BRITISH EARTHQUAKES

R.M.W. Musson
British Geological Survey, West Main Road, Edinburgh EH9 3LA
+44-131-650-0205   rmwm@bgs.ac.uk

Abstract

Although the UK is not a strongly seismic region, the study of earthquakes in Britain presents many interesting points. Firstly, British earthquakes are rather well documented through history, partly due to the intellectual development and literacy of the country, and partly because British earthquakes have always been so newsworthy that even minor events have been recorded in some detail. Secondly, as is rather typical for intraplate areas, the relationship between seismicity and geological structure is unclear. The definitely non-random spatial pattern of British earthquakes is clearly due to something, but among competing theories as to what that something is, no hypothesis is clearly the best. In this paper the subject of British seismicity is viewed from several angles. First, the sources that underlie the earthquake catalogue are discussed in order to give a clear indication of the limits on the completeness and accuracy of the data. General statistics for earthquake occurrence in the UK are then presented. This is followed by a description of British seismicity from region to region, with remarks on some key earthquakes of interest. Different hypotheses on the nature of the tectonic or geological control are then reviewed and assessed. Finally, the subject of active faults in the UK is discussed, from the point of view of general seismic hazard assessment.

Key words

Earthquakes, seismicity, UK, British Isles, seismotectonics, neotectonics, active faults, seismic hazard

Introduction

One day in the spring of 1992, an unusual enquiry was received by seismologists at the British Geological Survey in Edinburgh. A landowner near Perth, while tending his garden, unearthed a partly-buried carved stone pillar. This pillar, which appears to have been the base of a sundial or similar garden ornament, bore the inscription “Earthquake heard here, January 19, 1840”. The date was unfamiliar; existing compilations of British earthquakes listed no event on this date. Further investigation revealed that there was indeed an earthquake on this day, with magnitude of around 3.2 ML, which had been missed by previous compilers. The Perthshire Advertiser (30 January 1840 p3) carried a report from an anonymous correspondent in the village of Stanley, north of Perth, that the earthquake was heard (but not felt) there, and Musson (1993a) conjectures that the pillar was originally located in the garden of Stanley House, and looted from there after the house was destroyed by fire in 1887.

This anecdote can be viewed as symbolic of the story of British earthquakes. On the one hand they are generally small in world terms, as one expects in a purely intraplate environment. On the other, they are remarkably well documented. In what other country might one find a carved stone monument to an earthquake that was only 3.2 ML (local magnitude), and was only heard and not felt by the person who commissioned the memorial?

This high level of documentation partly reflects the rarity of occurrence of British earthquakes. When one does occur, it is a newsworthy event, and people are interested to record that they observed it, if only slightly. Thus it is not unusual, even in the historical record, to find observations of earthquakes at intensity levels as low as 2 EMS (European Macroseismic Scale), whereas in countries with a much higher level of seismic activity, weak earthquake shaking is sufficiently common that people do not bother to note it.

This has advantages in terms of the degree to which one can accurately determine earthquake parameters for historical events. There are two main approaches to estimating magnitude for pre-instrumental earthquakes from macroseismic data. The first is to relate the maximum intensity to magnitude; the second is based on the total felt area. If reports survive from only the damaged localities, the seismologist is largely restricted to using maximum intensity; however, this is a bad analogue for magnitude because it is heavily influenced by focal depth and local soil conditions. In contrast, total felt area is a very robust parameter which correlates strongly with magnitude and is unaffected by focal depth (Musson 1996a). Furthermore, with enough isoseismals, it is possible to estimate focal depth itself to some accuracy using the method first proposed by Kövesligethy (1906). Because the typical larger British earthquake, at least from the 18th century onwards, tends to be well-documented from the epicentre out to the fringes of observation, it is possible to construct a lengthy catalogue of British earthquakes with reliable parameters even for earthquakes in the pre-instrumental period (Musson 1994).

An exceptional case was the earthquake of 28 April 2007, in which a relative modest earthquake (4.3 ML) occurred at shallow depth close to the town of Folkestone, causing considerable damage concentrated in one area of the town, interruption of electricity supplies and disruption to travel. This was the first occasion in modern British experience of emergency procedures needing to be instated by local government as a result of a British earthquake. Since this paper was drafted in January 2006, the Folkestone earthquake is not included in any of the analysis or figures that follow.

In this paper, the seismicity of the UK will be discussed under three headings: firstly, the evolution of data sources and how this affects the resolution of UK seismicity over time. Secondly, the general and regional characteristics of British seismicity. Lastly, some account will be given of attempts to explain the distribution of earthquakes in the British Isles in relation to geology and neotectonics; and to active faulting. A further aspect, the history of the study of British earthquakes, has already been dealt with in depth in Musson (2004a).

British seismicity in relation to data sources

An understanding of the historical data from which our knowledge of British earthquakes is derived is important, in order to appreciate the limits on the earthquake catalogue. As discussed above, the relative rarity of earthquakes in the UK has paradoxically encouraged their documentation; but one must also consider the fact that the intellectually active culture of the inhabitants of the British Isles, and high literacy levels, has contributed considerably. Amongst other points of note can be cited: (i) the fact that documentation of local earthquakes was deliberately pursued as a scientific endeavour as early as the 1660s, with the founding of the Royal Society (Musson 2004a); (ii) that regional newspapers were published on a regular basis from around 1700 onwards (Musson 1986); (iii) variations in language and literacy sometimes impact directly on reporting patterns; for instance, the distribution of reports of the large Scottish earthquake of 8 November 1608 reflects the linguistic divide at that time between English-speaking (literate) and Gaelic-speaking (non-literate) areas (Musson 1989a).

Early records

Research into historical seismicity has to take into account the historical and intellectual environment of the times, and this extends to changes in language. In particular, the very word “earthquake” had a much wider usage in the past than is now the case. In particular, it was widely used to refer to landslips, landslides, cavern collapses, bogbursts and anything similar. One 18th century writer (Anon 1750) lists no less than nine kinds of earthquake, including one which turns out to be volcanic ash fall. Many 18th century writers on earthquakes (one might mention Burton 1734, 1737 as an example) give much weight to the Herefordshire event of 17 February 1571 as one of the “greatest” of English earthquakes – this was a landslip on the eastern side of Marcle Hill, the outlines of which can still be seen today. Even as late as the early 19th century, a major block slide at Lyme (Dorset) on 23 December 1839 was reported as being an “earthquake”.

Thus one must be wary of reports in monastic chronicles which simply state, “This year there was an earthquake”. Even if the same report appears in several chronicles, one cannot be sure what type of event is actually being referred to. One requires some extra clue; thus, “This year there was an earthquake throughout England” is clearly a real earthquake and not a landslip.

As a result, there is some doubt as to what is the earliest British earthquake on record. One can discount the many spurious pre-millenial earthquakes that originate in the work of Short (1749), including the notorious 811 St Andrews earthquake (which appears in the NOAA online earthquake database). These have long been recognised as fakes (Musson 2005a). The earliest contender is an entry in the Annals of Ulster (Balé and Purcell 2003 is the most recent edition) for 601 – “An earthquake in Bairche.” Bairche (or Ui Bairrche) was a kingdom in the south-east of what is now County Laois. The same source mentions an earthquake in Britain in 664 without details. A report in the Annals of Clonmacnoise (Murphy 1896) states that in 680 there was “an extreme great wind and earthquake in Ireland”.

There is no reason to suppose that any of these were definitely earthquakes; the conjunction with a storm in 680 makes it even more likely that that one was something else. More promising is the event in 684 (or 685 in some sources) which is variously described as occurring in Ireland, the Isle of Man, and Britain. This could be taken as a large earthquake with epicentre in the Irish Sea, but careful textual analysis by Dumville (1984) shows that all texts are derived from a lost original dealing with the Isle of Man, and that the references to other places are errors introduced by copyists. So this event could equally well be a rockfall somewhere in the Isle of Man. In 707 the Annals of Ulster record “Two earthquakes in the same week in the month of December in the northern part of Ireland”. This does sound as if it might be more likely to refer to a real earthquake. The same source also records events in October 721 (no place specified), 8 February 730 (no place specified) and 12 April 740 (Islay), any of which may or may not be genuine earthquakes.

The earliest English source is Florence of Worcester (Stevenson 1853-6) who records in 974 a tremendous earthquake all over England, and this is therefore the first event which is unquestionably a true earthquake. According to Goutoulas (1653) this earthquake threw down houses and killed people; whether Goutoulas was citing a source now lost or made this detail up one cannot tell.

For the next three centuries, English and Welsh chronicles become the most important source of information on British earthquakes. Undoubtedly Scottish chronicles were also compiled, but few survived the violence of the Reformation (Maxwell 1912). The principal surviving Scottish chronicle is that compiled at Holyrood (Stevenson 1853-6) which mentions only two British earthquakes (both English) and without details.

In general, entries in chronicles are greatly lacking in detail. Often only the date is mentioned. Thus only very few events can be located with any certainty. One technique that has been proposed is to use simple appearances of a notice about an earthquake in a monastic chronicle as necessarily indicating that the earthquake was felt in that monastery (Melville 1983). This allows one to build up maps that appear to be felt-area maps but which are in fact report-area maps. Confusing the two is highly dangerous; Houtgast (1992) relocates a North Italian earthquake to Central Europe by this means, on the basis of a map in Guidoboni (1983) which that author expressly notes is not a felt-area map. The problem is discussed at length by Musson (1998a) in which it is concluded that the assumption that a bare report in a monastic chronicle implies that the earthquake was felt at that locality cannot be sustained. In Musson (2004b) comparison is made between monastic reporting of earthquakes and reporting of storms, which demonstrates further that the presence or absence of mention of an earthquake in a particular chronicle is not a reliable indication that the event was or was not felt at the monastery in question.

An interesting case is the earthquake of 20 February 1247, almost certainly one of the largest earthquakes to originate in Britain. This has previously been assumed to have had an epicentre near Pembroke (e.g. Principia 1982) largely because it caused damage to St David's cathedral. As Woo (1991) argues, damage to single anomalous structures is a highly unreliable way of pinpointing the epicentre of an earthquake (one thinks of the concentration of media reporting on the 1997 damage to the basilica at Assissi, which was in fact on the edge of the felt area of the Umbria-Marche earthquake of that year). It is quite credible that the 1247 earthquake could have been a Caernarvon-area event similar to those of 1852 and 1984. On the other hand, it is equally possible that it really was a Pembroke event, a larger version of the 18 August 1892 earthquake. There is simply no way of resolving the issue. Stretching the scanty data further than they will go is not a solution. Similar cases are presented by the earthquakes of 15 April 1185 and 11 September 1275, which damaged Lincoln cathedral and St Michael's Glastonbury respectively. It would be most unsafe to propose epicentral locations for either event based purely on damage to single, rather anomalous, structures; the one a cathedral in unknown condition at the time of the event; the other a church perched on the top of a steep hill (Glastonbury Tor).

Late and post-medieval periods

After the 13th century, reporting of earthquakes declines, as the old style of compendious annal was gradually replaced by a more consciously historiographical style of writing, with a greater political focus and less interest in natural phenomena. In the 15th century, there is almost a total lack of recorded earthquakes for Britain; it is unlikely that it was a seismically quieter century than any other; rather, it is a reflection of the type of historical sources that were compiled at the time, and which have survived to the present day. However, from the mid 16th century onwards, the increase in literacy and spread of printing greatly encouraged the recording of earthquakes. The damaging earthquake of 6 April 1580 (Dover Straits), which killed two children in London, sparked a flurry of printed pamphlets describing the event; some of these are now lost, but several survive and are useful sources on the event (Ockenden 1936, Neilson et al 1984). One of these contains the first ever British earthquake catalogue (Fleming 1580).

However, throughout the 16th and 17th centuries, the preservation of reports on earthquakes is largely a matter of chance. Sources are often miscellaneous documents: letters, notes on flyleaves, pamphlets, memoranda, etc, which cannot be relied on to survive or become known. An earthquake felt in Antrim around the end of the 16th century is known only because Sir Thomas Molyneaux remembered hearing an old lady mention that when she was young she heard it recalled that years before she was born, Sir Hugh Clotworthy felt an earthquake in his house. Molyneaux put this recollection in a letter to his brother William in 1690; the letter survived and was published in 1841 (Wilde 1841); and only thus do we know of the earthquake. (It may have been a distant observation of the large Scottish earthquake of 23 July 1597).

On the other hand, no account has survived of the moderately large Scottish Borders earthquake of June 1668. That this earthquake happened at all can be construed from two later documents, but no contemporary record of it has been traced. This puts considerable limits on what one may deduce about the completeness of the historical record up into the late 17th century.

Macroseismic data from newspapers

At the beginning of the 18th century the situation changes significantly with the coming of the newspaper as a feature of British life. The use of newspaper data in historical earthquake studies is quite different from the use of manuscript data, for one simple reason. It is not the job of occasional writers to record every event that happens in a place, so whether they do or not is down to chance factors. In the case of newspapers, however, it *is* their job to record such day to day occurrences, and therefore, within certain limits, one has reason to expect to find data preserved. Thus, if an earthquake which was felt in Manchester is not reported as having been felt in Leeds by a newspaper published in Leeds, the absence of data does suggest that the effects of the earthquake in Leeds (if any) were limited and not news-worthy, a conclusion that cannot be so easily drawn with manuscript sources. Earthquakes are news, and newspapers make it their business to report news. This makes them a prolific and very useful source of data, so long as one bears in mind the limitations of newspapers as sources.

The biggest limitation is a historical one. Although printed newspapers in Britain are amongst the earliest in the world, they started on a limited basis in the more economically important towns and cities, and only gradually spread to the whole country. Also, the survival of complete runs of some early newspapers is not as good as one would like. Existing collections of some titles, particularly before the 1820s is fragmentary. In such cases, a paragraph describing an earthquake may sometimes survive through having been copied by one newspaper for which a good run of copies survives, from another newspaper now lost.

Early newspapers were quite unlike those that are published today. The typical 18th century newspaper consisted of four pages; most of the first and last pages were given over to advertisements, while the second page and part of the third would carry national and foreign news as received from London. Then one or two columns on page three would contain information about local occurrences, not as separate news items, but as a succession of short paragraphs. This would include any accounts of an earthquake being felt in the town of publication or in the vicinity, or the area served by the newspaper, or sometimes further afield. These accounts were sometimes simply copied from other newspapers, often without attribution. This means that a newspaper published on Friday 20 May, stating that "Last Tuesday an earthquake was felt in X", may be simply copying an item published in another paper on Friday 13 May - the date of the earthquake would therefore be 10 May not 17 May. This sort of confusion has often resulted in duplications in earlier earthquake catalogues.

During the early 19th century, and especially after 1853 when taxes on newspapers were eased, the typical local newspaper expanded in size, and the amount of information carried increased substantially. A typical local paper of the 1870s would have a page of national and foreign news extracted from the news agencies, a leader column which would be followed by local news for the town of publication, more local news for the surrounding area arranged by town or village, and then various features such as farming news, sports, fiction, household and so on, in addition to the advertisements which would completely cover at the very least the front page. In such cases, the principal parts of the paper for scrutiny for earthquake data are the local news items, which present first-hand accounts of the local effects of the earthquake. Items appearing via the news agencies rather than the paper's own staff are second-hand and tend to be less reliable. Useful information can sometimes appear in less expected places, for example, a diary column.

At this period, pages were large, type faces were small, headlines were small or non-existent, and pictures were generally absent. As a result, the sheer number of words in a single issue of a newspaper could be quite large. (In the absence of other forms of entertainment, having a lengthy newspaper to read was appreciated by many families.) Correspondingly, the amount of information printed on an earthquake when it occurred could be significant. Correspondents in the different villages would send in their accounts, and it is possible to compile reasonably full accounts of the macroseismic fields of the larger earthquakes from such reports.

In the course of the 20th century, the British local newspaper went into something of a decline. Many ceased to publish or were merged into other titles, and those that remained changed their style considerably. The modern style is characterised by far less text; space is bulked out by large headlines and pictures, as well as larger type sizes. The average report of a local earthquake in a modern newspaper will have a very large headline and one or more large pictures of such subjects as a local person who felt it, a seismologist, or a map of the places affected. The text itself will be short, with an interview with a seismologist forming the bulk, and perhaps one or two very short personal experiences, usually devoid of the sort of data from which one could assign intensity.

Accordingly, the quantity and quality of information about the effects of local earthquakes derived from newspapers declines in the second half of the 20th century. It is fortunate that after 1974 such data are preserved through regular macroseismic surveys of earthquakes conducted by BGS. The use and limitations of newspaper data in macroseismic studies is discussed in more detail by Musson (1986).

Early instrumental records of British earthquakes

The introduction of instrumental recording of earthquakes goes back to around 1900, and to some extent, even earlier. Primitive instruments built on the simple pendulum principle existed in Italy as early as the 18th century. The first inverted pendulum instrument was built in Scotland in 1840 (Forbes 1844). These early instruments had no way of measuring the time at which an event occurred, and the first clock-regulated instrument was that of Palmieri in Italy in the 1850s.

Some British earthquakes of the 19th century left instrumental traces on other devices also; the 1871 earthquake in the North Penninnes is supposed to have left a record on a magnetogram from Stonyhurst, though inspection of the record today does not reveal this. The 1884 Colchester earthquake was certainly recorded quite clearly on a magnetogram at Richmond.

The important development made by John Milne and his colleagues in Japan in the 1880s was the introduction of an instrument of known mechanical properties such that the movement of the ground could be calculated accurately from the instrumental trace. It was this that opened the door to useful instrumental studies in seismology. Instruments of the Milne type were deployed in Britain in the 1890s.

The first British earthquake to be recorded instrumentally was the Derby earthquake of 24 March 1903, which was detected on a Milne instrument at Bidston, an Omori at Birmingham run by Charles Davison, and the Wiechert instrument at Göttingen.

The main problem with early instrumental records of British earthquakes is that the instruments in use in the earlier part of the 20th century were designed primarily for the study of large earthquakes at teleseismic distances. They were very insensitive to high-frequency waves from small earthquakes at close distances. As a result, the traces, where they exist at all, are frequently so small as to be very difficult to read.

Since, for most historical British earthquakes, so few instrumental records exist, there is no possibility of using instrumental data to locate the epicentre or depth, which is better done from macroseismic data. However, measurements of the magnitude are possible, subject to uncertainties. The source of these uncertainties is threefold. Firstly, there is the difficulty of reading the maximum amplitude in the first place from such records. However, because magnitude scales with the logarithm of the amplitude, even an error as large as a factor of two in reading the amplitude will only change the magnitude value by 0.3. Secondly, the derivation of the ground motion amplitude from the seismogram trace amplitude may be uncertain if there is uncertainty about the transfer function of the instrument in question, or if the instrument was badly calibrated. Thirdly, if there is uncertainty in the position of the epicentre, there will be a problem in correctly estimating the epicentral distance; this is especially important if the amount of locational uncertainty is a significant fraction of the distance to the instrument (Neilson and Burton 1984).

One can add that, since final instrumental magnitudes are usually the mean of several station magnitudes, if there are only a few recordings this increases the overall uncertainty of the final value, since individual station magnitudes may be significantly affected by the radiation pattern of the earthquake.

Therefore, from 1900 to 1970, although instrumental recordings of British earthquakes are a very useful source, frequently one finds that one has to rely on macroseismic data still as providing the most accurate parameters.

The development of early instrumental monitoring has been traced in detail by Lovell and Henni (1999), in a report that lists information for all known stations in the British Isles that were operative before 1970. The total number of such stations is 51, but this includes a number that were run on an amateur basis, sometimes only for a very short period. For many stations, although it is known when they operated and which instruments were used, the seismograms have long been lost and even bulletins may not survive. The principal surviving collections of historical seismograms (held in BGS archives) are those from Aberdeen, Paisley, Edinburgh, Eskdalemuir, Durham, Bidston and Kew. It is very regrettable that the records from some important stations, such as Stonyhurst and Oxford, have long since vanished. (One seismogram alone survives from Oxford, and only because it was borrowed and never returned.)

Extensive use of surviving records, including bulletins where these are available, was made by Neilson and Burton (1984, 1985, 1988) in determining magnitudes for earthquakes between 1900 and 1970. The earliest British earthquake for which a seismogram exists is the 1903 Derby earthquake.

Modern instrumental data

The very start of specific UK-oriented earthquake monitoring using modern seismometers capable of detecting small-magnitude earthquakes was in 1967. At this date work began on LOWNET, a network of short period vertical component seismometers (Willmore Mk 2) deployed at seven outstations in Central Scotland and a central three-component set in Edinburgh (Crampin et al 1970). This network was formally declared open in January 1969; it was operated by the then Institute of Geological Sciences, now the British Geological Survey (BGS). This network proved very capable of detecting small earthquakes in Central Scotland (this was a period when a large number of mining induced earthquakes were observed in the Midlothian, Fife and Clackmannanshire coalfields) but its detection capacity for events in England and Wales was limited. Annual catalogues from LOWNET up to 1978 are given by Burton and Neilson (1980).

In response to a variety of factors, the network gradually expanded after its original foundation. The first expansion south came in 1976 when the Department of the Environment sponsored new stations near Stoke-on-Trent, Leeds, Leicester and Hereford. The occurrence of significant felt earthquakes in the Kintail area in 1974-5 resulted in extra stations also being deployed in the Kyle of Lochalsh area. Further additions to the network were made in an ad hoc manner in response to particular demands – stations in Shetland, north-east Scotland and East Anglia in connection with hydrocarbon developments in the North Sea; a dense local network in Cornwall and Devon to monitor the Hot Dry Rock geothermal project (Browitt 1991).

In 1989 the network was put on a new footing with the establishment of a Customer Group to sponsor the BGS Seismic Monitoring and Information Service. This Customer Group is led by what was then the Department of the Environment, now the Office of the Deputy Prime Minster, and includes members from the Government, Nuclear, Hydrocarbon and Water sectors amongst others. Under this project, the network was gradually expanded, in order to provide coverage over the whole country. This development, which combined specific local objectives with a recognition of the need to provide uniform coverage of the country, is detailed in a series of annual reports, of which the most recent at time of writing is Baptie (2005). Figure 1 shows the station distribution at five stages of development: 1970, 1980, 1985, 1993 and 2005.

The increase in geographical coverage and the increase in network density can both be seen in Figure 1. The current configuration has just over 140 stations, and allows the detection and location of any earthquake of magnitude 2.5 ML or greater anywhere in the UK, even in bad noise conditions. For many parts of the country detection capability is even better than this, giving data completeness down to 1.5 ML or lower.

In addition to an increased number of stations, the last ten years has seen the deployment of a greater number of low-gain instruments, strong motion recorders and broadband stations. A future plan is to shift the emphasis of the network to broadband recording with a large dynamic range. A problem with the network during the 1980s and 1990s was that most of the network consisted of instruments with relatively narrow dynamic range and high gain. The high gain setting allowed the detection of common small earthquakes, but less common larger earthquakes had a tendency to saturate the network.

The net result of BGS's seismic monitoring activity since 1969 is that the seismicity of the UK in the last 35 years is known in great detail, even down to quite small magnitudes. Thus for the instrumental period, we have a similar situation to the historical period, that although the seismicity itself is low in world terms, the data set available for study is quite rich because it extends to low magnitudes. It is interesting to note that the spatial pattern of minor earthquake activity from recent instrumental data, even if one takes only 5-10 years of data, very closely mirrors the long-term spatial pattern (which will be discussed in more detail later). The most significant deviation is that little or no recent seismicity in South Wales west of the Rhondda has been detected, though this area has been notable for strong earthquakes in the past.

The UK earthquake catalogue

The first catalogue to combine historical data and modern instrumental results in a systematic way is the published catalogue of Musson (1994). Although there is not, at the time of writing, any published update on this, the data file has been kept up to date with regards to earthquakes occurring post-1994, and also revisions to historical earthquake parameters, and the inclusion of some events only recently discovered, of which the largest is the 1650 Galashiels earthquake, around 3.5 ML in size (Musson 2004b). The file at present has some restrictions according to parameters, as follows. For the period before 1700 the intention is to include all known events of magnitude at least 4 ML; between 1700 and 1969, all events of magnitude at least 3.0 ML; and thereafter all events of magnitude at least 2.0 ML. In practice the file does include events smaller than these limits, but not systematically. Small events can be hard to catalogue for the historical period. For example, if a report states that "several shocks were felt over the course of the following week", does one attempt to add multiple parametric entries to reflect this or not? Because of the relative unimportance of very minor seismicity (especially aftershocks) one is inclined not to.

Parameters of historical earthquakes are estimated using the procedures described in Musson (1996a). The key parameter for determining magnitude is the area within the isoseismal 3 EMS; or isoseismal 4 EMS if 3 is poorly defined; or felt area if there are no isoseismals.

For small earthquakes one typically may have a list of a few places where the shock was felt, with little more detail preserved. In such cases the area defined by these places, with a small margin, is assumed to be the felt area and is assumed to be equivalent to isoseismal 3 EMS. One could argue in such cases that probably for the event to be reported at all the intensity must be more likely to have been 4 EMS, and therefore it would be more appropriate to use the equation for deriving magnitude from isoseismal 4 EMS. The difference would be, for example, for an event felt over 600 km^2, between a magnitude of 2.7 ML and 3.3 ML depending on whether one assumes intensity 3 or 4. It has been found in practice that the latter assumption leads to a discontinuity in the magnitude-frequency plot, suggesting that the former is more accurate.

This leaves the problem of events reported from one place only, for example, the earthquake of 12 July 1852 of which the sole original report merely states that the earthquake was felt as a slight shock at Kilmarnock. In such cases, where there is no felt area at all, the default is to set a magnitude of 2.0 ML.

Catalogue completeness

The issue of the limits to which an earthquake catalogue is complete is an important one, especially when investigating seismic hazard. There are two ways to estimate this; the first is by statistical analysis of the data, and the second is by consideration of the nature of the sources.

Statistical methods are mostly derived from the work of Stepp (1972) and assume that occurrence rates are constant over time. Thus if one finds that the average rate of occurrence of earthquakes greater than magnitude M is lower for the period 1500-2000 than it is for 1800-2000, this is because for part of the period 1500-2000 the data set is incomplete for magnitude M, and the missing events have lowered the overall rate. Using this principle, one can find the break point between the complete and incomplete parts of the catalogue.

The historical approach is to consider how well documented a period is for a particular area. If there are many good written sources after (say) 1700, one may conclude that after 1700 it is inconceivable that if an earthquake of magnitude M had occurred, it would not have been reported in such a way that today we would know that the earthquake had occurred. This is ultimately a judgement call, but it can be used to discriminate between regions that could never be analysed statistically because of few data. For instance, historical written sources for eastern Sutherland, Caithness and Orkney are much better than for western Sutherland; thus one can infer that the earthquake record is more complete for the former area even though it contains almost no earthquakes.

Taking magnitude 4 ML as a reference point, historical judgement suggests that the catalogue is probably complete as follows:

Region	*Start of completeness*
South England	1700
South-west England	1750
Wales	1750
North England	1750
South Scotland	1700
North Scotland	1850

Table 1 - Regional earthquake catalogue completeness in the UK

The actual number of observed events in each of these regions by 50-year intervals is shown in Figure 2.

A statistical analysis is shown in Figure 3 for the whole of mainland Great Britain, for magnitudes 4 ML and 5 ML. These plots show the average number of events per year for subcatalogues 1600-2005, 1605-2005, 1610-2005 … 1995-2005. Reading the graph from right to left, one has first a region where the average values fluctuate because the short subcatalogues are not representative. As the length of the record grows, the value becomes a more stable representation of long-term rates, and then starts a steady decline as one reaches the time period for which the catalogue is incomplete. For magnitude 4 ML this is about 1850. For magnitude 5 ML, about 1770.

In any analysis of completeness, one is trying to estimate how many earthquakes, and of what size, are missing from the record. In Britain one is in the unusual situation of actually knowing about a few cases of earthquakes that are missing (normally this is impossible – if one knew about them they wouldn't be missing). It appears that in June 1668 a largish earthquake occurred in the Scottish Borders, for which no documentary account whatever exists. That it happened can be inferred from two much later reports which use this date as a reference point in discussing later earthquakes. Historically, it makes no sense to refer to this date unless it was the date of another earthquake. One report is from Kendal, the other was written in Edinburgh but refers to Peebles, Kelso, Dumfries and Berwick. Thus this earthquake, for which no description exists, must have affected a wide area of North England and South Scotland and have had a magnitude of at least 4.4 ML (Musson 2004b).

A second similar case is the Ullapool earthquake of 1925, of which again there are no direct reports. When an earthquake occurred in the Dingwall area on 22 December 1925, it was mentioned that people who felt this earthquake in Strathpeffer compared the shock with one felt “several weeks ago” (so late November – early December) which was stronger in the Loch Broom district (North

Star, 26 December 1925 p4). This suggests an earthquake with a magnitude of at least 3.5 ML in north-west Scotland, for which no immediate record has survived. It is not mentioned by Dollar (1950).

I have also heard it rumoured that an earthquake of some strength was felt in western Scotland in the 1960s but it is hard to know what weight to put on this.

Now that the basis for the data on British earthquakes has been described, and the limitations on the data laid out, one can now move on to discuss the patterns of activity that are revealed.

British earthquakes in time

As is well known, the frequency of earthquakes is inversely proportional to their magnitude such that

$$\text{Log } N = a - b\, M \qquad (1)$$

where N is the cumulative number of events above magnitude M. This is known as the Gutenberg-Richter law. Here this relationship is studied for an area comprising the mainland of Great Britain, the Orkneys, the Irish Sea and the immediate waters off the east coast of Britain (Figure 4). In order to get the most out of the data, the catalogue is divided into four periods and the data normalised accordingly. Magnitude 3.0-3.9 is analysed with respect to the period post 1970; 4.0-4.9 for the period after 1850; 5.0-5.5 from 1770 and 5.6 and above from 1550. This is a smaller area than was used in Musson (1994) and with more pessimistic assumptions about completeness. However, the results are not dissimilar. The constants a and b in equation (1) are determined to give the result:

$$\text{Log } N = 3.59 - 1.01\, ML \qquad (2)$$

This is shown in Figure 5. The fit is made using least squares, as this is probably more appropriate than maximum likelihood given the fact that the graph is a composite over different historical periods. There is no indication that a truncated exponential relationship (as in Cornell and Vanmarcke 1969) would provide a better fit.

One can now compute the magnitude of earthquakes with different return periods (return period is the inverse of the annual probability of occurrence) for Great Britain as follows:

Return period (years)	*ML*
1	3.5
10	4.5
100	5.5

Table 2 - UK earthquake return periods

The corresponding values given in Musson (1994) were 3.7, 4.7 and 5.6 ML. Note that the analysis above is performed for the catalogue after removal of aftershocks (by hand) so applies to main shocks only.

In Figure 5 there is a prominent bump in the curve around 5.0 ML, which is also evident in Musson (1994), where it was explained as being most likely due to a greater number of earthquakes in the magnitude range 4.8-5.2 ML occurring by chance in the historical period of observation. More recent work (Musson 2005b) casts doubt on this. Figure 6 shows, for the area in Figure 4, the number of earthquakes occurring within magnitude bands 4.0-4.1, 4.2-4.3, etc, for the period since 1850. These are discrete values rather than cumulative ones. For the magnitude range 4.2-5.2 the distribution is completely flat, and there is no indication that an earthquake of exactly magnitude 4.2 ML is more likely than one of 5.2 ML. The probability of this arising by chance, with a number of earthquakes this large, is extremely remote. This pattern is not uniform across the country; South Wales-Hereford in particular seems to produce more earthquakes 4.8-5.2 ML than it does 4.1-4.7 ML. The figures for the period since 1850 are seven and one respectively. This is simply not compatible with a straightforward magnitude-frequency relationship of the form of equation (1). This behaviour is really rather peculiar. It suggests something approaching characteristic earthquake behaviour, although the strict characteristic model as originally proposed by Schwarz and Coppersmith (1984) does not seem to be appropriate (as it implies complete rupture of faults).

Depth of British earthquakes

Depth determinations for modern instrumental earthquakes are not always well constrained, and those outside the network may be very poorly constrained. Depth determinations for historical earthquakes are derived from macroseismic observations (the pattern of spacing of isoseismals), and this method often produces uncertain determinations. However, for events since 1970 for which both instrumental and macroseismic depth measurements have been possible, agreement is generally good, and there is no systematic bias of macroseismic depths towards deeper or shallower estimates.

Figure 7 shows a cross-section of British seismicity from roughly south to north along the long axis of the island, 490 km long and 336 km wide. There is a tendency for seismicity to be shallower in the south, especially in Cornwall and Devon. Average depth becomes slightly deeper as one progresses north, with some pockets of notably deeper events in south and north Wales. Then in southern Scotland the seismicity is again shallow, becoming deeper to the north. Focal depths, specifically of the best-located modern earthquakes, were studied by Baptie (2002), who divided the country into six regional areas, and found the shallowest mean depths in north-west Scotland and Cornwall, with means of 5.7 km and 6.8 km. However, the dataset used is dominated by small-magnitude events, and these may not be representative of the pattern for larger earthquakes.

One thing that is significant about Figure 7 from a practical point of view is that there does appear to be a correlation between magnitude and depth. This is shown more clearly in Figure 8. Only two earthquakes are both larger than 4 ML and have depths of 5 km or less. It is clear that larger earthquakes are much more likely to nucleate in the crystalline basement than the upper crustal layers.

The depth at which earthquakes can nucleate is controlled by the depth of the brittle-ductile transition, which limits the possibility of brittle failure. The depth of this depends on factors such as geothermal gradient, mineral composition, the presence of fluids and the tectonic strain rate. Information on crustal structure in the UK dates back to the LISPB refraction experiments in the 1970s (Bamford et al 1978). A study by Whittaker and Chadwick (1984) identified three broad zones with depth, of which the lowest, consisting probably mostly of heavily tectonised metamorphic gneisses, is characterised by ductile deformation. The top of this layer, and therefore the limit to the seismogenic zone, is considered to vary between 11 and 22 km. However, some earthquake depths do exceed this, especially in Wales, western Scotland, and an area around the Humber. This may be due to lower geothermal gradients causing brittle behaviour even in the lower crust.

When one combines the observed depth distribution with the small rupture sizes to be expected for earthquakes of the typical magnitude experienced in the UK, it becomes clear that the likelihood of any British earthquake producing surface rupture is extremely small. No surface rupture has ever been reported in any British earthquake in modern or historical times. Stewart et al (2001) review the evidence for palaeoseismic surface rupture in Scotland and conclude that the evidence that has been advanced in other studies for such rupture is unreliable.

Regional character of British seismicity

The character of UK seismicity varies considerably from region to region. Before discussing the relationship of seismicity to geology, it will be appropriate to give some attention to discussion of the spatial pattern of earthquakes in the UK. This is shown in Figure 9.

It has been hypothesised (Ove Arup 1993) that the distribution of earthquakes in the UK can be considered to be entirely a chance artefact, and that seismicity is in fact random over the whole country. This hypothesis can be disproved by the application of statistical tests for random distribution (Musson 2000a). The variation in seismic behaviour between different parts of the country is real and significant, and therefore leads to questions as to why this should be so. The different regions will now be discussed in turn, working roughly north to south.

1) Scottish Highlands

The Western Highlands of Scotland comprise one of the more actively seismic areas in the UK. Seismicity is quite strong within a zone reaching from Dunoon to Ullapool. Within this, however, the pattern of epicentres is to some extent clustered. One can observe an almost linear band of activity from Dunoon, through Oban to Moidart, running NNW-SSE. A second cluster near Fort William is also sublinear with a north-east-south-west trend, paralleling the Great Glen, but south of it. A third cluster is found around Kintail, which was very active in the 1970s, but some historical epicentres are also found here. A smaller cluster near Ullapool saw some swarm activity in 1987.

How much this pattern is influenced by poor locations is hard to say. Some historical earthquakes in this region are clearly very poorly constrained, especially early events such as the large 1597 earthquake, which is plotted in Figure 9 round about Glen Garry, but could in truth easily be further north, south or west. An earthquake in 1817 was felt in Glasgow and Edinburgh and Inverness, but in the absence of any reports from the epicentral area, the location of it is greatly uncertain. The epicentre of the 28 November 1880 earthquake (5.2 ML) has been disputed, and the subject was reviewed by Musson (1989b). Principia (1982) placed the epicentre west of Coll, and this is found also in e.g. Ambraseys (1988). It now seems clear that the epicentre was between Oban and Inveraray, but this is not very precise. Even in the 20th century, the earthquake of 16 August 1934 was considered by Dollar (1950) to have an epicentre near Inverness (largely as a result of jumping to conclusions); Principia (1982) placed it somewhere in the wilds of Strathconon Forest. Musson (1989a) concluded that a series of small events in the Torridon area shortly following this earthquake were in fact aftershocks, and relocated the event to Torridon; thus the epicentre has “drifted” from the east coast to the west coast in subsequent studies.

One can overestimate this problem. If one restricts consideration to events post-1975, which should be mostly well-located, the spatial pattern is not dissimilar. Furthermore, many historical determinations are not subject to much doubt. One can consider examples such as the sequence of small but strongly-felt events in the Strontian area in 1809, which clearly had epicentres close to Strontian. The 26 December 1946 Lochaber earthquake (4.1 ML) is located from macroseismic data, but these are sufficiently detailed and copious that one can locate the earthquake with some confidence to Glen Roy.

It has been suggested, e.g. by Browitt et al (1976) that historical earthquake epicentres in Scotland “migrate” to valleys because this is where the population is, and hence where the felt reports are. To some extent this is obviously true; a historical earthquake with an epicentre in a completely remote area can only be known from reports from the nearest settled area, and if the earthquake is not strong enough to be felt over a wide area, the nearest cluster of villages will appear to be the location of the event. That said, in many cases the data are sufficiently good that one can see that this effect is not occurring. One might suppose that epicentres are naturally drawn to the Great Glen due to considerations of human geography; but the macroseismic locations of the 1839 Invergarry and 1858 Stratherrick earthquakes are both east of the great Glen.

Of particular note are the Inverness earthquakes of 1816, 1890 and 1901. What relation these have to the other seismicity of northern Scotland is hard to say. There has been a historical tendency to assume that the Great Glen Fault (GGF) must be active because there are earthquakes in its vicinity, and therefore that the earthquakes in its vicinity must have occurred on it, because it is an active fault – a circular argument. Thus Davison (1924) assumes that the 1888 Invergarry earthquake must have occurred on the GGF, although careful scrutiny of the data would have shown this to be clearly not the case. The 1934 “Inverness” earthquake has already been mentioned; likewise the 1946 Lochaber earthquake, which is categorically stated to be a GGF earthquake by Dollar (1947) solely on the grounds of rough proximity, and the assumption that the GGF is “active”.

The three earthquakes of 1816, 1890 and 1901 occurred just to the south-west of the city of Inverness, and all three had substantial aftershock sequences, which aid in the accurate location of the main shock. The magnitudes were 5.1, 4.5 and 5.0 ML and damage from the two larger ones was considerable. It was reported of the 1816 earthquake that it was fortunate that at the moment it occurred, the streets were deserted, otherwise there would certainly have been casualties from the shower of bricks and tiles that fell. The 1901 earthquake caused a crack to appear in the canal towpath at Dochgarroch – this is evidently a ground failure effect rather than a fault trace, but even ground failure effects are uncommon in British earthquakes. The epicentre of the 1901 earthquake appears to be at Dochgarroch, and this is actually on the trace of the GGF, but it is also at the point where a north-south trending fault splays off to the north, according to Horne and Hinxman (1914), and the 1816 and 1890 epicentres are to the north and compatible with this fault. As will be discussed later, north-south and east-west trending faults are more likely to be reactivated given the prevailing stress direction, so this minor fault, if it is present at seismogenic depths, is probably more likely to be the causative feature of the Inverness earthquakes than the GGF. However, even if this fault were directly responsible for the Inverness sequences, this does not mean that the GGF is uninvolved, as will be discussed further below.

Since 1901 there have been no earthquakes around Inverness at all. Given that three significant earthquakes have occurred here in historical times, it seems not unreasonable to expect that a fourth may occur at some future date. Such a future event should settle the question as to the cause of the Inverness earthquakes; on the other hand, given the growth of the city since 1901, the human consequences may be unfortunate.

Outside the main zone of activity in the Western Highlands, there is some more or less diffuse seismicity in the Inner Hebrides-Kintyre area, from Mull to Arran. The 4 March 1999 Arran earthquake (4.0 ML) was the last event at time of writing to have been widely felt in Scotland. The epicentre was off the south of Arran.

The rest of Northern Scotland has very little seismicity. The Outer Hebrides are aseismic; there is very little activity in Sutherland, Caithness, the Orkneys and Shetlands (Musson 1998b). All of the Grampians, Aberdeenshire, Buchan etc is almost completely without known seismicity; the Aviemore earthquake of 28 August 1995 (2.7 ML) which was felt by many, was unusual in occurring in such an area.

2) Central Scotland

A characteristic feature of much British seismicity is the occurrence of earthquake swarms. An earthquake swarm is defined as a series of earthquakes, usually small, with no clear main shock. The classic example is the Vogtland area on the German-Czech border (Grünthal 1989). Swarms may contain hundreds of events extending over a period of weeks or months. In Britain swarms vary as regards their character, and may be quite short. The Ullapool sequence of 1987 has already been mentioned; it lasted for about a month, in which time ten events over 2.0 ML were recorded, including two 3.0 ML shocks and one 2.9 ML. The Manchester earthquake sequence of October 2002, while having what could technically be regarded as a main shock, was very swarm-like in character (Baptie and Ottemöller 2004).

The seismicity of Central Scotland is dominated by two swarm centres that have been active repeatedly: Comrie and the Ochil Hills. The first of these was important for the history of seismology.

The Comrie activity is known mostly through historical data; considering that at its height scores of shocks were being felt in Comrie village per day, for a protracted period, the total number of events that would have been recorded had modern instrumentation been available must have been immense. The typical Comrie swarm includes one or two large earthquakes, so it could be debated whether these are true swarms or just large earthquakes with abnormally intense foreshock and aftershock sequences (Kárník 1992 pers. comm.).

The history of Comrie activity can be summarised as follows:

8 November 1608 – a large earthquake widely felt in Scotland, which caused damage at Perth, most likely originated at Comrie, but due to the early historical period no records are available closer than Perth. No information is therefore available as to whether it was followed by more activity; a strong earthquake in Scotland on 9 March 1622 may have been related, but this later event is so badly documented that there is no way to locate it.

7 September 1801 – swarm activity started at Comrie in 1788, and this time the activity was chronicled by two local ministers, so a good account of the sequence is preserved (Milne 1842-4). The activity reached a climax with a 4.6 ML event in 1801, which was felt between Inverness and southern Scotland, and was reputed to have occasioned the collapse of part of a barn near Edinburgh, killing two labourers who were sleeping there and injuring some others.

23 October 1839 – after the 1801 earthquake activity at Comrie decreased to a low, intermittent level, and then at the beginning of October 1839, the shocks resumed, with a crescendo over the space of three weeks to a 4.8 ML event on 23 October – the largest Comrie event known. This was followed by extremely intense swarm activity until 1846, with two more shocks exceeding 4 ML in magnitude. The earthquakes were intensively investigated by two amateur seismologists in Comrie village, and a committee formed by the British Association for the Advancement of Science (Milne 1842-4, Musson 1993b).

After the decline of activity in 1846, seismicity in Comrie has remained at a low, intermittent level. An apparent resumption in the 1860s led to the reconvening of the BAAS committee, but this turned out to be a false alarm.

The Ochil Hills swarm activity has been quite different in character. The first earthquake we know of that was clearly in this location occurred on 30 April 1736. The main historical period of swarm activity for the Ochil Hills was between 1900-1916, studied by Charles Davison (1916, 1924) from macroseismic data (and it is a matter of speculation how many of the events he noted as swarm earthquakes were, in fact, quarry blasts). While one could argue that the Comrie activity is not true swarm activity because, for instance, in the 1788-1801 period one can distinguish the 7 September 1801 event as a clear mainshock, the same is not the case for the Ochils. The largest of the Ochil Hills earthquakes never reached 4 ML in magnitude; the larger events were between 3.1 ML and 3.7 ML.

A further brief swarm occurred between 1979 and 1980, the largest event being again, a mere 3.2 ML.

Davison (1924) believed that the Ochil Hills earthquakes occurred along the Ochil Fault that bounds the southern edge of the range of hills; but of course, this is also where the settlements are. Instrumental locations for the 1979 earthquakes show locations further north, in the hillsides around Glendevon. This is another example of the migration of macroseismic locations towards settled areas.

In 1940 two earthquakes (3.3 and 3.7 ML) took place near Stirling (the second was nearer to Kilsyth). Arguably, these could be regarded as related to the Ochils activity. Elsewhere in Central Scotland, small earthquakes have occurred under Glasgow from time to time, the most recent being in 1964.

3) Southern Scotland

Southern Scotland is an area of very low seismicity, with the exception of Galashiels, where events took place in 1650, 1728 and 1844, and the area around Dumfries and Lockerbie, where a number of earthquakes, all minor, have occurred over the years.

4) Northern England

The pattern of earthquake activity in Northern England is quite distinct. A narrow zone of relatively strong earthquake activity (up to 4.9 ML) runs along the Pennines, all the way from the Peak District up to the Scottish Border, where it stops abruptly. Two strong earthquakes have occurred at the Border, a relatively recent event on 26 December 1979 (4.7 ML), with epicentre near Longtown (north of Carlisle), seems to have resembled closely an earlier event in 1650 (Musson 2004b).

Wensleydale seems to have been repeatedly active in historical times: in 1768, 1780 and 1933. Other focal points have been Weardale, Kirby Stephen, Skipton, Todmorden, and the Peak District itself. Derbyshire is noted for a sequence of three events in 1903, 1904 and 1906, of which the first two were widely felt, and was also the probable location of strong shocks in 1575 and 1683. The pattern of seismicity around the southern end of the Pennines is quite diffuse. The record of modern seismicity is contaminated by mining-induced events; it can be hard to distinguish, especially for older events, what is mining-related seismicity and what is natural seismicity in a mining district. One assumes, for instance, a natural origin for the 17 March 1816 Mansfield earthquake (4.2 ML), but recent low-magnitude seismicity in the Mansfield area has been of a mining character. On the other side of the Peak District, the possibility of earthquakes within the Manchester area had been demonstrated by the 14 September 1777 event (4.4 ML), and a smaller event near Salford in 1931, but this was not really a precedent for what occurred in 2002, when a 3.9 ML event in the eastern part of the city was followed by over a hundred aftershocks, many of them felt.

West of the Pennines, the situation in the Lake District is unclear. The Lake District dome has been subject to a scattering of seismicity, mostly events less than 4 ML (the exception being the 4.1 ML earthquake in the Northern Lake District in 1901). A number of small earthquakes have occurred around the edges of the dome, but these have tended to be shallow, and therefore some of them have been rather strongly felt locally. The extreme case is the 15 February 1865 earthquake (Musson 1998c); this event can be accurately located close to the village of Rampside, just east of Barrow-in-Furness. This village was heavily damaged, with liquefaction effects (sand fountains) in the saturated sands on the foreshore. Yet 10 km away the earthquake was hardly perceptible. This illustrates the difficulty of determining earthquake magnitude from maximum intensity. The Barrow earthquake had a epicentral intensity of 8 EMS; so using maximum intensity as an analogue of magnitude would class this event as one of the largest British earthquakes; yet when the magnitude is calculated from felt area, it is less than 3 ML.

The key event for the Lake District is the earthquake that occurred on 11 August 1786; this had a magnitude of 5.0 ML, making it the largest Northern England event with the possible exception of the poorly-understood 26 February 1575 earthquake. The epicentre was offshore from Whitehaven, and is sufficiently close to the major Lake District Boundary Fault (LDBF) that marks the western edge of the Lake District to raise questions as to whether the LDBF can be considered “active”. As is so often the case, the LDBF is not a simple structure, but consists of a fault zone of anastamosing traces. A small earthquake (< 3 ML) on 17 November 1755 caused strong shaking at the village of Irton, south-west of Whitehaven; this event seems therefore to have been shallow, and occurred within the LDBF zone. Given a narrow definition of what constitutes an “active” fault, the 1755 event is perhaps more significant for the LDBF than the offshore 1786 event, even though the latter is much larger. However, whether such a definition is appropriate will be considered in the final section of the present paper.

In contrast, the north-east of England is an area of extremely low seismicity, from Northumberland down to York. The exception is an earthquake with estimated magnitude of 4.4 ML offshore from Whitby, which may be related to a belt of seismicity extending offshore from Flamborough Head, which will be touched on later.

5) Wales and the Marches

The seismicity of Wales has been discussed in detail elsewhere (Musson 2005b). There is a distinct division of Welsh seismicity into four areas: Snowdonia, the rest of North Wales, Central Wales, and South Wales. It is convenient to take the latter zone as extending across the English border and bending up north-eastwards through Herefordshire and Shropshire.

Snowdonia is one of the most concentrated active seismic areas in the UK. The largest known onshore British earthquake occurred here on 17 July 1984 (5.4 ML), with a well-located epicentre near Yr Eifel in the Lleyn Peninsula. Very similar earthquakes occurred on 9 November 1852 and 7 October 1690. The poorly-documented earthquake of July 1534 was probably similar, and it could be conjectured that the earthquake of 20 February 1247, most likely one of the strongest Medieval British earthquakes, also originated in this area. The 1852 and 1984 earthquakes were felt in all of Scotland, England, Ireland and Wales, which is unusual for a British earthquake. The 1690 earthquake may have been felt in all four, but there is no extant report from Scotland. 1247 was felt in all four. The 1852 and 1984 earthquakes, both very well documented, caused little damage, indicative of deep foci. The depth of the 1984 earthquake is instrumentally determined at 18 km (Turbitt et al 1985). The 1984 earthquake actually caused more damage to poorly-maintained chimneys in Liverpool than it did in the epicentral area.

In addition to these characteristic large earthquakes at intervals of around 150 years, lesser earthquakes also occur, including two that were over 4.5 ML in 1903 and 1940. The larger events seem to have occurred on the south-western side of the Snowdonian Massif, though all except 1984 are not very well located. More recent small-magnitude seismicity, on the other hand, is also observed on the north-west side of Snowdonia, and there is historical data for small earthquakes occurring on the north-east side. A point of interest concerning the 1940 earthquake is that it was the last British earthquake to date that caused fatalities, with a death toll of two (Musson 2003).

Away from Snowdonia, seismicity in the rest of North Wales tends to be diffuse and characterised by events not exceeding 4 ML in magnitude.

Central Wales is characterised by low seismicity. Some small earthquakes have occurred, especially around Newtown and Brecon, but the coasts of Cardigan Bay south of Harlech show very little seismic activity.

South Wales is another story. Here strong earthquakes (around 5 ML) are relatively common; in fact, disproportionately so. According to the Gutenberg-Richter relationship, as discussed previously, for every unit decrease in magnitude, there should be around a tenfold increase in frequency. This does not happen in South Wales; there are more recorded events around 5 ML in magnitude than around 4 ML. This is true over an area that runs not only along the South Wales coast, but curves up through Herefordshire and into Shropshire. Swansea seems to have been particularly affected, with notable earthquakes close to the city in 1727, 1775, 1832, 1868 and 1906. The first and last of these were the strongest: the 1906 earthquake had a magnitude of 5.2 ML and caused considerable damage to houses (mostly to chimneys and plaster). Since 1906, the only earthquake near Swansea has been a 3.2 ML in 1930.

In south-east Wales, there has been recent seismicity in the Newport area, especially in the 1970s, but south-west Wales appears as almost a complete blank on maps of modern instrumentally recorded earthquakes (i.e. since 1970). This is unusual, because for the most part the spatial distribution of the seismicity of the last thirty years matches very well the long-term pattern. In south-west Wales this correspondence breaks down, and one would not suspect from modern data that this part of the country was subject to strong earthquakes, which it has been in the past. Two key earthquakes are those of 18 August 1892 and 3 November 1893. The first (5.1 ML) had an epicentre near Pembroke, the second (5.0 ML) near Carmarthen. These two earthquakes make an interesting comparison. They were located quite close in both space and time, but the distribution of effects was totally different. The Pembroke earthquake was felt strongly to the south of the area of maximum effects, being widely perceived in Cornwall and Devon, but was little felt in North Wales. The Carmarthen earthquake was the complete reverse of this (Musson et al 1984). All else being equal, it appears that this can only be explained by differences in the source effects on the radiation pattern.

Across the border from eastern Wales into England, the same pattern can be seen, with magnitude values of 5.2, 5.3, 4.8 and 5.1 ML for earthquakes in 1863, 1896, 1926 and 1990 in Hereford and Shropshire, and a lack of events in the 4.0-4.5 ML range.

6) English Midlands and East of England

There are no very clear divisions between the seismicity of the Welsh Marches to the west, the Peak District/South Yorkshire to the north, and the English Midlands. For the most part, central and south-east England are areas that are not very active seismically, but there is not a sharp transition between high and low seismicity, and there are patches of notable activity in what might be considered relatively stable areas.

Thus, for example, even when mining-related seismicity is removed from the Staffordshire area, one can still see some significant earthquakes 4.5 < ML < 5.0, notably the 1916 Stafford earthquake and the 2002 Dudley earthquake.

Particularly prominent is a concentration of seismicity around Leicester, where nine earthquakes > 3 ML have occurred in the last 250 years. The largest of these took place on 9 February 1953; this had a magnitude of 5.3 ML and caused damage in the Derby-Loughborough area, including damage to Blackbrook reservoir (making it one of three British earthquakes to have caused damage to dams, the other two being the 1839 Comrie and 1979 Ochil Hills earthquakes). A child in Derby suffered a fractured skull.

East of this area, one can make out a line of epicentres (which may or may not be a coincidence) of sporadic events as far as the vicinity of Norwich, where a magnitude 4.0 ML event occurred in 1994. An earthquake in 1480 caused damage in Norwich, but little else is know about this earthquake.

In Lincolnshire there is little seismicity besides a 4.2 ML event on 1 August 1755, but we know that Lincoln cathedral was badly damaged by an earthquake in 1185. Was this a strong local earthquake with epicentre similar to that of the 1755 event, or was it an offshore earthquake similar to the 1931 North Sea event, as first proposed by Davison (1931)? The latter explanation seems inherently more likely, since the largest British earthquakes for which magnitudes can be estimated are all offshore events. Possible evidence for a strong local Lincoln event in 1185 can be adduced from folklore relating to destroyed villages in Lincolnshire and Nottinghamshire, if any weight can be put on this. Archaeoseismic investigations in Lincolnshire would be interesting.

Essex is a county completely devoid of earthquakes except for one – which ironically was the most damaging British earthquake in the last 400 years. The infamous Colchester earthquake of 22 April 1884 was not especially large (4.6 ML) but it was shallow (3 km) and occurred below a populated area. The epicentre was near the village of Peldon, south of Colchester, and in Peldon and the surrounding villages and farmhouses, damage was quite severe, amounting to an intensity of 8 EMS. There exist numerous photographs and drawings of the damage; however, some of the former were faked to make the damage appear worse than it was (Musson 1990). A few people were hurt by falling stones, but the only fatalities were an old woman in Wivenhoe, on her death bed, whose "end was hastened by the shock", and a woman in Manningtree who was so distressed by the whole occurrence she threw herself in the River Stour a few days later (Musson et al 1990, Musson 2003). A lurid book-length account of this earthquake by Haining (1976) is quite unreliable. (For example, Haining relates how in London the earthquake toppled a lamp in Bow Street Magistrates Court, causing a fire. In fact, the fire in Bow Street occurred after the earthquake, was caused by a servant girl upsetting a candle in an inn, and did not affect the Magistrates Court.) One conclusion to be drawn from the 1884 earthquake is that, even in the most aseismic places, an earthquake at least up to 4.5 ML can occur, and can cause damage if it is shallow enough – even if this is a rare event.

The rest of the Home Counties have little seismicity of note, though one should mention two small earthquakes in 1750 (8 February, 2.6 ML, and 8 March, 3.1 ML) with epicentres in Central London (the first near Leadenhall Street, the second near Lambeth, within an accuracy of a few kilometres). Some minor damage was caused. It is interesting to speculate on the consequences of a recurrence of a similar event today in the same place, and whether a larger event could occur. By analogy with Colchester 1884, one cannot rule out something like a 4.5 ML at the very least.

In addition, the Chichester area seems to be another location for possible swarm-like activity; a small swarm occurred in the 1830s, and there was some seismicity in this area in 1963 and the early 1970s.

7) South-west England

The south-west of England is characterised by seismicity that never seems to rise much above 4 ML in magnitude. The most active areas are western Cornwall, especially around Penzance and Helston, the Launceston-Tavistock area, and the north coast of Cornwall extending up into North Devon. The seismicity of Cornwall and Devon has been studied exhaustively by Musson (1989c), and the seismicity of the peninsula is also discussed in Musson (2000b).

While Somerset has seen a few small earthquakes, most recently a 3.1 ML double event near Bridgwater in 2004, Dorset and Wiltshire are very inactive.

As mentioned previously, one must be careful with discussions of the 11 September 1275 event. Some popular publications attribute this to a Cornish source, on the grounds that it destroyed Mount St Michael (near Penzance). In fact, what was destroyed was St Michael’s on the Mount, which is the name of the church on Glastonbury Tor. But putting an epicentre in Somerset (as in Ambraseys and Melville 1983) is unlikely to be correct. Damage to one very anomalous construction perched on a steep hill is not indicative that the earthquake source nearby. Musson (1994) suggests a possible epicentre near Chichester on the basis of what is stated concerning places where the earthquake was felt. The annals of Osney (Luard 1868-9) state that people were killed by this earthquake – this is the only contemporaneous record that refers to deaths in a medieval British earthquake.

8) Offshore seismicity

Seismicity offshore is also strongly localised. The area of the Viking Graben and South Møre Basin, in the northern North Sea, is perhaps the most active seismic area in Europe north of the Alps. The question is sometimes asked whether the strong seismicity here has anything to do with hydrocarbon extraction, but it is clear that seismicity in the Northern North Sea long preceded oil extraction. An earthquake on 4 January 1879 can be identified as having occurred in this area because it was felt on both sides of the North Sea, in Norway and Shetland. With other reports from the same period for which there are only reports from one coast, one cannot be sure if these are significant earthquakes some distance offshore or very minor ones close to land.

The earthquake of 24 January 1927 (5.7 ML) was the second largest British earthquake of the 20th Century. It was felt over most of Scotland and down the east coast of England as far as Norfolk, and in Western Norway. It caused no damage to buildings (Musson et al 1986).

South of the Viking Graben, the Central Graben system of the North Sea appeared inactive until recently; but until recently seismic monitoring was insufficient to detect any but the largest events occurring so far from land. Here there is rather more evidence that at least some seismic activity may be induced (Ottemöller et al 2005).

There is a marked belt of seismic activity off the east coast of England from Flamborough Head, stretching south-eastwards down to a latitude of 53^{o} N. The most notable earthquake here was the 7 June 1931 event; with a magnitude of 6.1 ML it is the largest British earthquake for which magnitude can be estimated. It was felt over a very extensive area: more or less the whole of Great Britain, eastern Ireland, and parts of northern France, the Low Countries, northern Germany, Denmark and western Scandinavia (Versey 1939, Neilson et al 1986). Damage occurred along the east coast of England, but was minor in character. One woman died in Hull from a heart attack.

Two of the strongest British earthquakes, with magnitudes between 5.5 and 6.0 ML, had epicentres in the Dover Straits. The dates of these two events were 21 May 1382 and 6 April 1580, and they seem to have been very similar in all respects. The latter of the two is, of course, the better documented; it was felt strongly in London, where there was some damage, and two children were killed in Christ Church Newgate. Repairs to the damage to the tower of St Peter's Broadstairs are still visible (Figure 10). Effects on the French side of the Straits were as bad, if not worse, with damage at Calais, Boulogne and elsewhere. An unknown number were killed on the continent. Contrary to some reports (e.g. Varley 1996) the earthquake did not cause a tsunami, although the earthquake shaking was enough to disturb the water in Kentish harbours. Reports of coastal inundation and ships sinking probably relate to the effects of a storm the following year (Melville et al 1996). A lesser earthquake, but still with magnitude probably > 5 ML, occurred in this same area in 1449 (Musson 1994, Melville et al 1996). Since 1580, there have been only two much smaller earthquakes in this area, in 1776 and 1950 (4.1 and 4.4 ML respectively). Evidently, if the 1382 earthquake can repeat itself in 1580, it must be possible that it could repeat itself again in the future; given the increase in urban development in south-east England and the Pas de Calais, one must consider that the consequences of such a repeat event could be considerable.

Moving westwards, there has been a certain amount of seismicity in the English Channel, including historical events up to 5.1 ML in magnitude, but the greatest concentration has been in the Channel Islands, and close to France. Since the 1980s, a cluster of events has been observed west of the Scilly Isles. Although one must wonder whether these are real earthquakes or man-made events (for example, munitions disposal), it seems most likely that these are real earthquakes (Walker 1994, pers. comm.)

The Irish Sea is not very seismically active; however, one strong earthquake (5.1 ML) did occur in 1843 with an epicentre between the Isle of Man and Morecambe Bay. Otherwise the only event of note is an earthquake in 1951 just off the coast of County Wicklow. This was felt over a wider area than its instrumental magnitude of 3.7 ML would suggest, giving it a macroseismic magnitude of 4.4 ML. It is the strongest Irish earthquake on record, indeed, practically the only Irish earthquake of any significance. The aseismicity of Ireland is very notable, and not due to lack of data. On 14 December 2005 a 2.8 ML earthquake in the Irish Sea was felt in Wicklow, with a not dissimilar epicentre to the 1951 event (Bukits 2006) – the distance between the two epicentres is 27 km, but the 2005 event is much more accurately located.

For the most part, the seas around Scotland west of the Viking Graben have almost no seismicity. There is palaeoseismic evidence for large earthquakes on the continental slope north and west of the UK (Musson 2006), and ambiguous (and weak) evidence for an anomalously large event having occurred in historical times (Musson 2004b, 2006), analogous to the 1927 Grand Banks (Newfoundland) earthquake, which was a case of a large (>7 Mw) earthquake on a passive margin in an area otherwise aseismic.

Seismicity of Britain in relation to geological structure

The quest for an understanding of how the distribution of seismicity in the UK relates to geological structure has been a long and unfruitful one, starting with the various studies by Davison (1924), who was the first British seismologist to understand the relationship of earthquakes to faulting, after formative studies by, for instance, Reid (1910); although it was realised by David Milne as early as the 1840s (Milne 1842-4) that a connection existed between earthquakes and faults. The British Isles have a complex geological and tectonic structure as a result of a complex geological history involving several orogenic phases. If one looks first at the situation in terms of gross crustal structure, one can divide Britain into a number of geological provinces or terranes. These are shown in Figure 11, adapted from Chadwick et al (1996) after Whittaker et al (1989). It is immediately apparent that there is no correlation between terranes and seismicity. Although one could argue that these crustal blocks are roughly homogeneous in terms of crustal properties, including styles and distribution of faulting, this is not reflected in any homogeneity of seismicity. Some other factor, or factors, is affecting the distribution of seismicity.

The problem can be summarised neatly with one question: why is Ireland so aseismic? The absence of Irish epicentres from maps of the seismicity of the British Isles is nothing to do with absence of data; documentary sources for Irish seismicity are at least as good as those for, say, Wales, and modern instrumental monitoring also demonstrates a remarkable lack of seismic activity in Ireland. Indeed, even a source as ancient as Ware (1662) remarks on the absence of earthquakes in Ireland. Yet geologically, many structures are common to both Ireland and Britain, and there is no obvious discontinuity that could explain why the seismicity stops at the Irish Sea.

In the last decades, therefore, there has been a search for some factor X that could be considered as the control on the distribution of seismicity. This is far from being a merely academic pursuit. Consider the situation that would arise if it were planned to build some sensitive facility at Stonehaven, close to where the Highland Boundary Fault (HBF) intersects the east coast of Scotland. If one supposes that the high seismicity that has been concentrated around Comrie in historical times is due to the HBF in some directly causative way, it seems perfectly likely that similar seismicity in the future could occur anywhere along the HBF, including at Stonehaven. However, if the Comrie seismicity is due to some other factor unique to Comrie, future swarms should recur at Comrie and not at Stonehaven. The difference in terms of hazard is considerable, and the implications in terms of required antiseismic design could amount to millions of pounds.

The general difficulty with all work on this subject is that it has been of the order of looking for correlations between seismicity and some other factor, with the success of the hypothesis “the controlling factor for British seismicity is X” being judged by how well X shows a spatial correlation with a map of epicentres. No factor shows a perfect correlation; and if any did, it would have no predictive power. If one could identify some factor X such that wherever X occurred there were earthquakes, and wherever X was lacking there were no earthquakes, there would be no places where one might infer that earthquakes could occur in the future even though they have not done so in the past (c.f. Stonehaven in the preceding illustration). In such a case, one might simply say that earthquakes occur where they have occurred. What one needs ideally is a mechanistic theory, the success of which can be judged other than by spatial correlation.

Some proposals can now be examined, concentrating on modern studies.

Palaeogene-Neogene deformation

A study by Muir Wood (1989) of Eocene and post-Eocene tectonics in Western Europe identified four phases when intraplate deformation was concentrated along linear zones stretching from the Mediterranean to the mid-Atlantic ridge; Muir Wood (1989) considers these zones to have acted as sub-plate boundaries within the Eurasian plate. These are shown in Figure 12. Each resulted in significant deformation in the British Isles in the following four phases:

1. Middle-late Eocene: southern and south-western England
2. Early Oligocene: eastern and northern England and most of Scotland
3. Late Oligocene: southern England, western Wales, eastern Ireland, Hebrides
4. Late Miocene: all of Britain except eastern England and eastern Scotland, all of Ireland except Munster

It is easy to pick out features that appear to be convenient as explanatory variables. One could argue that areas subject to active deformation in the late Miocene continue to be active, and this explains nicely why eastern Scotland and north-eastern England are so inactive. But it fails to explain the absence of seismicity in Ireland or south-west Scotland. Muir Wood (1989) writes that “The large late Miocene/Pliocene inversion on the north-western edge of Wales is still a nest of seismicity”. However, a comparable inversion developed at the same time on the southern margin of Lough Neagh, and this shows no signs of continued seismicity at the present day. One has to suppose either that the absence of seismicity in Northern Ireland is a temporary phenomenon, or that the presence of late Miocene inversion structures is not a good predictor of modern seismicity in the UK. There is no strong mechanical explanation for Miocene deformation being a controlling factor in modern seismicity (though it seems reasonable to suppose that areas that were weakened at that time may continue to be weak), and ultimately this is an explanation that “works only when it works”.

In one of the first studies to attempt a seismic source model for the whole of Britain for the purposes of computing seismic hazard, Ove Arup (1993) used Muir Wood’s (1989) outline to define three categories: areas excluded from all four deformation phases, areas included in one or two phases, and areas included in more than two phases or included in the final phase. It is shown in Musson (2000a) that this model does not predict contemporary seismicity very accurately.

To examine the effects of multiple phases of activation, Figure 13 is provided. Heavy shading covers the areas present in three out of four phases, lighter shading indicates two out of four (obviously, some allowance should be made for the fact that none of these boundaries are in actuality sharp or well-defined). Nowhere scores four out of four, since the phases 1 and 2 do not overlap. The most activated areas are Cornwall, south Devon-Dorset, Sussex and Skye-Lewis. These are far from being notable for relatively high seismicity today.

In fact, there is in general the same problem with combinations of explanatory variables as there is with single ones. To say that seismicity occurs where factor X is present and also factor Y suffers from the same problem that inevitably one finds some places where X and Y are present but there is apparently no seismicity, and one cannot demonstrate whether this is a temporary exception or a piece of counter-evidence.

Glacial rebound

The idea that isostatic recovery following deglaciation might trigger seismicity goes back as far as Hobbs (1927) in North America, and Kolderup (1930) proposed this as an explanation for seismicity in the northern North Sea. It was invoked by Versey (1939) as an explanation for the 7 June 1931 Dogger Bank earthquake, but on no good evidence.

It was noted by Musson (1996b) that there is a remarkable correlation between the distribution of seismicity in Scotland and the distribution of ice cover during the last glacial advance, the Loch Lomond interstadial, shown in Figure 14 (Dawson 1992, Bowen et al 2002, Clark et al 2004). Since focal mechanisms in the area show predominantly strike-slip faulting, indicating that the principal stress direction is horizontal, it cannot be argued that isostatic rebound is the driving force behind Scottish seismicity, but one could hypothesise that the present distribution of seismicity in Scotland is a shadow of a former period of enhanced seismicity immediately following deglaciation, when isostatic recovery was much more rapid than it is now. Under this hypothesis, one could argue that structures that were reactivated after deglaciation have continued active to the present day, where oriented favourably with respect to the present stress regime. But while this might tentatively explain why seismicity in Scotland is localised chiefly to the west coast between Dunoon and Ullapool, it does nothing to explain the presence or absence of seismicity anywhere else in the UK.

Muir Wood (2000) extends the concept of deglaciation seismotectonics considerably by proposing a pattern of stress interference between radial strain fields due to post-glacial rebound and forebulge collapse on the one hand, and the general tectonic maximum horizontal stress on the other. He proposes a pattern of quadrants as shown in Figure 15. The inner zone is characterised by radial extensive strain associated with a rising rebound dome, and the outer ring characterised by radial compressive strain associated with a sinking forebulge. In the plane of the maximum horizontal stress direction, owing to the effect of the stress interference on the Mohr-Coulomb failure criterion, one expects high seismicity in the outer sector and low seismicity in the inner sector. Normal to the maximum horizontal stress direction the pattern is reversed.

Figure 16 shows the effect of this in the UK; the zones are adapted from Figure 4 in Muir Wood (2000) and overlain on UK seismicity above 4 ML. While the theory is attractively ingenious, as Figure 16 shows, it does not match events very well. In the first instance, it is not clear why the forebulge for a rather small ice sheet should be so much larger than the rebound dome, unless it has just been drawn that way to take in all the seismicity of England. Secondly, and rather obviously, what should be a seismic forebulge to the north-west of Scotland is completely inactive, as Muir Wood (2000) also notes. Since 1995, seismic instrumentation in the extreme north and north-west of Scotland has been greatly upgraded, and since 1999 a network has been operating in the Faroes, and not even microseismicity has been detected in this very quiet area. Thirdly, the aseismic and seismic sectors of the rebound dome in Scotland don’t really match the actual distribution of epicentres very well, even allowing for the fact that the boundaries in Figure 16 can only be approximate.

Major fault systems

In a wide-ranging study of seismotectonics in the UK, Chadwick et al (1996) examined spatial correlations of a number of geological and geophysical variables (gravity anomalies, magnetic anomalies, heat flow, depth to Moho, etc.) with the distribution of British seismicity, in order to look for spatial correlations. In general the results were negative. However, some interesting findings did emerge, notably that there seemed to be a tendency for the larger British earthquakes to occur in the footwall blocks of major fault systems.

It has often been the case, since Davison (1924) and earlier papers by Davison referenced therein, that attempts have been made to attribute British earthquakes to major mapped fault structures. However, the fact that an epicentre falls on the surface trace of a fault does not have the significance that Davison placed on it when one realises that the fault is dipping, and the earthquake focus is some 15 km below the surface, well away from the actual fault plane. This means that if this co-location is significant, it cannot be in terms of a simple thrust-fault reactivation model on major dipping structures (these events also have, where they can be determined, fault plane solutions indicating strike-slip faulting on near-vertical faults).

If there is a significance, it must be in terms of these major fault systems acting as general crustal weaknesses. Thus one can conjecture that a minor north-south strike-slip fault is more likely to be reactivated by the regional stress field if it is situated in a zone of crustal weakness near a major fault zone than if it occurs in stronger crust away from such zones. Thus, returning to the discussion of faulting in connection with the Inverness earthquakes, whichever fault was responsible, it is arguable that crustal weakness due to the presence of the Great Glen Fault zone contributed to the reactivation. This idea is supported by the fact that there is a discernible tendency for earthquakes to cluster around fault intersections, such as the Iapetus Thrust with the Pennine Fault, and the Variscan Front Thrust with north-west-trending transcurrent faults (Chadwick et al 1996). Similarly, there are cases where earthquakes seem to cluster between converging faults: Moine Thrust and Great Glen Fault, Moine Thrust and HBF, Church Stretton Fault and Variscan Front (Chadwick et al 1996).

In Figure 17, adapted from Chadwick et al (1996), the proposed corridors around major crustal faults are shown by the stippled area (not fully extended either west or east of Britain). There is a similar problem to what has been seen before; there are cases where the seismicity does seem to be associated with these zones of weakness, but many places where the crustal faults have no seismicity associated with them. Therefore the presence of major crustal fault structures does not seem to be a good predictor on its own, although it may be a contributive factor.

It may be worth remarking that overlaying Figures 13 and 17 is no more illuminating; the problem is that all tectonic hypotheses based either on Palaeogene deformation or major crustal structures imply that there should be significant seismicity in eastern Ireland and the Outer Hebrides, which there simply isn’t.

A further finding of Chadwick et al (1996) was that a significant positive correlation appears to exist between seismicity and areas of Precambrian and Palaeozoic faults, and a corresponding negative correlation between seismicity and areas of Mesozoic extensional faulting (except in the northern North Sea). This may be due to the fact that reactivation of the moderate to steeply-dipping normal faults of the Carboniferous and Mesozoic basins would require a regime of basin inversion, which does not hold for the UK at the present (Chadwick et al 1996).

Mantle processes

A recent approach of interest derives from new data that have become available from seismic tomographic studies of the upper mantle. It is known that substantial uplift of Britain and parts of the adjacent regions took place during the Cenozoic (e.g. Brodie and White 1994). Recent studies based on seismic tomography show an anomalously hot, low-density region in the underlying lithosphere and asthenosphere beneath Britain, down to at least 200 km (Goes et al 2000). It is suggested by Bott and Bott (2004) that this low-density upwelling is the cause of Cenozoic uplift, in contrast to previous theories involving underplating (Cox 1980).

The approximate outline of the hot upper-mantle region under Britain is given by Goes et al (2000) from both P and S-wave tomography, the former giving a broad zone about 200 km wide running from Cornwall to the Solway; the latter a larger area including western Scotland and much of the English Midlands. Bott and Bott (2004) remark that the combined anomaly zones cover most of the seismicity of mainland Britain, a good result considering that the results of Goes et al (2000) are not well resolved. Bott and Bott (2004) therefore propose that the distribution of seismicity is linked to the presence of thermally weakened crust and the associated uplift. Further, they suggest that, given the lack of resolution in Goes et al (2000), the distribution of seismicity is actually a better indicator of the mantle anomaly than the seismic tomography data (Figure 18). This is not an unreasonable position to take, but it reduces again to saying that the predictor variable for seismicity is where earthquakes happen to have occurred.

Since Bott and Bott (2004), a further paper has appeared by Arrowsmith et al (2005), which presents similar tomographic data of the hot upwelling beneath Britain, but with higher resolution than in Goes et al (2000). This reveals much more structure to the mantle anomaly, expressed as P-wave velocity anomalies, which can now be seen to extend down as far as 300 km in northern Scotland. Unfortunately it does not confirm Bott and Bott's (2004) speculation that the distribution of seismicity is a good guide to the detailed structure of the anomaly. Arrowsmith's et al (2005) results show in particular that the anomaly extends further to the north and west than shown by Goes et al (2000) and to be less pronounced in the south, and lacking in much of the English Midlands. In fact, they find a good correlation (not entirely unexpected) between the P-wave velocity anomaly and high gravity anomalies. A simplified plot of the low P-wave velocity anomaly is shown in Figure 19. Arrowsmith et al (2005) interpret the feature as being related to an arm of plume material extending from Iceland, which, given the increased strength of the anomaly north-west of Scotland, seems plausible.

Arrowsmith et al (2005) also assert that there is evidence that the distribution of earthquakes in Britain is related to the anomaly, specifically, that earthquakes are concentrated around the edges of the anomaly. One might find some support for this in Figure 19 – except that given the complex shape of the low P-wave velocity area, the amount of the Britain that is approximately "close" to the edges of the anomaly is rather large. One can also plainly see edges of the anomaly with no seismicity. So as a predictor, this mantle anomaly is not very useful. One can also remark back to the good agreement of the anomaly with gravity data, and recall that gravity anomalies were already studied by Chadwick et al (1996) and found not to give very good results with respect to predicting the distribution of seismicity.

Geology or geometry?

One possible reason for the lack of success in finding some geophysical or geological parameter that can be interpreted as controlling the distribution of seismicity in Britain is that the principal control may not be a crustal property at all. Instead, it might be necessary to consider the interaction of crustal units from a geometric perspective.

The following analogy, while slightly whimsical, and certainly not exact, may be useful. In 1965 a game called Booby Trap was marketed by Parker Bros in North America and Waddingtons in the UK. This consisted of a shallow wooden tray, which could be traversed by a wooden bar that extended the width of the tray. This bar was controlled by a strong spring between the bar and one end of the tray. A rod could be used to draw back the bar, compressing the spring; if the rod was released, the bar would be propelled by the spring back down the tray. The playing pieces consisted of a large number of small plastic cylinders of different diameters, with small handles attached. At start of play, the bar was pulled back as far as possible, and the tray was filled with the playing pieces, shuffled around randomly. When the bar was released, the playing pieces were compacted against the far end of the tray, and held under compression by the force of the spring.

Play consisted of attempting to remove pieces from the tray without causing the bar to move. Players scored more points according to the size of the piece removed, and were penalised if the piece they removed caused a readjustment of the remaining pieces and the movement of the bar. Although to a cursory inspection it would appear that all the pieces were closely wedged together, in fact some of the pieces would form a stable network (stress arches) that took all the stress, the remaining pieces being merely infill. The skill of the game lay in recognising which pieces were "load-bearing" and which were "infill", and removing only the latter.

This is not by any means a close analogy of intraplate seismicity, but there are some points in common. The playing tray is in effect a horizontally compressive stress regime, containing a number of different physical units (the playing pieces). Similarly, the British Isles is subject to a compressive stress regime from mid-Atlantic spreading, and is composed of a number of different geological units of different rheological strengths. The overall physical properties of the assemblage of playing pieces in the tray of the Booby Trap game can be considered as entirely homogeneous, inasmuch as the tray is a regular shape, the base of it is evenly smooth, and so on. Yet in play, marked heterogeneities could develop such that quite large areas could turn out to be "infill" – pieces that were not actually under significant strain, since the bar was entirely restrained by other pieces. These areas acted rather like aseismic regions, since pieces could be removed safely without causing the "earthquake" of the bar moving. What determined the location of these areas was entirely the geometric arrangement of the pieces. The chains of "strong" pieces formed because of geometrical interactions between them, not because of any property of the pieces, or of the gaps (faults) between them. Issues such as differences in fault strengths are obviously omitted from this analogy, which should not be stretched too far – its purpose is only to demonstrate the influence of geometric patterns on behaviour under loading.

One can imagine that similar interactions could occur in an intraplate area like the UK. While a primary division of the British Isles can be made into different terranes as shown in Figure 11, there are other distinct blocks that are likely to function as units that "jostle", to use a term from Chadwick et al (1996). One can cite as an example the structure of the Pennines, where cohesive blocks like the Alston and Askrigg Blocks form the backbone of the upland ridge. It is rather notable that in Southern Scotland, which has very little seismicity, repeated small earthquakes occur in a restricted area which happens to be directly north of where the Pennine Chain meets the Iapetus suture. It seems probable that these are due to a stress concentration where the Pennines are acting as a rather scaled-down rigid indenter impacting on the Southern Uplands.

The concept of a rigid indenter might also be invoked to explain the concentration of seismicity at the northern apex of the Midlands Microcraton, and possibly elsewhere. For instance, the 2002 Dudley earthquake (4.7 ML) occurred as a strike-slip event (presumably left-lateral) on a roughly north-south fault on the western margin of the Midlands Microcraton (Baptie et al 2005), consistent with a slight northward movement of the Midlands Microcraton relative to adjacent blocks.

The Pennines and the Midlands Microcraton are examples where such geometric interactions of blocks are relatively straightforward to recognise, but the same principle may be at work elsewhere, and building up a complete pattern may be an impossible task, since one lacks complete information. A first attempt at this was made by Chadwick et al (1996), and the resulting model is reproduced here as Figure 20. In this, incipient relative block kinematics are shown qualitatively by arrows of different size; areas of relatively higher strain rates are shaded. Chadwick et al (1996) stress that these movements should be thought of as slight rotational readjustments involving only very small movements – not systematic displacements. This is not active deformation.

The predicted areas of high strain rate in Figure 20 do all coincide with seismic activity, but one would not predict from Figure 20 alone that north-west Wales should be one of the most seismically active parts of Britain. While it seems probable that this seismic activity is associated with the Lleyn Shear Zone, it appears, at least from earthquakes since 1800, that seismicity in north-west Wales is concentrated on the flanks of the Snowdonia Massif, and one can speculate that some sort of interaction between these two features is occurring. A contrary hypothesis would be that previous large earthquakes associated with a north-west Wales source, but which cannot be located accurately because of poor macroseismic data, were actually strung out along the Lleyn Shear Zone: this would include the earthquakes of 1690 and 1534, and perhaps 1247 also. Since there is no possibility of ever being able to locate these past events more accurately, it seems impossible to resolve the question other than by waiting a few hundred years for the next large earthquake in north-west Wales; however, it is worth noting that low-magnitude seismicity does appear to be concentrated on the west and north sides of Snowdonia.

This model offers a potential explanation of the lack of seismicity in Ireland. Ireland being near to the passive margin, it could be argued speculatively that there is less reason for internal readjustments between crustal blocks in responsive to the maximum compressive stress direction; either due to there being less lateral confinement, or because, in effect, Ireland is in the strain shadow of Great Britain.

Active faults in the UK

Some mention has already been given to cases where earthquakes in the UK have been attributed to particular faults, and also the relationship between seismicity and major shear zones. However, the question of determining “active faults” in the UK is contentious and needs separate discussion.

As mentioned, the first attempt to link earthquakes in the UK by naming specific faults as causative features, following the publication of elastic rebound theory by Reid (1910), was the work of Davison (1924). Davison’s identifications are generally not very well thought out, since he neglects to take into account depth of focus and 3D fault geometry, and he has a tendency to conclude that any major fault vaguely in the vicinity of an earthquake epicentre must be the causative feature.

This set a pattern which one encounters frequently in the literature, especially the popular literature, where one finds assertions that such-and-such an earthquake was caused by such-and-such a fault, on no more evidence than that the fault is a mapped structure somewhere near the epicentre. No account is made of the uncertainty of the epicentral location, the depth of focus, and the possible presence of other faults, perhaps with no surface expression, within the volume defined by the uncertainty in the location of the earthquake. Sometimes these associations lead to mislocations of many tens of kilometres, as in the case of the earthquake of 16 August 1934, already discussed.

There are very few historical earthquakes that one can speculate with any sort of confidence about associations with named, mapped faults. One of these is the aforementioned 1865 Barrow-in-Furness earthquake, which was extremely shallow, and can be very accurately located because of detailed macroseismic data; it is highly likely that it was produced by the Yarlside Fault, near the village of Rampside (Musson 1998c).

The prospect for associating earthquakes with specific faults becomes much better after the introduction of modern instrumental monitoring, as it is then possible to have focal mechanisms as additional evidence. The first use of this was made by Assumpcao (1981), who used focal mechanism data to associate the Kintail earthquakes of 1974 with the Strathconon Fault. Even after 1970, few such assignments could be proposed. It is stated by Blenkinsop et al (1986) that the 1984 Felindre earthquake occurred on a specific fault in the Felindre Basin, south-east of Newtown. Redmayne and Musson (1986) suggest that the Loch Long Fault was responsible for the Ardentinny earthquake of 1985. However, in recent years, improvements in accuracy of location have greatly improved the possible resolution of causative fault structures; the 2002 Dudley earthquake can be assigned with some confidence to one of the bounding faults of the South Staffordshire coalfield (Baptie et al 2005), and the fault responsible for the Manchester swarm of the same year can be imaged, even though it does not correspond to a mapped structure (Baptie and Ottemöller 2004).

But are these active faults, as the term is generally understood? The widely used definition of USEPA (1981) is that any fault that has produced an earthquake in the last 10,000 years is an active fault. This is a profoundly unhelpful definition. How many active faults are there in the UK? From the previous paragraphs one might answer that very few can be identified. But the question is not “how many can be identified?” but “how many are there?”. One can approximate an answer to this question by looking at a complete seismicity map of the UK (e.g. Figure 9), counting the epicentres and making some allowance for some events sharing the same source: the answer is at least several hundred. Every earthquake in the catalogue occurred on some fault or other; the fact that in most cases the causative fault cannot be identified does not stop it existing, or being classed as an active fault according to the USEPA definition (Musson 2005c).

This is very dangerous when it comes to seismic hazard assessment. It is often assumed to be a requirement in seismic hazard that any active fault requires special treatment in a seismic source model. Thus, typically, any fault that can be identified as active according to the USEPA definition is individually modelled, as in Frankel et al (1996). This gives rise to the following situation: consider two similar earthquakes, one of which can be associated with the trace of a known fault, and the other cannot. Common practice would be to treat the first as occurring on an active fault, and the other as part of background seismicity. The result would be a concentration of hazard near the first earthquake, while the second event is blurred away over a wide area. This is completely unrealistic, as there is no physical difference between the two earthquakes or between the two faults that caused them. The only difference is that the first fault is mapped and the second one isn’t. But “being mapped” is not a fault property that affects its capacity for seismogenesis or its potential contribution to hazard. Any practice that results in the “mapped-ness” of a fault having a large impact on the hazard results is rather unsatisfactory. But this is what follows from the USEPA definition. One has a choice: one can model only the known active faults, which leads to an unreal dichotomy, or one can attempt to invent parameters for every unknown fault for every earthquake in the catalogue, which is obviously a very unsatisfactory approach.

Much better is to scrap the idea of active faults altogether, and consider instead “controlling faults”, defined as those faults that are the main expressions of continuing deformational activity, e.g. graben-bounding faults in areas of present extensional tectonics (Musson 2005c).

It is now possible to solve the problem by stating that there are no controlling faults in the UK, since there is no continuing active deformation. The reactivation of faults is a different process. If we assume, for instance, that the Loch Long Fault was responsible for the 1985 Ardentinny earthquake, this does not imply any necessity for the Loch Long Fault to continue to produce earthquakes in the future; it may do so, but equally, activity may shift onto neighbouring faults. There is no need for the Loch Long Fault to continue seismic, unlike, say, the North Anatolian Fault which will remain seismically active for millions of years because of its role in interplate motion.

Thus to return to, for example, the Lleyn Shear Zone – there is no evidence that the dipping fault plane of this feature is currently active in any sense of the word. But, as a result of blocks “jostling” or some other reason, faults in the footwall block are being reactivated. Not just one fault, but evidently several. None of these faults individually are controlling faults, but as an assemblage (of an unknown number of faults in total) they are expressive of a process of readjustment or stress relief along the line of the Lleyn Shear Zone.

The same is probably true of other major fault zones such as the Great Glen and Highland Boundary Faults. Thus it would be incorrect to state that the Great Glen Fault is an "active fault", but it does appear to act as the locus for an assemblage of faults productive of minor earthquakes, none of which are of uniquely important status.

Conclusions

The seismicity of the UK has often been classified as low-to-moderate. It is not so high that people generally consider British earthquakes to be a threat, but it is sufficient that earthquakes need to be considered in the design of sensitive facilities such as nuclear power plants. Indeed, it was the British nuclear industry that supported a vast improvement in the understanding of British seismicity during the 1980s, including both the investigation of historical seismicity and improvements in instrumental monitoring, and the British nuclear industry continues to support seismology in the UK through the ODPM-led Customer Group that funds the UK seismic network.

As a result of this investment, and the various studies the history of which is recounted in Musson (2004a), we know as much about the past earthquake record of the British Isles as we are ever likely to. Despite this, the controlling factors for the distribution of British seismicity remain obscure. The search for simple geological or geophysical parameters to explain the spatial distribution of epicentres has not been successful, if regarded impartially. The British Isles is a classic example of intraplate seismicity, and it is likely that a complex pattern of fault reactivation related to the 3D geometry of strong and weak crustal blocks and their kinematic interaction is the true key to British seismicity.

Acknowledgements

I would like to express my thanks to all my colleagues in BGS with whom I have had the pleasure of discussing these issues over the years, and especially to Andy Chadwick. The monitoring and study of earthquakes in the UK is supported by the Customer Group led by the Office of the Deputy Prime Minister. This paper was supported by the Natural Environment Research Council, and is published with the permission of the Executive Director of the British Geological Survey (NERC).

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Figures

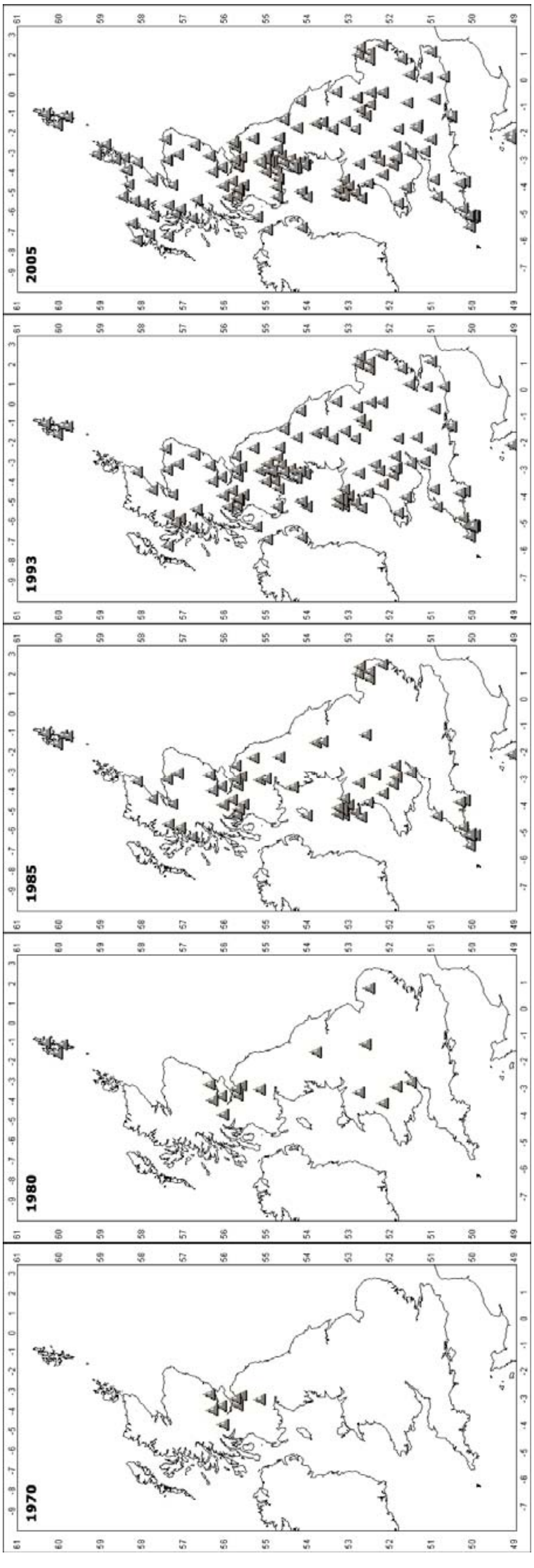

Figure 1 - Stages of development of the UK monitoring network

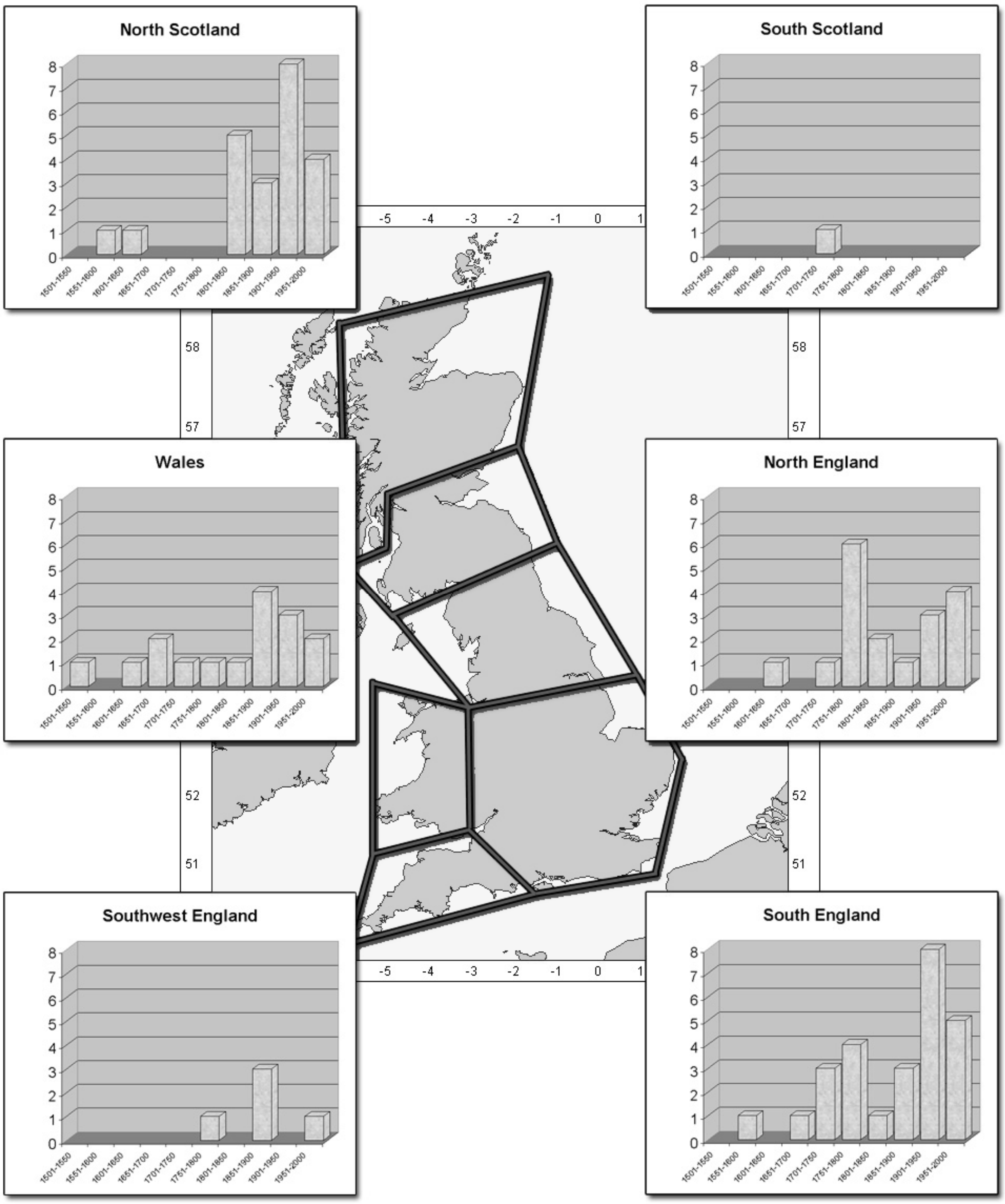

Figure 2 - Events of 4 ML and over by 50-year interval, by region

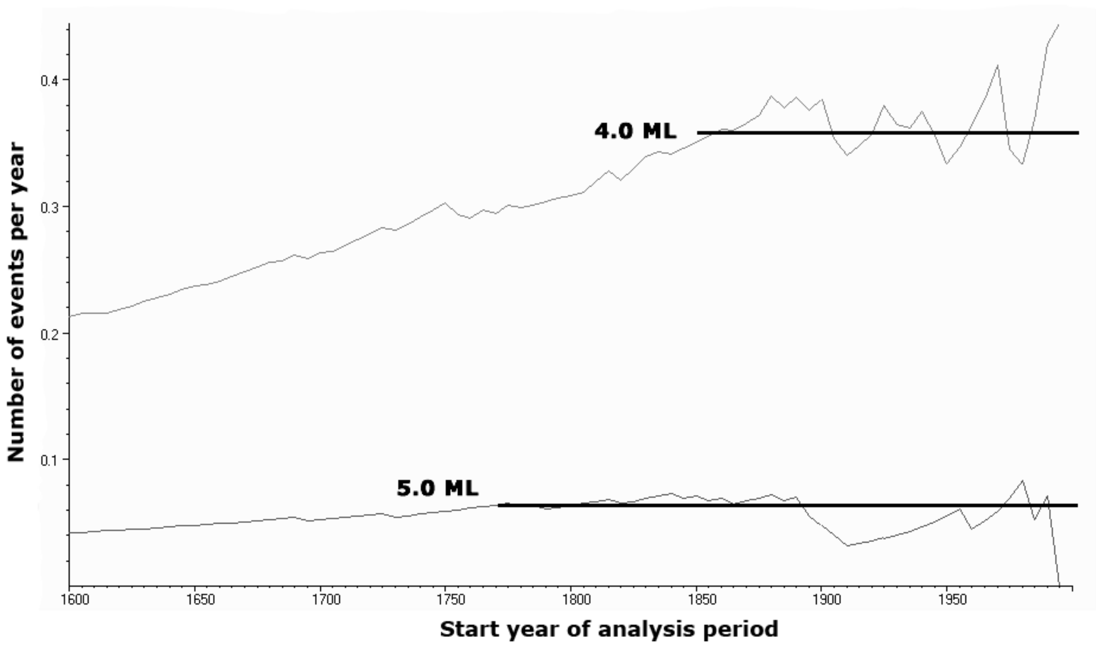

Figure 3 - Completeness analysis for the UK catalogue

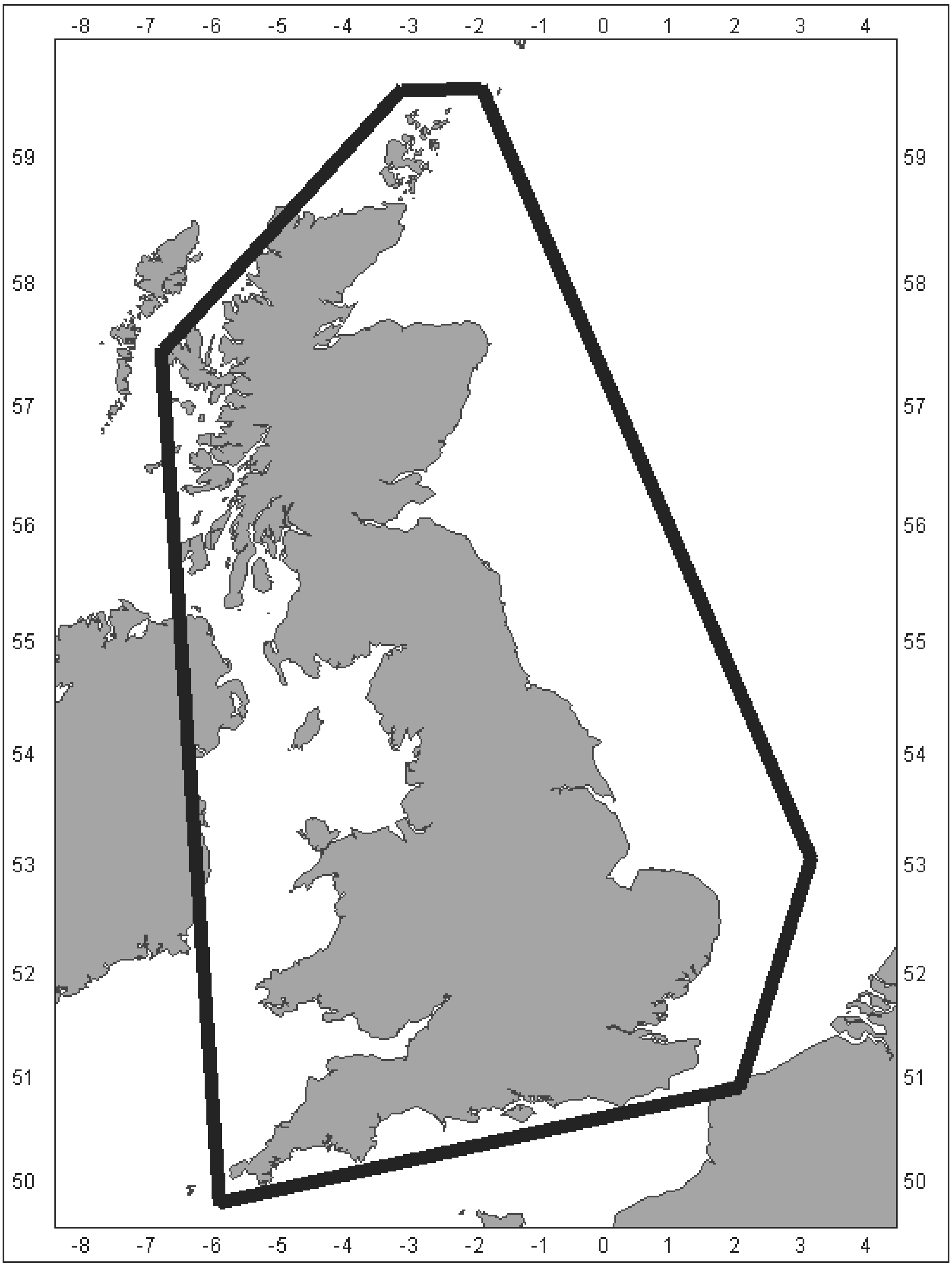

Figure 4 - Area for Gutenberg-Richter analysis

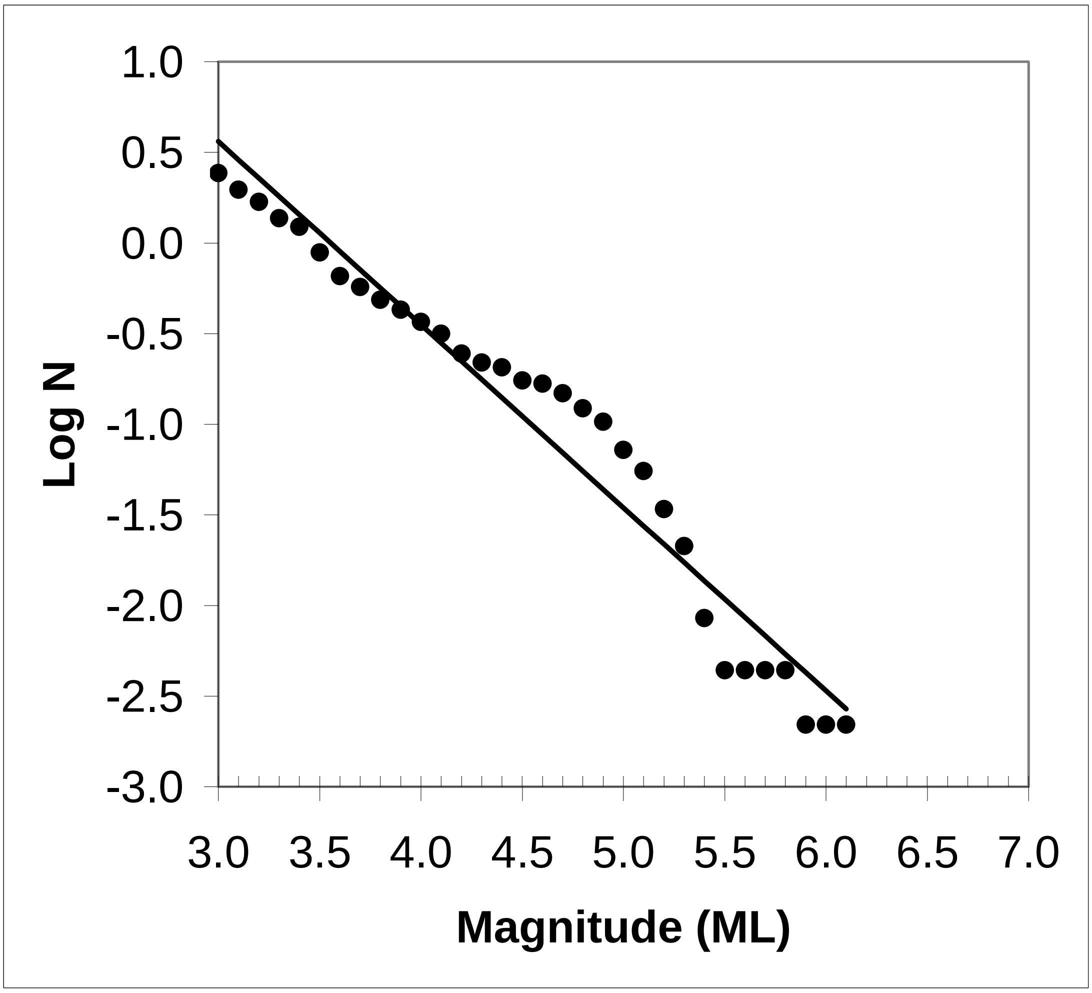

Figure 5 - Gutenberg-Richter plot for Great Britain; line shows best-fit to equation (1).

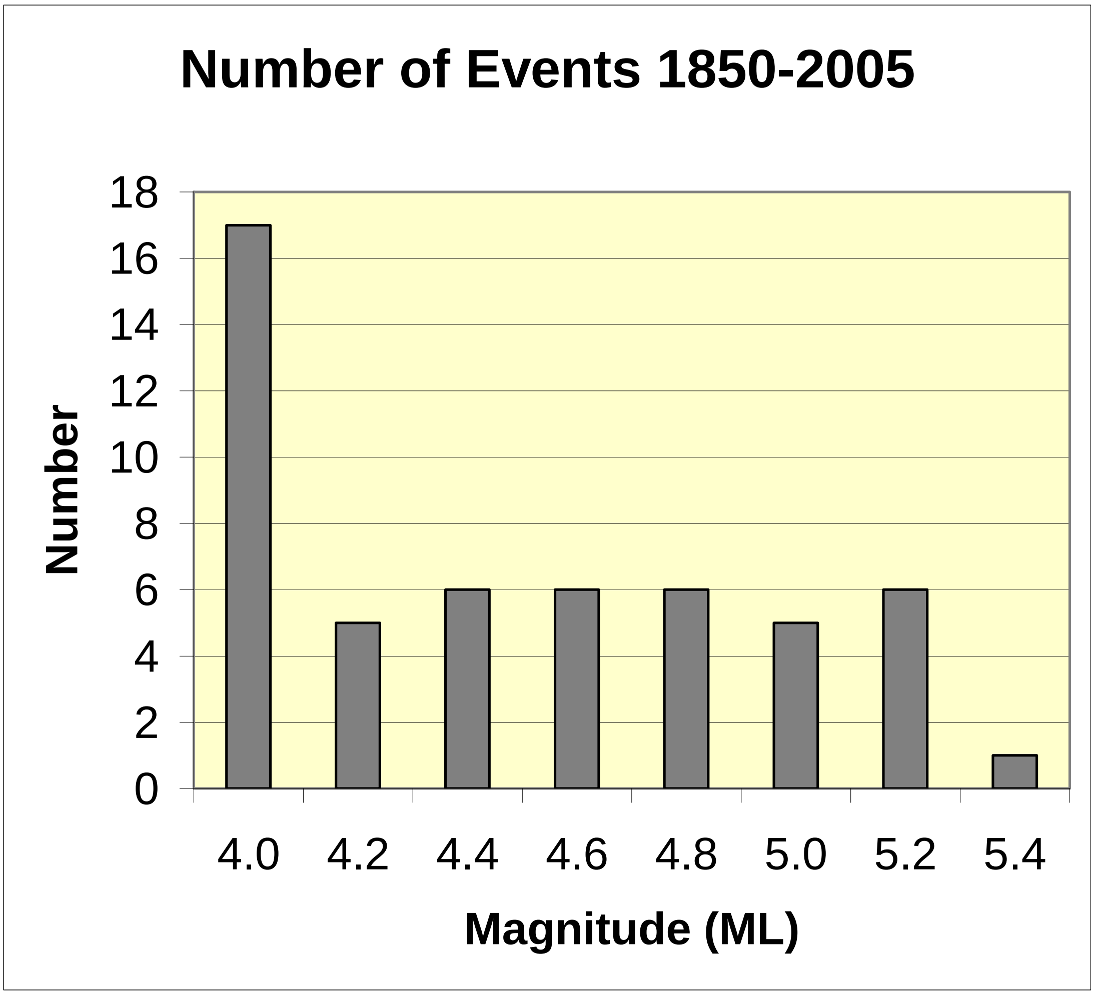

Figure 6 - Discrete number of earthquakes by magnitude for the period since 1850.

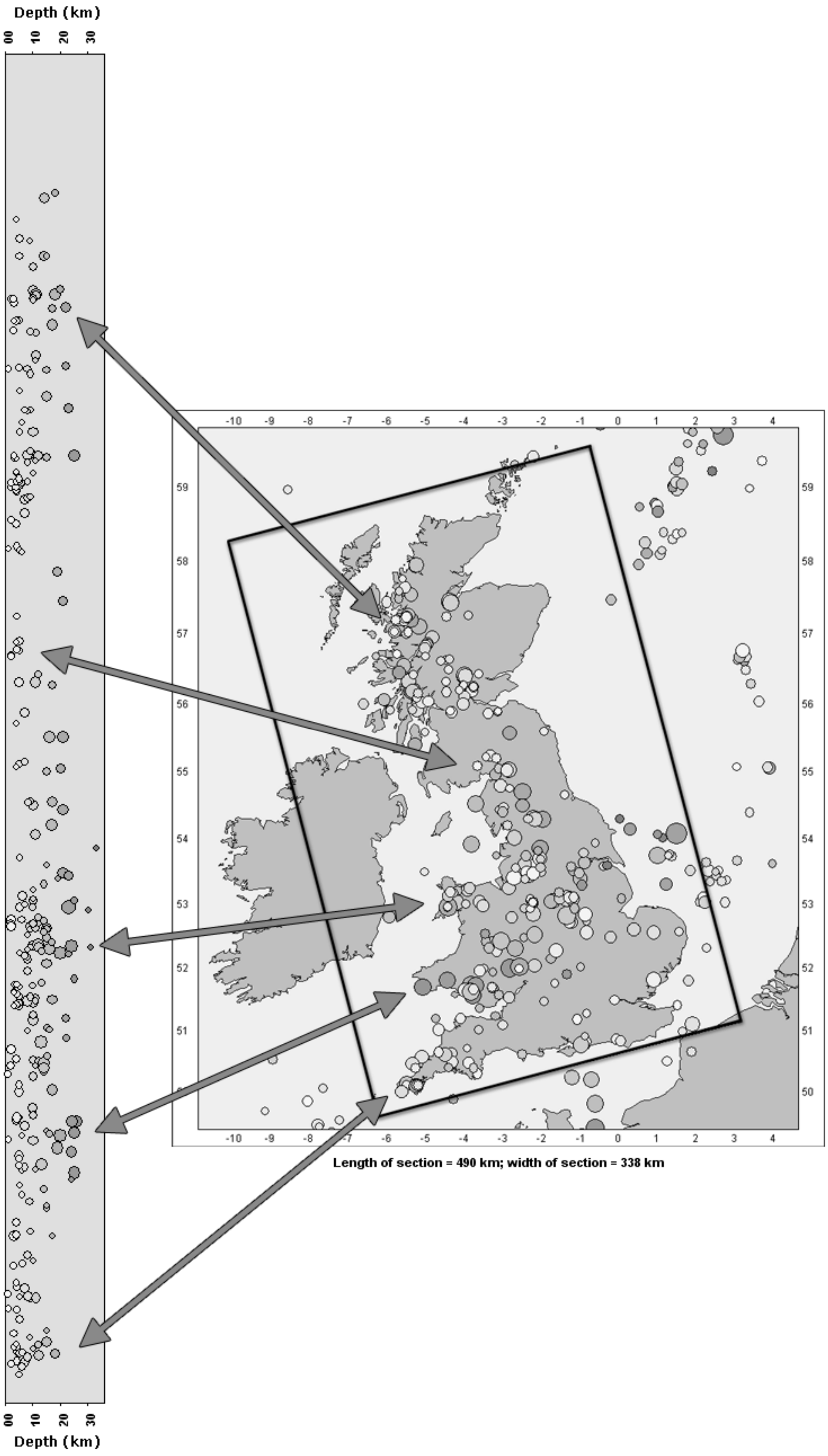

Figure 7 - Cross-section of British seismicity 1700-2005 (south to north; no vertical exaggeration). Symbol size proportional to magnitude; shallower events are paler).

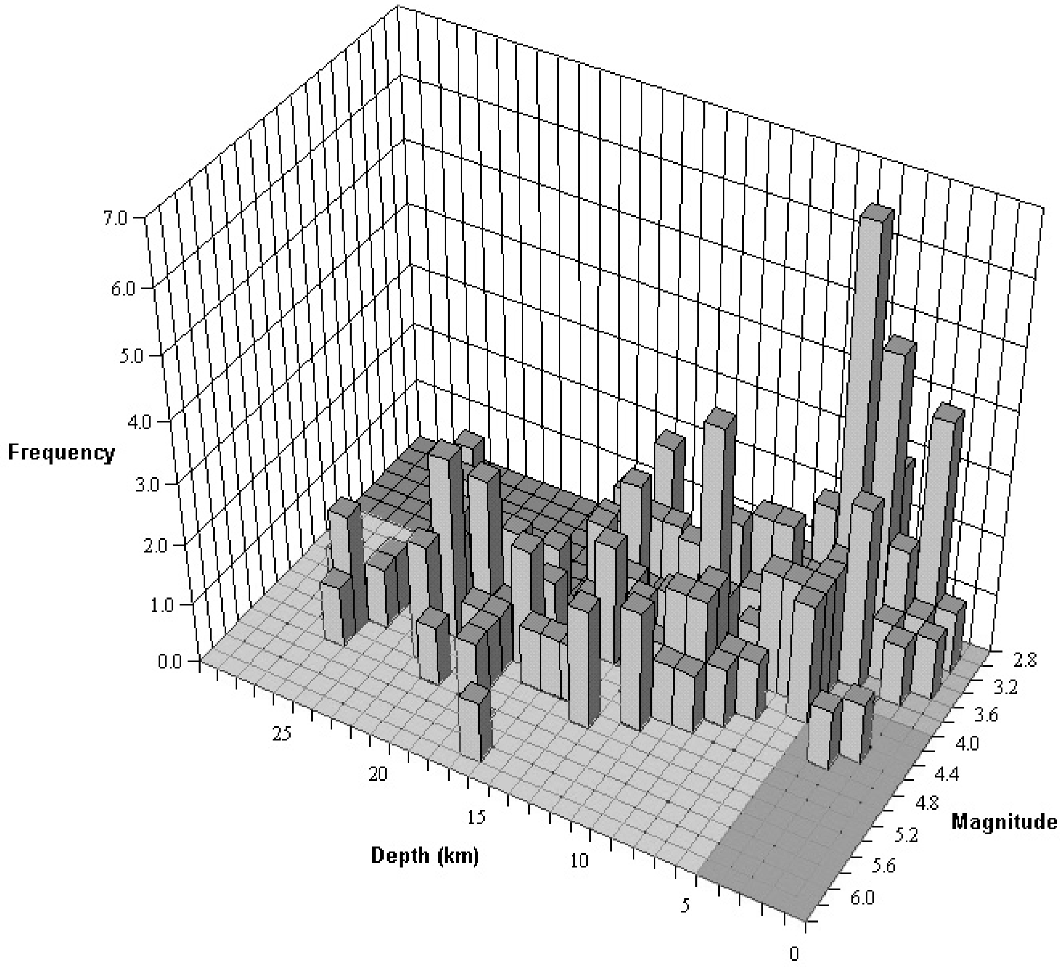

Figure 8 - Magnitude-depth analysis for British earthquakes (area as in Figure 4).

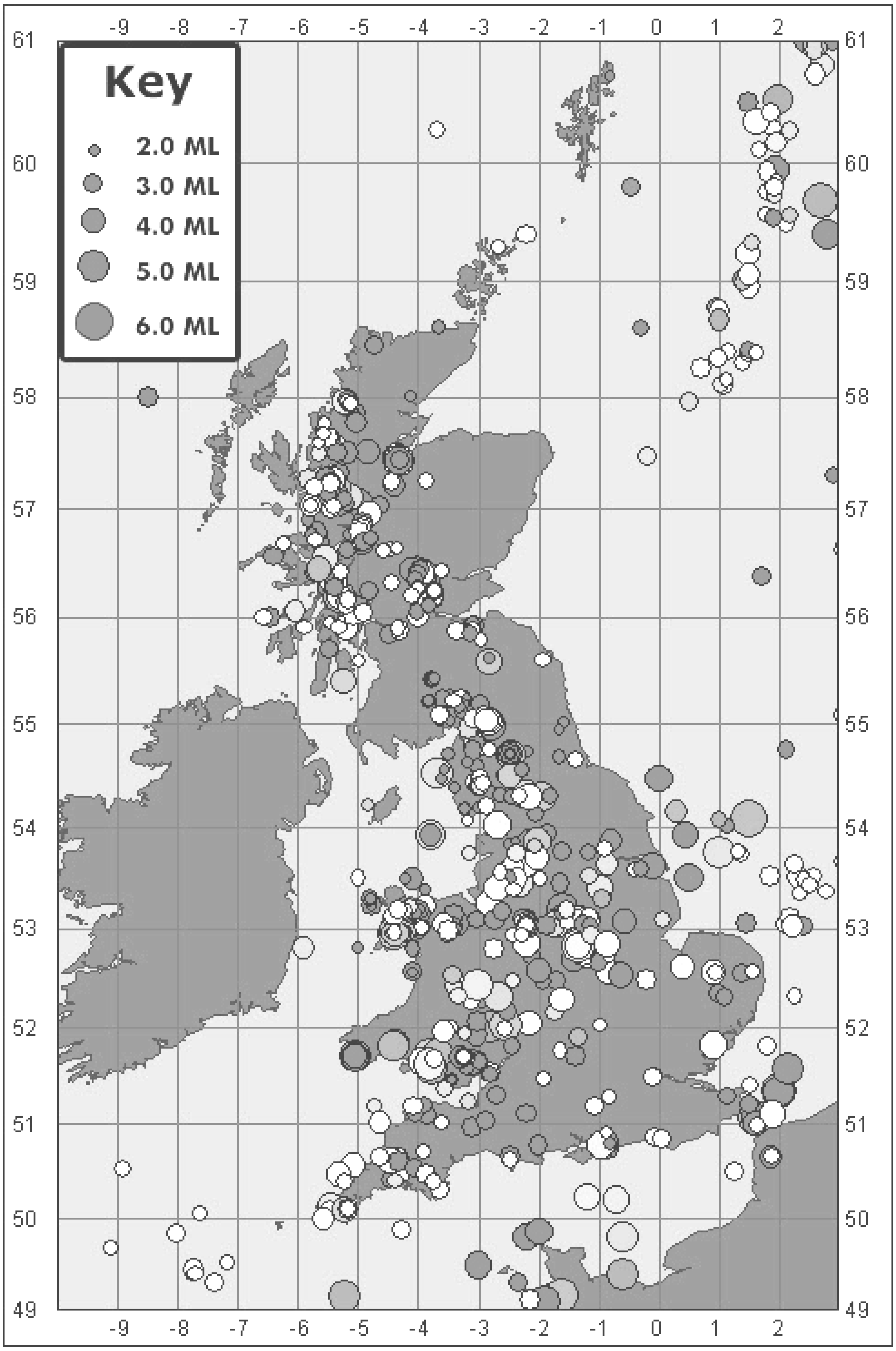

Figure 9 - The seismicity of the UK (symbol size relates to magnitude, paler symbols indicate shallower events). All locatable events > 2 ML to end of 2005 are plotted, with the exception of some Irish earthquakes < 3 ML.

Figure 10 - Tower of St Peter's, Broadstairs, showing repairs to the damage from the 1580 earthquake (two darker bands of stonework running down from top of tower).

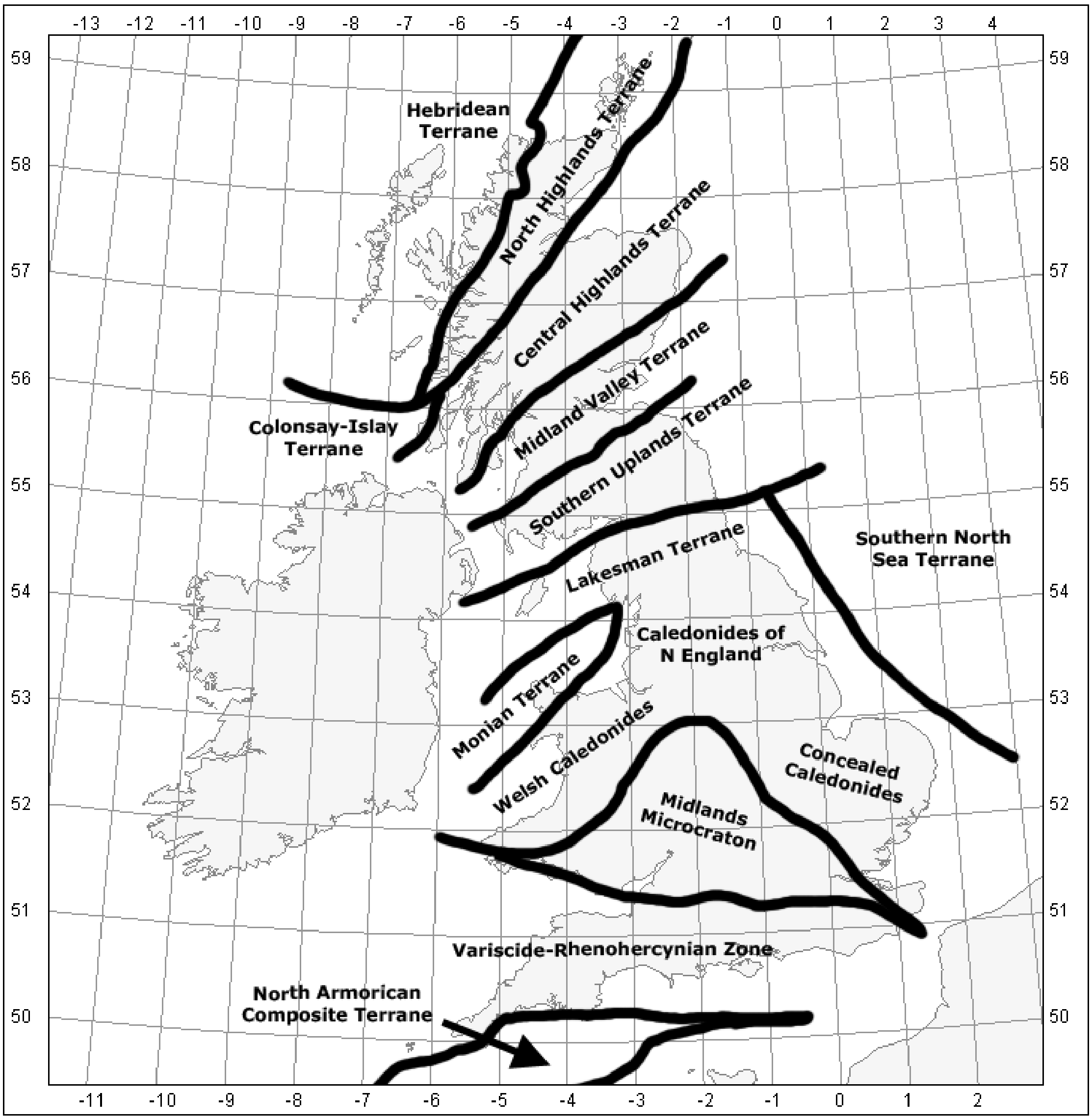

Figure 11 - British terranes, after Chadwick et al (1996)

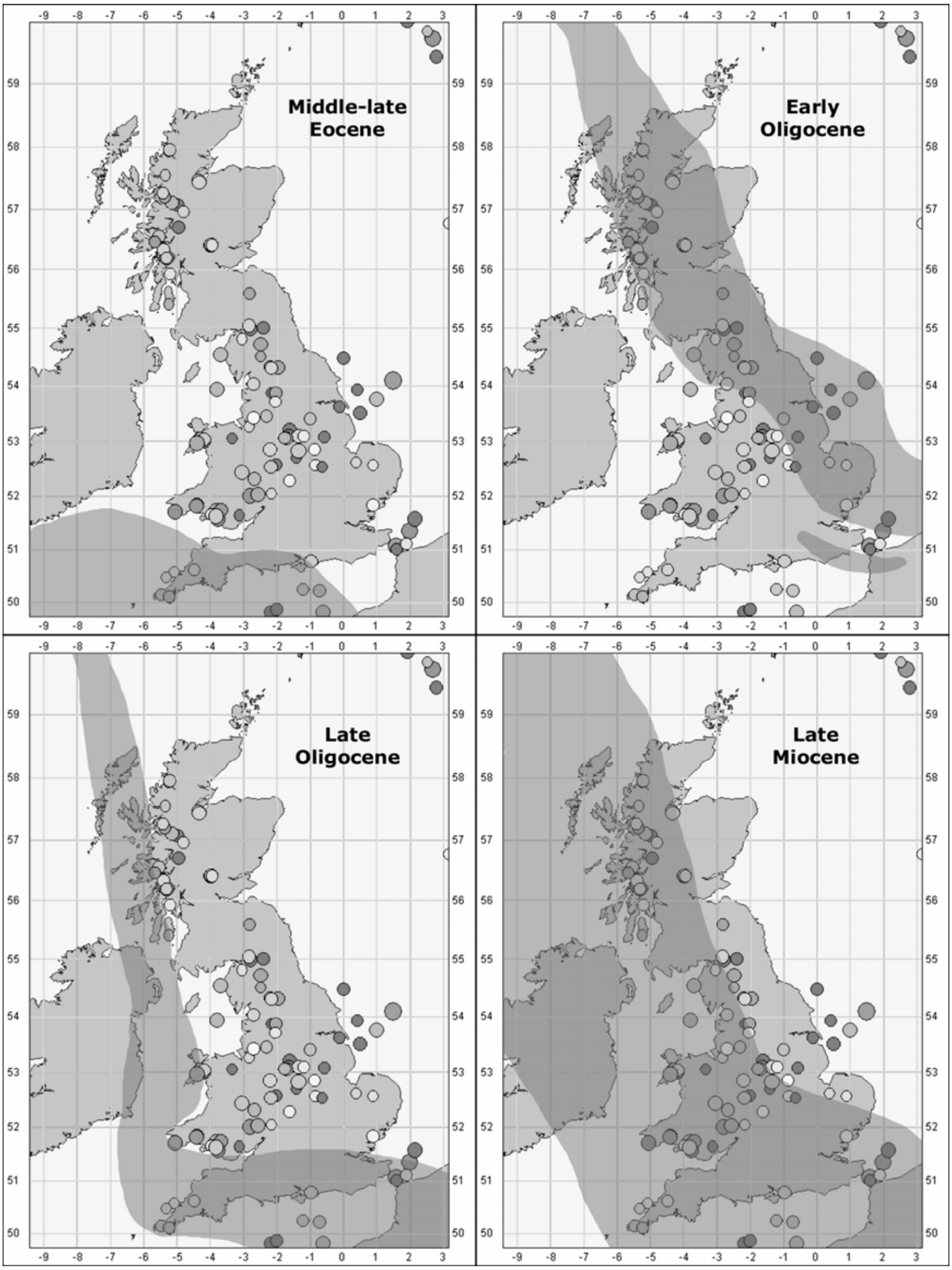

Figure 12 - Palaeogene-neogene sub-plate boundaries, after Muir Wood (1989). Shading: areas of deformation at different epochs; circles: earthquakes > 4 ML.

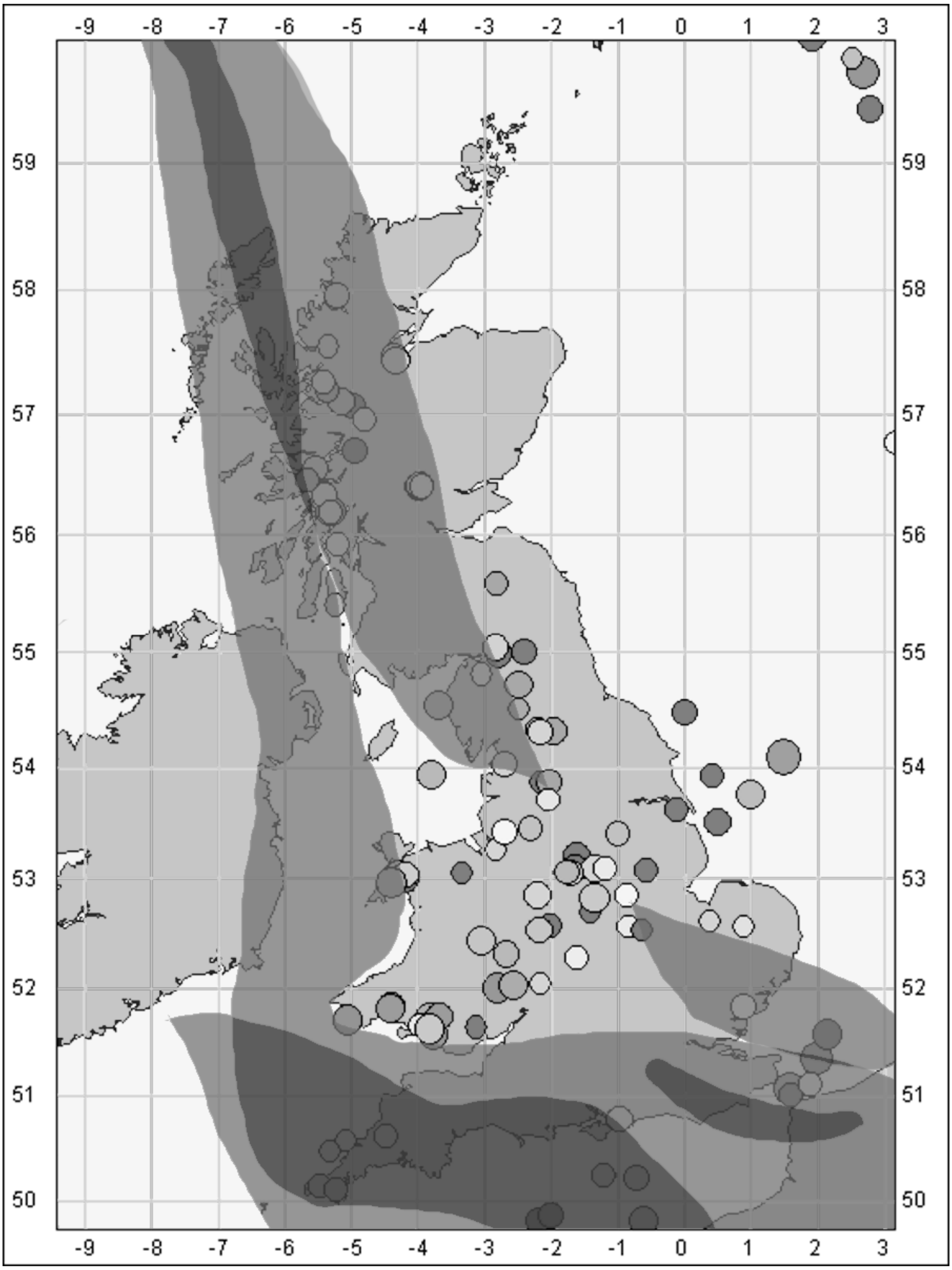

Figure 13 - Areas of multiple palaeogene-neogene activation in Figure 12. Shading: areas of deformation as in Figure 12, but superimposed; circles: earthquakes > 4 ML.

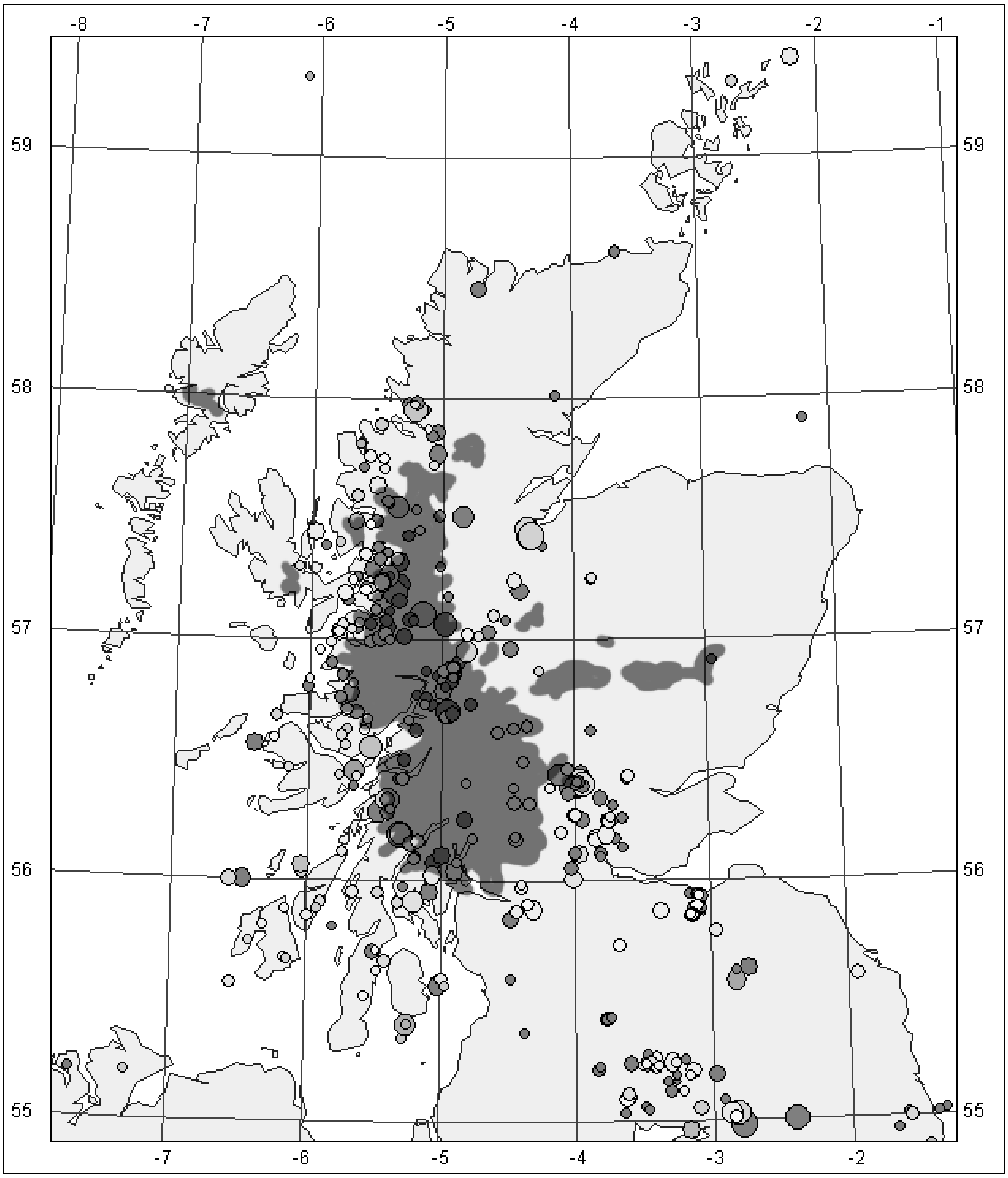

Figure 14 - Scottish seismicity compared to the Younger Dryas ice limits in Scotland (shaded area).

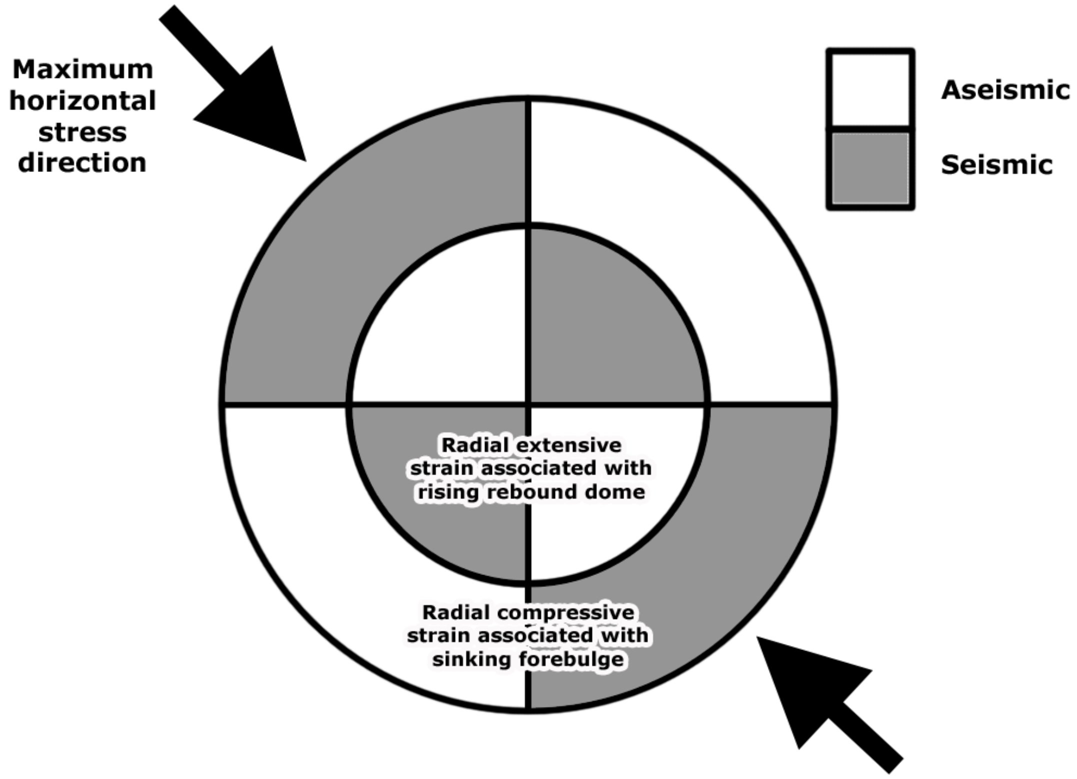

Figure 15 - Quadrant scheme for deglaciation seismotectonics (redrawn and simplified from Muir Wood, 2000)

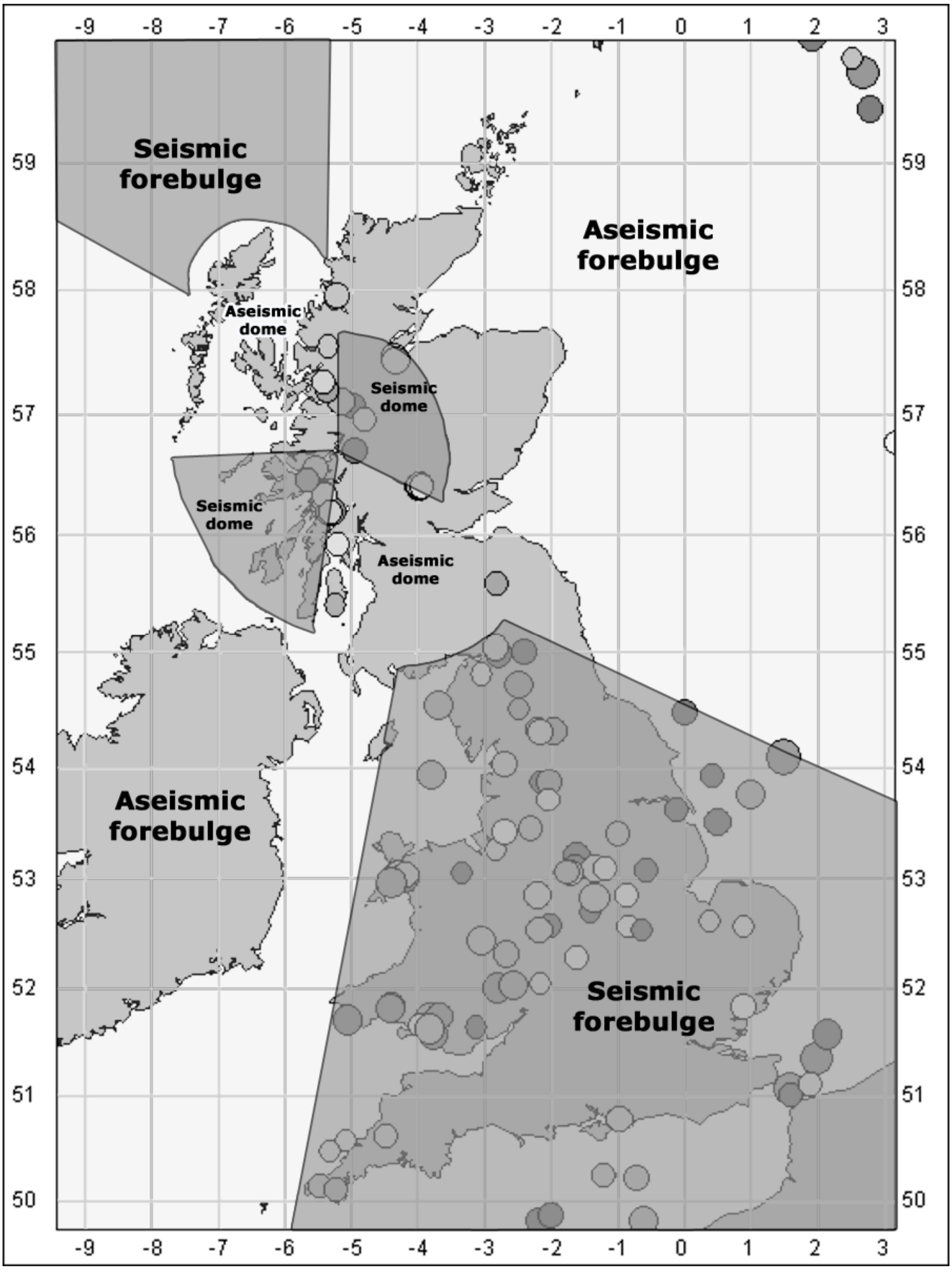

Figure 16 - Application of deglaciation quadrants to the UK (after Muir Wood, 2000, with earthquakes >4 ML added)

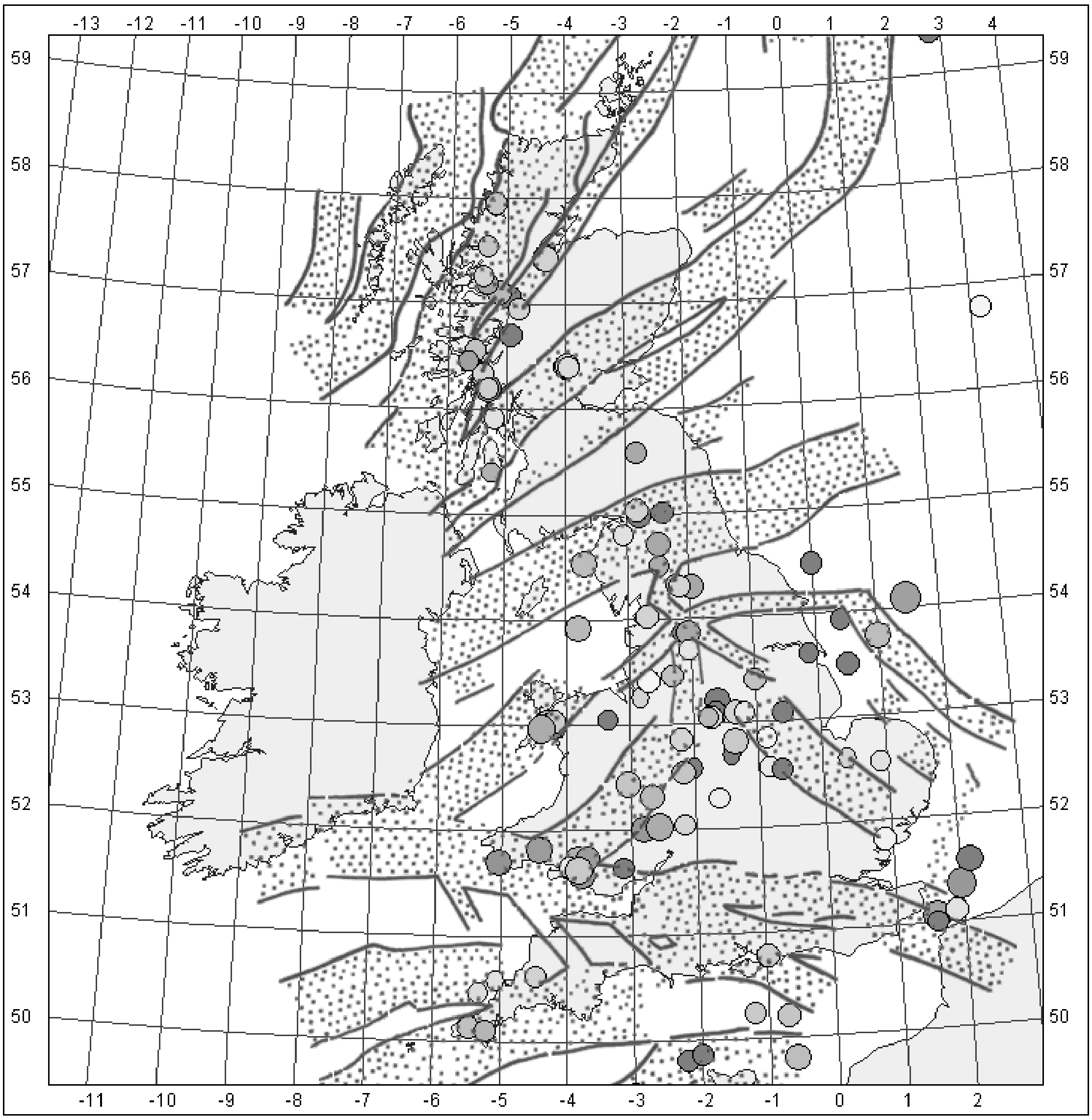

Figure 17 - "Corridors" around major fault systems in the UK, after Chadwick et al (1996)

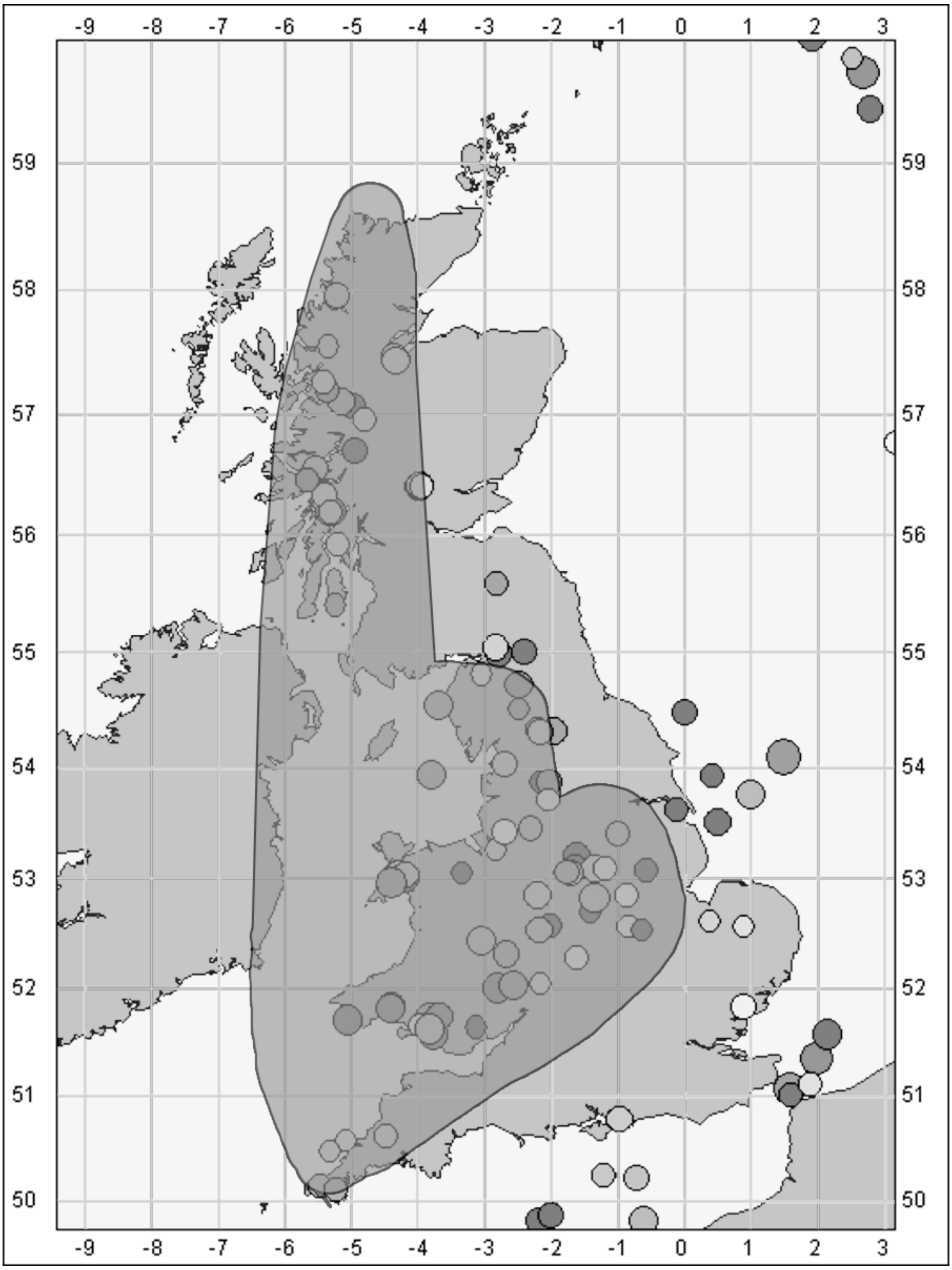

Figure 18 - Mantle anomaly (shaded area) and seismicity (shaded circles), after Bott and Bott (2004)

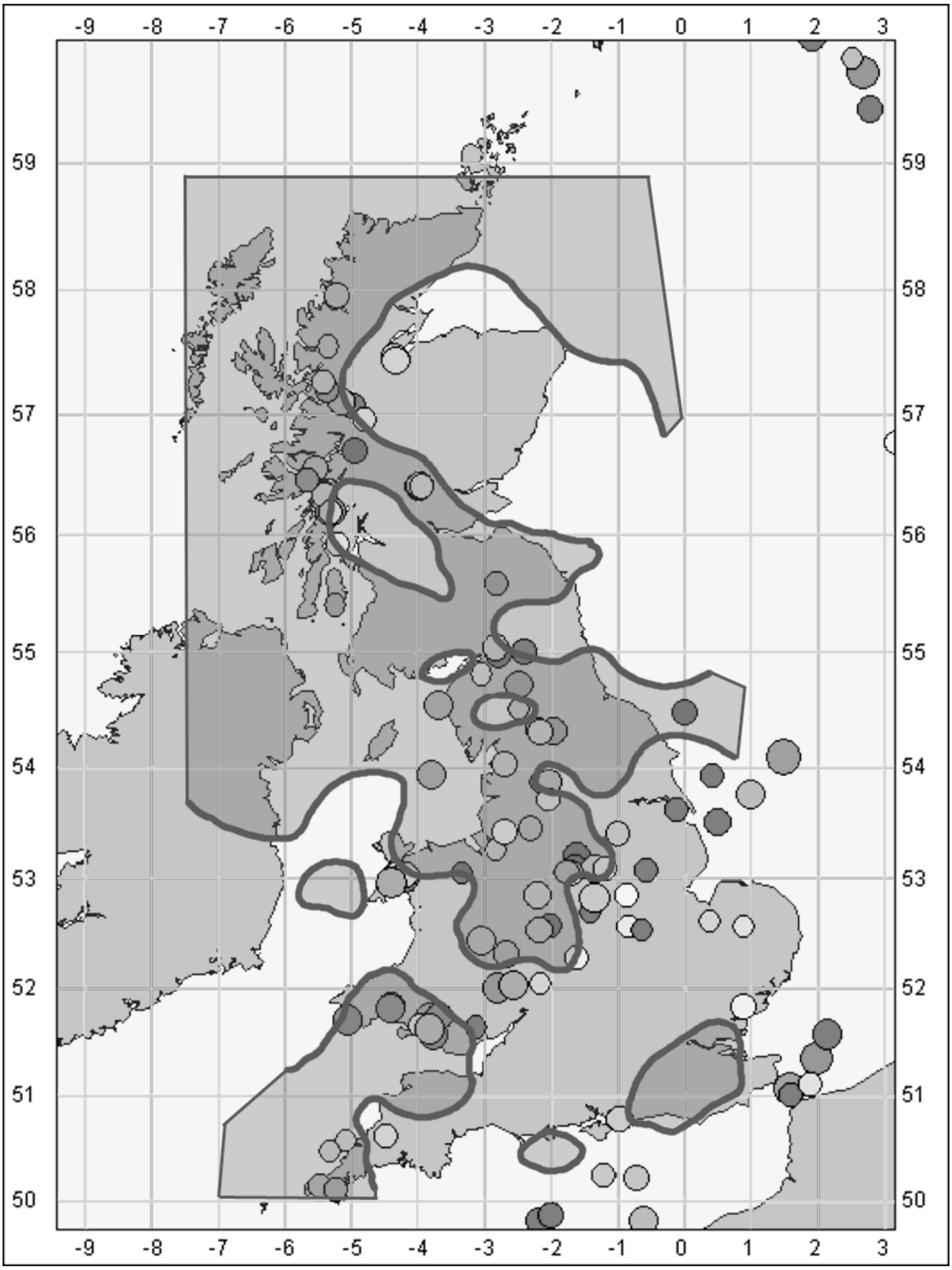

Figure 19 - Extent of P wave anomaly, after Arrowsmith et al (2005); thick line = edge of anomaly; thin line = edge of data

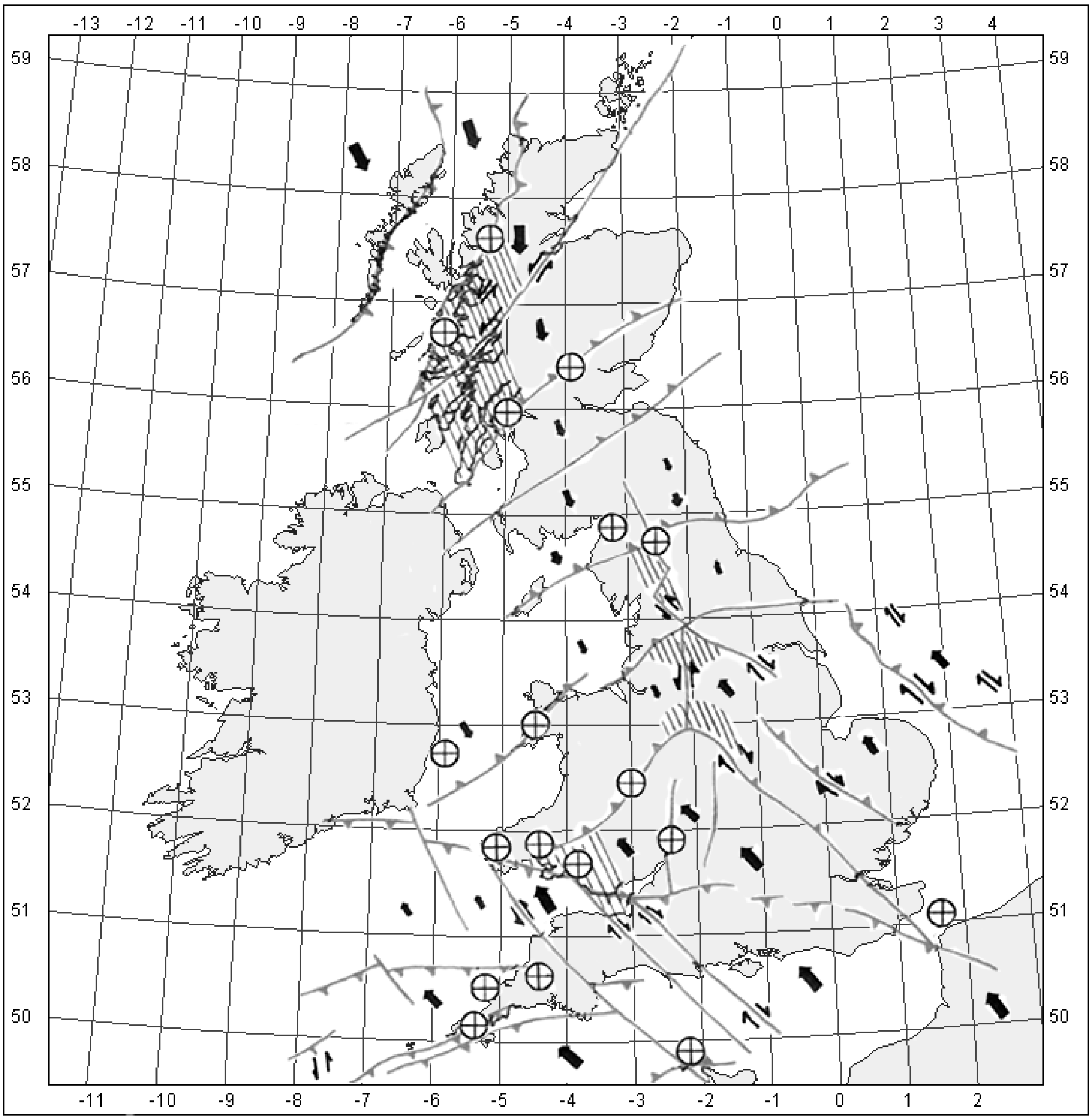

Figure 20- Kinematic model for the UK, after Chadwick et al (1996); shaded areas = high strain rate; single arrows = incipient block motion (relative); double arrows = strike-slip reactivation of transcurrent fault or thrust fault sub-parallel to incipient block relative motions; circles = distributed reactivation of thrust fault (dominantly strike-slip)