The 31st of August 1819 Lurøy earthquake revisited

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Bungum, H. & Olesen, O.: The 31*st* of August 1819 Lurøy earthquake revisited. *Norwegian Journal of Geology,* Vol. 85, pp. 245 - 252. Trondheim 2004. ISSN 029-196X.

The 31*st* of August 1819 Lurøy earthquake has, for a long time, been considered as the largest known historical earthquake in Fennoscandia, possibly even on the mainland of northern Europe. Triggered by a recent claim that the magnitude (M_S) of this earthquake should be 5.1 rather than 5.8, which implies a reduction in radiated energy by about a factor of 10, we have reassessed the original macroseismic observations together with the published literature. Our conclusion is that a M_S magnitude of 5.8 for the 1819 Rana region earthquake still remains a reasonable, justifiable and defendable estimate. Such a magnitude is moreover consistent with present hazard models for the region, which include the possibility that magnitude 6+ earthquakes can occur today in the most seismically active areas in Norway, such as the coastal parts of western Norway, Nordland and the Oslo rift zone. Rock avalanches and landslides, potentially triggered by earthquakes, could moreover generate tsunamis in fjords and lakes and constitute the greatest seismic hazard to society in Norway.

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Introduction

Earthquakes differ from other environmental loading phenomena such as ocean waves in that the largest events have very long return periods, typically of the order of 100 years at plate margins but possibly thousands of years in more stable continental regions such as Fennoscandia (e.g. Johnston & Kanter 1990; Bungum et al. 2005). Given that a reasonably complete record of instrumental recordings on a global basis dates back only to the early 1960s, this clearly creates particular problems in terms of acquiring sufficiently long periods of observation for the purpose of recurrence estimation. This, in turn, emphasizes the importance of historical observations, which for earthquakes is termed macroseismic or intensity data, and is essentially concerned with how earthquakes have been felt by people and their effects on both the built and the natural environment.

Historical records from Norway are very scarce from earlier than the late 17th century while more systematic reports on natural phenomena like earthquakes date back only to the late 19th century. The few large historical earthquakes that have been reported are therefore of great importance for understanding the seismotectonics and seismic potential in this region. Consequently, the recent papers by Kebeasy & Husebye (2003) on the 1759 Kattegat earthquake and by Husebye & Kebeasy (2004a) on the 1819 Lurøy earthquake are commendable. There are, however, some problems with both these papers that already have given rise to one critical comment (Wahlström 2004), with a reply by the authors (Husebye & Kebeasy 2004b). The present discussion is concerned with the Husebye & Kebeasy (2004a) paper (hereafter called H&K).

In the following we will discuss specific parts of the H&K paper, concentrating on the topics that are most important for the magnitude assessment. The main point here is that H&K reduce the surface wave magnitude M_S for the 1819 earthquake from 5.8 to 5.1 (and the local magnitude M_L to 4.8), which implies a reduction in radiated energy by about a factor of 10. First, however, we find it appropriate to provide some comments on the development of magnitude assessments of earthquakes in Norway. We agree fully with H&K that reported magnitudes for historical earthquakes, in Norway and elsewhere, have had a tendency to be overestimated. One reason for this is that historical reports are often unduly 'colourful' in their descriptions, and that earlier scientific assessments have not filtered such effects well enough. Moreover, different assessment criteria may have been used, magnitudes may have had a tendency to be 'inherited' indiscriminately from earlier reports, and in addition, different and poorly defined magnitude scales may have been employed. As an example of earlier reported magnitudes we note that Karnik (1971) had magnitude (M) 6.0 for the 31 August 1819 earthquake and 6.4 for the 23 October 1904 Oslofjord earthquake, which is another historical earthquake of importance in Norway. Husebye et al. (1978) reported M 6.0 for both the 1819 and the 1904 earthquakes, while Husebye et al. (1979) had M 6.0 and 6.4, respectively, for the same two events. These were commonly accepted values at that time, and it was also common that the magnitude scale was not well defined.

Table 1.

Equation		(A) From Muir Wood & Woo (1987)		(B) From Muir Wood & Woo (1987); Muir Wood (1989)		(C) From Muir Wood (1989)	
No	Based on	R (km)	M_S	R (km)	M_S	R (km)	M_S
1	A_{III} log linear	630	5.7	850	5.9	800	5.9
2	A_{III} non linear		5.8		6.2		6.2
3	A_{IV} log linear	450	5.6	530	5.7	530	5.7
4	A_{IV} non linear		5.8		6.0		6.0

Table 1. M_S magnitudes for the 1819 Nordland earthquake as calculated using Eqs.(1) to (4), based on different estimates of felt areas, here expressed in terms of felt radii assuming a circular area. For column (A) the numbers are taken from Table 3.4 in Muir Wood & Woo (1987); for column (B) the numbers are derived by the present authors based on a felt area figure which was published both as Fig. A7.2 in Muir Wood & Woo (1987) and as Fig. 4 in Muir Wood (1989); for column (C) the numbers are taken from the text on page 234 of Muir Wood (1989).

Magnitude scales and assessments

In an effort to correct for some of the above-mentioned inconsistencies, a major reassessment of historical earthquakes in Norway was conducted by a UK-Norwegian team (Muir Wood & Woo 1987; Muir Wood et al. 1988), building on a similar effort in the UK (Woo & Muir Wood 1986; see also Ambraseys (1985a,b). The approach adopted was to go back to the original and primary information as a basis for a complete reassessment of the most important historical earthquakes. Magnitudes were tied to the well-defined M_S scale, based on a correlation between M_S and felt area derived from a set of events for which both instrumental records and intensity data were available. Since this is important for the present discussion of the H&K paper, we quote here the relations developed by Muir Wood & Woo (1987) between M_S and felt area A_{III} for intensity level *III*, using a log-linear and a non-linear relation, respectively:

$$M_S = -0.36 + 1.0 \cdot \log A_{III} \tag{1}$$

$$M_S = 0.95 + 0.69 \cdot \log A_{III} + 0.0006\sqrt{A_{III}} \tag{2}$$

For intensity level IV, the corresponding relations from Muir Wood & Woo (1987) read:

$$M_S = 0.90 + 0.81 \cdot \log A_{IV} \tag{3}$$

$$M_S = 1.57 + 0.63 \cdot \log A_{IV} + 0.0007\sqrt{A_{IV}} \tag{4}$$

The scatter (standard deviation) was not provided for these relations, but from the plots it appears to be at least 0.2 magnitude units, which is common for almost any regression on magnitudes.

It is noteworthy that this new magnitude scale based on felt areas for intensity levels *III* and *IV* led to a major downward adjustment for some of the larger historical earthquakes, with 5.6 for the 1759 Kattegat earthquake (Muir Wood 1989; see also Kebeasy & Husebye 2003; Wahlström 2004), 5.8 for the 1819 Lurøy earthquake and 5.4 for the 1904 Oslofjord earthquake. These adjustments demonstrated that earlier estimates were too high, and the question now is whether a further downward adjustment of these magnitudes is needed, as claimed by H&K.

Eqs. (1) to (4) can all be used for deriving magnitude estimates, and in order to offer some insights into sensitivities we provide in Table 1 an overview of the results for the 1819 earthquake when combining the four relations with different estimates for felt areas (or radii, given a circular shape). The three columns use values taken from (A) a table in Muir Wood & Woo (1987), (B) a figure in the same report (reproduced also in Muir Wood, 1989), and (C) the text from Muir Wood (1989). The values are seen to be in the range 5.6 to 6.2. Note that a typographical error in Muir Wood (1989) has been corrected in that an R_{IV} of 350 km in the paper should be 530 km (R. Muir Wood, pers. comm. 2004), which is now used in column (C).

Before discussing the H&K paper we would like to offer some additional evidence supporting the Muir Wood & Woo (1987) felt area based magnitude scale. This was based on a correlation between instrumental and macroseismic data, employing 27 A_{III} and 25 A_{IV} values for instrumental M_S values between 4.0 and 6.1. The frequency-magnitude distributions based on this scale are stable and reasonable, and moreover, consistent with instrumental M_L values derived for more recent earthquakes, after accounting for the difference between the magnitude scales (Bungum et al. 1986; Muir Wood et al. 1988; Bungum & Selnes 1988; Bungum et al. 1998; Lindholm & Bungum 2000). Recently, there have been two new assessments of the magnitude for the 1904 Oslofjord earthquake, reported

Fig. 1. Isoseismal map with intensity data points (redrawn from Muir Wood 1989) for the earthquake near Lurøy, Nordland on 31 August 1819. Average radius of perceptibility is about 800 km.

by Bungum et al. (2004). The first, based on a new genetic algorithm inversion of the macroseismic field (Pettenati & Sirovich 2003), calibrated through a number of western Norway earthquakes (Pettenati et al. 2004), has given a moment magnitude (M_W) of 5.4-5.5, in addition to a new location, depth and focal mechanism. The second, based on a reassessment of instrumental data, has given a new M_S value also of 5.4, in addition to a location which is fully consistent with the new macroseismic solution. These independent magnitude estimates strongly support the felt area based M_S value of 5.4 from Muir Wood & Woo (1987).

Intensity observations and their implications

Here we will concentrate on a few main points. H&K repeatedly claim that Muir Wood & Woo (1987) and Muir Wood (1989) have based their magnitude of 5.8 exclusively on A_{III}. The fact is that it is not specifically stated how the value is derived, but it is clear, from Muir Wood & Woo (1987), that A_{IV} also has been used in the assessment. Consequently, Muir Wood (1989) is using a magnitude range (5.8-6.2). Such details, however, are of lesser importance. What counts more is the underlying basis as provided in Table 1, in terms of observations and models that can be used in an independent assessment. Table 1 actually provides three sets of felt area assessments for both intensity levels (*III* and *IV*) which, together with a log-linear and a non-linear relation in each case, provide 12 different magnitude estimates. These are all in the range 5.6 to 6.2.

While we find no support for a R_{III} value as low as 350 km, as claimed by H&K, we agree with them that a R_{III} value of above 800 km may be on the high side, thereby reducing the confidence for a magnitude of 6.2 (see Table 1). This affects the intensity value of *III* for the Stockholm region. However, since Ambraseys (1985a) does include in his compilation a reported shaking from Tierp village, about 100 km to the north of Stockholm, we cannot totally rule out the Stockholm observations. This also shows that the statement by H&K that the Lurøy earthquake was not felt outside Stockholm to the south of 61°N is not correct. It could be noted in this respect that it is often difficult to distinguish R_{III} from R_{II}, which may explain that the magnitudes as derived from A_{III} in Table 1 are somewhat higher than those based on A_{IV}. The latter values are centred on 5.8.

Wahlström (2004) argued that the macroseismic map of Kjellén (1910) was the most convincing argument against the downgrading of the magnitude of the 1819 earthquake. Even though the map does not include the 1819 observations in northern Norway, Kjellén (1910) still listed this earthquake as one of the five largest earthquakes *in Sweden*, and it was reported to have been felt from Tornedalen in the north to Stockholm in the south and Trondheim to the west. He estimated the epicentre to be located within a well-known seismicity zone located along the western coast of the Bay of Bothnia (within Norrland). Consequently, Kjellén (1910) did not utilise the detailed observations from the Nordland area reported by Keilhau (1836), even though he referred to the latter publication when evaluating other events in his catalogue. For the 31st August 1819 event Kjellén refers exclusively to reports by Ehrenheim (1824) and Moberg (1893). Ehrenheim had moved the earthquake one year forward whilst Moberg did not include the northern Norway observations in his account. Kjellén (1910) obviously did not realise that the heavy shaking reported from both the Lurøy-Rana and the Bay of Bothnia areas could in fact originate from the same earthquake. This conclusion is supported by the fact that the Lurøy-Rana observations are neither included in his list of seven Norwegian events reported felt in Sweden nor mentioned in his extensive catalogue of Swedish earthquakes (414 events) which also includes reported effects from Norway. The claim by Husebye & Kebeasy (2004a,b) that the Swedish reports were strongly influenced by the extraordinary reports from the Lurøy region is consequently unfounded. Their statement that the many observations of strong shaking from several Bothnian villages and towns (e.g. Lycksele, Umeå, Torneå, Haparanda, Oulu, Raahe and Kalajoki, Fig. 1) on 31 August 1819 can be attributed to small local earthquakes is similarly unfounded, in accord with the conclusions reached in other studies (e.g. Ambraseys 1985a; Muir Wood 1989; Wahlström 2004). Kjellén (1910) did, on the contrary, regard the Lurøy earthquake swarm reported by Keilhau (1836) as a local phenomenon (see notes 2 and 5 on pages 69 and 140, respectively, in Kjellén 1910).

The cluster of observed intensity levels *IV-VI* around the Bay of Bothnia demonstrates therefore a consistent pattern (Kjellén 1910; Muir Wood 1989; Wahlström 2004), with a distance of 500-600 km from the Lurøy-Sjona area. A conservative R_{III} estimate of 630 km fitting the Gulf of Bothnia observations implies a magnitude of 5.7-5.8 (Table 1). Consequently, we remain with values in the range 5.6 to 6.0, which we consider to be a reasonable confidence range. We therefore conclude that our best estimate for the magnitude of the 1819 earthquake will be 5.8 ± 0.2. This magnitude rests on two pillars: the quality of the magnitude (M_S) vs. felt area correlation, and the quality of the estimated felt areas for the 1819 earthquake as given in Table 1. The first of these questions has already been discussed, including the additional support that has come from the new analyses of the 1904 Oslofjord M_S 5.4-5.5 earthquake. The second question, which is a main point in the H&K paper, is discussed in the following paragraphs.

The downgrading of the intensities in the H&K paper for the 1819 earthquake is based on two parallel arguments. Firstly, considerable attention is given to a lack of reports from places like southern Helgeland and Lofoten. H&K call this negative evidence which it is not; negative evidence is, as pointed out by Wahlström (2004), when an earthquake is reported not felt, and not when an earthquake is not reported at all. H&K's use of 'outliers', in particular in their response to Wahlström (Husebye & Kebeasy 2004b), is similarly not supported by the usual statistical understanding of this concept.

The two best reports of the earthquake effects in 1819 are included in descriptions of natural history by contemporary vicars in the Rana (Hemnes) and Saltdal areas (Heltzen 1834; Sommerfeldt 1827, respectively). Their accounts were also published in newspapers in Oslo a few days after the earthquake (Muir Wood 1989) and consequently were not written several years later as indicated by H&K. If these two parson naturalists had not had this specialized interest in Earth science, reports from these two areas might have been just as sparse as the reports from Helgeland and Lofoten. The distance from Lurøy to southern Helgeland is almost the same as from Lurøy to Saltdal, about 100 km. When H&K emphasise the absence of reports from southern Helgeland and Lofoten, they consequently need to refer to contemporary natural history reports lacking descriptions of the 1819 earthquake from these two particular areas. It is also difficult to envisage how an earthquake could destroy the roof of a house in Saltdal, trigger a rock avalanche to the north of Bodø and cause a chimney to fall down in Overhalla (Muir Wood 1989), and at the same time not be felt in other areas located at the same distance or closer to the epicentre (Fig. 1). Moreover, when H&K admit to knowing of reports from Brekken, Overhalla and Stadsbygd they refer to these places as 'obscure'; if so, why are places like Saltdal, Lurøy and Hemnes less 'obscure'? Most of the population in Norway lived in rural areas in 1819 and not in towns and cities.

Our point is, simply stated, that a lack of reported observations cannot be applied in assessing the magnitude of an earthquake in 1819 in northern Norway, not least since Scandinavian newspapers were not issued to the north of Trondheim in 1819. H&K's comparison with a 1962 earthquake is, in this context, out of place. Wahlström (2004) provided a viable explanation for the missing reports of earthquakes from the Helgeland and Lofoten areas. They were simply regarded as local earthquakes with intermediate magnitudes and were not distinguished from similar earthquakes occurring from time to time in these two areas. Husebye & Kebeasy (2004b) neglected to comment on this proposed explanation in their response to Wahlström (2004). We therefore conclude, even if R_{III} admittedly carries a considerable uncertainty, that H&K have not brought forward any reliable evidence that can be used for reducing R_{IV}, and thereby also the magnitude of the 1819 Lurøy earthquake.

Near-field reports and local magnitudes

H&K further claim that the local near-field reports have been over-interpreted, and they have consequently reduced the estimated intensities from the Lurøy, Rana and Saltdal districts (Muir Wood 1989) from *VII* to *VI*. They have used this as an argument for downgrading the magnitude, where one of their points is that the rock avalanches in the steep mountains were caused by heavy rains and heavily increased wave amplification due to the steep topography. Their lack of reference to the abundant literature on earthquake-induced sliding is, in this context, a complication since their claims thereby remain largely unsupported. It is well known that the combination of dynamic loading from earthquakes and high joint (pore) water pressure greatly increases the probability of rock slope failures, which makes it hard to understand how the landslides in 1819 could not be related to the earthquake. Empirical relations (e.g., Keefer 1984; Rodriguez et al. 2000) indicate that a M6 earthquake under average conditions may trigger landslides up to a distance of 50-100 km. Similar effects from a M5 earthquake may be limited to 10-15 km.

Fig.2. Storstrand farm in the bay Utskarpen, looking east. The location is shown in Fig. 3 by a circle labelled with number 3. The depression located between the shore to the left and the forest to the right was caused by the landslide triggered by the magnitude 5.8 earthquake in 1819 (Heltzen 1834; Muir Wood 1989). The farmhouses were moved away from the scarp after the slide event. The terrain was levelled with a bulldozer in the 1960´s, but the scarps of the landslide are still visible (shown by the arrows). The bay to the left in the photograph was filled up with sediments after the landslide so that boats could not reach the shore. This shallow sea floor was uplifted above sea level during an aftershock in 1819. Muir Wood (1819) ascribed the uplift to piling up of material behind a rotational slump in the marine clay.

Given that the soils may have been saturated, as claimed by H&K, and moreover that the earthquake most likely had a shallow focus (Hicks et al. 2000a), the presence and levels of landsliding are readily explained as earthquake triggered. Heavy rains may cause superficial landslides but additional loading is usually needed to trigger deep-seated slides (cutting approximately 50 m into marine clay and forming an up to 15 m high scarp, Fig. 2) as in the relatively flat terrain at Utskarpen (Heltzen 1834; Keilhau 1836; Muir Wood 1989). Utskarpen is situated about 50 km from Lurøy where H&K located the epicentre. We would also like to add that three weeks of overcast skies and rain (Heltzen 1934) can not be regarded as extreme weather conditions along the coast of Nordland; this is, indeed, not abnormal. Earthquake accounts from the 19th century were often accompanied by meteorological reports. H&K's scepticism to the 'inflated' Lurøy reports does not include these weather observations and they have even added an extra week and claim that the rain lasted for four weeks. The intensities of *VII* should therefore be maintained in these particular cases. The report saying that "the ground heaved so fiercely that he fell down several times" (Heltzen 1834; Muir Wood 1989) points further to an intensity of *VII*. H&K argue that observations of people experiencing problems whilst walking appear in recent reports (Aasvik 1985) but, on the contrary, they actually originate from the initial reports on the earthquake (e.g. Heltzen 1834). This adds to the long list of H&K's imprecise and even incorrect use of the voluminous literature on the Lurøy earthquake.

Kebeasy et al. (2003) simulated the seismic wavefield responses in the Lurøy and Rana areas using a 3D finite difference scheme and concluded that shallow earthquakes (~5 km) below these two areas could cause a wavefield amplification of more than 20 due to the rough topography. Independent of the question of the reliability of that simulation we note that the Lurøy and Rana (Hemnes) areas are located about 50 km apart and there are no reports of two separate large magnitude earthquakes on 31st of August 1819. All reports indicate one large event followed by several smaller earthquakes. We therefore conclude that the many reported rock avalanches in the Lurøy, Træna and Rana areas and even as far north as Bodø can not be attributed to extreme wavefield amplification.

Helzen (1834) described spectacular sand-blows many decades before such structures were generally attributed to large-magnitude earthquakes by the geological community. Liquefaction structures are, according to most palaeoseismological studies (Obermeier 1996; Grünthal 1998), common at intensity values of *VII*. Jibson (1996) and Grünthal (1998), moreover, report that deep-seated, large landslides occur within levels of *VII-IX*. Hence, there are good arguments to maintain the reported intensity values of *VII* within the Rana, Lurøy and upper Saltdal region.

What H&K write about waves in the fjords is similarly confusing and leads us nowhere in terms of assessing the magnitude. The waves in question can only be of two types; either they are tsunamis (which we assume is what H&K mean when they use the word 'tidal waves') generated by subaqueous surface faulting (which is not likely in this case, even for a shallow focus earthquake) or by a landslide that either reaches the water or is fully contained under water (which is possible in this case), or they are so-called seaquakes. A seaquake is a well-known phenomenon tied to P waves through the water, but invariably limited to the epicentral region. Such seaquakes can be very strong and are even known to have caused the sinking of ships (Hove et al. 1982). The sinking of the schooner 'Henrietta' in 1894 (Muir Wood and Woo 1987), as mentioned by H&K, could have been related to a seaquake, in which case no 'exceptional sea waves' should be expected. It is the same phenomenon which is often reported as 'white sea', e.g. from the small Meløy earthquakes in 1978-79 (Bungum et al. 1979).

Our final comments with respect to the 1819 magnitude concern H&K's new local magnitude of M_L 4.8 (see Alsaker et al. 1991, who developed the M_L scale for Norway). Firstly, the M_L vs. felt area relation used (Almhjell et al. 2001) is not reported in a peer-reviewed publication and is therefore not easy to evaluate. One point here is that the slight nonlinearity in the M_S vs. felt area relations found by Muir Wood & Woo (1987) should be expected to be a lot stronger for M_L, because of a nonlinearity between M_L and M_W/M_S (e.g.,

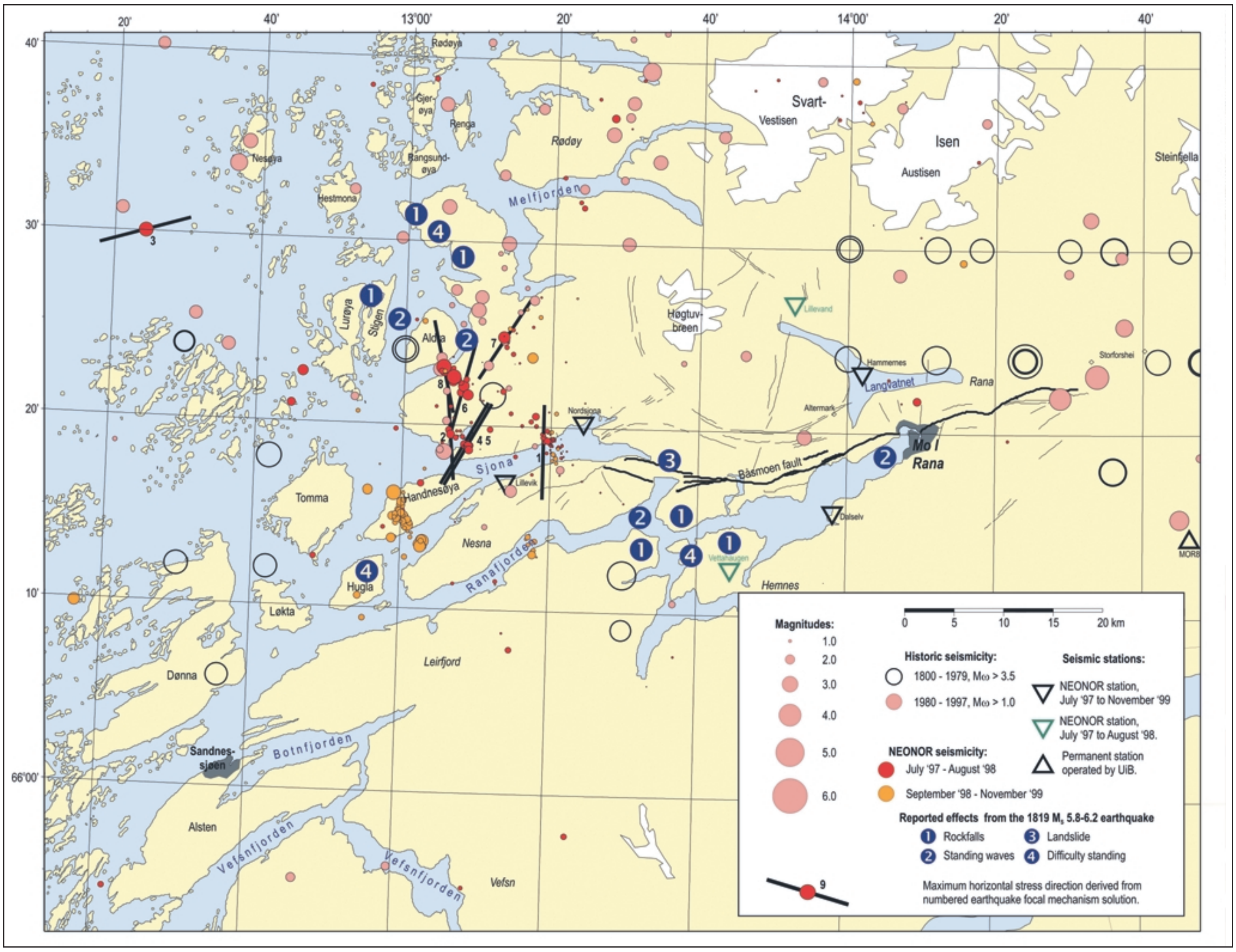

Fig. 3. Historical earthquakes (open circles), recent instrumental locations (pink circles) and new microseismicity locations (red and yellow circles) in the Rana area. Reported effects from the 1819 earthquake are shown by the numbered circles. A total of more than 300 earthquakes occurred during the period 1997-1999 in the Nesna-Sjona-Lurøy area which was also the focus of the 1819 earthquake. Horizontal compressive stress directions from eight focal mechanism solutions in this area are shown by the black bars. The focal plane solutions show that the area is strongly affected by present day extension. From Hicks et al. (2000b); see also Hicks et al. (2000a). The E-W extension is also supported by an observed negative deviation in the order of 1-2 mm/year from the regional uplift pattern in the Nesna-Sjona area, as reported by Olesen et al. (1995, 2004) and Dehls et al. (2002). This local subsidence is observed in two independent datasets, using levelling and permanent scatterers techniques, respectively.

Bungum et al. 1992). A linear extrapolation above the range covered by the regression may therefore lead to a significant underestimation of moment or surface wave magnitude. Secondly, we find that a R_{III} value of 350 km (and 3/4 of that for the short axis), as used in the M_L assessment, is not consistent with the observations available from the 1819 earthquake, even if (as mentioned above) 800-850 km may be on the high side. Incidentally, when using 350 km for R_{III} and an axis ratio of 3/4 in the Almhjell et al. (2000) formula we get an M_L value of 4.9 and not 4.8 as used by H&K.

Seismic hazard implications

We cannot conclude this communication without also commenting on the effects that a reduced magnitude for the 1819 earthquake would have, as claimed by H&K, on the seismic hazard in the region, namely a 25% reduction. This may be the case for H&K's hazard analyses, but certainly not from the point of view of the developers of the zoning maps (Bungum et al. 1998; Bungum et al. 2000) behind the new seismic design code for Norway (Norsk Byggstandardiseringsråd 2004).

The two main reasons for this can only be mentioned very briefly here. Firstly, we note that the assessment of maximum magnitude has a much wider perspective and is usually only weakly affected by a single earthquake. The model for maximum magnitude behind the zoning maps (Bungum et al. 1998) is, however, uniform all along the Norwegian continental margin, containing values (logic tree based) ranging all the way up to M7. These values are consistent with the fact that passive continental margins and aborted rift zones are the intraplate areas that experience the highest seismicity levels globally (Johnston & Kanter 1990; Johnston et al. 1994; Bungum et al. 2005). We therefore also strongly disagree with the H&K conclusion that a magnitude 6+ earthquake cannot occur in Norway today. Substantial additional support for this can be found from the fact that the length and offset of the Late Holocene Berill Fault in Møre & Romsdal, southern Norway (Anda et al. 2002) reveal an earthquake magnitude of 6.1-6.5 (Olesen et al. 2004). The fault is obviously not associated with the deglaciation period, as is the case for the other postglacial faults in northern Fennoscandia (Lagerbäck 1990; Dehls et al. 2000; Kuivamäki et al. 1998), and it therefore constitutes very good evidence for predicting the possibility of a future magnitude 6+ earthquake in the seismically most active areas of Fennoscandia.

Secondly, we note that the seismic activity rates in the different zones are quite robust and also smoothed in order to avoid strong local variations, as is also seen from the smoothness of the resulting hazard contours. Among the basis arguments for this assessment is the large number of earthquakes found along the coast of Nordland. Hicks et al. (2000a) recorded, for instance, a total of more than 300 earthquakes during the period 1997-1999 in the Lurøy-Sjona area (Fig. 3), and they also found that the activity rate from these events corresponds to a return period of 16 years for M_L 4 and 2300 years for M_L 6. This is consistent with the long-term seismicity in the region. The return period for M_S 6 for a large zone surrounding all of Norway and its entire offshore areas is, by comparison, about 100 years (Bungum et al. 2000). The effect on the hazard from a reduced 1819 magnitude should therefore be expected to be substantially lower than 25%, but probably varies with exceedance frequency. Hence H&K's claim on this point is both unfounded and unsubstantiated.

We would also like to stress that although the level of seismicity in Norway is stable, the societal vulnerability to earthquakes has increased enormously over historical time. One only has to compare the population (5,000) and infrastructure of the Rana region of 1819 (very few roads, no industry and mostly one-storey wooden houses) with the population (30,000) and infrastructure of the same areas today (a major road system, railways, hydropower plants, bridges, tunnels, smelters and tall buildings) to realize how many more people and constructions are at risk. The 1992 M 5.8-5.9 earthquake in Roermond, the Netherlands, for example, cost Dutch society a sum of 100 million Euros (van Eck & Davenport 1994; Berz 1994) and illustrates the importance of carrying out state-of-the-art seismic hazard analyses also in Norway. The fact that this country now has building design codes both for offshore and onshore constructions illustrates that this has also been recognized by its authorities.

Conclusions

It would appear from this detailed review of the 1819 Lurøy earthquake that the main purpose of the H&K paper has been to reduce earlier magnitude estimates for this event. To this end, H&K present a series of arguments that could possibly support a reduced magnitude, whereas reported observations and factors that lend support to a different conclusion are given less weight. Our main conclusion from this re-assessment of the data and reports is that a M_S magnitude of 5.8 for the 1819 Rana region earthquake still remains a reasonable, justifiable and defendable estimate, which is also in agreement with the conclusions of Wahlström (2004). The question of the magnitude for this earthquake will in any case have only marginal influence on seismic hazard models for this region.

We also conclude that magnitude 6+ earthquakes can occur today in the most seismically active areas in Norway, such as the coastal parts of western Norway, Nordland and the Oslo rift zone. Rock avalanches and landslides, potentially triggered by earthquakes, could moreover generate tsunamis and constitute consequently the greatest seismic hazard to society in the populated fjord regions of western and northern Norway.

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