Earthquake activity in Fennoscandia between 1497 and 1975 and intraplate tectonics*

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The systematic collection of macroseismic data on earthquake occurrence in Fennoscandia began in the 1880's when the use of questionnaires was initiated. Prior to that time the macroseismic information is fragmentary and incomplete, although the pioneering works of Keilhau (1836), Thomassen (1888), Kjellén (1903), Kolderup (1913) and others have preserved the essential data at least on the largest earthquakes. Using the pertinent macroseismic and seismograph data available, we have constructed seismicity maps for Fennoscandia covering the time interval 1497 to 1975. The earthquake activity is subdivided in 3 primary zones: the Western Norway, the Telemark-Vänern, and the Bothnian seismicity zones; and 3 secondary zones: the Lappland, the Norwegian Shelf, and the Norwegian Sea seismicity zones. The latter two zones are not part of Fennoscandia but were included to bridge the gap between the well-defined interplate seismic activity along the Mid-Atlantic Ridge and the relatively diffuse intraplate seismicity pattern of Scandinavia itself. The on-going tectonic processes causing the earthquakes are discussed in terms of plate driving forces and the geological history of Fennoscandia. There is in a number of areas, such as along the Bothnian coast of Sweden, some correlation between pertinent geological and geophysical information and observed seismicity, even though the possible causal connections are not quite clear.

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The systematic collection of macroseismic data on earthquake occurrence in Fennoscandia began in the 1880's when the use of questionnaires was initiated. Macroseismic data prior to that time have to be extracted from newspaper reports and similar sources and are therefore less reliable, although the larger earthquakes are affected less in this respect. The oldest known report on seismic activity in Fennoscandia dates back to 1497 (Kjellén 1903, 1909) when a relatively large earthquake was felt in Sweden. The oldest reports for Norway and Finland date back to 1612 (Keilhau 1836, Kolderup 1913) and 1610 (Renqvist 1930), respectively. The classical analysis of available macroseismic observations is provided by Båth (1956), who published an excellent earthquake catalogue for Fennoscandia covering the period 1891–1950. Båth omitted earthquakes occurring prior to 1891 in order to obtain a homogeneous data base.

Instrumental observations of Fennoscandian earthquakes date back to 1904 and 1905 when the first mechanical pendulum seismographs were installed in Uppsala and Bergen. Due to their low magnifications (around 400 and 200) these instruments did not contribute much new information about the seismic activity in the area under consideration. However, in the period 1955–1965, the Fennoscandian seismograph network was expanded and the instrumental quality vastly improved by installation of modern, high-gain electromagnetic seismographs with magnification from 15,000 to 150,000. In 1971 another generation of instruments was introduced with the large aperture Norwegian Seismic Array (NORSAR) in southeastern Norway (Bungum, Husebye & Ringdal 1971, Bungum & Husebye 1974). The Fennoscandian network of stations (Fig. 1) represents a vast improvement in the capability of monitoring the seismic activity in this area; although we will demonstrate that it is not adequate for a complete seismo-tectonic study of local earthquakes.

The available macroseismic observations have been used among others by Kjellén (1903), Kolderup (1913), Kvale (1960), and Båth (1953, 1972) in correlating the seismic activity with postulated tectonic features such as faults related to the Caledonian mountain belt and the

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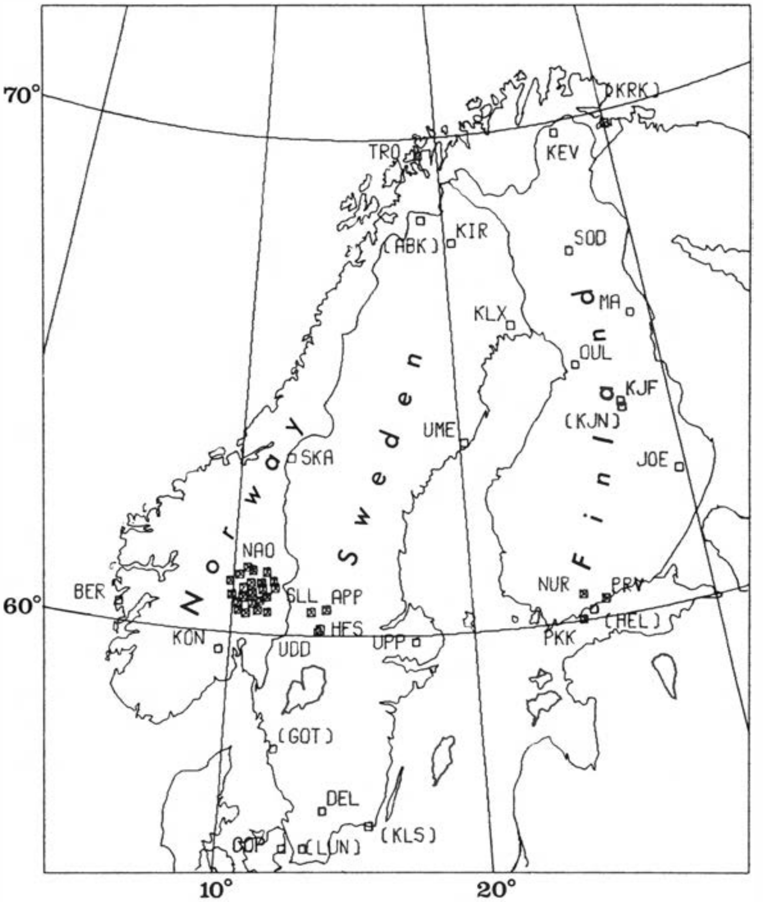

Fig. 1. Geographic locations of seismic stations in Fennoscandia. Stations no longer in operation (in 1976) are in brackets, and array stations are marked with crosses. The full station names are: ABK – Abisko, APP – Äppelbo, BER – Bergen, COP – Copenhagen, DEL – Delary, GOT – Göteborg, HEL –Helsinki, HFS – Hagfors, JOE – Joensuu, KEV – Kevo, KIR – Kiruna, KJF & KJN – Kajaani, KLS – Karlskrona, KLX – Kalix, KON – Kongsberg, KRK – Kirkenes, LUN – Lund, MA – Maaselkae, NAO – NORSAR (Norwegian Seismic Array), NUR – Nurmijärvi, OUL – Oulu, PKK – Porkala, PRV – Porvoo, SKA – Skalstugan, SLL – Stollet, SOD – Sodankylä, TRO – Tromsø, UDD – Uddeholm, UME – Umeå, UPP – Uppsala.

uplift of the Fennoscandian land mass in the Tertiary, and to stress release associated with the recent glacial rebound. More recently Husebye, Gjøystdal, Bungum & Eldholm (1975b) presented an analysis, based on instrumental observations, of the seismicity of the Norwegian and Greenland Seas and adjacent continental shelf areas. In that paper, attention is drawn to the relatively high earthquake activity in some areas off the mid-oceanic ridge, and in particular to areas on the continental shelves and to a seismicity zone in the northern Lofoten Basin (e.g., Fig. 5). Moreover, suggestions of symmetry between regions of high seismicity in western and northern Norway with similar regions in eastern Greenland were also discussed as these areas were juxtaposed prior to the opening of the Norwegian Sea in the early Tertiary. Aki & Husebye (1974) have discussed the stress distribution in the northern Lofoten Basin in view of reported strike-slip motion associated with an earthquake here. In addition to these investigations, Miyamura (1962), using Båth's (1956) macroseismic data, studied earthquake recurrence, i.e., the relationship between number of earthquakes observed in a given period and the magnitudes of the earthquakes (see also Karnik 1969, 1971). This type of problem is of great importance in earthquake hazard analysis and recently a detailed study of this kind, using the seismicity data presented here, has been undertaken by Husebye & Ringdal (1976).

The purpose of this work is twofold. The first part is devoted to a review of all macroseismic and instrumental information available on earthquake occurrence in Fennoscandia with special emphasis on Norway. The second part discusses these data in the context of intraplate tectonics.

Earthquake occurrence in Fennoscandia

A considerable amount of macroseismic data and to some extent instrumental records are available on the seismic activity in Fennoscandia. With the notable exceptions of Kolderup (1913) and Båth (1956), the construction of systematic and comprehensive earthquake catalogues has seldom been attempted. We have carefully scrutinized all known published information and also prepared earthquake catalogues for the time period 1497 to 1976. Our efforts, including considerations of the homogeneity and quality of the original data, fell into 3 parts; the macroseismic material for the period 1497–1890, macroseismic and to a small extent instrumental data for the period 1891–1950, and the instrumental and to a small extent macroseismic data for the period 1951–1975. This work, including earthquake lists, earthquake intensity decay with distance, depth of focus, seismic hazards as inferred from macroseismic observations, aftershock sequences, etc., has recently been presented in a comprehensive report by Husebye, Gjøystdal & Bungum (1975a). These data, together with additional data from Finland and Sweden, which we believe may contribute to a better understanding of the seismo-tectonic processes characteristic of Fennoscandia, will be presented in the form of seismicity maps in this paper.

Interval 1497–1890

The main macroseismic data sources are the works of Keilhau (1836), Thomassen (1888, 1890), and Kolderup (1913) for Norway, the works of Kjellén (1903, 1909) for Sweden, and the work of Renqvist (1930) for Finland. It should be pointed out that the seismic activity is relatively higher in Norway and Sweden than in Finland, as evident from Fig. 2 and also Figs. 3 and 6.

We remark further that probably the strongest earthquake in this period is the one occurring at Lurøy in N. Norway (66.4 N, 12.8 E) on 31 August 1819, with a magnitude of approximately 6.0. Although the earthquake caused only very minor damages, it was reported that stones were rolling down from nearby hills and local rivers became muddy for several days (Keilhau 1836). Another unusual feature of this large earthquake was its long sequence of 'aftershocks' occurring in the Lurøy area in the subsequent 10 years.

Interval 1891–1950

The classic work here is that of Båth (1956); his earthquake catalogue for the above interval is both homogeneous and of high quality. This catalogue has generally served as a standard reference for seismic activity in Fennoscandia, and the corresponding seismicity map is shown in Fig. 3.

The largest earthquake occurring in this period is the one in Oslofjorden (59.2 N, 10.5 E) on 23 October 1904, with a magnitude of approximately 6.0. The earthquake was widely felt in southern Scandinavia (Fig. 4) but no serious damage was reported. Like the large Lurøy earthquake of 1819, also the Oslofjorden earthquake was accompanied by a sequence of 'aftershocks' which are plotted in Fig. 4 together with intensity expressed through isoseismal lines (Kolderup 1905, Svedmark 1908, Austegard 1975).

Interval 1951–1975

In general, macroseismic observations are inferior to seismograph records of earthquakes in seismicity studies. The reason is that from the seismogram information one can more easily derive estimates for hypocenter location, energy release or magnitude, focal mechanism

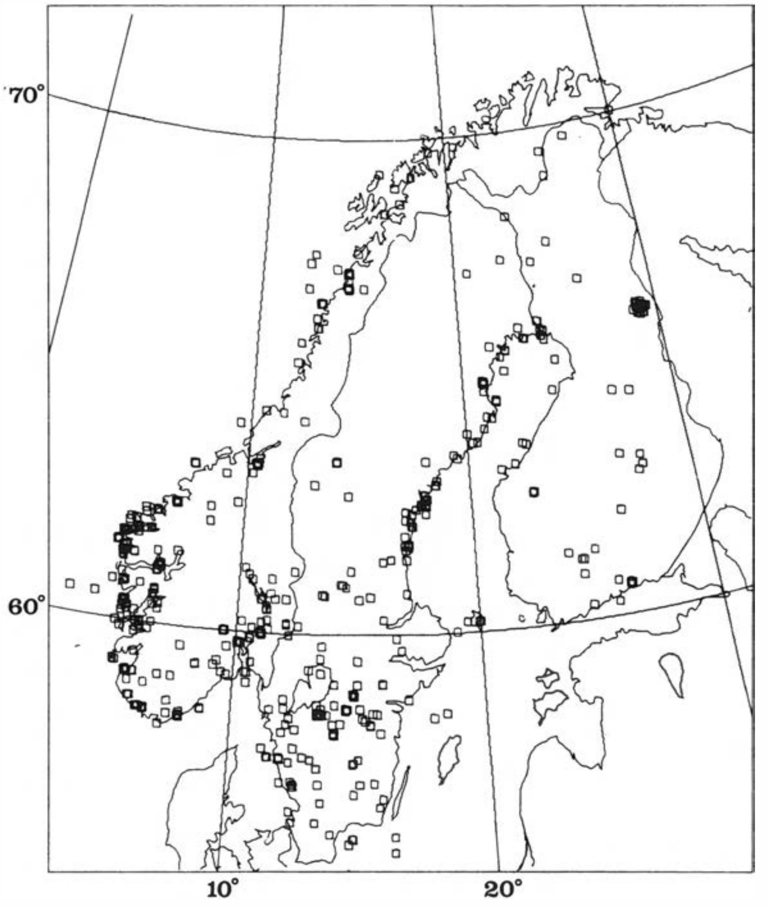

Fig. 2. Seismicity of Fennoscandia for the time period 1612–1890 as inferred from the macroseismic material collected and published by Keilhau (1836), Thomassen (1888, 1890), Kjellén (1903, 1909), and Kolderup (1913). In some cases the given epicenters are randomly moved within the uncertainty limits so as to reveal repeated earthquake occurrences in the same location. The same technique has also been applied to Figs. 3–4 and 6–8.

and wave attenuation. However, macroseismic data are still routinely collected at seismological observatories (Kvale 1959, Sahlstrøm & Båth, 1958, Nilsen 1965) although publication of this material has become rather infrequent in recent years.

The Fennoscandian seismograph network was extended and vastly improved in terms of instrument quality during the period 1955–1965 and we should now be in a position to undertake a more detailed study of the seismicity of Fennoscandia which is characterized by the relatively infrequent occurrence of small earthquakes. Strangely enough, macroseismic information is still highly esteemed simply because a comprehensive and systematic earthquake catalogue primarily based on local seismograph records has, to our knowledge, not been published. This may be attributed to shortcomings of the local seismograph network, and we would here point to the large station separation, the poor azimuthal coverage for earthquakes in the coastal areas of Norway, and the difficulties in obtaining large numbers of original seismogram records (which are stored in three different countries). With regard to the latter, it means that the crucial phase identification or interpretation of the seismic records is

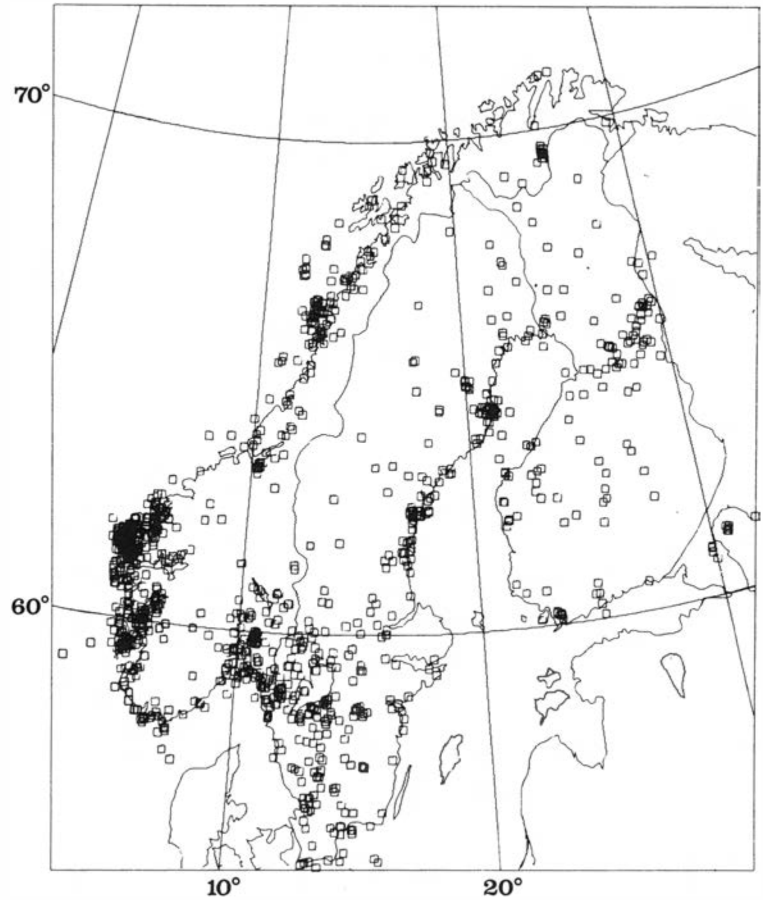

Fig. 3. Earthquake occurrence in Fennoscandia for the time period 1891–1950 as published by Båth (1956).

often based on second-hand information, i.e., as given in local seismic bulletins. The most serious problem, however, is that of discrimination between the relatively few (tectonic) earthquakes and the very many artificial events from quarry blasts, naval activities in the adjacent seas, etc. These shortcomings of the seismograph network are, however, not critical for earthquakes with magnitude greater than 4.0–4.5, because these events are also recorded by stations outside Fennoscandia. Actually, such data were used by Husebye et al. (1975b) when discussing the seismicity of the Greenland and Norwegian Seas and the adjacent coastal areas (e.g., Fig. 5). More recently, Dahlman, Ohlson & Slunga (1975) have published a seismicity map for Fennoscandia which is based on a detailed study of instrumentally recorded earthquakes in the interval 1968–1972. In addition, Noponen (1971–1975) has reported on a routine basis both earthquakes and large chemical explosions occurring in Fennoscandia. All macroseismic and instrumental data available to us have been used in constructing the seismicity map presented in Fig. 6 covering the interval January 1951 to December 1975.

The information available to us on earthquake occurrence in Fennoscandia has been summarized in Figs. 2–7, which are the most comprehensive collection of seismicity maps published for this area. We note in passing that the macroseismic and instrumental data essentially give the same pattern of the seismic activity, as Figs. 2, 3, and 6 clearly demonstrate.

The assessment of seismic hazard or potential earthquake damage on human constructions like houses, nuclear power plants, etc., is mainly based on past records of large earthquakes. A

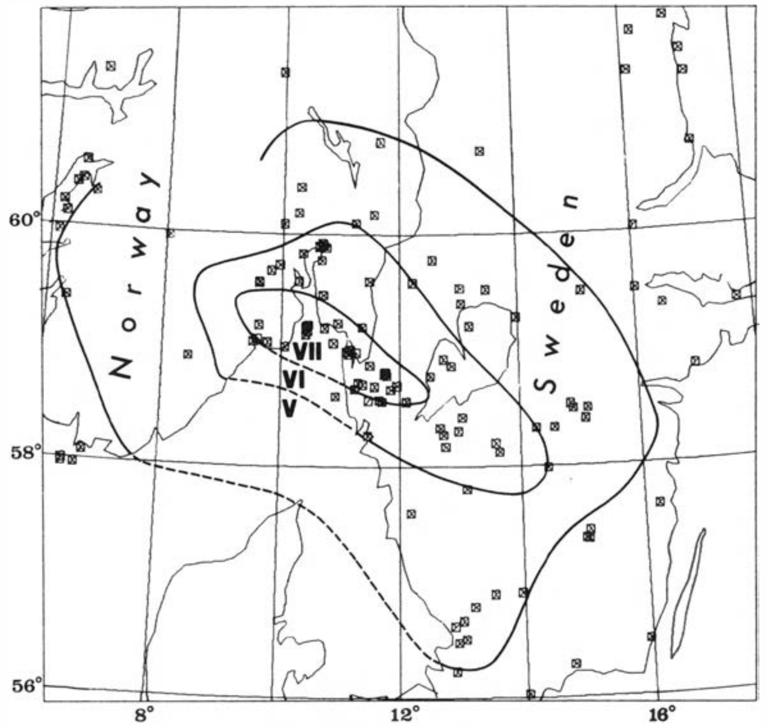

Fig. 4. Macroseismically located earthquakes in southern Norway and Sweden for the time period 1904–1910, i.e., for six years following the large Oslofjord earthquake in 1904. The isoseismals for this event are also given (Austegard 1975), with roman numbers indicating intensity levels.

critical parameter in this respect is the space-time relationship between the largest earthquakes within the region of investigation. It is peculiar that most of the largest shocks (Fig. 7) are restricted to the period 1863–1913, while two other strong earthquakes took place between 1819 and 1834. More recently, some relatively large earthquakes have been located instrumentally off the western coast of Norway. There is no indication of spatial migration in the distribution of the largest earthquakes. Neither do we find any clear regularity in the occurrence of large events in specific areas. In other words, the very largest earthquakes observed seem to be isolated phenomena. This points towards exceptionally long recurrence intervals, i.e., a large time lag between two large events occurring within a tectonically uniform area. As it is considered unlikely that any major earthquakes occurring after 1600 are left unreported, the recurrence interval appears to be larger than 350 years for earthquakes magnitude M larger than around 6.0. Specifically, the mentioned case study for risk analysis (Husebye & Ringdal 1976) gave an estimated recurrence time of 400 years for earthquake intensity $I_0 = 8$ (corresponding to $M \sim 6.5$) based on macroseismic observations for the S. Sweden area and using Gumbel's (1958) extremal value statistics.

Precision of estimated earthquake focal parameters

In order to assess objectively the seismicity information for Fennoscandia presented in Figs. 2–6, we have to comment on the precision of the estimated earthquake parameters.

The location precision using macroseismic observations is probably better than 30 km for most small and medium sized events. However, the epicenter precision often decreases with increasing intensity, and this applies in particular

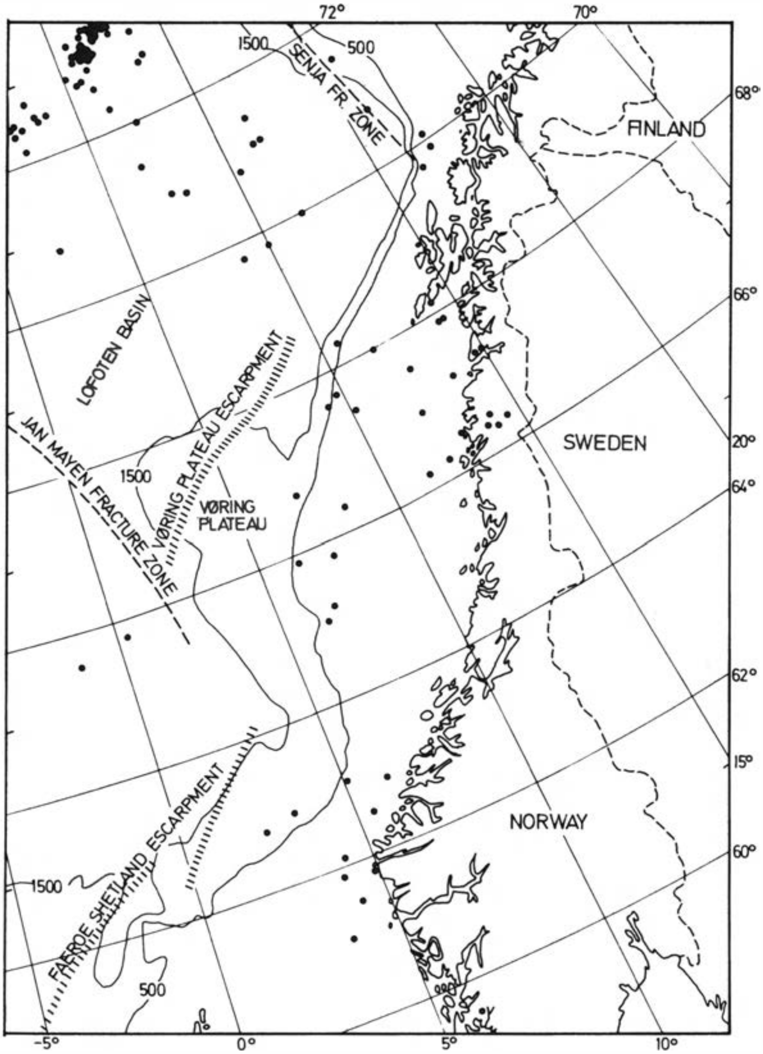

Fig. 5. Instrumentally recorded earthquakes in Fennoscandia and the Norwegian Sea for the time period 1955–1973, as predominantly reported by U.S. Geological Survey. The figure is redrawn from Husebye et al. (1975b).

to earthquakes occurring prior to 1890. Moreover, biased errors can be severe and cannot be estimated. For instance, earthquakes in coastal areas may occasionally have been grossly mislocated due to lack of seaward observations. In case of epicenter determinations based on instrumental data, the precision increases with increasing event magnitude, although biased errors may be severe, in particular when the azimuthal coverage is poor. For example, epicenter locations as estimated by different seismological agencies for a particular event sometimes exhibit distance separations in excess of 150 km.

Focal depths as calculated from macroseismic data depend critically on assumed values for the geometric damping. Husebye et al. (1975a) have calculated focal depths of Fennoscandian earthquakes, and assuming relatively low attenuation values, they found that 55–80 % of the shocks probably have depths less than 20 km. This gives slightly shallower events than previously reported by Båth (1956). Indirect evidence from instrumental observations like P and S wave phase identifications supports the macroseismic results. Lately, Båth (1975) has reported a few very shallow earthquakes in Sweden using R_g-dispersion as a depth criterion.

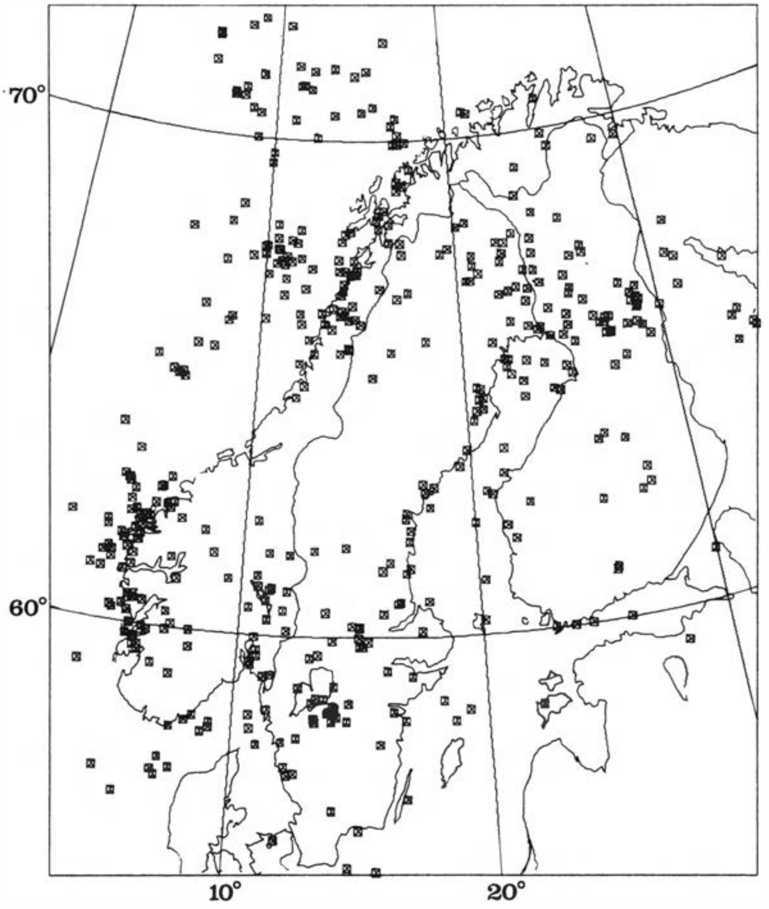

Fig. 6. Instrumentally and macroseismically located earthquakes in Fennoscandia for the time period 1951–1975, using all available published solutions. When several agencies have reported on the same event, the solutions have been checked against each other. The catalogues used in analysis have been critically examined in order to remove possible explosions.

Maximum observed macroseismic intensity may be wrong by one unit so that the corresponding uncertainty in magnitude is probably around 0.5 units. On the other hand, as no reliable P-wave amplitude-distance curve has been established for Fennoscandia, the magnitude cannot be estimated from instrumental observations with sufficient precision. The earthquake magnitudes in this paper have been calculated from the macroseismic data according to the following formula:

$$M = -1.33 + 1.78 \log_{10} R - \\ -0.89 \log_{10} \left[10^{\frac{2(I_0 - 2)}{\gamma}} - 1\right] + 0.61\, I_0$$

where R is the radius of perceptability and I_0 is the maximum intensity. This is the same formula as the one previously used by Båth (1953), who used a value of $\gamma = 6$ for the attenuation parameter. It has been argued recently (Austegard 1975, Husebye et al. 1976, Husebye & Ringdal 1976) that $\gamma = 4.5$ is a more reasonable value for Fennoscandia, so we have used this new value for γ and recalculated all macroseismically derived magnitudes. For $I_0 = 3$, this decreases (as compared to $\gamma = 6$) the M values by 0.2 units and for $I_0 = 6$ by 0.4 units, leading to a smaller disperson in the magnitudes and consequently somewhat higher values for the slope (b-value) of the frequency-magnitude

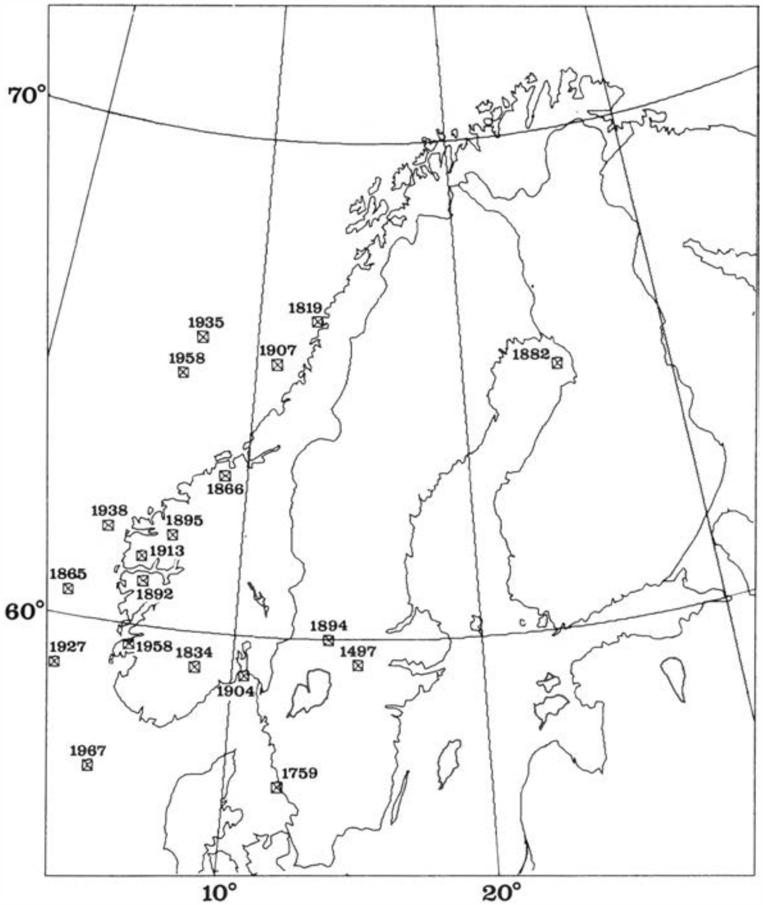

Fig. 7. Fennoscandian earthquakes for the time period 1497–1975 and with magnitude greater than 5.0. The year of occurrence is given for all the events.

distribution. This relative decrease in seismic magnitude for the larger earthquakes seems reasonable in view of their macroseismic effects, since no serious damage has been reported throughout historical times. For example, Holm & Lande (1976) have carefully scrutinized Swedish historical records, the oldest dating back to 1164, but they could not find any evidence of reported earthquake damage.

In a few cases small earthquakes reported may actually be caused by cracking of the soil during extreme low temperatures, meteorological phenomena, etc. The problem of false alarms is quite severe for instrumental data as the number of artificial explosions recorded is much larger than the number of recorded earthquakes, and it is sometimes difficult to discriminate between the two types of seismic sources. However, a simple quantitative test for discriminating between earthquakes and explosions is to check the time of occurrence of the seismic event population as a function of local time of day. Explosions are usually confined to the prime working hours while the earthquake distribution should be flat, as discussed by Steinert, Husebye & Gjøystdal (1975).

The most essential information needed for seismo-tectonic studies is the focal mechanism solution. This type of dynamic parameter which is commonly used for deducting the intraplate stress field (Sbar & Sykes 1973) is not available for any Fennoscandian event, the main reason being that no large-magnitude event has occurred within this area in recent years.

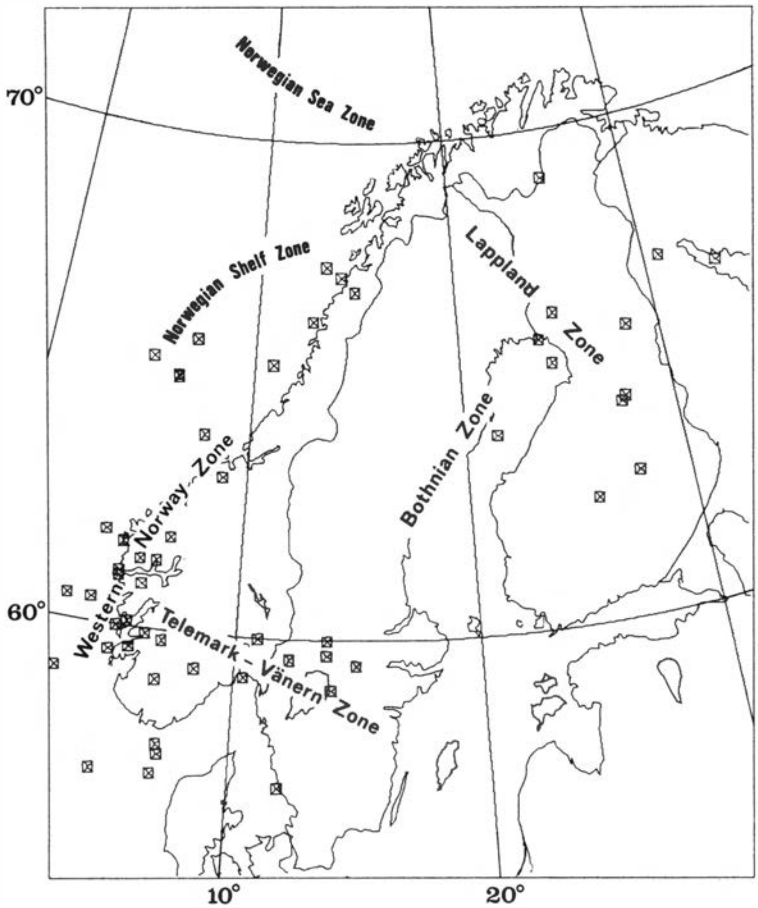

Fig. 8. Fennoscandian earthquakes for the time period 1497–1975 and with magnitude greater than 4.5. An outline is also given of the seismicity zones defined on p. 63–65. The 3 southernmost zones are most active and therefore called primary, and the 3 northernmost ones are the secondary zones.

Stress generating mechanisms and possible interpretation of Fennoscandian seismicity data

Stress generating mechanisms

It is now recognized that thermal convection within the mantle is primarily responsible for mountain building, volcanism, and seismicity at plate margins as formulated by the concept of global tectonics (Isaacs, Oliver & Sykes 1968). In contrast, observed tectonic activity *within* plates, of which the seismicity of Fennoscandia is an example, cannot directly be explained by processes at or related to plate boundaries. In this respect, recent progress in plate tectonics on mechanisms of stress accumulation in the lithosphere (Turcotte & Oxburgh 1976) and the relative importance of different plate driving mechanisms (Forsyth & Uyeda 1975, Solomon, Sleep & Richardson 1975) are likely to give a better understanding of intraplate distribution of stresses. The usefulness of these concepts has already been demonstrated in studies of earthquakes within a number of plates (e.g., Mendiguren 1971, Forsyth 1973, Sbar & Sykes 1973, Sykes & Sbar 1973, Fitch, Worthington & Everingham 1973, Smith & Sbar 1974), and this approach could therefore be relevant to an improved understanding of the Fennoscandian seismicity. We will therefore briefly review recent results on stress generating mechanisms, geological and geophysical information pertinent to the area, and then in this context discuss our seismicity data as presented in the previous section.

There are many sources of stress in the lithosphere, and a convenient subdivision of stress generating mechanisms is: stresses tied to the driving forces of present plate motions; and remnant stresses reflecting the different tectonic cycles undergone by Fennoscandia through geological times. With respect to plate motion the most important driving forces are, according to Forsyth & Uyeda (1975), the gravitational ridge push, the mantle drag force, and the descending slab pull. The latter is not relevant here, as there is no evidence of Mesozoic or Cenozoic subduction in Fennoscandia.

The ridge push is an important plate driving force, and is physically explained in terms of the gravitational push exerted by the upwelling material at mid-oceanic ridges (Orowan 1964, Forsyth & Uyeda 1975). With the opening of the Norwegian Sea in the Early Tertiary, this force is likely to have created compressive stresses within the plates on either side of the mid-oceanic ridge. Intraplate earthquake studies imply dominant horizontal stresses with maximum compression approximately perpendicular to the oceanic ridge axis (Mendiguren 1971, Forsyth 1973, Sykes & Sbar 1973). In this respect the Norwegian and Greenland Seas are atypical as the spreading flow-lines are parallel over the entire area (Talwani & Eldholm 1977), while the Mohns Ridge and Knipovich Ridge have different strike directions.

The mantle drag force is connected to a phenomenon by which the asthenosphere is either passively resisting or alternatively dragging the lithosphere. This ambiguity cannot be properly resolved before data on mantle flow rates become available. The relative importance of the mantle drag force on local seismicity is very difficult to assess although plates having a predominant continental structure move relatively slowly, and this applies in particular to the Eurasian plate (Forsyth & Uyeda 1975). This again points towards the possibility of greater viscosity under the continents than under the oceans, i.e., the former has no clear low velocity zone in the asthenosphere in contrast to oceanic areas (Alexander 1974).

There is also another aspect of plate motion, namely, generation of so-called membrane stresses as discussed by Turcotte (1974). The generation mechanism here is related to changes in the principal radii of curvature when the assumed rigid plates move over the surface of the imperfectly spherical earth. Oxburgh & Turcotte (1974) have discussed the evolution of the East African Rift zone in terms of membrane stresses. The same mechanism may have been instrumental in formation of the central rift system and graben structures in the North Sea, i.e., a consequence of the predominantly northward migration of the western part of the Eurasian plate following the break-up of Pangaea in Permian times (Dietz & Holden 1970). On the other hand, H. Ramberg (1971) has simulated structures resembling rift valleys and also oceanic ridges utilizing dynamic models in which a buoyant body rises in the mantle. The rising body would spread laterally below the crust and accordingly tension fractures tend to develop.

In order for remnant stresses in the lithosphere to be significant, stress accumulation over geological times is required. The presence of ancient mountain chains is evidence against stress relaxation over periods which are of the order of 10^7–10^8 years and possibly even 10^9 years (for references, see Eisbacher & Bielenstein 1971, Ranalli & Chandler 1975). For example, the two dominant zones of earthquake occurrence in Great Britain (Lilwall, 1976) are coincident with the remnants of the Caledonian and Hercynian mountains in this region. Another example is the gravity-high associated with the Oslo Graben (I. B. Ramberg 1972, 1976), indicating that elastic stresses in this area may have been preserved since Permian times. As long as the lithosphere behaves elastically, which is a reasonable assumption for relatively small stresses, the problem is linear and stresses are additive. Thus the present state of stresses in the Fennoscandian lithosphere may be the result of several periods of stress generation and possibly also of stress relaxation, inasmuch as this continental part of the Eurasian plate has been through several tectonic cycles throughout geological times. For the much younger oceanic lithosphere of the North Atlantic Ocean, the state of stress should in principle be considerably simpler as this area has gone through a well-defined cycle and has a relatively uniform composition and structure.

The types of stresses discussed in the previous section are tied to past movements of the Eurasian plate and the opening of the Norwegian Sea and the Arctic Ocean. We cannot ignore, however, possible stress effects due to loading and unloading of the lithosphere connected to the Tertiary uplift of western Fennoscandia, sedimentation off the Norwegian coast, and recent glacial episodes. The corresponding changes in the vertical component of stress will result in horizontal stresses if the rock behaves elastically. However, when erosion or sedimentation occurs the rock temperature changes, so a complicating factor here would be thermal stresses as discussed by Turcotte & Oxburgh (1976) and Haxby & Turcotte (1976). The thermal stresses, being tensional, tend to cancel the overburden compressional stresses. An example here is the extreme variation in crustal thickness occurring at continental margins, i.e., this feature may contribute significantly to the state of stress in these areas. It should also be noted that the observed Fennoscandian glacial rebound is inconsistent with the hypothesis that it is due to viscous flow (Jeffreys 1975).

Geological and geophysical information pertinent to Fennoscandia

The development of the Fennoscandian shield has involved successive orogenic cycles which have been described in considerable detail by many geologists (e.g., O. Holtedahl 1960, Zachrisson 1973). The Caledonian cycle began with the geosynclinal sedimentation off western Norway in late Precambrian times and the subsequent folding period was completed at the end of Silurian. Moreover, H. Ramberg (1966) suggests on the basis of centrifuged dynamic model studies that the rise of basal-gneiss culminations in the Caledonides of Norway (and other orogenic belts) represents a buoyancy phenomenon in response to an unstable stratification in the crust in the geosynclinal region. The development of the interesting geological province, the so-called Oslo Graben, began in late Carboniferious time (approximately 300 m.y.b.p.) (Oftedahl 1960, I. B. Ramberg 1976). A less pronounced graben structure is hypothesized in the lake Vättern area (Lind 1972, see also Fig. 9).

In early Tertiary the opening of the Norwegian Sea started (Talwani & Eldholm 1977), and during this period an uplift of western Fennoscandia took place (Torske 1975). Moreover, since the end of the last ice age, about 10,000 years ago, a glacial uplift of Fennoscandia has taken and is still taking place. Based on both geological and geodetic data, Mørner (1975) suggests that the Fennoscandian uplift has a double nature: one glacic-isostatic factor that continuously decreased with time and dies out some 2–3000 years ago; and one tectonic factor that remains constant (see also O'Connell 1976).

The post-Permian geological history of the North Sea and Barents Sea, adjacent to the relatively stable Fennoscandian platform, is different. Tectonic features of interest here are a postulated triple-junction in the Skagerrak Sea, the Polish-Danish furrow with the Fennoscandian Border zone forming the boundary between the stable Baltic Shield or platform, and, at that time, the rapidly subsiding intracratonic Permian basin of the North Sea (e.g., see Ziegler 1975, Whiteman, Naylor, Pegrum & Rees 1975). The Barents Sea has also been geologically mapped in recent years (for references see Demenitskaya & Levin 1970, Renard & Malod 1974, Eldholm & Talwani 1977). We note that the seismic activity in both the North Sea and Barents Sea is very weak (Husebye et al. 1975b), indicating that the stress distribution in these sedimentary basins is different from that in Fennoscandia.

Evidence on tectonic movements such as locations of exposed fault lines and border lines between dominant geological provinces in Fennoscandia is displayed in Fig. 9. In the case of Finland, where available tectonic information is relatively abundant, the picture is somewhat blurred as Tuominen, Aarnisalo & Søderholm (1973) and Kukkamäki (1963) present a very large number of geological lineaments and fault lines. The structural trends however, are dominantly parallel and perpendicular to the Caledonian folding axis, even though the Finnish rocks are of Precambrian age.

Numerous seismic gravity and magnetic surveys and some heat flow measurements have been undertaken in Fennoscandia (e.g., see Der & Landisman 1972, Massé & Alexander 1974, Aki, Christoffersson & Husebye 1977, I. B. Ramberg 1976, Åm 1975, Swanberg et al. 1974 and the references therein), but this kind of information has seemingly only an indirect bearing on the seismic activity. Moreover, the upper mantle P-velocity structures for the Baltic Shield as recently derived by King & Calcagnile (1976) are typical for shield areas, i.e., there is no strong evidence for a low velocity zone in the asthenosphere (Alexander 1974). Also the available in-situ stress measurements for Fennoscandia indicate a somewhat complicated stress pattern, and relevant data are included in Fig. 9 (Hast 1969, 1973, Myrvang 1975, Ranalli & Chandler 1975).

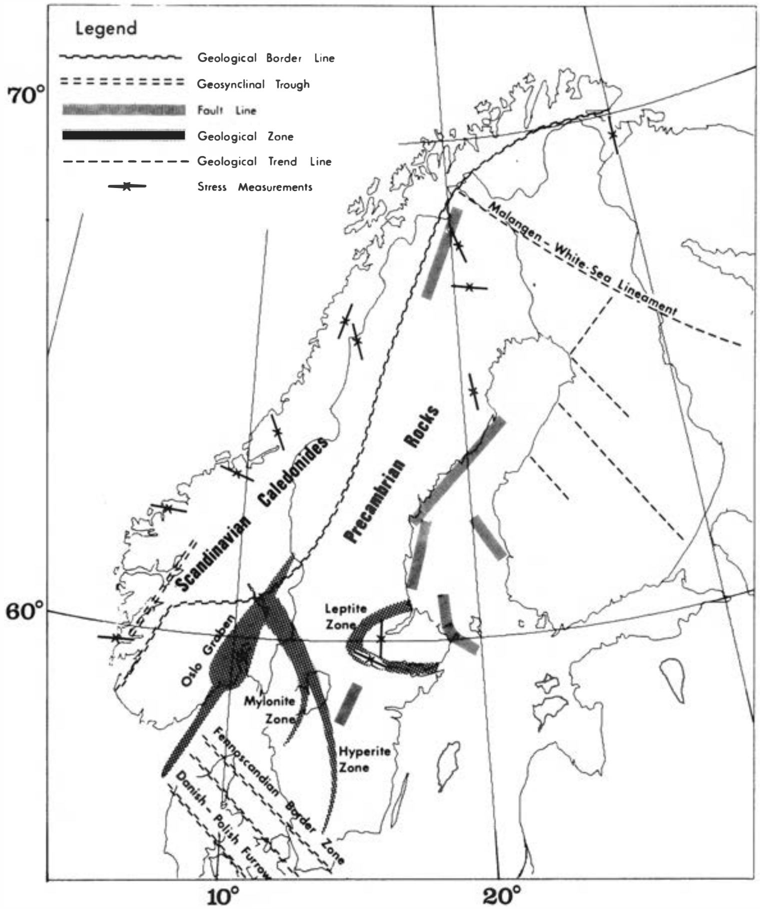

Fig. 9. Sketch map showing geological and tectonical trends and structures which are considered relevant for the present Fennoscandian seismicity study. The in-situ stress measurements plotted are taken from Hast (1973) and Ranalli & Chandler (1975). The Oslo Graben continues both northward and southward as the Oslo Rift zone. The mylonite, hyperite, and leptite zones are redrawn from Stephansson & Carlsson (1976), while the fault lines in the Baltic Sea are redrawn from Flodén (1973). The Vättern graben structure in S. Sweden (Lind 1972) and the Holocen Pärve fault in N. Sweden (Lindqvist & Lagerbäck 1976) are also shown. The geological trend lines in Finland are after I. B. Ramberg (1973), while the Malangen-White Sea lineament is after Tuominen et al. (1973).

Earthquake occurrence in Fennoscandia and parts of the Norwegian Sea

In the previous sections we have discussed the tectonic forces likely to be dominant in generating stresses which subsequently cause the Fennoscandian earthquakes and also available evidence on past and present tectonic movements in this area. This constitutes the context in which we will discuss our seismicity data.

The seismicity pattern of Fennoscandia is somewhat diffuse, which is normal for intraplate earthquake occurrence (Sbar & Sykes 1973, Smith & Sbar 1974), in contrast to the seismic activity along plate boundaries like the mid-oceanic ridge in the Norwegian Sea. For several reasons we found it convenient to subdivide the Fennoscandian earthquakes in three primary zones (see Fig. 8): namely, the western Norway zone, the Telemark-Vänern zone, and the Bothnian zone; and three secondary zones: namely, the Norwegian Sea zone, the Norwegian

Shelf zone, and the Lappland zone. These seismicity belts are convenient for a geographical, and to a large extent also tectonic, subdivision. Notice that these zones, which comprise almost all observed seismic activity, are to the south roughly bordered by the Fennoscandian Border zone and the Danish-Polish Furrow (Fig. 9) and to the north by the Malangen-White Sea lineament of Tuominen et al. (1973). Tectonically the Norwegian Sea zone is not a part of Fennoscandia, but it is included here to bridge the gap between earthquakes on Mohns Ridge and Knipovich Ridge (see Husebye et al. 1975b) and intraplate earthquakes within Scandinavia.

The Norwegian Sea seismicity belt is unusual as it hardly has any counterpart in other areas of the mid-oceanic ridge system in the North Atlantic. Lazareva & Misharina (1965) have reported strike-slip fault plane solutions for one event in this zone and one on the Knipovich Ridge. The puzzling earthquake occurrence in the Norwegian Sea belt and within the Lofoten Basin has been discussed in some detail by Aki & Husebye (1974) and Husebye et al. (1975b).

The Norwegian Shelf zone (Fig. 4) is also anomalous insofar as few if any other North Atlantic shelf areas exhibit a comparable seismic activity. The unique feature here is the young, thin oceanic crust westward and the old, thick continental crust including sedimentary layering of 6–10 km thicknesses to the east (Talwani & Eldholm 1972). The tectonic evolution of passive continental margins has been discussed by many scientists (e.g., see Sleep 1971, 1973). In this respect the dominant tectonic forces considered to be important are those associated with the cooling of the newly founded adjacent oceanic lithosphere, and the subsequent sedimentary loading. Also the 'hot creep' process suggested by Bott (1971), by which a thinning and thickening occurs of the continental and oceanic crust respectively, may be of some importance. However, the suggested evolutionary processes should generate similar stress patterns at other continental margins in the Norwegian and Greenland Seas, while the seismic activity there actually is very modest as compared to that in the Norwegian Shelf zone. In other words, tectonic forces not directly associated with the evolution of the continental margin itself may be of some importance in this particular case. We would here point to the Tertiary uplift of Fennoscandia which may have included the continental shelf area as well (Talwani & Eldholm 1972).

The western Norway seismicity belt (Fig. 8) is confined to the coastal areas in northern Norway; the earthquake activity is weak in the central part (Trondheim area), while the epicenter distribution becomes more dispersed south of 62°N. One branch seemingly follows the Caledonian Synclinal Trough (Fig. 9), while another branch follows the coast. The west coast area of Norway is relatively prominent seismically as discussed by Husebye et al. (1975b). A characteristic feature here is that this zone is within the Caledonides and its strike direction is parallel to the folding axis; we take this as evidence for the importance of remnant or locked-in stresses from the above folding period. Another feature is that most of the epicenters are confined to the coastal areas where the relief is relatively pronounced, which in turn may indicate additional loading stresses and/or a zone of weakness. On the other hand, we did not find a clear correlation between the observed seismicity and the basal culminations and related tectonic features in the Scandinavian Caledonides as discussed by H. Ramberg (1966). Finally, the Caledonides in Scotland are relatively seismic active (Lilwall 1976), also pointing towards the importance of the Caledonian folding in present-day earthquake occurrence in western and northern Norway.

The Telemark-Vänern seismicity zone, which represents a geographical envelope of earthquake epicenters in this area, is typified by graben structures. The most prominent one is the Oslo Graben (for geological and geophysical descriptions, see Oftedahl 1960, I. B. Ramberg 1976, Aki et al. 1977), which is the northern chain in Stille's (1925) Mittelmeer-Mjøsen zone. According to Lind (1972), the lake Vättern area (58°N, 15°E) is a minor graben, while a number of NS-striking faults have been localized in the lake Vänern area. The above features, together with the mylonite and hyperite zones (for details, see Stephansson & Carlsson 1976), are included in Fig. 9. A comparison with the previously presented seismicity data in Figs. 2, 3, and 5 gives some clustering of epicenters around the mentioned structures. This points towards a causal connection between these structures and earthquake occurrence, i.e., release of locked-in stresses. However, in view of the relatively large uncertainties in epicenter locations we refrain from a more detailed discussion of earthquake occurrence in this particular area and available geotectonic information.

The Bothnian seismicity belt, parallel to the Caledonian folding axis, goes from the Gävle area (ca. 60.7° N) to the northern end of the Gulf of Bothnia. The latter area has the most pronounced earthquake activity, which supports the assumption that there is a correlation between the relatively strong glacial uplift here and earthquake occurrence in this particular area (Kjellén 1903, Båth 1953). Also this zone is characterized by block faulting, and Flodén (1973) gives geophysical evidence in support of dominant fault lines in the coastal areas south of Ørnskøldsvik (ca. 63.5°N), as shown in Fig. 9. A number of other fault lines further to the east (Fig. 9) have been hypothesized mainly on seismic profiling (see Flodén 1973, and Stephansson & Carlsson 1976), but in this case there is no clear correlation with observed seismic activity. It is noteworthy that Lundqvist & Lagerbäck (1976) have reported Holocene tectonic movements along the so-called Pärve fault in Swedish Lappland (Fig. 9), although the seismic activity in this area is modest. The Bothnian zone appears to be separated from the Telemark-Vänern zone by an area of exceptionally low seismicity which is coincident with the leptit zone shown in Fig. 9. The Lappland seismicity zone is only weakly defined by the data available to us, but a series of earthquakes in 1973/74 was entirely within this zone (Porkka & Korhonen 1975). It should also be noted that this area has always been thinly settled, so the macroseismic information available would necessarily be scarce. Evidence in support of this zone is the pronounced Malangen-White Sea lineament based on Nimbus satellite imagery photography (Tuominen et al. 1973).

The above seismic zones account for nearly all seismic activity in Fennoscandia during the last five hundred years. The most notable exceptions are some activity near Kristiansand in southern Norway, a weak zone along the west coast of Sweden, and the dispersed earthquake occurrence in central and southern Finland. In the latter case, Pentillä (1963) and Teisseyre, Pentillä, Tuominen & Vesanen (1969) attempted zoning of the seismicity data. Observed trends, although not very pronounced, are dominantly parallel or perpendicular to Caledonian folding axis as is the case of the mentioned trends in geological observations. In certain areas there is geological evidence of deep faults or fractures, for example, in Jämtland as discussed by Strømberg (1974), while no significant earthquake activity has been reported for this area.

Finally, we would remark that the seismic activity in Fennoscandia, which is typical for intraplate earthquake occurrence, in some areas exhibits a reasonable correlation with the geotectonic-geophysical information available. However, other areas like the Fennoscandian Border Zone and graben structures in the North Sea exhibit a negligible seismicity. Consequently, we have refrained from a too detailed comparison between observed seismic activity and available geophysical data, because the causal connection between these two types of manifestations of past and present tectonic movements is likely to be complicated. The very reason for this may be that Fennoscandia has been through several tectonic cycles in geological times, which in turn is reflected in the present presumed complex stress distribution. Another complicating factor is that the observed seismicity zones necessarily are broad due to the relatively imprecise epicenter locations. However, in areas where the observational data is relatively abundant, as is the case for central Asia and California, numerical pattern recognition techniques have proved very useful in joint interpretations of seismicity and geotectonic information (for references, see Gelfand et al. 1972, 1976). On the other hand, an improved understanding of the on-going seismic activity in Fennoscandia and at the same time a better assessment of dominant stress sources requires at the present stage primarily focal mechanism studies of earthquakes occurring within this area.

Conclusions

The seismicity of Fennoscandia is typical for intraplate regions, being characterized by relatively small magnitude earthquakes of infrequent occurrence. The probably two largest earthquakes reported had magnitudes around 6.0 and took place near Lurøy, N. Norway, in 1819 and in Oslofjorden in 1904.

The seismicity information presented in Figs. 2–8 covering the time interval 1497–1975 indicates a surge in seismic energy release during the period 1863–1913.

The seismicity of Fennoscandia is mainly restricted to three geographical zones: the Western Norway, the Telemark-Vänern, and the Bothnian zones.

The correlation between available geotectonic and geophysical information and seismic activity is with few exceptions not very clear, and this is also typical for intraplate earthquake occurrence in general.

The on-going tectonic processes causing the earthquakes in Fennoscandia are discussed in the context of plate driving forces and the geological history of the region. A more detailed discussion here must await the availability of fault plane solutions for Fennoscandian earthquakes.

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