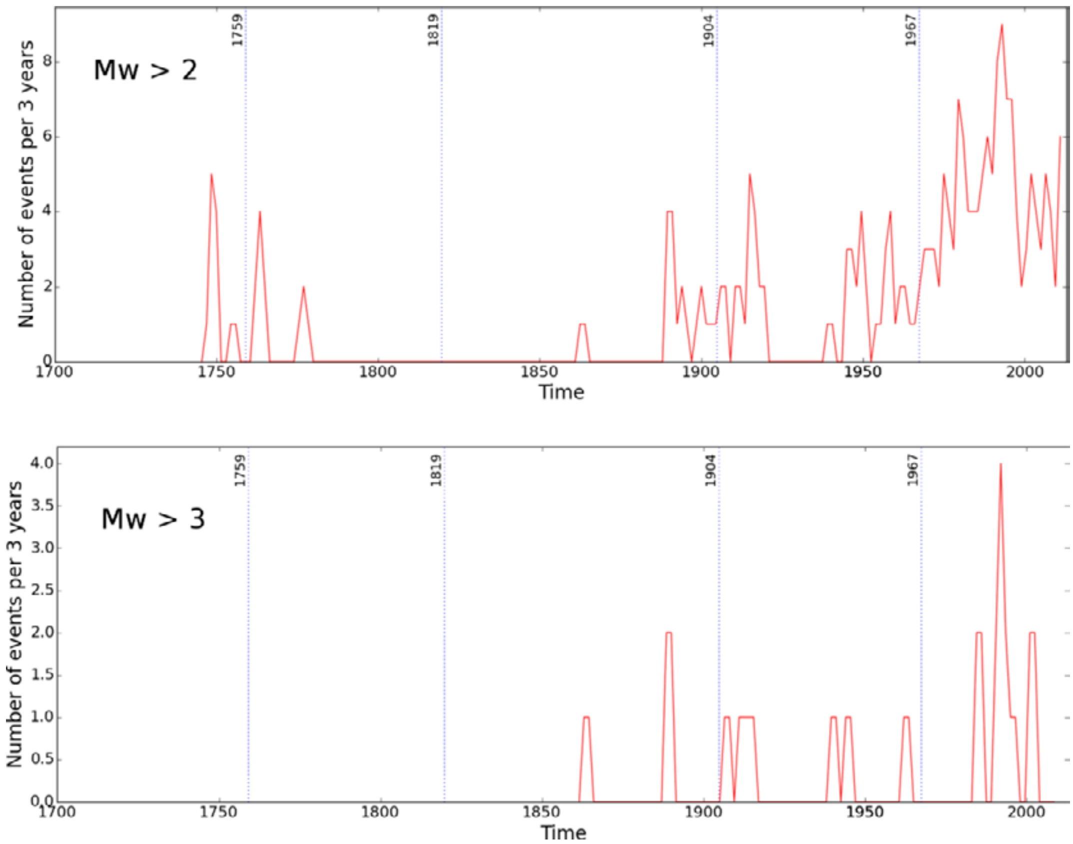


Figure 8. Temporal variation in the number of earthquakes in a sliding three year period in the Burträsk area, northern Sweden from 1700 to 2012. Homogenized magnitudes. All events above magnitude 2 (upper) and 3 (lower). Large Fennoscandian events are indicated by blue dotted lines.

In summary we see that the apparent temporal variations in seismicity are mostly affected by human activity and interest, and with little chance to draw solid conclusions on the behavior of the Earth's crust. Only for the very largest events does there appear to be some level of significant non-stationarity, as discussed in Section 3.3. Due to the low sensitivity of early seismic networks and the low seismicity rate we may have a larger chance to find variations that could be related to the crust itself if we study small events in a small area. Such studies may provide very relevant results, but may also only be applicable to the studied area.

4.2. Spatial variations

That there are spatial variations in the Fennoscandian seismicity has been clear since at least the start of more systematic collection of earthquake data in the 1870s. Kjellén (1910) published a seismicity map of Sweden, based only on pre-instrumental data, which is very similar to the seismicity distribution we observe today, except that he had no observations from the endglacial faults of northernmost Sweden. We have discussed spatial variability above and will not discuss it in any detail here, but in the maps in Figure 2 and 4 we see that on the large scale there are significant spatial variations in seismicity both in Fennoscandia and within Sweden. Any determination of earthquake rate will therefore be significantly affected by the actual region used. This would seem to suggest that we should use detailed zonation when analyzing recurrence relations. However, as pointed out by Slunga in SKI (1992) A2 and as is visible in the maps in Figure 2, the correlation between the location of the larger (larger than M4, red and blue in Figure 2) events and the areas of highest seismicity is not that strong, except for at the Norwegian west coast. It is therefore necessary to use large regions to represent the seismicity in order to characterize it appropriately. In addition, although the Kjellén (1910) map is very similar to today's seismicity map, it is not unlikely that there is non-stationarity in the Swedish seismicity as well. Intraplate seismicity in the central USA and China has shown strong spatial non-stationarity when viewed over longer time scales (Stein et al., 2009; Liu et al., 2011).

4.3. Recurrence times

As discussed above, estimating recurrence times for large earthquakes is not only very difficult but may even be inappropriate for stable continental interiors such as the Baltic Shield. Disregarding for a moment controversies about magnitudes, we have in the vicinity of Sweden had the 1759 Kattegatt M5.6, the 1819 Lurøy M5.8 and the 1904 Oslofiord M5.4 events. We could also add the 2004 Kaliningrad M5.0 event to the list, as it occurred in a similar geologic context, but should probably disregard the events off the Norwegian west coast due to the different tectonic setting and distance to Sweden. These four events would point to a recurrence period of a little less than 100 years for events larger than magnitude 5 in an area much larger than Sweden. The uncertainties in such an estimate are large, but can be quantified. Looking in more detail at the locations of these events, there are no reported events of magnitude 4 or larger within 80 km of the Kattegatt, Oslofiord or Kaliningrad events, except for the almost simultaneous 4.8 event in Kaliningrad. The area of the Lurøy event is more active and has had six events larger than magnitude 4 within 80 km since 1819, but none larger than 5. The radius considered here is large and reflect the lack of events in the regions. If the radius was instead restricted to the actual size of the causative fault lengths, on the order of 10-20 km, only one or two of the Lurøy events in 1958 would be considered close to 1819 event, and they are one magnitude smaller. We thus see no recurrence of larger events in the data.

Further information on recurrence times could potentially be obtained by studying smaller events. If we use the magnitude homogenized catalog for Sweden, Finland and the Baltic referred to above, we find three event pairs with magnitudes between 4 and 5 within 50 km of each other. None of the events are, however, from the instrumental period, the most recent is a 1902 M4.1 event in central Finland which occurred 36 km from a M4.0 event in 1626. The uncertainties in epicentral location of these events are significant as they are based on macroseismic observations and considering that the events have fault sizes of up to perhaps a kilometer, it is very uncertain if one is a repeat of the other. The same problems apply to the other two event pairs. Going to even smaller magnitude events we have to restrict ourselves to the period after 1970 when instrumental locations became better constrained. Investigating events of M3 or larger we find very few within a distance of each other that would correspond to the fault size. In most cases these are then either aftershocks or parts of a swarm of events occurring very close in time. These events also tend to occur in the well known areas of higher seismicity rate, such as Västergötland, the Oslo graben and the Burträsk region.

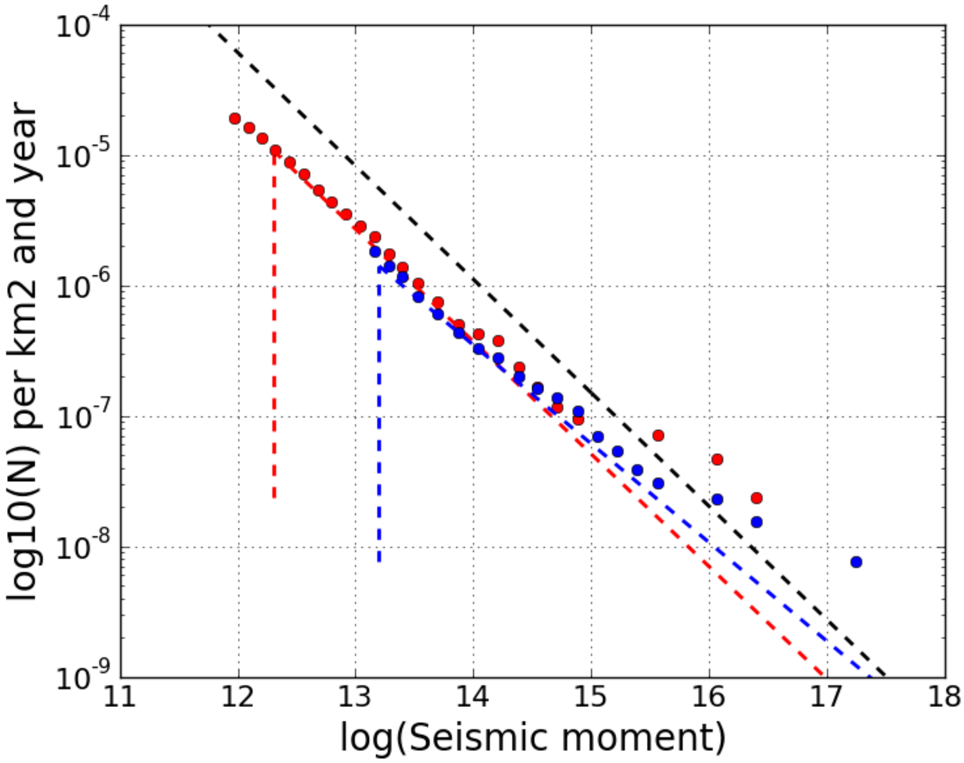


Figure 9. Frequency-moment distributions for an area encompassing Sweden, Finland and the Baltic Sea region. Homogenized magnitudes, see text, converted to seismic moments. Vertical red and blue dashed lines show moments of completeness. Red circles and dashed line: Events from 1970 to 2012 with magnitude greater than or equal to 2. b'-value 0.86±0.04. Blue circles and dashed line: Events from 1875 to 2012 with magnitude greater or equal to 3. b'-value 0.76±0.06. Black dashed line: the SKI (1992) seismicity function, b' = 0.87.

By looking at frequency-magnitude, or frequency-moment, distributions we can study the recurrence rate problem from another perspective. Again, the low seismicity rate makes it difficult to obtain results that do not have large uncertainties. We used two time periods of the data described in Section 4.1 above, the homogenized Sweden, Finland and Baltic catalog. In Figure 9 we show frequency-moment distributions for the two subsets of the data, one with events from 1970 to 2012 with magnitudes greater or equal to 2 (red) and another with events between 1875 and 2012 with magnitudes equal to or above 3 (blue). The number of events has been normalized to the area (917,163 km^2) and years, 42 and 137 respectively. We note that the data are close to each other in the range of overlap, but that the longer time series seems to lack some smaller events, i.e. it is not complete at lower moments. At the high moment end we see that the higher moments fit somewhat better to the long period regression line, indicating that the shorter time period was much too short to observe all events. This agrees with the similar conclusion reached by Slunga in SKI (1992) A2, where he showed that the 1904 earthquake does not agree with the regional frequency-magnitude distribution unless the circular area around the event has a radius of at least 500 km (which, incidentally, includes the high seismicity area of western Norway). The b' value of the short time series, 0.86±0.04, is similar to the one used by SKI (1992) whereas the long time period b'= 0.76±0.06 is lower. We note in Figure 9 that all three regression lines under-predict the largest magnitude events, but we keep in mind that for the black line we do not know the area used in the calculation in SKI (1992), which significantly influences the offset.

5. Implications for the results of SKI (1992)

The purpose of this report is to review the current status of paleoseismology and historical earthquakes in Fennoscandia and to assess how new results would affect the envelope ground response spectra in SKI (1992). Quantifying the effect on the spectra requires a full calculation which is outside the scope of this report. Here we will instead discuss qualitatively how the spectra may be affected by the new data.

Since the time of SKI (1992) there is not just more data available. As alluded to in the Introduction, methodologies for PSHA for nuclear facilities have evolved, and so have recommendations from the IAEA (e.g. IAEA, 2010). Especially the handling of uncertainties in the parameters has come under scrutiny and these are nowadays almost exclusively accounted for using logic trees, where branches are assigned probabilities either through statistical analysis or expert judgment. It is clear that SKI (1992) has deficiencies in the uncertainty methodology. Some effects of the methodological development has caused quite some discussion, with recent PSHA studies frequently resulting in appreciably higher design ground motions than previous assessments from the 1970s and 1980s (e.g. Bommer & Abrahamson, 2006). This has been explained as mostly due to neglect of ground motion variability, i.e. variations in ground motion for equal size earthquakes at the same distance from the epicenter, in the earlier studies (Bommer & Abrahamson, 2006).

5.1. Accounting for paleoseismicity

We saw in Section 3.1 that a dozen Fennoscandian endglacial earthquakes have been analyzed to the extent that they have all the attributes necessary to be included in a seismic hazard analysis: occurrence time, location and magnitude. There are more potential endglacial faults described in the literature but these still need further analysis. How then should these paleoearthquakes be included in a PSHA? The only PSHA studies that we know of today that have considered endglacial faulting (EGFs) are the studies mentioned in the introduction for the Finnish power companies Fennovoima and TVO/Fortum. As EGF rupture is believed to be confined to a relatively short time period at the end of the latest glaciation, triggered by the deglaciation stress field, and as EGFs from previous glaciations have not been identified (e.g. Lagerbäck & Sundh, 2008), these projects decided to not include the EGF events in the recurrence relations but instead use them to argue for a larger maximum magnitude (M_{max}) than would otherwise have been used.

If we use this approach with SKI (1992) it would imply a significant increase of the assumed maximum magnitude, from moment magnitude 6.3 (scalar seismic moment $M_0 = 3 \cdot 10^{18}$ Nm) to some magnitude from Table 1 above. Which of those magnitudes to use is an open question as the EGFs did all occur around local deglaciation, when the Earth's crust had a significant addition of horizontal stresses, which varied spatially, that has since been relaxed by land uplift. The stress state today is not the same as it was 10,000 years ago, nor are the stressing rates similar. One could therefore argue that earthquakes the size of the EGFs are improbable currently. On the other hand, the EGFs tell us that earthquakes of that size can actually take place in Fennoscandia and that should be reflected in the value of M_{max}. Influenced by the existence of the EGFs, the Fennovoima study used four different M_{max}, 5.5, 6, 6.5 and 7 in four different branches in the logic tree with very different probabilities. If the SKI (1992) methodology of calculating occurrence rates is unchanged, with b' = 0.87, uniform spatial probability of occurrence and the same areal normalization, then an increase of M_{max} to magnitude 8 implies a maximum moment of $\log(M_0) = 21$, six orders of magnitude higher than the anchoring point and corresponding to an epicentral density of approximately $8.2 \cdot 10^{-13}$ events per square kilometer and year. This is likely to have very little influence on the results.

What if the EGFs are included in the occurrence relation, the SKI (1992) seismicity function? Such an approach involves a number of choices on how to perform the calculations. As most of the confirmed EGFs are in northern Fennoscandia it may make sense to use a zonation of the region, so that occurrence rates and maximum magnitudes can vary spatially. This implies using different b''s and different anchoring densities for different regions, and hence these would vary when calculating the occurrence rates at different distances from the relevant sites, possibly making the probabilities for the ground response spectra different at different locations. Incorporating the EGFs would affect the calculation of b' itself, as we would add a number of data points to the far right in the frequency-moment relation curve, leaving a gap between events of $\log(M_0) \sim 17$ and 21. The temporal normalization used would also be an issue, we would have to use a 10,000 year time range for the EGFs but only have other earthquake data for a few hundred years. Finally, in order to be consistent the EGFs would also have to be considered when estimating M_{max}, implying an M_{max} slightly larger than the Pärvie 8.0 event in order to accommodate the possibility that the Pärvie event was not the largest possible. How this approach would affect the final spectral probabilities is non-trivial to assess without doing the calculations. Using the maximum likelihood algorithm of Aki (1965) for the calculation of b' implies that the very largest events have little influence on the actual b' value. In addition, the temporal normalization needed for 12 events approximately 10,000 years ago will of course lower the probabilities significantly for these events. How to merge data sets with different occurrence rates at different times has been studied by e.g. Kijko et al. (2016) but the time periods necessary here have not been used before. It is however likely that the limited occurrence period of the EGFs a long time ago and their limited spatial extent will make their effect on the spectral probabilities in SKI (1992) of limited significance.

Finally, if a modern PSHA was to be carried out the various ways to include EGFs could be assessed through the use of a logic tree approach. However, even in such circumstances it would be very difficult to estimate the probabilities for the different branches in a scientifically rigorous manner.

5.2. The SKI (1992) seismicity function

The seismicity function used in SKI (1992) was developed using Fencat data up to approximately 1987. As we have pointed out above, there are a number of questions about exactly how the epicentral density function at the anchor point and the b'-value were obtained, questions which make it difficult to repeat the calculations. Almost 30 years later we have significantly more data available to us, not only in Fennoscandia but also in northern Germany and Poland, areas which were, perhaps, not taken into account in SKI (1992). These new data are not only instrumentally well located earthquakes that have occurred since 1987, but there have also been a number of studies of historical events, both individual events, as discussed in Section 3.2, and in larger regions, notably the European earthquake catalog by Stucchi et al. (2013) and Grünthal et al. (2013), which documents the years 1000 to 2006. We showed above in Section 3.2 that there have been very few events in Fennoscandia since 1987 of the seismic moment used in SKI (1992) to define the anchor point, which could suggest that there is little gain in analyzing the new data. However, we do not know how many events were actually used to define this point in the pre-1987 data so even a few more could be significant. In addition, estimation of the b'-value critically depends on a large number of events with a range of magnitudes, which the new data could provide. With the increased detection levels of modern seismic networks we also have data in a larger range of magnitudes available to us today, which can provide better uncertainty estimates on the b'-value.

We note again that the use of a frequency-moment distribution, and a b'-value, such as in SKI (1992) is very unusual. The usual methodology also uses seismic moments, but converts these to moment magnitudes before estimating the a- and b-values from a frequency-magnitude distribution $\log(N) = a - bM$. As the Hanks & Kanamori (1979) moment magnitude scale is linear in $\log(M_0)$, $\log(M_0)$ will also have a linear relationship to the logarithm of frequency, assuming linearity of the frequency-magnitude distribution. The moment magnitude scale used in SKI (1992) is based on the seismic moment vs. magnitude relationships developed by Slunga et al. (1984), which deviates from linearity for moments above approximately 10^{14} Nm. The linear relationship between the logarithms of frequency and moment assumed in SKI (1992) may thus be suitable for larger magnitude events, according to Hanks & Kanamori (1979), but is not consistent with the underlying magnitude data. This is clear from Figure 7 of Report No. 1 in SKI (1992), where the estimates by Slunga, from SKI (1992) A2, show significant non-linearity. The fact that the SKI (1992) seismicity function and the Slunga distributions agree approximately at the anchor point may be due to the definition of the anchor point. We note that this magnitude – moment conversion seems to overestimate the rate of large events as compared to the Slunga distribution. However, looking back at the frequency-moment distributions in Figures 5 and 9, for the modern SNSN and two extractions from Fencat, respectively, we see that the SKI (1992) seismicity functions seems to be conservative for events smaller than about 10^{17} Nm but that for larger events it may underestimate the rate. This conclusion, however, critically depends on the definition of the SKI (1992) seismicity function, and that we do not know.

As outlined in Section 2, the issue of zonation is discussed at some length in the Appendices of Report No. 1, SKI (1992), but only briefly touched upon in the main section of the report. Modern PSHA tends to spend quite some effort on zonation, trying to identify zones of uniform seismicity rate. This is difficult for Fennoscandia as we have seen above, in spite of clear spatial variation in the rate of smaller events the larger events do not always occur in the areas of higher seismicity. In addition, the possibility that the seismicity is non-stationary in space and time needs to be considered when drawing up seismic source areas. Zonation may unduly increase the estimated hazard in some areas while decreasing it too much in other. How this would affect specific sites will then depend on the location of these relative the seismic zones. SKI (1992) assumes a constant, identical rate for the analysis of Ringhals and Barsebäck, and in the comments extends this to Oskarshamn and Forsmark. The validity of this assumption, and the effects a zonation would have on the hazard at these sites is unclear and need more quantitative investigation.

5.3. Ground motion prediction equations

As ground motion prediction equations (GMPEs), or attenuation relations, for Fennoscandian type crust were rare in the mid-1980s, SKI (1992) used spectra from Japanese earthquakes to study the decay in peak-ground-acceleration (PGA) with distance for different frequencies. Considerable amounts of work have been carried out since then on GMPEs for many different regions, using a variety of approaches. Since GMPEs can now be counted by the hundreds, recent work has focused on choosing the most appropriate GMPEs for a specific region (e.g. Delavaud et al., 2012). For stable continental regions like Fennoscandia there are now suggestions for which GMPEs may be most suitable (e.g. Delavaud et al., 2012, Douglas et al., 2013). None of these GMPEs have, however, been developed specifically for Fennoscandia. Under the auspices of the recent Finnish nuclear industry projects, significant effort has been put into the development of a Fennoscandian GMPE (Vuorinen, 2015). In Figure 10 we show an example of the new Fennoscandian GMPE modeled for the recent 2016 Bottenviken M4.1 earthquake. We see that the model nicely fits the observations, and that the Figure also well illustrates the problems with a Fennoscandian GMPE for seismic hazard purposes: (i) there is very little data at close distances and (ii) the 2004 M5.0 Kaliningrad event is the only large event during the time period of reasonably large seismic networks (and that event is far away from seismic stations), so we cannot develop GMPEs for larger magnitude events. There is ongoing work to calibrate the Vuorinen (2015) model to events from other SCR areas, which may extend it to higher magnitudes.

We conclude that there are a number of GMPEs available that are much more suitable to use for Fennoscandian conditions than the models in SKI (1992), and that there, in addition, has been a lot of work also on including uncertainties in GMPEs. How these new models would affect the SKI (1992) hazard estimates require a more thorough quantitative comparison.

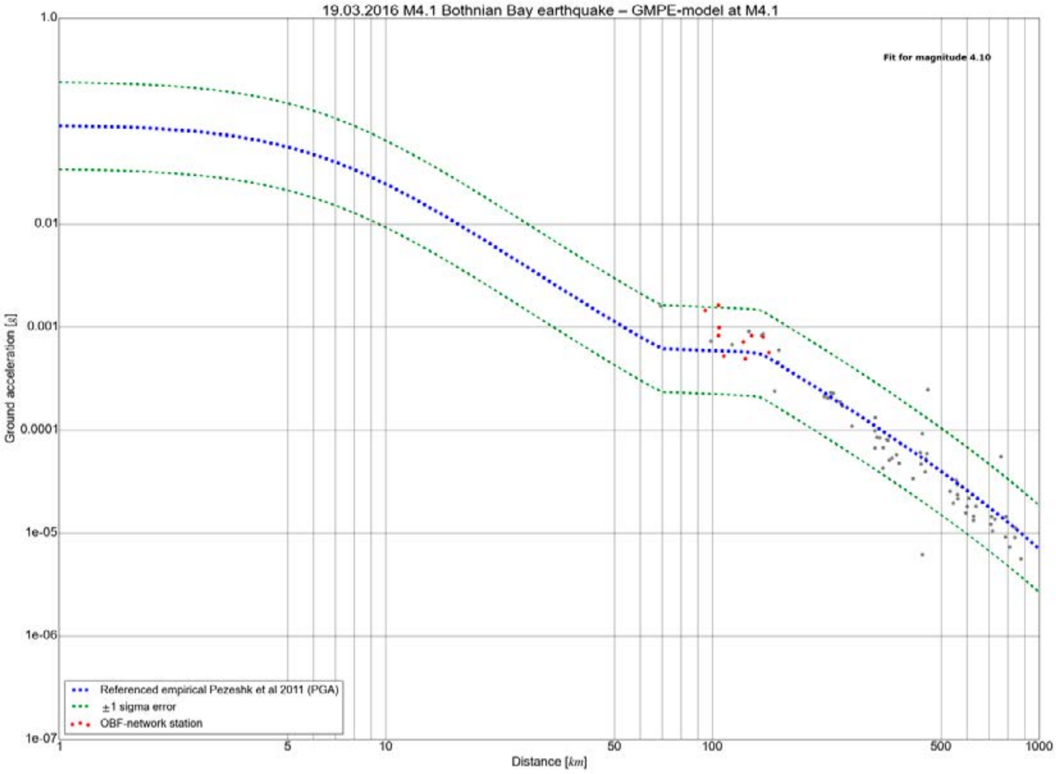


Figure 10. Ground motion prediction model (GMPE) for a Fennoscandian earthquake of magnitude 4.1 (blue line) and the one sigma uncertainty (green lines). Measured peak ground accelerations (grey and red circles). Model by Vuorinen (2015).

5.4. Further work

Advances in data acquisition, hazard methodology and international recommendations has made SKI (1992) outdated. That said, with the low seismicity rate and lack of larger events in Fennoscandia it is not obvious that a modern PSHA would provide significantly more accurate results than SKI (1992). It may, however, be able to add more significant uncertainty estimates to the results. If a new PSHA is initiated for Sweden we emphasize the importance of including expertise with a strong background in the field of intraplate seismicity, as the nature of intraplate seismicity is different to the more commonly studied seismicity in plate boundary or in actively deforming regions.

Globally, there has been relatively little research focus on intraplate seismicity, largely because of the generally low hazard there and the difficulty of collecting large data sets due to the low seismicity rates. Our understanding of intraplate seismicity has therefore evolved only slowly. However, the subject has gradually received more attention and observations of both seismicity and deformation in intraplate regions have increased significantly in the last decade. There are a number of important questions which are still open, and we recommend further study in fields such as:

- Paleoseismology. There is significant uncertainty and discussion on the level of paleoseismicity in Sweden as a whole, and in southern Sweden in particular. It would be very valuable if the various observations and claims could be classified in a uniform approach, such as done, for example, in the Norwegian NEONOR project (e.g. Olesen et al., 2013).
- Stationarity of earthquake occurrence. The assumption that seismicity rate is constant in time and space has a major effect on hazard estimation and may not be well founded in the data.
- The maximum possible magnitude, M_{max}. The assessment of M_{max} is difficult in all tectonic settings, and this is further complicated in Fennoscandia by the very large earthquakes that occurred at the end of the latest deglaciation.
- Ground motion prediction. There is ongoing development of ground motion prediction equations for Fennoscandia and this should be further pursued, both in terms of including data from similar tectonic regions with larger earthquakes and in terms of estimates of uncertainties on the predicted motions. It is also important to further study the transmission of high frequency ground motions over large distances, where observations are now available from the latest Fennoscandian magnitude 4+ events.

6. Conclusions

The Swedish Radiation Safety Authority (SSM) initiated this project to update and assess the earthquake data which forms the basis of the currently used envelope ground response spectra for Swedish nuclear power plants in SKI Technical Report 92:3 (SKI, 1992). SSM especially emphasized the need to include information on paleoseismic events, which are not included in the SKI (1992) data. We have reviewed the literature on the earthquake data underlying SKI (1992) and note that it is unclear from SKI (1992) exactly which data were used and how the epicentral density function at the anchoring point and the b'-value were obtained. This makes it difficult to assess how an updated earthquake dataset would influence the results in SKI (1992).

The term paleoseismic data is used here to mean earthquake data prior to the first written record of an earthquake in 1375, i.e. data based on indirect (geological) observations. In Fennoscandia such data is very scarce and consists almost exclusively of earthquakes that are inferred to have ruptured around the time of local retreat of the Weichselian ice sheet some 10,500 to 9,500 years before present. We have reviewed the literature on these so called post- or endglacial earthquakes and find that 11 events in Fennoscandia have been documented such that classification (they are in fact earthquakes), location, occurrence time and magnitude for each event have been estimated at a reasonable level of confidence. The events are listed in Table 1 in this report. There are a number of proposed additional events for which more thorough investigations are needed before they can be included in the current inventory.

Since the earthquake data extraction for SKI (1992), from the joint Nordic earthquake catalogue Fencat in 1987, there has been an approximate five-fold increase in the number of instrumentally detected and well analyzed earthquakes. In addition, a number of projects have worked on cleaning the historical earthquake catalogue from non-earthquake events such as frostquakes, lightning and various types of blasting and also reassessing the magnitudes. During this reclassification of historical events even events that are very large in a Fennoscandian perspective, such as the 1894 magnitude 5.1 event now inferred to be a frost event, have been reclassified and thus significantly affect the statistics of large earthquakes. The data show significant temporal variation in seismicity rates, much of this is, however, related to variations in human activity and interest in earthquakes and to variations in instrumental detection levels. For the very largest events, there appears to be some significant level of non-stationarity, but statistical confidence is relatively low because of the rather small number of events. The limited amount of data even for smaller events, and, especially for older data, possible errors in the catalogue such incorrect identification of other phenomena as earthquakes, means that statistically assessing stationarity is difficult. Spatial variation in seismicity in Sweden is apparent from the data, with higher rates in the Lake Vänern region, along the Baltic north coast and along some of the endglacial fault scarps. However, there is not a clear association of events larger than magnitude 4 to the regions of highest seismicity rates.

This project was not intended to quantify the effect of the updated earthquake data on the response spectra of SSI (1992). As it is unclear how the data underlying SKI (1992) were extracted, even more qualitative comparisons to the updated database will be somewhat uncertain. We find that if the methodology used in SKI (1992) to calculate occurrence rates is followed, the paleoseismic events are likely to have very little influence on the results. Including the paleoseismic events in the SKI (1992) seismicity function would require additional decisions on areal zonation and temporal normalization, because they mostly occurred in the north and approximately 10,000 years ago. We conclude that irrespective of the approach taken to include these events in the SKI (1992) function their effect on the spectral probabilities is likely to be very limited. The reassessment of the historical earthquake data, and the large quantities of new instrumental data, is likely to affect the SKI (1992) seismicity function. We attempt a calculation of a similar function using the updated database and find, keeping in mind the uncertainties in the definitions of zones and time periods in SKI (1992), that the SKI (1992) seismicity function may overestimate the rate of events smaller than about 10^{17} Nm but may underestimate the rate of larger events.

Since the production of SKI (1992) there have been significant developments in earthquake data acquisition and analysis, hazard methodology and international recommendations for seismic hazard assessment of nuclear facilities. This has made SKI (1992) outdated. A modern PSHA would treat instrumental, historical and paleoseismic data differently to the approach in SKI (1992) in terms of understanding of intraplate seismicity, wave propagation in shield areas and through the introduction of logic trees to evaluate various scenarios in rates, areal zonation, maximum magnitudes and attenuation effects. However, with the low seismicity rate and lack of larger events in Fennoscandia it is not obvious that a modern PSHA would provide significantly more accurate results than SKI (1992). It may, however, be able to add more significant uncertainty estimates to the results. If a new PSHA is initiated for Sweden we emphasize the importance of including expertise with a strong background in the field of intraplate seismicity, as the nature of intraplate seismicity is different to the more commonly studied seismicity in plate boundary or in actively deforming regions.

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Appendix 1: Literature review of post-glacial paleoseismic events in Sweden

Introduction

The term ‘post-glacial fault’ refers to preexisting bedrock structures that were reactivated by a combination of tectonic and isostatic stresses either during or after deglaciation (Stewart et al., 2000). These features have also been known as glacially induced faults (Lund, 2015) or endglacial faults (Lindblom et al., 2015). The presence of post-glacial faults in Fennoscandia has been known for decades (Kujansuu, 1964; Lundqvist and Lagerbäck, 1976), and an abundance of scientific literature exists on the subject. Geological and geophysical interest in post-glacial faults stems from their relative rarity on a global scale (Stewart et al., 2000; Lund, 2015) and their importance in understanding the paleoseismicity of the region. Given the relatively short period during which seismicity has been recorded instrumentally, an understanding of the paleoseismicity, documented in the geologic record, is important to understand the long-term seismic hazard in Sweden. This literature review seeks to compile geologically-derived location, timing, and magnitude values for seismic events following deglaciation. Historically-derived records of seismicity are discussed elsewhere.

The report is divided into two sections. The first section describes each of the scarps believed to result from post-glacial faulting and describes the geology and methods used to determine the location, timing, and magnitude of each event. These paleoseismic events are compiled into a summary table (Table A1) which is organized according to the availability of geologic data and thus the certainty of the timing and magnitude values. The second section describes proposed paleoseismic events that did not lead to surface rupture of post-glacial faults. Due to the high degree of uncertainty associated with such proposed events, not all of them are discussed in detail. Rather, the geology and methods of three proposed events are discussed along with their associated uncertainties. Proposed paleoseismic events that are not associated with a visible scarp are not included in the summary table because there are significant uncertainties related to their location, timing, and magnitude.

Scarps

Scarps discussed here (Fig. A1) are those mapped by Mikko et al. (2015) using high-resolution elevation data derived from light detection and ranging (LiDAR) (Lantmäteriet, 2015). The mapping carried out using the LiDAR imagery has refined the mapping previously carried out using aerial photographs (Lagerbäck and Sundh, 2008). In the LiDAR imagery, scarps cut across glacial landforms or sediments suggesting that they are younger than the glacial deposits (ie post-glacial). In northern Sweden, however, large areas have undergone very limited glacial erosion leading to the preservation of older glacial landforms (Lagerbäck, 1988; Lagerbäck and Robertsson, 1988). Thus, relying on geomorphology alone has serious limitations regarding age estimates of fault rupture. Additionally, in the absence of stratigraphical data across the faults, the assumption has been made that the entire scarp (both length and height) formed during a single seismic event. This has been confirmed for only the Lansjärv and Röjnoret faults and even then only at a few locations along the scarps. Trenching across the Bollnäs scarp also indicates a single faulting event, but the fault itself has not been confirmed in the bedrock. Such assumptions affect the magnitude estimates of paleoseismic events.

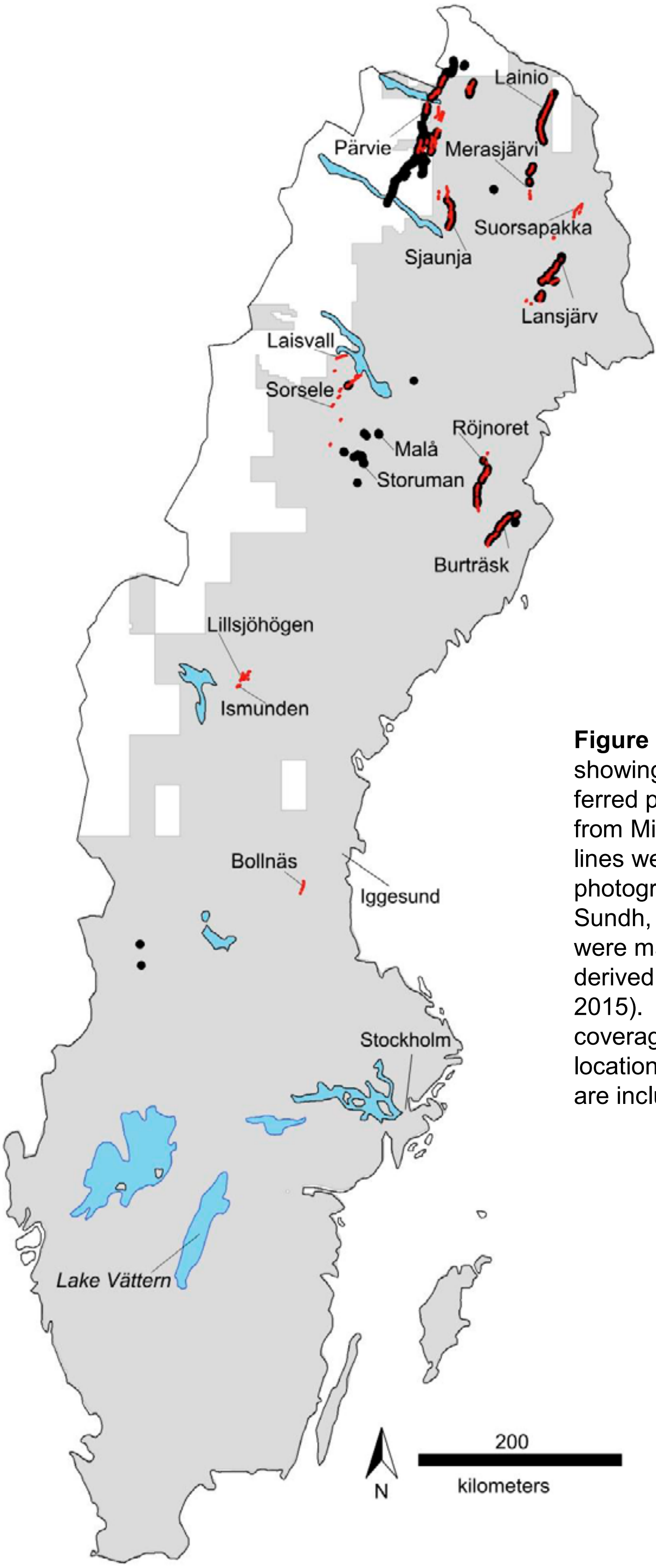

Figure A1. Map of Sweden showing the locations of inferred post-glacial fault scarps from Mikko et al. (2015). Black lines were mapped using aerial photographs (Lagerbäck and Sundh, 2008), and red lines were mapped using LiDAR-derived imagery (Mikko et al., 2015). The extent of LiDAR coverage is shaded gray, other locations mentioned in the text are included.

The discussion of the scarps is not organized by geography or size. Rather, they are presented in a more pedagogical format. The first example outlines the ambiguity of the cross-cutting relationships. The following examples provide information about what has been determined through geologic investigations of the best studied scarps. The scarps with no field investigations are presented last.

Pärvie

The Pärvie fault lies about 30 km west of Kiruna. It strikes NE-SW for some 150 km. Along this main segment, the scarp faces west and has a vertical offset of about 10 m. Due to the overhanging scarp, movement is interpreted to be reverse (Lundqvist and Lagerbäck, 1976). Recent mapping of the Pärvie system using high-resolution elevation data derived from light detection and ranging (LiDAR) (Lantmäteriet, 2015) has shown that the fault system is significantly more complex than shown to be through aerial photographic interpretation (Mikko et al., 2015). Numerous shorter segments, often facing east, were discovered in the LiDAR imagery.

The cross cutting relationships associated with the Pärvie fault present some ambiguity with regard to its age (Lundqvist and Lagerbäck, 1976, Lagerbäck and Sundh, 2008). Some segments of the scarp cut across glacial landforms, indicating that the scarp is younger than the glacial deposits. Other segments are overlain by undisturbed glacial landforms or cross cut by glacial channels, indicating that the scarp is older than deglaciation. This has been interpreted to indicate only partial deglaciation at the time of fault rupture (Lundqvist and Lagerbäck, 1976, Lagerbäck and Sundh, 2008). An additional complicating factor in northern Sweden is that many glacial landforms pre-date the late Weichselian glaciation and have been preserved beneath cold-based ice (Lagerbäck, 1988; Lagerbäck and Robertsson, 1988). Thus, adjacent glacial landforms may differ in age by tens of thousands of years.

The only Quaternary stratigraphic data across the Pärvie fault comes from hydroacoustical data collected from Lake Torneträsk. Despite the scarp being visible both north and south of the lake, faulted sediments were not detected in the hydroacoustical data set (Vogel et al., 2013). Deeper reflection seismic data, however, confirm the presence of a the fault to a depth of 8 km (Ahmadi et al., 2015).

Given the cross cutting relations, it is likely that at least some segments of the Pärvie fault were active after the latest deglaciation which occurred in the Abisko region about 9,500 years before the present (Berglund et al., 1996; Kullman, 1999).

All published estimates of paleoseismic magnitude for the Pärvie fault assume a single event that ruptured to the surface along the entire length of the fault. Since the geology does not necessarily support this, these magnitudes should be considered upper limits. Arvidsson (1996) calculated a magnitude of $M_w=8.2\pm0.2$ using available seismologic data. This number was refined to $M_w=8.0\pm0.4$ by Lindblom et al. (2015) using more up to date seismologic data. The instrumental record indicates that the Pärvie fault is seismically active (Lindblom et al., 2015) .

Lansjärv

The Lansjärv fault is adjacent to the settlement of the same name or about 40 km northwest of Överkalix. The scarp is visible in aerial photographs for about 50 km and strikes

NE-SW. The movement was reverse with a vertical displacement of generally 5-10 m but a maximum of 20 m (Lagerbäck, 1990; 1992).

About 10 m below the highest post-glacial shoreline, a trench across the Lansjärv fault, revealed faulted bedrock overlain by faulted tills. The uppermost till was overlain by undisturbed littoral sediments (Lagerbäck, 1990; 1992). This stratigraphy indicates that rupture occurred after deglaciation but before the regression of Baltic waters. Lagerbäck (1990; 1992) suggests faulting occurred about 9,000 years before the present. More recent chronostratigraphical work in Norrbotten, however, indicates deglaciation about 10,500 years ago and an initial shore displacement rate of 9 m/100 years (Linden et al., 2006). Thus, fault rupture can be bracket to between 10,500 and 10,350 years before present.

Arvidsson (1996) used instrumental seismicity data to calculate a magnitude of Mw=7.8±0.2 for the rupture of the Lansjärv fault. Again, a single event was assumed, but at Lansjärv a single event was confirmed by the geology (Lagerbäck, 1990, 1992). The instrumental record indicates that the Lansjärv fault is seismically active (Lindblom et al., 2015).

Soursapakka

The Soursapakka scarp lies 10 km east of Tärendö and about 40 km northeast of the Lansjärv fault. It was recently discovered in LiDAR-derived imagery by Mikko et al. (2015). Like Lansjärv, the Soursapakka scarp strikes NE-SW for about 17 km and faces predominantly west. Alhough the Soursapakka scarp may be a continuation of the Lansjärv fault, no scarps are in the interveneing 40 km despite the availability of LiDAR imagery (Mikko et al. 2015).

No stratigraphic data exist from the Soursapakka scarp, but it cross cuts glacial landforms. Thus it is believed to be post-glacial. Deglaciation occurred about 10,500 years before present (Linden et al., 2006).

There is no published magnitude estimate for the rupture of the Soursapakka fault, but the instrumental record indicates that the fault is seismically active (Lindblom et al., 2015).

Sorsele

The Sorsele fault lies 25 km north of the village of the same name. It strikes NE-SW and was originally mapped as only 2 km long using aerial photographs (Ransed and Wahlroos, 2007; Lagerbäck and Sundh, 2008). Recent mapping using LiDAR imagery has added several new segments and lengthened the scarp to more than 40 km, albeit discontinuously (Mikko et al., 2015). Vertical displacement of the fault is about 1.5-2 m (Lagerbäck and Sundh, 2008).

Trenching of the fault confirmed reverse faulted bedrock overlain by faulted till and lacustrine silt (Ransed and Wahlroos, 2007). Based on this stratigraphy, fault rupture occurred after deglaciation which occurred sometime after 10,000 years before present (Hughes et al., 2015). Due to its location above the highest post-glacial shoreline there is no minimum limiting age for rupture of the Sorsele fault.

There is no published magnitude estimate for the rupture of the Sorsele fault, but it may be seismically active (Lindblom et al., 2015). Seismic station coverage is sparse in this area.

Laisvall

The Laisvall scarp lies about 15 km southeast of the village of Laisvall and 25 northwest of the Sorsele fault. It strikes ENE-WSW for about 11 km, and the 5 m high scarp faces northwest (Mikko et al., 2015). The Laisvall scarp is located near the edge of the current LiDAR extent (Mikko et al., 2015). Thus, it may be extended as new LiDAR data become available.

No stratigraphic data exist from the Laisvall scarp, but it cross cuts glacial landforms. Thus it is believed to be post-glacial. Deglaciation occurred after 10,000 years before present (Hughes et al., 2015).

There is no published magnitude estimate for the rupture of the Laisvall fault, but it may be seismically active (Lindblom et al., 2015). Seismic station coverage is sparse in this area.

Burträsk

The Burträsk scarp lies 5 km east of the village of Burträsk. It strikes NE-SW for about 45 km (Mikko et al., 2015). The scarp faces west and is generally 5-10 m high with a maximum of 15 m of relief (Lagerbäck and Sundh, 2008).

Although trenching did not reveal a direct relationship to a bedrock fault, all of the tills are reported to be chaotic and disturbed (Lagerbäck and Sundh, 2008). Assuming that these disturbances are a result of rupture along the fault, then this must have occurred after deglaciation which occurred shortly before 10,000 years ago (Hughs et al., 2015).

Assuming that a single event created the Burträsk scarp, then the magnitude of that event can be estimated using the dimensions of the surface rupture and empirical relationships derived from historical earthquakes (Wells and Coppersmith, 1994). Stewart et al. (2000) used these relationships to estimate a magnitude of Mw=7.1. The instrumental record indicates that the Burträsk fault is seismically active (Lindblom et al., 2015).

Röjnoret

The Röjnoret fault lies about 5 km west of Boliden. The west-facing scarp strikes approximately N-S for about 60 km (Mikko et al., 2015) and is 5-10 m high (Lagerbäck and Sundh, 2008).

Trenching across the scarp revealed reverse faulted bedrock overlain by multiple tills, all of which were faulted (Lagerbäck and Sundh, 2008). Thus the fault ruptured to the surface post-glacially. Lagerbäck and Sundh (2008) suggest that the lack of soil development beneath deposits interpreted to have slid down the scarp during rupture may indicate that the event occurred shortly after deglaciation. Deglaciation occurred shortly before 10,000 years ago (Hughes et al., 2015).

Stewart et al. (2000) estimate the magnitude of the earthquake associated with the rupture of the Röjnoret fault to Mw=7.1 based on the dimensions of surface rupture and the empirical relationships of Wells and Coppersmith (1994). The Röjnoret is referred to as the Bastuträsk fault by Stewart et al. (2000), and instrumental records indicate that it is seismically active (Lindblom, 2015).

Bollnäs

The Bollnäs scarp lies about 2.5 km west of the town of Bollnäs. It strikes N-S for about 12 km (Malehmir et al., 2015), faces east, and has a relief of about 5 m (Smith et al., 2014).

Trenching in the area did not reach bedrock to confirm the post-glacial fault origin of the scarp. Nevertheless, faulted till and glaciolacustrine clay were observed in trenches across the scarp. These faulted glacial sediments were overlain by undisturbed post-glacial silt (Smith et al., 2014). This stratigraphy suggests fault rupture shortly after deglaciation but prior to the regression of Baltic waters. Numerous landslides in till in the Bollnäs area, interpreted to be seismically triggered, have been radiocarbon dated to shortly after deglaciation. The oldest of these and most likely to approximate the timing of fault rupture is 10,180 years before present, and deglaciation is estimated to have occurred about 10,670 years before present (Smith et al., 2014).

The magnitude of the earthquake associated with the Bollnäs scarp was estimated at Mw=6.2 by Smith et al. (2014) using the empirical relationships of Wells and Coppersmith (1994). A slightly more conservative value, Mw=6.1, was obtained by Malehmir et al. (2015) using the empirical relationships of Leonard (2010). Although there has been some measureable seismicity in the area, it is not clearly related to the Bollnäs fault (Smith et al., 2014; SNSN, 2016).

Ismunden and Lillsjöhögen

The Ismunden fault lies about 30 km east of Östersund. It strikes generally NE-SW with the majority of scarp segments facing southeast. The nearby Lillsjöhögen scarp strikes notably more N-S and faces east (Mikko et al., 2015). Vertical displacement along both scarps ranges from 2 to 8 m (Berglund and Dahlström, 2015). The scarps extend discontinuously for more than 20 km.

Because the scarps cut across glacial lineations, they are interpreted indicate post-glacial fault rupture (Berglund and Dahlström, 2015). Deglaciation occurred about 10,000 years ago (Hughes et al., 2015). No published stratigraphic data exist from these scarps.

Based on the empirical relationships of Wells and Coppersmith (1994), Berglund and Dahlström suggest a magnitude of 5-6 based on the length of visible scarps and a magnitude of 6.5-7.5 based on average and maximum heights. The final estimate presented by Berglund and Dahlström (2015) is about Mw=7. The instrumental record indicates that minor seismicity may be associated with the Ismunden and Lillsjöhögen faults (Berglund and Dahlström, 2015; SNSN, 2016).

Lainio

The Lainio scarp lies about 1 km north of the village of Lainio. It extends for about 50 km and the strike changes from NW-SE in the south to NE-SW in the north. The scarp faces west and has 10-20 m of relief (Lagerbäck and Sundh, 2008).

No stratigraphic investigations have been carried out along the Lainio scarp. Thus, all age estimates are based on cross cutting relationships seen in the geomorphology. As at Pärvie, this is problematic because of the presence of preserved glacial landscapes in northern Sweden (Lagerbäck, 1988). Along the Lainio scarp, the geomorphic relationships are also ambiguous with regard to age. At one location a glacial meltwater channel is constrained by the scarp suggesting that the scarp pre-dates deglaciation. At another location rock fall deposits from the scarp appear not to have been affected by glaciation, suggesting that the rockfall and scarp are younger than deglaciation (Lagerbäck and Sundh, 2008). Despite conflicting evidence, Lagerbäck and Sundh (2008) suggest that the fault ruptured just prior to deglaciation. Deglaciation in the area occurred between 11,000 and 10,000 years ago (Hughes, et al., 2015).

Assuming that the entire length of the scarp ruptured at once, the magnitude associated with seismicity would have been Mw= 7.1 according to Stewart et al. (2000) and the empirical relationships of Wells and Coppersmith (1994). The instrumental record indicates that the Lainio fault is seismically active (Lindblom et al., 2015).

Merasjärvi

The Merasjärvi scarp lies 3 km east of the village of Merasjärvi or 25 km west of Junosuando. In aerial photographs, the west facing scarp is visible for some 8 km striking NNE-SSW (Lagerbäck and Sundh, 2008). Mapping of the scarp using LiDAR imagery has significantly lengthened it to a discontinuous length of about 30 km (Mikko et al., 2015).

No stratigraphic data exist to constrain the timing of fault rupture, and the scarp lies in an area of preserved glacial landforms (Lagerbäck, 1988). Lagerbäck and Sundh (2008) suggest a post-glacial age due to the ‘extremely fresh appearance’ of the scarp, but the authors admit that glacial erosion in the area has been limited. Local deglaciation occurred about 10,000 years ago (Hughes et al., 2015).

Stewart et al. (2000) estimated a seismic event of magnitude Mw=6.3 using the relationships of Wells and Coppersmith (1994) and assuming a single event along a long 9 km scarp. Given that the scarp has been lengthened significantly using LiDAR imagery (Mikko et al. 2015), this magnitude is likely a minimum value. The instrumental record indicates that the Merasjärvi fault is seismically active (Lindblom et al., 2015).

Sjaunja

The Sjuanja fault lies about 30 km west of Gällivare. It strikes approximately N-S for 40 km, and the majority of scarps face east. The Sjuanja fault was considered part of the Pärvie system by Lagerbäck and Sundh (2008), but Mikko et al. (2015) suggest that it be considered separate because it lies 50 km east of the main Pärvie fault.

The Sjuanja scarp cuts across streamlined glacial landforms and glaciofluvial deposits. It is assumed to be post-glacial in age, but no stratigraphic data exist. Deglaciation less than 10,000 years ago (Hughes et al., 2015).

There are no published magnitude estimates for seismicity associated with the rupture of the Sjaunja fault, but the instrumental record indicates that it may be seismically active.

Proposed seismic events without surface rupture

If paleosesimic events are not associated with surface rupture, then the location, timing, and magnitude of the events must be based on evidence other than fault scarps. Often such claims of paleoseismicity refer to ‘disturbed sediments’ that are interpreted to be seismically induced, or ‘seismites’ (Mörner, 1996). While seismicity can certainly create many different types of sediment disturbances (Obermeier, 1996), there are also a variety of other mechanisms that can create similar disturbances, particularly in a glacial environment.

The margins of temperate glaciers are tremendously dynamic geological environments. The terrain is unvegetated, and sediments are often saturated with meltwater. Mass movements are common (Benn and Evans, 2007, p. 262), and sedimentation rates can be extreme (Benn and Evans, 2007, p. 289). Both mass movements and rapid sedimentation can create increased pore pressures that lead to water-escape or liquefaction structures (Collinson and Thompson, 1989, p. 45) that are often pointed to as evidence of paleoseismicity. Sediments supported by ice, such as esker deposits, are prone to collapse as the ice melts (Benn and Evans, 2007, p. 247, p. 243. Such collapse creates faults within the sediments that are entirely unrelated to paleoseismicity. Additional faulting or folding of glacial sediments is done by the flow of the ice itself (Benn and Evans, 2007, pp. 249-255).

Given the variety of ways in which glacial sediments may be disturbed, such disturbances are not necessarily indicative of paleosesimicity. Thus, this report does not address every claim of paleoseismicity based on disturbed sediments. Rather, the location, timing, and magnitude of three proposed seismic events, without surface rupture, are examined as illustrative of the entire set of proposed events.

Stockholm

The same proposed event is referred to as Stockholm 3 (Mörner, 1996) and Stockholm 4 (Mörner, 2005). Here, it is referred to as Stockholm 3. Although Mörner and Tröften (1993) describe a 1 m high and 4 km long scarp, such a feature was not identified as cutting glacial deposits in LiDAR-derived imagery (Mikko et al., 2015). Thus, the remaining indicators of paleoseismicity are two types of sediment disturbances (Mörner, 1996). First there are exposures of varved glaciolacustrine clay that indicate faulting, folding, and liquefaction (Mörner, 1996). Such features are not uncommon, but neither are they particularly diagnostic of paleoseismicity. Similar features have been described by Lagerbäck et al. (2005) along the coast of Uppland and attributed to landslides. The second type of sediment ‘disturbance’ described by Mörner (1996) is a relatively thick and coarse-grained layer in the varve sequence.

For decades, such varves have been interpreted as indicating the sudden drainage of ice dammed-lakes (DeGeer, 1940). Mörner (1996), however, suggests that such deposits are seismically induced turbidites. This dramatic re-interpretation of the Swedish varve chronology is not supported by the scientific literature. The possibility that drainage varves may be seismites was first proposed as one of three possibilities to explain magnetic intensity variations in sediment cores by Mörner (1980). Despite a lack of evidence to support such an hypothesis, these thicker coarser varves were referred to as 'an excellent register of ground accelerations' by Mörner (1982) in an unreviewed conference abstract. Subsequently, Mörner (1985) expands on this claim to include many more drainage varves as 'evidence' of paleoseismicity. Thus, both types of sediment disturbances claimed to indicate paleoseismicity are easily explained by other processes, and there are no data to support preference of the seismite hypothesis over other processes.

The proposed age of Stockholm 3 event, derived from varve correlations, is 10,430 years before present according to Mörner (1996). Varved sediments can be dated by correlating the relative thicknesses of a series of varves to an established varve sequence. Absolute ages can then be derived if the established chronology is either radiocarbon dated or linked to a Holocene varve sequence (Strömberg, 1989; Cato, 1998). Such a correlation, between the proposed Stockholm 3 event and an established varve chronology, has not been published. With regard to dating, Mörner (1996) cites Mörner and Tröften (1993) and Mörner (1985), neither of which contain varve correlations. Mörner (1985), however, cites Mörner (1978) with regard to dating, and Mörner (1978) does not include varve correlations either. Without correlating the varve thicknesses to an established chronology, the age of the deposits cannot be resolved beyond deglacial in age.

Another age for this event proposed by Mörner (2005) is '~10,000' ^{14}C years before present. Mörner (2005) does not discuss the origin of this date, nor provide a reference for it. The lack of a reference, lack of an associated error, and the use of '~' makes this date unreliable because its origin is completely unknown.

Mörner and Tröften (1993) describe the dimensions of the proposed scarp, which is not visible in LiDAR (Mikko et al., 2015), and state 'The corresponding magnitudes (referring to two proposed events) are likely to have been in the order of 6-7.' No references to empirical relationships of historical events are provided, and the values appear to be entirely ad hoc. For the same event, Mörner (1996) suggests that the magnitude exceeds 8 based on the distance from the epicenter to areas of disturbed sediments. There are three problems with such an approach. First, without a visible scarp, the location of the epicenter is entirely unknown. Second, given the lack of chronological control it is impossible to link disturbed sediments to the same time event. Third, the magnitude distance relationships used by Mörner (1996; 2003 p. 39) do not correspond to the relationships (Galli and Meloni, 1993) which Mörner (1996; 2003 p. 39) cites. For a given distance that seismites are found from the epicenter of an earthquake, Mörner (2003 p. 39) suggests magnitudes substantially higher (Fig. A2) than what empirical historical data indicate (Galli and Meloni, 1993; Galli, 2000) despite citing some of the same sources.

To summarize, 1) processes other than paleoseismicity may explain the sediment disturbances; 2) the deposits in question are uncorrelated and undated; and 3) magnitude values are either made up or do not reflect the data that they cite.

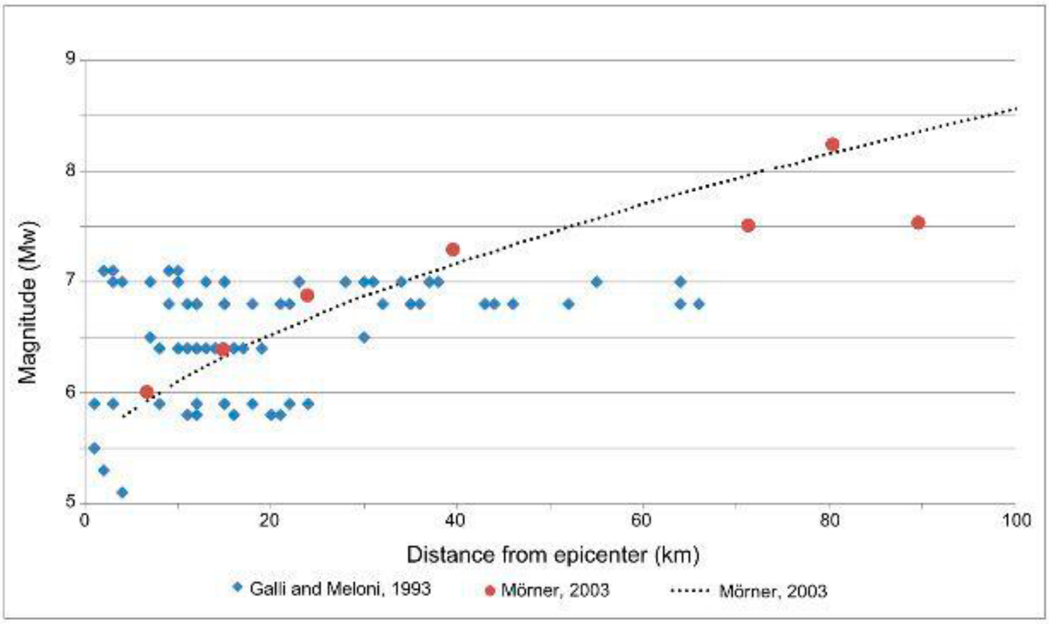

Figure A2. Blue diamonds show the relationship between distance from epicenter to seismically disturbed sediments and magnitude (Galli and Meloni, 1993). The red circles define the black line used by Mörner (2003) to derive magnitudes of proposed paleo-seismic events. Although Mörner cites Galli and Meloni (1993) his line is plotted within the field of blue diamonds as opposed to just below it. Thus, Mörner (1996; 2000; 2003) overestimates magnitudes of proposed paleoseismic events based upon the distribution of disturbed sediments. For example, if sediments are seismically disturbed to a maximum distance of 20 km an earthquake epicenter, then Galli and Meloni (1993) suggest a magnitude of about 5.7 (below the blue diamonds). For the same 20 km maximum distance, Mörner (1996; 2000; 2003) suggests a magnitude of about 6.5 (along the black line).

Iggesund

The Iggesund paleoseismic event is proposed to have occurred along the east coast of Sweden south of Hudiksvall. Although Mörner (2000) suggests that the Boda caves were created when the bedrock was ‘blown up’ by paleoseismicity, it is now widely believed that these large pieces of displaced rock were moved by glacial ice (Lagerbäck et al., 2005; Lagerbäck and Sundh, 2008). According to Mörner (2000), the remaining indicators of paleoseismicity are related to sediment disturbances and include landslides, liquefaction structures, and the widespread presence of a thick and coarse-grained varve. Mörner refers to this varve as -424 which is described by Strömberg (1989) as an anomalous feature. Not only is the varve particularly thick and coarse-grained, but it can also be identified in several different drainages. Deposition of such a layer cannot be explained as the result of the drainage of an ice-dammed lake. Strömberg (1989) suggests that it results from high meltwater discharge along the margin of the retreating ice sheet caused by exceptionally warm and/or rainy conditions. Nevertheless, thick coarse-grained varves have not been demonstrated to indicate paleoseismicity as described above.

According to Mörner (2000), the Iggesund event occurred 9663 years before present. This age is supposedly derived from varves, but no correlation of varve thickness to an established chronology is provided. On the contrary, correlating the geology to the varve sequence seems to be entirely ad hoc. Mörner (2000) states:

> At Iggesund harbour a number of trenches were dug in order that the varves could be recorded and counted. A lower varve sequence, which included disturbances, liquefaction structures, and injection structures, is separated from an upper sequence by a graded turbidite. This turbidite is assigned to varve -424 with a precision of <±5 varves.

First, it is impossible to correlate varves that have been disturbed in such a way. Second, there is no mention of the number of varves or their thicknesses, and no varve correlation diagrams, such as those in Strömberg (1989), are included.

The dating of a nearby landslide is equally unsatisfying. Refering to sediments collected from within the landslide scar, Mörner (2000) states:

> The -424 varve is missing from this basin and the succession begins with some decimeters of silt, followed by typical post -424 varves. The earthslide, therefore is considered to have formed at the -424 event.

Assigning the landslide to an event that does not exist in the stratigraphy is nonsensical. Assuming the rest of the stratigraphic description is correct, the most that can be claimed is that the landslide occurred after deposition of varve -424 and before the regression of Baltic waters.

Purportedly, the sediment disturbances at Iggesund were caused by a seismic event of magnitude 8 based on the distance from the epicenter to areas of disturbed sediment (Mörner, 2000). Such reasoning suffers from the same flaws as that used to assign a magnitude to the Stockholm 3 event. First, the location of the epicenter is unknown. Second, the lack of chronological control does not allow the sediment disturbances to be linked to a single event because the varves have not been correlated. Third, the magnitude distance relationships used by Mörner (2000; 2003 p. 39) do not correspond to the empirical relationships (Galli and Meloni, 1993) which Mörner (1996; 2003 p. 39) cites.

To summarize, 1) processes other than paleoseismicity may explain the boulder caves and the sediment disturbances; 2) the deposits in question are uncorrelated and undated; and 3) magnitude value does not reflect the data that is cited.

Vättern

Post-glacial fault rupture beneath Lake Vättern has been proposed by Jakobsson et al. (2014). Hydroacoustical data from the south and north ends of the lake show features that have been interpreted as collapse structures in late and post-glacial sediments. These structures do not extend into the deeper glacial stratigraphy, and no faulted glacial sediments or bedrock are visible in the hydroacoustical imagery. Despite a proposed vertical off set of up to 13 m, no scarp is visible in the bathymetry (Jakobsson et al., 2014). Additionally, no scarps were observed in LiDAR imagery around the lake (Mikko et al., 2015).

The disturbed sediments are overlain by undisturbed clay deposits that contain a pollen assemblage similar to the Younger Dryas-Preboreal transition zone. Since the disturbed sediments must pre-date this period, they are assigned an age of 11,500, essentially at the Pleistocene-Holocene boundary (Jakobsson et al., 2014).

The collapse structures are up to 13 m deep and are found at the same stratigraphic level at both ends of Lake Vättern some 80 km apart. Thus, Jakobsson et al. (2014) suggested that these are the dimensions of fault rupture and calculated a magnitude of Mw=7.5 based on empirical relationships of fault rupture dimensions to historical earthquake magnitudes (Wells and Coppersmith, 1994). While this method has been applied to numerous faults in Sweden, they have visible scarps associated with surface rupture. In the Vättern case, use of such relationships is inappropriate given the fact that no rupture has been observed either in the bathymetry or in the stratigraphy.

Concluding remarks

According to the literature, there are twelve scarps in Sweden that appear to cross cut glacial sediments. These 12 features often include multiple segments and complex geometries. Most of these structures are in the northern part of the country, but examination of LiDAR-derived imagery has revealed previously unknown scarps in central Sweden. For all of these features, fault rupture is interpreted to have occurred around the time of deglaciation generally between 10,500 and 9,500 years before present. Magnitude estimates for the seismic events associated with fault ruptures range from as low as Mw=6.1 to as high as Mw=8.2. Review of the literature has also revealed the complete lack of stratigraphic information relating to most of the scarps. Stratigraphic information would not only help constrain the timing of fault rupture, but also provide information critical to magnitude calculations. The assumption that each scarp formed as a result of a single event remains untested on seven of the twelve scarps.

Despite the vast body of literature related to proposed paleoseismicity in Sweden in the absence of surface rupture, great uncertainty surrounds many of these claims. Sediments may be disturbed in a number of different ways in a glacial environment, and they do not necessarily indicate paleoseismicity. The assigning of ages and magnitudes to proposed paleoseismic events defined by disturbed sediments is often ad hoc or attained through misuse of published empirical data. Without the presence of a scarp, significant uncertainties exist regarding the location (ie occurrence), timing, and magnitude of proposed paleoseismic events. Thus, they are excluded from the current inventory.

Name	Location	Approximate length (km)	Timing (years before present)	Magnitude (Mw)	Reference for magnitude estimate	Seismic-ally active
Lansjärv	40 km northwest of Över-kalix	50	After de-glaciation 10,500 but before re-gression 10,390	7.8	Arvidsson, 1996	yes
Sorsele	25 km north of Sorsele	40	After de-glaciation < 10,000	no data	none	maybe
Röjnoret	5 km west of Boliden	60	After de-glaciation 11,000-10,000	7.1	Stewart et al., 2000	yes
Bollnäs	2.5 km west of Bollnäs	12	After de-glaciation 10,670 but before peat accumulation in landslide scars 10,200	6.1	Smith et al., 2015	no
Burträsk	5 km east of Burträsk	45	After de-glaciation 11,000-10,000	7.1	Stewart et al., 2000	yes
Pärvie	30 km west of Kiruna	150	During de-glaciation 9,500	8.0	Lindblom, 2015	yes
Lillsjöhögen & Ismunden	30 km east of Östersund	20	After de-glaciation < 10,000	7.0	Berglund and Dahl-ström, 2015	maybe
Lainio	1 km north or Lainio	50	During de-glaciation 11,000-10,000	7.1	Stewart et al., 2000	yes
Soursapakka	10 km east of Tärendö	17	After de-glaciation <10,500	no data	none	yes
Merasjärvi	3 km east of Merasjärvi	30	After de-glaciation 11,000-10,000	6.3	Stewart et al., 2000	yes
Sjaunja	30 km west of Gällivare	40	After de-glaciation < 10,000	no data	none	maybe
Laisvall	15 km southeast of Laisvall	11	After de-glaciation < 10,000	no data	none	maybe

Table A1. Literature derived location, timing, and magnitude information for mapped post-glacial fault scarps in Sweden. Yellow indicates that the scarps have been trenched to bedrock to confirm a single post-glacial rupture. White indicates that trenching did not reach bedrock, but a single rupture is indicated. Blue indicates that no stratigraphic data exist from the scarps. See text for further references.

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The Swedish Radiation Safety Authority has a comprehensive responsibility to ensure that society is safe from the effects of radiation. The Authority works to achieve radiation safety in a number of areas: nuclear power, medical care as well as commercial products and services. The Authority also works to achieve protection from natural radiation and to increase the level of radiation safety internationally.

The Swedish Radiation Safety Authority works proactively and preventively to protect people and the environment from the harmful effects of radiation, now and in the future. The Authority issues regulations and supervises compliance, while also supporting research, providing training and information, and issuing advice. Often, activities involving radiation require licences issued by the Authority. The Swedish Radiation Safety Authority maintains emergency preparedness around the clock with the aim of limiting the aftermath of radiation accidents and the unintentional spreading of radioactive substances. The Authority participates in international co-operation in order to promote radiation safety and finances projects aiming to raise the level of radiation safety in certain Eastern European countries.

The Authority reports to the Ministry of the Environment and has around 300 employees with competencies in the fields of engineering, natural and behavioural sciences, law, economics and communications. We have received quality, environmental and working environment certification.

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