Paper No. 144, Bull. ISET, Vol. 11. No. 2. June 1974, pp. 38–54

SEISMIC ZONING OF FENNOSCANDIAN SHIELD

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INTRODUCTION

The major earthquake zones of the world are found in (i) the Circum-Pacific belt, (ii) the Alpine-Himalayan belt and (iii) the mid-oceanic ridges of the Atlantic, Indian and Pacific oceans. These earthquake-belts which cover large segments of the Earth, have been studied in detail by various workers, but little attention has been paid to earthquake occurrences on continental shields and platforms. These areas are assumed to be stable with very low earthquake probabilities for causing significant damage to life and property.

Systematic records of earthquake occurrences in shield regions have been available only since recent times. Before the establishment of seismological stations it was only the earthquakes responsible for damage to life and property which were studied and documented. Many earthq uakes which may have occurred in inaccessible or thinly populated regions failed to arise any curiosity for their study. This lack of data has supported the general belief that shield regions are seismically stable and pose no earthquake hazard.

Shallow earthquakes, which are mostly responsible for damage at or near ground surface, represent the mainfestations of the tectonic processes now in action in the crust and upper mantle. The tectonic processes also control the development of the geotectonics and physiography of the region. The crustal structures and physiographic features of a region can thus be utilised for the demarcation and identification of seismotectónic belts for seimic regionalisation and preparation of seismic zoning maps for design and construction of earthquake resistant structures.

EARTHQUAKE OCCURRENCES IN FENNOSCANDIAN SHIELD

Fennoscandian shield shows frequent earthquake activity, but in general these are of low magnitude and small in number. The biggest earthquake reported during the last hundred years was the one on the 23 October 1904 in the Oslo region. Systematic macroseismic studies (by questionnaires and field studies) of all earthquakes felt in Norway have been made since 1887. The first comprehensive study of Norwegian earthquakes was published by C.F. Kolderup in 1913. The studies of C.F. Kolderup, N.H. Kolderup, T. Birkeland, A. Kvale (all Norway), E. Svedmark, K.E. Sahlstrom, M. Bath (all Sweden), H. Renquist and E. Pentilla (Finland) and other Seismologists and Geologists in these countries have helped in systematic compilation and analysis of the earthquake data. Fig. 1 shows the map of Norway and Sweden by Kolderup (1913) showing the regions of large seismicity for the period 1887–1911. Fig. 2 shows the frequency of earthquake epicentres in the Fennoscandian shield compiled by Sahlstrom (1939) for the period 1600–1925. Fig. 3 shows the location of earthquake epicentres in Fennoscandian from 1891–1950 based on the catalogue of earthquakes compiled by Bath (1956). The earthquake data shown in fig. 1–3 are based upon macroseismic information taken from earthquake reports and other documents.

With the establishment of seismological stations and installation of the more sensitive instruments, earthquakes are now being recorded with greater accuracy. Fig. 4 shows the location of earthquake epicenters based on instrumental determination from 1951–1969

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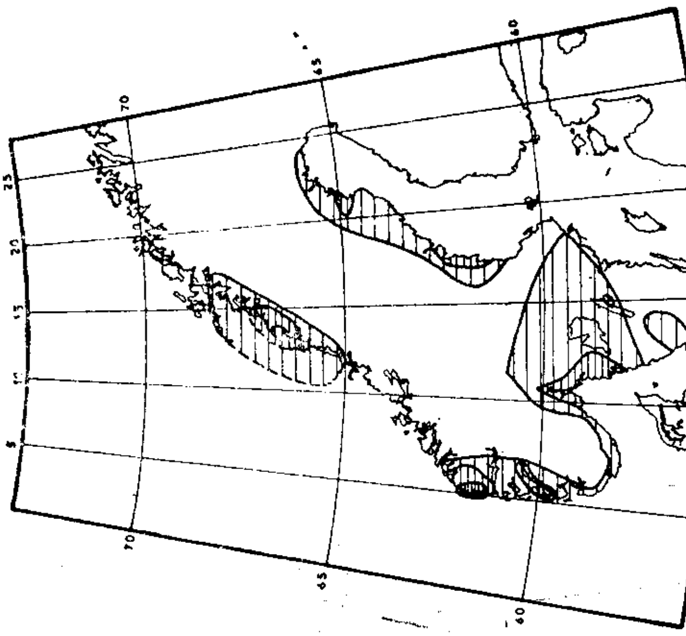


Fig. 1. Map of Norway-Sweden showing areas where Seismicity was greater from 1887 to 1911 by Kolderup (1913)

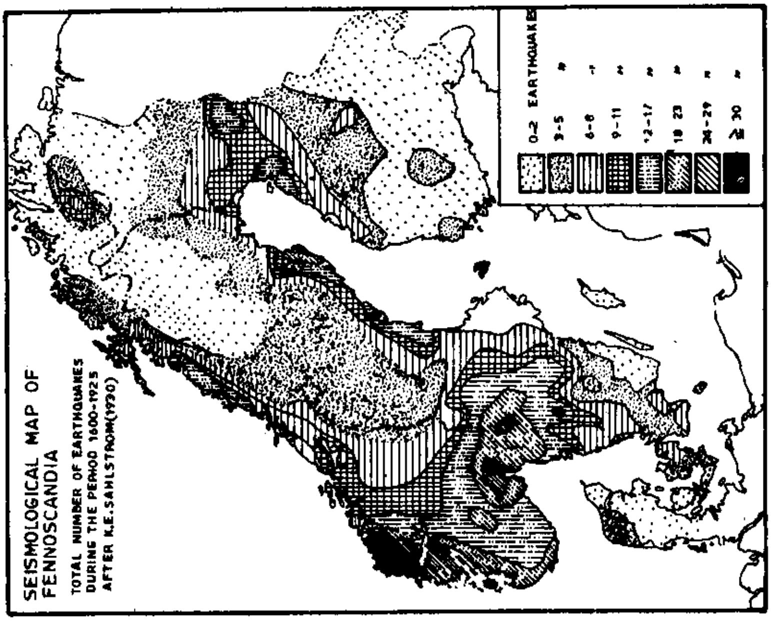


Fig. 2

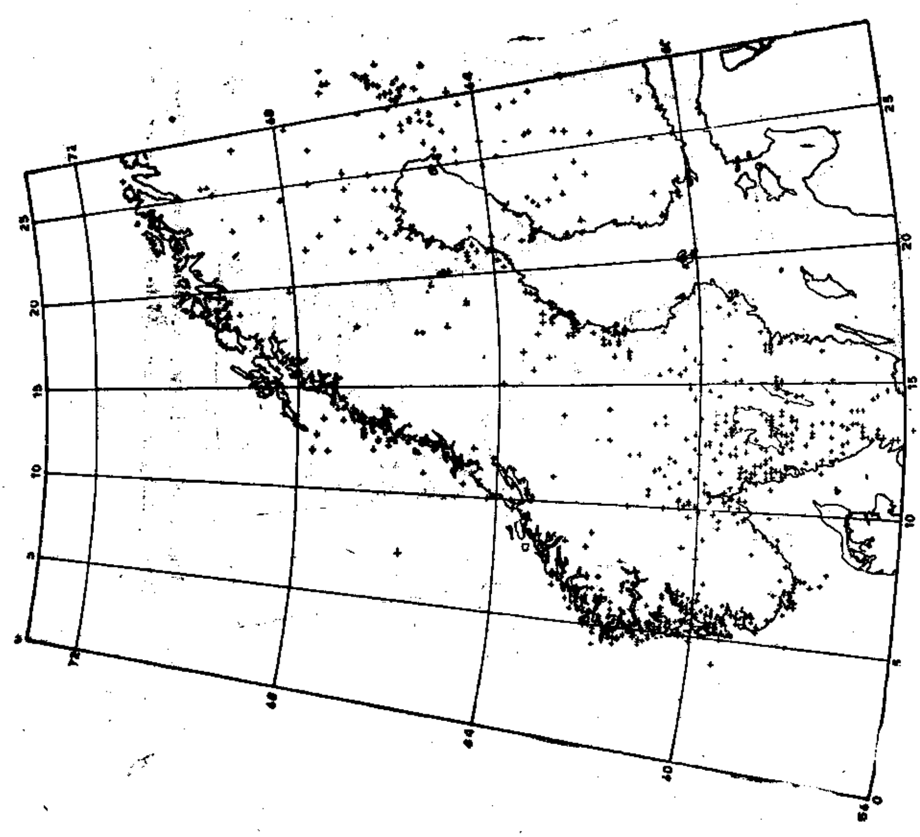


Fig. 3. Earthquake Epicentres in Fennoscandian Shield (1891–1950) based on the Catalogue of Earthquake compiled by Bath (1956)

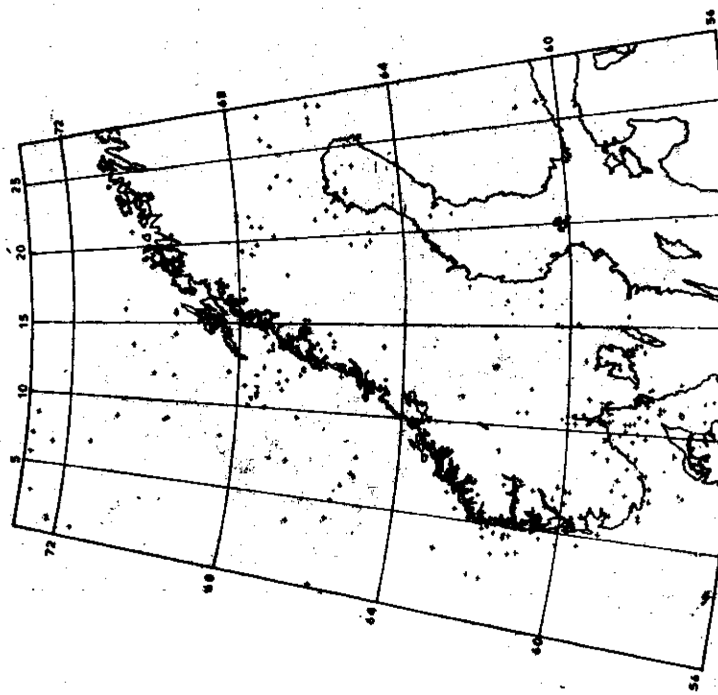


Fig. 4. Earthquake Epicentres in Fennoscandian Shield (1951–1969)

as published in the bulletins of the various seismological stations. These epicenters also broadly lie in seismic belts as shown in Fig. 1–3.

As more data on wave velocity and crustal structure are now available, the location of the epicenters and the probable depths of 311 earthquakes from 1958–1969 recorded at various seismological stations were recalculated. (Srivastava and Sellevoll, 1971). Fig. 5 shows the location and probable depth of these earthquakes. Fig. 6 shows the thickness of the crust used in the analysis.

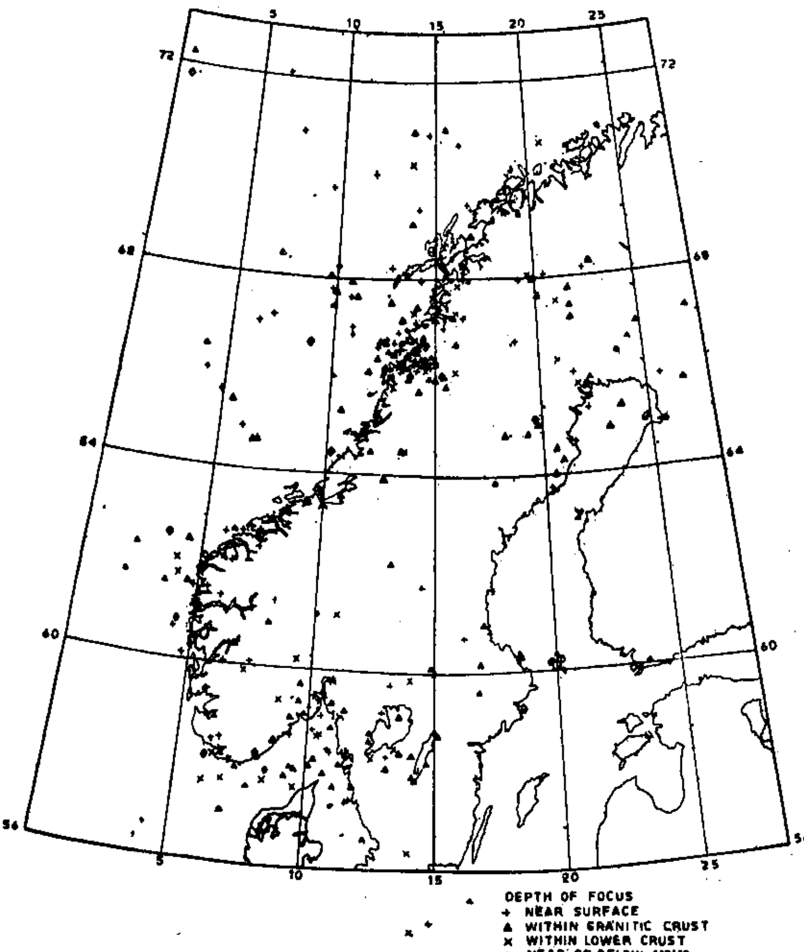


Fig. 5. Location and Depth of Focus of 311 Earthquakes (1958–1969) Recalculated for a Three Layer Crustal Structure of the Fennoscandian Shield (Srivastava and Sellevoll, 1970)

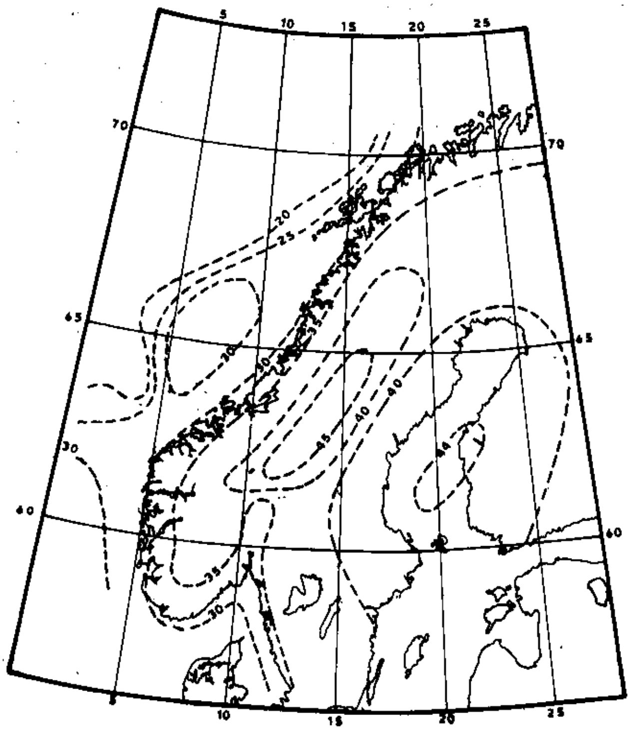


Fig. 6. Thickness of the Earth's Crust (in km) in Fennoscandia used in the Analysis for Evaluation of Depth of Focus of Earthquakes

SIZE OF EARTHQUAKES

For evaluation of seismic potentialities quantitiative estimates of the size of the earthquake and their frequencies are needed.

In 1935 C.F. Richter defined the magnitude of shallow earthquakes as

$$M = \log_{10}\frac{A}{A_0} \qquad \text{...... (i)}$$

where M is the magnitude of the earthquake, A, is the maximum amplitude recorded by a Wood-Andersen Seismograph (natural period 0.8 sec, almost critical damping and static magnification 2800) at a distance of 100 km from the centre of the disturbance and A_0 is an amplitude of one thousand of a millimeter. In practice the recordings are made at different distances and then extrapolated to a distance of 100 km from the centre of disturbance. Due to the inhomogenities and presence of discontinuity surfaces in the Earth's crust, variations in the magnitude values at different seismological stations for the same earthquake are observed. However, an average value of M determined from a number of stations gives a good estimate of the magnitude (Gutenberg and Richter. 1966).

The energy released in the form of seismic waves does not originate at a point source, as in the case of an explosion, but originates in a volume of rock which is proportional to the magnitude, hence the rock volume is greater for larger earthquakes than for small shocks (Tocher 1958, King and Knopoff, 1968, Housner 1970). The use of magnitude scale in the absence of other criteria is thus a convenient way of classifying earthquakes according to the size of the earthquake source and hence the approximate area affected by strong movements.

In the absence of any instrumental recording the severity of the ground movements is evaluated from a study of the macroseismic effects due to the earthquake by an intensity number (Roman numbers). The Modified Mercalli Intensity (MM Intensity) Scale is generally used. It ranges from I (ground notion not felt by anyone) to XII (total damage). Richter (1958) has given a rough correlation between MM Intensity and earthquake magnitude "for ordinary ground conditions in metropolitan centres of California" (table I). The intensity gives an idea of the severity of the ground motion, the degree of damage of the existing structure and the behaviour of the prevailing ground conditions. However as the evaluations of intensity are based on the local conditions, psychological factors and personal judgement of the investigators, considerable disagreement is generally observed in intensity data. The intensity thus does not provide a precise measure of the size of the earthquake. However, intensity survey usually include the area over which the earthquake was reported to have been felt, which help in evaluation of magnitude (Gutenberg and Richter, 1966). Such estimates, though only providing a rough estimate of the magnitude, are helpful in the absence of instrumental records.

TABLE—I

Rough correlation of magnitude with MM. Intensity (Richter, 1958)

Magnitude	2	3	4	5	6	7	8
Maximum Intensity	I–II	III	V	VI–VIII	VII–VIII	IX–X	XI

Upper Bound of Magnitude : The frequency of shallow earthquakes are reasonably well described by the equation (Gutenberg and Richter 1966) :

$$N = AN_0 e^{-M/B} \qquad \ldots\ldots (2)$$

where N is the number of earthquakes per year having magnitudes equal to or greater than magnitudes M in area A. N_0 is the number of earthquakes with M greater than or equal to zero. As the earthquake occurrence is not constant over the years N_0 is considered to only indicate the average seismicity. B is the distribution parameter of large versus small earthquakes. For purposes of plotting frequency distribution, equation (2) is put in the form.

$$n = -\frac{dN}{dM} = \frac{1}{B} A N_0 e^{-M/B} \qquad \ldots(3)$$

$$\log_{10} n = a - bM \qquad \ldots(4)$$

$$a = \log_{10}\left(\frac{1}{B} AN_0\right) \quad b = \frac{1}{2\cdot 3\ B}$$

where (ndM) is the number of earthquakes having magnitudes between M and $M + dM$. Equation (4) plotted as a straight line gives a plot of earthquake frequency versus magnitude (Fig. 7 and 8). The data of earthquake from 1904 to 1946 for the entire world (Fig. 7) fit this equation closely for $b = 0.9$ and $N_0 = 2.5/\text{yr/mile}^2$, except that above $M = 8$ the observed frequency drops off as shown in figure 7 (Gutenberg and Richter, 1966), and goes to "zero" at bbout $M = 8.7$ which may be taken as an upper bound of the size of earthquakes (Housner, 1970). This upper bound can be considered to specify a limit and the expectation of occurrence of earthquakes having magnitudes greater than this could be considered to be negligible. Assuming that the expectation of the occurrence of an earthquake with this upper limit in the world to be the same in all parts of the world, the upper bounds of magnitude for a region can be evaluated from the relation.

$$M = 7.1/b - [\log_{10}(A) + \log_{10}(b) - a]/b \qquad (5)$$

Data on the magnitude of earthquakes in Fennoscandian shield based on instrumental records is not available for a very long period of recent historical time. The magnitude

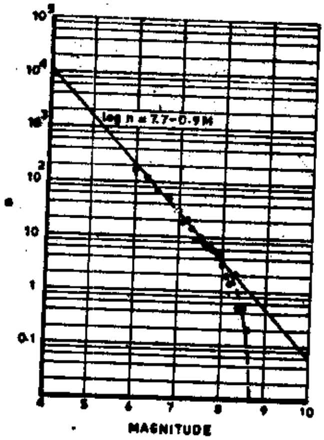


Fig. 7. Mean Annual Frequency Distribution of World Earthquakes (1904–1946), after Gutenberg and Richter (1966)

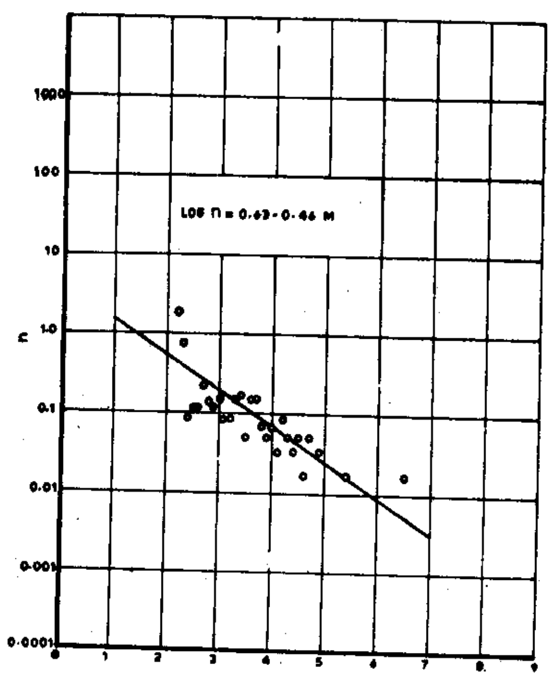


Fig. 8a. Mean Annual Frequency Distribution of Earthquakes in Oslo-Lake Vattern Seismotectonic Belt (1891–1950)

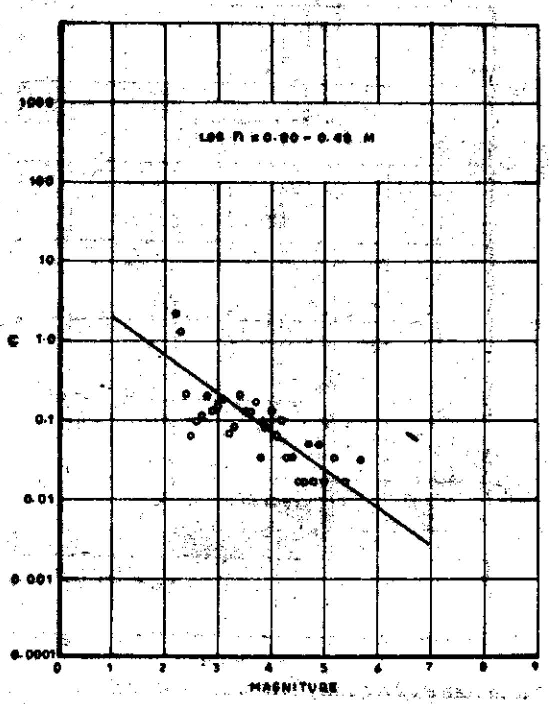


Fig. 8b. Mean Annual Frequency Distribution of Earthquakes in Alesund-Stavangar Seismotectonic Belt (1891–1950)

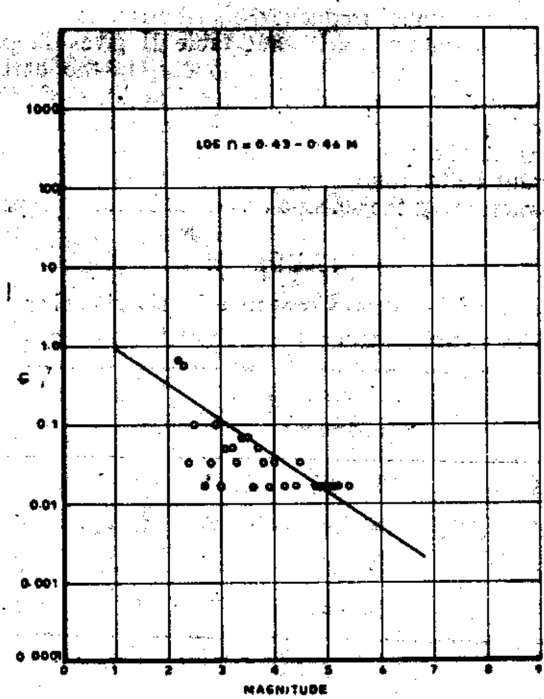


Fig. 8c. Mean Annual Frequency Distribution of Earthquakes in Alesund-Tromso Seismotectonic Belt (1891–1950)

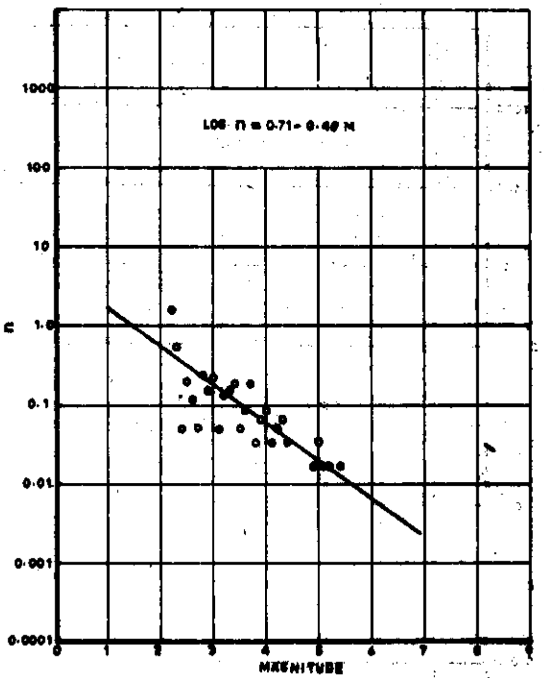


Fig. 8d. Mean Annual Frequency Distribution of Earthquakes in Bothnian Seismotectonic Belt (1891–1950)

of earthquakes from 1891 to 1950 were evaluated by Bath (1953, 1956) from macroseismic data. Fig. 8 shows the mean annual frequency distribution of earthquakes in the seismotectonic belts of the Fennoscandian shield, and table II gives the parameters 'a', 'b' and 'M' as the upper bounds of magnitudes. The frequency of the earthquakes as shown in fig. 8 have a considerable scatter of individual points. The lines of best fit for all the points in a seismotectonic belt have been obtained by the method of least squares, neglecting values of 'b' exceeding 0.9. The values of 'M' in table II are not predictions of maximum magnitudes in the different belts, and only indicate the upper bounds of magniutde in a probability sense characterising the seismic risks and status of seismicity of the belts.

TABLE—II

Parameters 'a' and 'b' of frequency magnitude distribution and upper bounds of magnitudes in the seismotectonic belts of the Fennoscandian Shield

BELT	Area Sq. Miles	'a'	'b'	M (Upper bound of magnitudes)
OSLO-L. VATTERN	100.000	0.63	0.45	6.8
ÅLESUND-STAVANGER	90.000	0.80	0.48	6.7
ÅLESUND-TROMS	80 000	0.43	0.46	6.5
BOTHNIAN	230.000	0.71	0.48	5.7

GEOTECTONIC AND PHYSIOGRAPHIC FEATURES AND PROBABLE TECTOGENESIS OF FENNOSCANDIAN SHIELD.

Theories of "oceanfloor spreading" and "plate tectonics" have been applied to show that the occurrences of earthquakes are related to fractures and faults along the boundaries between "plates of the lithosphere" which demarcate the Earth's major seismic belts (Isack, Oliver and Sykes, 1968). However, as the lithospheric plates are considered rigid and relatively strong and do not undergo any significant folding, distortion or stretching "Plate tectonics" offers no direct explanation for the occurrence of earthquake within the plates forming continental shields.

It is generally considered that the Fennoscandian shield formed a low and flat region during the Tertiary. Recent investigations show that the continental margins of the Fennoscandian shield probably extended up to the middle of the Voring plateau during Lower Tertiary from which it was separated (63 mil. years) from Greenland, initiating the sea spreading of the Norwegian-Greenland sea (Talwani and Eldholm, personal communication-1971). Åm (1970) also considers the inner part of the Voring plateau to be of continental origin. Either the shield extended as a landmass upto this continental margin or a sedimentary basin existed in between. The later appears to be more probable as the existence of thick sedimentary cover is indicated in the continental shelf region by recent surveys, and part of these sediments may be of Tertiary age (Sellevoll and Sundvor, 1971).

In the Upper Tertiary, Fennoscandian sheild was subjected to an oblique and unsymmetrical uplift which raised its present western and north-western parts in the Scandinavian peninsula higher than its eastern part in Sweden and Finland (0. Holtedahl, 1960). The remains of these peneplained plateau like surfaces are found at various elevations, but in some cases it is difficult to decide to what extent these surfaces represent the pre-uplift Teritiary landmass (0. Holtedahl 1960). It is assumed that the uplift of the shield occurred along faults which traverse the Lower Cretaceous Andoy deposits and probable faults along the Norwegian Channel and the inner parts of the continental shelf. Along these faults marginal channels are believed to have been carved out by later erosions, though the existence of such major faults off the coast has not been indicated in recent investigations across the Norwegian Channel (Sellevoll and Aalstad, personal communication-1971).

During the Quanternary age the Fennoscandian shield under-went a major glaciation separated by milder interglacial periods and shows evidence of uplift in late and post glacial periods. The late glacial and post glacial marine limits (Fig. 9), isobase for uplift since the begining of Littorina (5500 B.C.) time (Fig 10), and recent uplifts as indicated by tide gauge measurements, relevelling and historical excavations (Fig. 11) indicate higher uplifts in central parts of the Scandinavian peninsula and the region around the Bay of Bothnia when compared to the outer parts of the shield. The uplift gradients during the present time vary in the various parts of the shield with low gradients along the coastal region in the west and in the Oslo region while higher values are found for the central part of the Scandinavian peninsula and Finland.

Fjords and "strandflaten" are noted to be extensively developed along the west and north-west margins of the shield compared to other parts. Fjords form "relatively long and narrow, often curved or branching embayments with more or less steep sides and a considerable depth" with "a rock thresholed at the entrance (or thresholds separating trough like parts of fjord complex farther inland)." J. W. Gregory (1913, in O. Holtedahl, 1960) consider fjord-topography a "result of a fairly recent (Tertiary) tectonic breaking up of coastal districts." Due to the existence of the rock thresholds and the role of glaciers in sculpturing various landforms in these parts of the shields, fjords are considered by various workers to be the result of glacial erosion along pre-existing joints and other zones of weaknesses (O. Holtedahl, 1960, H. Holtedahl, 1960). Strandflats are low, uneven rocky foreland lying party above and partly below the sea level with steep escarpment like sides

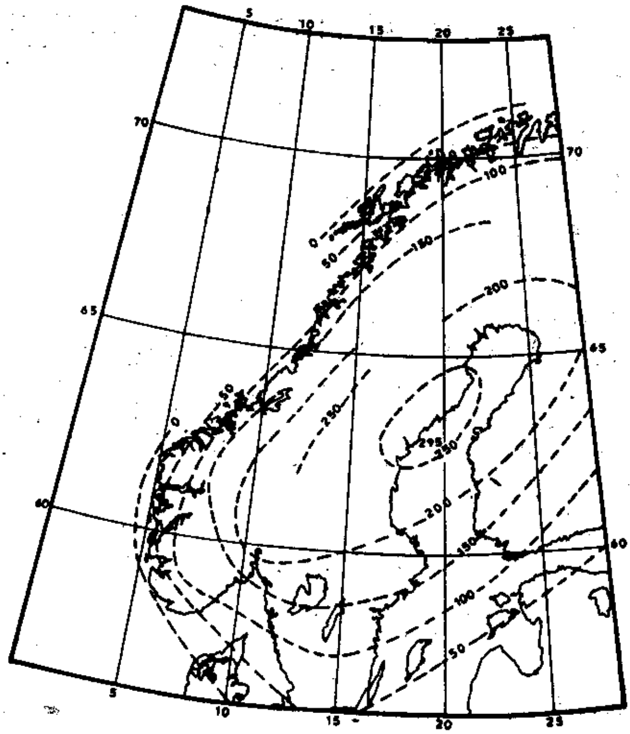


Fig. 9. Late and Post Glacial Marine Limits in Metres in Fonnoscandia (After Granlund 1949)

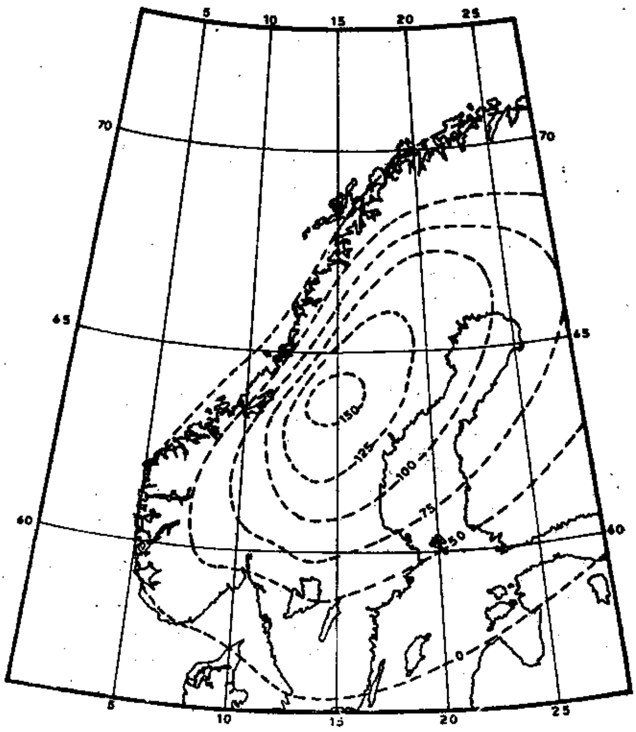
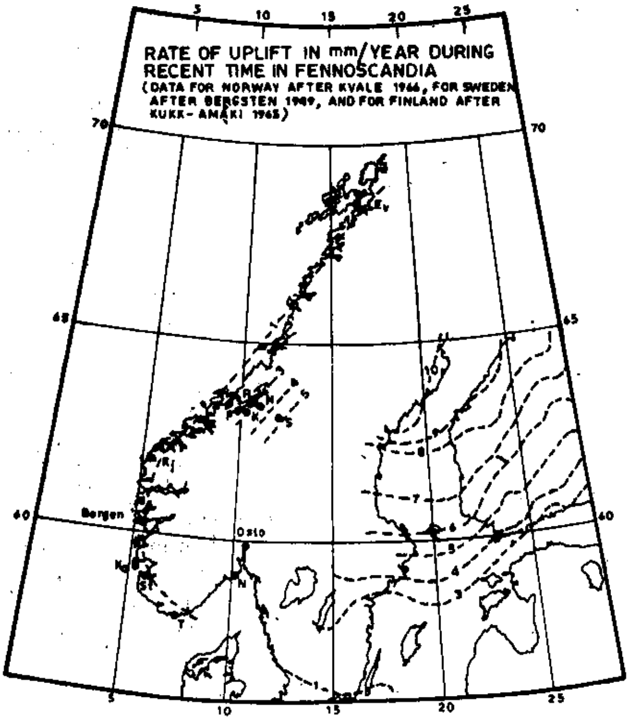


Fig. 10. Isobases for Uplift in Metres since beginning of Littorina Time 5500 B. C. (After Granlund 1949)



Ev–Evenskjder 0·6 mm/year (1945–49) Tide gauge records; He–Heimsjo 1·2 mm/year (1928–64) Relevelling; F–Forve 2·6 mm/year (1928-64) Relevell ng; K–Krakmo 3·3 mm/year (1928–64) Relevell'ng: R Rotvoll 3·1 mm/year (1924–64) Relevelling; H–Hell 3·7 mm/year (1926–64) Relevelling Kj–Kjlsdal 1·1 mm/year (1930–59) Tide gauge records; Bergen–0·1 mm/year (1883–1959) Tide gauge records; 1 m since 1150 AD uplift of piers; Ka–Karmoy 2m from 900 (?) AD uplift of boat houses; St–Stavangar 0·9 mm/year (1882–1950) Tide gauge records; T–Tregde 0·6 mm/year (1928-50) Tide gauge records; N–Nevlunghavn 2·2 mm/year (1927–59) Tide gauge records: Oslo 3·6 mm/year (1886–1950) Tide gauge records

Fig. 11

rising from the base which are found to occur as characteristic physiographic landforms outside western coastal mountains and fjords. These are considered by various workers to be the reasult of marine erosion, subareal denudation or their combinations during the various stages of the Pliestocene (H. Holtedahl, 1960).

Figure 6 shows the thickness of the crust-expressed by structural contours of the probable depth of Moho (P_n – 8.2 km/sec) in the Fennoscandian shield. These indicate lesser crustal thicknesses along the "Oslo-Lake-Vattern" region, the Norwegian Channel, the "Stavanger-Alesund-Tromso" coastal region and the Continental Shelf region due to pronounced uplift of the Upper Mantle. A minor undulation is also indicated in eastern Sweden following the Bay of Bothnia. The axes of maximum uplifts tend to follow zones of large crustal thickness (downwarps) whereas lesser uplift gradients are observed along zones of lesser crustal thicknesses (upwarps).

The geotectonic and physiographic history of the Fennoscandian shield appears to be genetically related to the undulations of the Moho and consequent flexuring of the crust. In accordance with Clooss (1939), tenet "Hebung-Spaltung-Vulcanismus" (rising-fissuring-volcanism) uplifting of the mantle causes upwarping and stretching in the crust and melting of rocks due to relief of pressure at the base of the crust Growing tension could lead to the formation of fractures in the crust extending from the interior to the surface along which the crustal blocks in the axial parts of the upwarp could subside and molten material could be injected. This hypothesis may explain the formation of rifts and other depressions. The geomorphological evidence from the folding, faulting and metamorphism within the rifts and other zones of subsidences indicate that compressive forces have a major role in the development of such a tectonic framework, which will produce thermo dynamic and metamorphic changes and inhibit the development of fractures in the Upper Mantle and lower crustal layers. This suggests that initiation of fractures due to uplift of the upper Mantle, begins in the upper layers of the crust, similar to the development of cracks in a plate under compression. Such fractures may follow or open pre-existing foliations, joints, and planes of weaknesses if these are aligned parallel to the axial trend of the upwarp. When such initial fracture zones extend deep into the crust, differential movements start due to sliding of crustal blocks or movement of magma along them. However, as the compressive forces would be horizontal in the axial parts of the upwarps, strike slip movements will predominate in such regions, whereas areas in the downwarps probably lying along zones of diverging convection currents will undergo greater vertical and dip slip movements. The main depressions develop along the axial part of the upwarp due to lagging behind of the axial blocks during the uplift of region, and subsidiary blocks forming rifts and horsts can develop on the limbs of the upwarp along either the initial fracture or subsidiary transverse fractures resulting from surface and subsurface adjustments (Tipnis and Srivastava. 1968).

The development of the Norwegian Channel and other depression and channels in the continental shelf along the axial parts of the upwarps, probably resulted form such a process. The major fjords were carved out by erosion along the median as well as transverse fractures and other zones of weaknesses developed in the axial parts and the limbs. The rock thresholds at the entrance of the fjords and inside are probably the results of differential uplift of the crustal blocks. The strandflat, which lie within the axial parts of the upwarps bordering the channels also appear to be the result of differential uplifts of blocks with modifications by later or simultaneous erosion and denudation.

The foregoing discussion suggest that uplifts of the mantle and fractures of the crust along narrow belts is responsible for the dismemberment of the continental crust into "lithospheric plates" and crustal blocks—the later due to their assimilation or consumption in the mantle or lagging behind the general uplift result in the development of sedimentary basins and other depressions. Such a dismemberment of continental shields, platforms and peneplained mountain ranges completes the cycle of progressive development and transformations of the physiographic and geotectonic divisions of the landmass.

SEISMOTECTONIC BELTS OF FENNOSCANDIAN SHIELD

In seismic and seismotectonic studies, linear arrangement of epicenters is usually considered to indicate the trend of seismotectonic belts. The pattern of earthquake occurrence in the Fennoscandian shield shown in Fig. 6-8, indicate that the epicenters form clusters and zones extending over broad belts. Many fault systems lying in these epicentral belts are taken to indicate close association with earthquake occurrences (Kvale, 1960). However, it should be kept in mind that the occurrence of earthquakes in a region is a result of geotectonic processes operating at present within the crust and upper mantle. Thus the orientation and trend of the fault lines in relation to the prevalent geotectonic frame-work of the crustal structure has to be kept in mind when establishing such associations. With the deformations of the crust over large areas, new fractures and faults releasing seismic energy could be produced in any part of the belt or movements may be initiated along pre-existing faults and fractures. Demarcation of seismotectonic belts thus should be carried out on the basis of the regional crustal structure, rather than on the trend of the fault lines and other seismotectonic lineaments of the region.

The distribution of epicentres in the Fennoscandion shield show four major seismic belts known as (i) the Oslo-Lake Vattern, (ii) the Ålesund-Stavanger, (iii) the Ålesund-Tromso and (iv) the Bothnian seismotectonic belts.

The Oslo-Lake Vattern seismotectonic belt lies over a pronounced crustal upwarp and covers Oslo, Kattegat, Skagerrack and eastern Swedish regions. This belt has shown the highest seismic activity in the past. The major earthquake occurrence is found along the axial parts of the upwarp. Two major seismotectonic lineaments following the Skagerrack and Kattegat regions with seismic disturbances in the lower crust are also found. Another major zone of seismic disturbances is noted along central parts of southern Sweden, with activity varying from near surface fractures to deeper locations in the lower crust. The trends of the epicenters in the whole of this belt show very close associations with the physiographic features which are major geotectonic depressions. The fractures in these depressions show close association to the trend of the crustal upwarping of the region.

The Ålesund-Stavanger seismotectonic belt covers the area along the Norwegian channel and the western Norway. This belt has the maximum number of earthquakes in the past and shows comparable seismicity with the Oslo-Lake Vattern belt. The epicentral distribution of earthquakes in the past (1891–1950) shows greater concentration of epicenters on the land ward side, however the size of these earthquakes has been smaller. The earthquakes on the coastal zone in the Norwegian channel were of greater intensity and recent earthquake occurrence in these parts of the belt indicate a deeper level of activity. This is probably due to the median fractures in the axial part of the upwarp becoming mobile by tectonic creep and stable sliding, which inhibits large strain build up near the surface. In addition seismic disturbances in the upper granitic crust and near the surface have clustered along the sides of the upwarp in the crustal parts and in the interior where often considerable strain has been build up.

The Ålesund-Tromso seismotectonic belt also lies over a crustal upwarp and covers the coastal and continental shelf region. The sides of this elongated belt show greater activity when compared to the central parts of the continental shelf. The seismotectonic situation of the coastal parts of this belt is similar to that of the Ålesund-Stavenger belt. The outer edge of the belt has a crustal thickness of about 18–20 km and probably forms the outer limit of the Fennoscandian shield. It shows pronounced activity which is related to the maximum uplift of the Moho. A. major seismotectonic lineament trending NW-SW is situated south of the Lofoten Islands. This zone probably represents an axis along which the continental shelf has subsided forming crustal depression in between the coastal parts and the outer margins of the shield.

The Bothnian seismotectonic belt covering parts of Sweeden and Finland around the Bay of Bothnia does not show high upper bound of earthquake magnitude. The western parts of this belt lies over a minor crustal upwarp compared to its lower western parts in Finland.

The seismotectonic belts described above show close association of crustal upwarps of the crust, and their surface physiographic features developed due to the operative geotectonic processes related to such upwarping. The depth of focus of the earthquakes is noted to increase towards the axial parts of the upwarps because as strain builds up at lesser depths, its build up is inhibited by tectonic readjustment and creep. As the earthquake occurrence cover large belts of the upwarps, the fault lines and other lineaments which are unrelated to the prevalent directions of the principal upwarping of the crust, have to be excluded in the identification of seismotectonic feature and demarcation of seismic zones.

SEISMIC ZONING OF FENNOSCANDIAN SHIELD

Earthquake hazards in shield regions are in general assumed to be negligible. However, it has been noted that earthquakes in such regions (e g. 1904 Oslo earthquake in Norway, 1967 Koyna, 1968 Bhadrachalam, 1970 Broach and other earthquakes in Indian Paninsular shield) were responsible for considerable damage.

The codes of practices for earthquake resistant design and construction of structures require delineation of seismic zones with different levels of seismic risks in terms of intensity and duration of strong motions which would have to be resisted by the structures. Such demarcations are made so that adequate safeguards are taken in regions of higher seismic risk and expenditure is saved in regions of lower seismic risks.

Data on earthquakes and their associated geologic and tectonic features is utilised in the preparation of seismic zoning maps. In general two approaches are followed. In the first approach, in regions where either data on active tectonic feature is not available or correlation of tectonic features with earthquake occurrence are considered to be inconclusive, zones are outlined based on past earthquake records by enveloping epicentral tracts of different magnitudes and corresponding intensities. If data on earthquake occurrences are available for sufficient duration the epicentral distribution can be considered to demonstrate the trend and extent of the seismotectonic belts, and statistical probabilities for occurrence of earthquakes of various magnitudes in different parts of the region can also be taken into consideration. However, if earthquake data is not available for a long period, seismic zones are then drawn surrounding centres of known earthquake disturbances, which does not take into consideration the seismotectonic set up of the region.

In the second approach, which has been followed for seismic zoning of Fennoscandian shield, seismotectonic setup is given greater significance. With installation of greater number of seismological stations more data on earthquake occurrence is now available, which has helped in establishing associations with geotectonic features and identification of seismotectonic belts for demarcation of zones with probable centres of disturbance likely to produce earthquakes. The trend and extent of the seismotectonic belt with highest seismic risk as manifested by magnitudes and frequency of earthquake occurrence in the belt then expresses the trend and extent of the highest seismic zone in the region. Zones with lower order of seismic risks surrounding the highest zone can then be drawn based on the seismic risks due to likely occurrence of earthquakes within the zone and the effects which could results due to earthquakes in the adjacent zones.

In preparing seismic zoning maps it is meaningless to demarcate zones with intensity of strong ground motion less then that corresponding to MM Intensity V, as the probability of occurrence of such an intensity at any place on the surface of the earth can not be ruled out, and greater than that corresponding to MM Intensity IX as the intensity of strong ground motion does not show very large increase for higher intensities and the great damage results mostly due to longer duration of strong shaking. Thus seismic zones corresponding to MM Intensities "V and less", VI, VII, VIII and "IX and above" provide sufficient information for design and constuction of usual structures.

Four distinct seismotectonic belts-named as Oslo-Lake Vattern, Ålesund-Stavanger, Ålesund-Tromso and Bothnian belts—are noted in the Fennoscandian shield, and earthquake epicentres in each of these belts are distributed along crustal upwarps which appear to be genetically related with the earthquake occurrence. The upper bound of magnitudes in these belts, based on the frequency magnitude distribution of earthquake occurrence of the period 1891–1950 are shown in Table II.

The above indicates that the first three belts possess comparable seismic status with probabilities of occurrence of an earthquake with upper bound of magnitude from 6.5 to 6.8. The largest earthquake in these belts was the 1904 Oslo earthquake, which is considered to have a magnitude 6.5 and intensity to MM Intensity VIII. Similar intensities were also observed during earthquakes of comparable magnitude 1967 Koyna, 1968 Bhadrachalam and 1970 Broach earthquakes in Indian Peninsular shield. Thus the maximum seismic risk within these belts can be considered to be less than that due to an earthquake of magnitude 6.5 and probable corresponding MM intensity VIII. The enveloping curve following the trend of these belts, defined by structural contours of Moho representing the upwarping of the crust, and their lateral extent, limited up to the physiographic and geotectomic features (considered to be genetically related to the upwarping and consequent seismotectonic activity) is shown in figure 12 as seismic zone IV. The epicentral distribution in these belts is noted to be mostly confined within this zone.

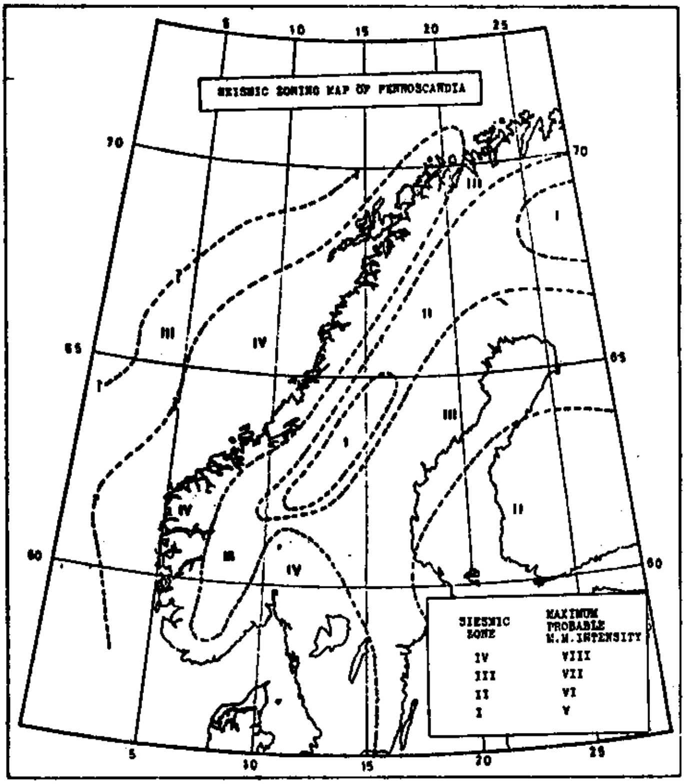


Fig. 12

The Bothian belt shows lower upper bound of magnitude. The crustal structure in this belt do not show very pronounced undulations. However the epicentres are noted to have greater concentration in the region of upwarps, and these parts are likely to have higher probabilities of earthquake occurrance. Other parts of this belt show lesser earthquake activity. The Bothnian belt thus is divided into two zones, an higher zone over Bothnian bay and eastern Sweden, and a lower zone in its south-eastern part in Finland and Uppsala region as seismic zones III and II respectively. Zone I has been demarcated along the crustal downwarp in the central region of Norway-Sweden and the and the region west of Kola Peninsula which have indicated lowest seismicity in the past. The later region probably also lie along a downwarp extending from Kola Peninsula where greater depths of Moho are indicated (Panasenko, in Penttila, 1969).

The above seismic zones have been drawn to indicate the regional pattern of seismic risks. The probable upper bounds of magnitude and corresponding likely maximum MM Intensity in each of the zones are shown in Table III. In case of important and special structures the location of active fault zones and other seismotectonic elements at the site will have to be taken into consideration, as higher intensity of ground shaking occur near them.

TABLE—III

Probable upper bounds of magnitude and likely maximum MM Intensity in the various seismic zones of Fennoscandian Shield.

Seismic zone	Probable upper Bound of Magnitude	Likely Maximum MM Intensity
IV	6.5–5.8	VIII
III	5.7	VII
II	5	VI
I		V

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