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Analysis of twenty microseism storms during the winter of 1987–1988 and comparison with wave hindcasts

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ABSTRACT

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In this paper an attempt is made to establish a quantitative relationship between the spectra of secondary microseisms recorded at Menai Bridge and the parent sea-wave spectra in the sea areas in the western approaches. Secondary microseisms have double the frequency of the parent waves. The sea-wave spectra were obtained from a hindcasting model. The problem was simplified by restricting the consideration of the microseismic energy to that associated with periods of 7 s or over and relating this to the energy spectra of sea waves of period 14 s or over. In this way only the strongest storms (usually only found in the North Atlantic) were taken into account. The hindcasting model gives a complete frequency-directional distribution of the sea-wave energy and so in coastal areas the energy moving normal to the coast can be ascertained. In the event, a workable relationship was found between the microseism spectra and those for waves in the Bristol Channel area and also the area bounded by the north coast of Ireland and the south-west coast of Scotland, including the North Channel. From the theory due to Longuet-Higgins it was possible to calculate the coastal reflection coefficients from the relationship by putting in the areas for these two regions. These were found to be between 0.15 and 0.25, which are too high for these coastal regions. It is probable therefore that these two regions act as monitors for larger sea areas.

An investigation was also made into the existence of primary microseisms (these have the same frequency as the parent waves). The results confirmed a previous finding that, with the vertical component, the energy is about 100 times less than that of the secondary.

1. Introduction

It has been known for a long time that there is a close connection between microseisms of frequency 1–0.1 Hz and sea waves caused by ocean storms. This was noticed by Wiechert (1905) and in 1941, Bernard found that there was a two-to-one relationship between wave and microseism periods. A sound theoretical basis was given to the relationship by Longuet-Higgins (1950) who showed that the microseisms were formed by the pressure set up at the sea bottom by the interference of waves moving in opposite directions. This could happen in the middle of a storm area where the wind could change in direction or by reflection off coasts.

Although many instances were found where coastal wave spectra were very similar to corresponding microseism spectra at double the frequency (see, for example, Darbyshire, 1950) and it is often possible to identify the onset of a microseism storm with a particular depression, nevertheless there has hardly been any work done to establish experimentally quantitative relationships between the microseism spectra and the parent wave spectra. This is due mainly to the relative paucity of wave spectral measurements compared with those of microseisms. Indeed, to make a meaningful comparison, it is necessary to know the wave energy distribution over both frequency and direction. Such wave measurements have been very rare up to the present but no

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doubt they will be made in the future by satellite techniques. In the meantime, recourse has had to be had to the hindcasting of wave spectra from the wind data as given by meteorological charts. Various hindcasting (and forecasting) models are available but the author has developed a model that can be easily programed even on a microcomputer. It has been described by Elliott (1987). It converts a grid of wind speeds and directions into a similar grid of wave frequency-directional spectra. The length of the grid square can be chosen to meet particular circumstances but it is usually taken to be 100–300 km. The result enables one to see immediately if there are any wave energies moving in opposite directions in the squares covering the storm area and also enables one to resolve the wave energy perpendicular to a coast.

With this method, it was decided to investigate every microseim storm during the autumn–winter of 1987–1988 where the energy associated with a period of 7 s and above exceeded a certain value. The period limit was chosen so that weaker storms could be ignored and so ease the identification of the source storm. The microseims were recorded at Menai Bridge, Anglesey.

2. The measuring system

Microseims have been recorded at Menai Bridge since 1982 after a lapse of some years. The seismographs were moved to a new site a few hundred metres away in February 1987 as the previous site had become too noisy. The seismographs used are of the Willmore type which have been modified to record microseims. The response is constant from 10 to 0.1 Hz, dropping to 97% at 0.05 Hz. It is therefore constant over the range of frequencies under investigation. The instruments measure ground velocity and the K factor is 490 V m^{-1} s^{-1}. The signal is fed to an amplifier of gain 5 and then to an analogue–digital converter. The output of this is such that one unit corresponds to 2×10^{-3} V and therefore corresponds to 8×10^{-7} m s^{-1}. Using variance as a measure of energy, 1 unit of variance then corresponds to $0.5 \times 64 \times 10^{-14}$ m^2 s^{-2}. (The number 3.12×10^{14} which will be used in Figs. 3–8 is the reciprocal of this.)

Although both a vertical and a north–south seismograph was available, all the observations were taken from the vertical one.

3. Analysis of microseismic storms

Records were taken continuously from February 1987. The outputs were recorded on magnetic tape. These were read and the results analysed with a Fast Fourier Transform program with a micro-computer. The records were usually only analysed at twelve-hourly intervals but this interval could be reduced to six or three when there were interesting developments. There was little activity of interest until September 1987 but after this there were several interesting storms right up to March 1988.

Before any investigation of the origin of microseisms, it is necessary to obtain refraction charts for their propagation across the N. Atlantic. The basis for these charts has been described by Darbyshire (1987), the microseism velocity being dependent on the ocean depth. Charts for 7 s and 8 s microseisms are shown in Figs. 1 and 2 and apply to microseisms approaching the British Isles. The rays are made to propagate outwards from the recording station at equal angular intervals. When the angle between two adjacent rays exceeds the initial angle there is divergence and when it is less there is convergence. By the principle of reversibility, microseisms originating in the converging and diverging zones would accordingly converge or diverge at the recording station.

In studying the onset of microseisms, it was decided, as mentioned above, to limit the investigation to microseisms with energy in the spectral band between 7 and 10 s period (0.14–0.1 Hz frequency). Shorter-period microseisms tend to be present at all times with varying intensities and so it is very difficult to isolate their origin. Furthermore, the investigation was further limited to cases where the energy above 7 s was 100 or more on the digitizer scale, (3.2×10^{-12} m^2 s^{-2}). In the event this quantity varied from 100 to 2000. Plots of the energies for eight different time sequences, involving twenty distinct storms, are shown in Figs. 3–6. It will be convenient to discuss those of

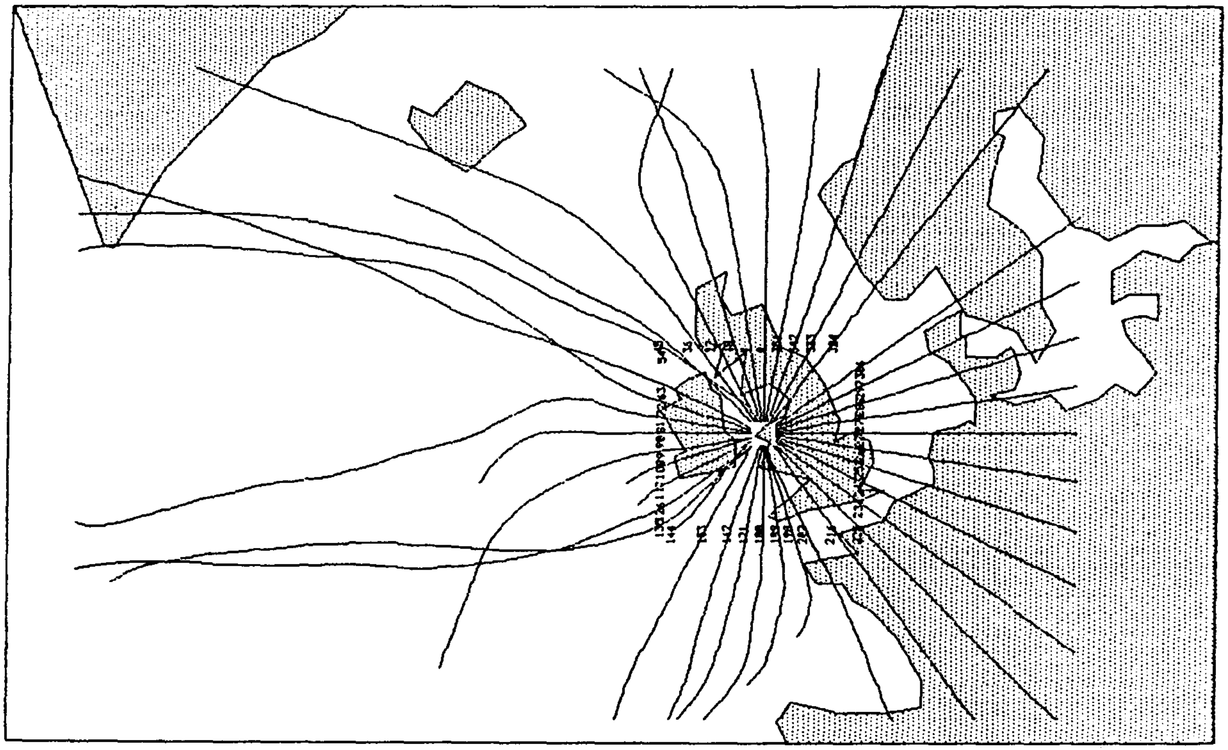

Fig. 1. Refraction of 7 s microseisms approaching the British Isles.

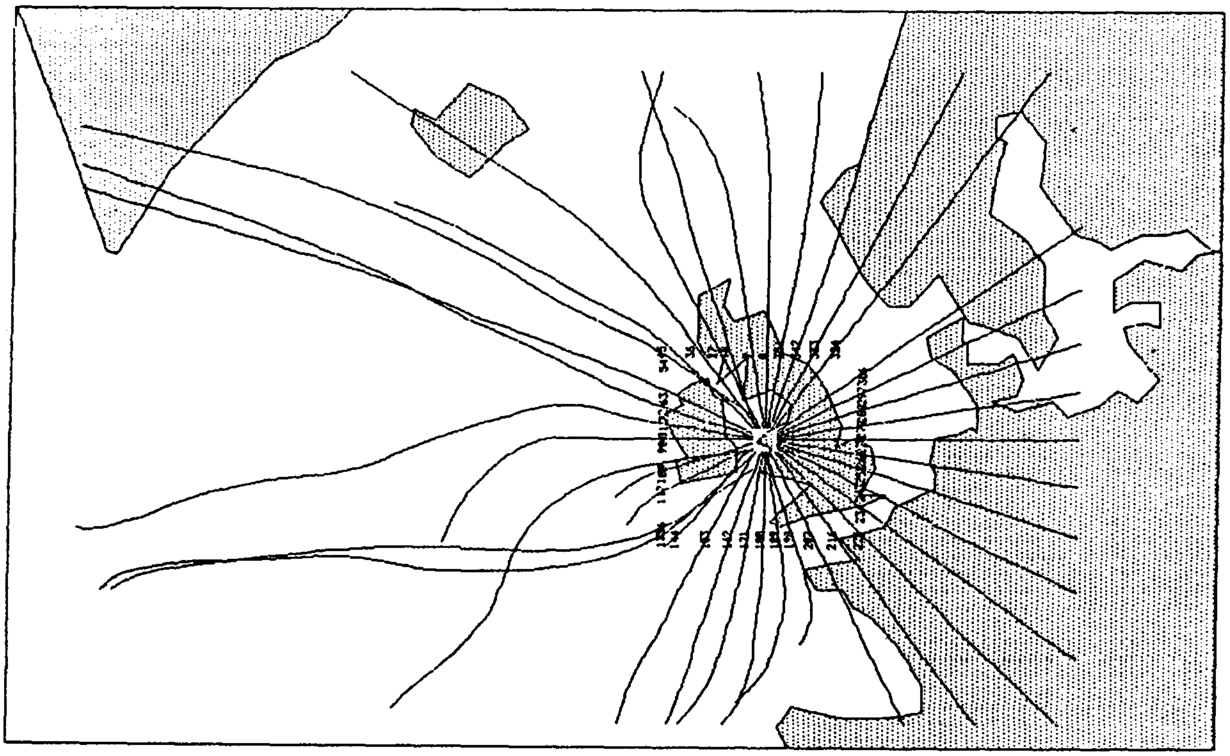

Fig. 2. Refraction of 8 s microseisms approaching the British Isles.

September–November 1987 and December 1987–January 1988 separately. The results are shown in Figs. 3–8, which have three sections. The lower one shows the wave activity in the Bristol Channel area and 1 unit on the vertical scale corresponds to $1/700\ m^4$. The second section applies to the North Channel and here 1 unit corresponds to $1/900\ m^4$. The top section shows the microseismic energy and 1 unit corresponds to $1/3.12 \times 10^{14}\ m^2\ s^{-2}$. The divisor for the third

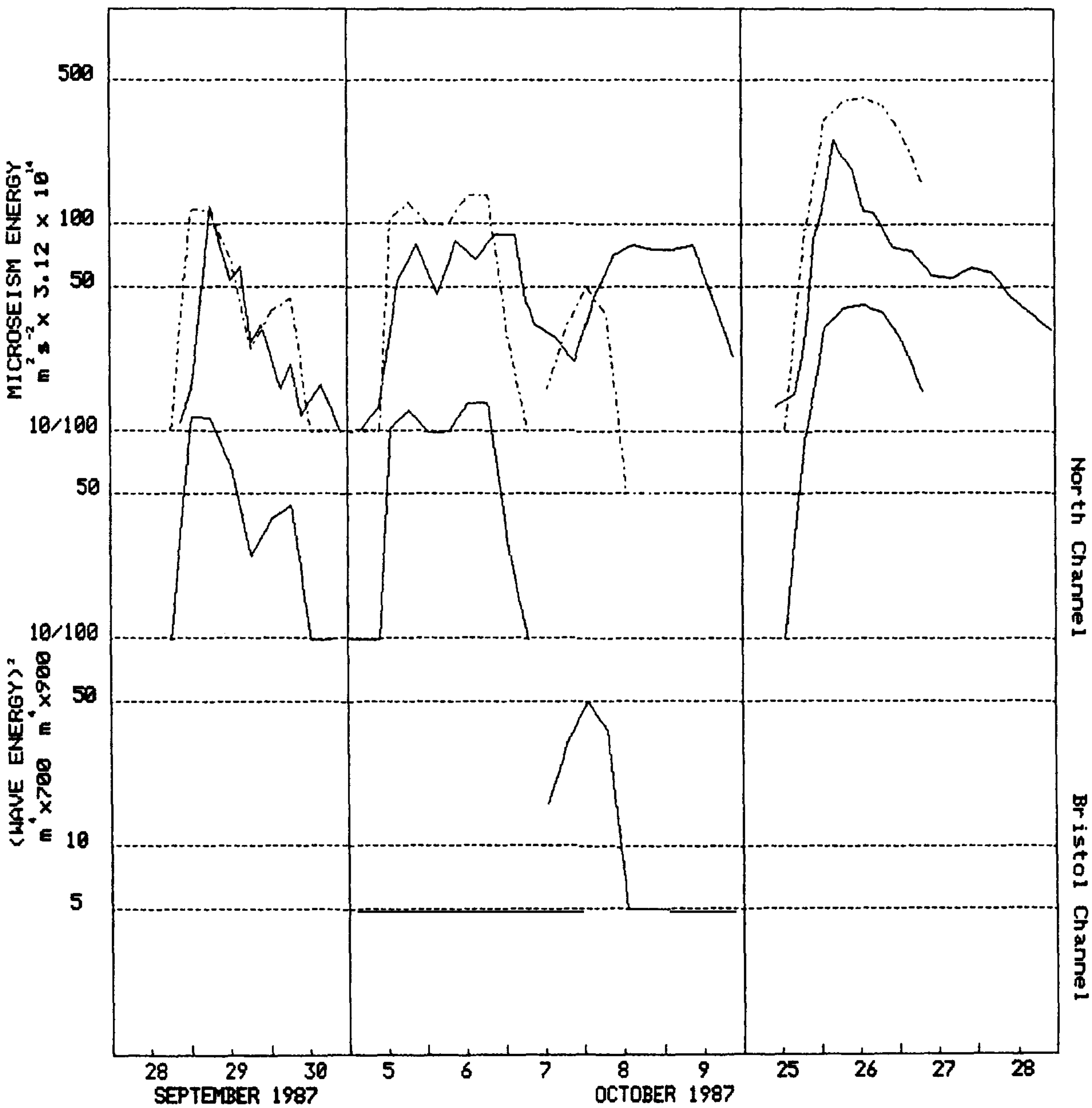

Fig. 3. Comparison of microseism energies and the square of hindcast wave energies for September and October 1987. (Microseism energies for 7 s and above. Wave values for 14 s and above.) The broken line in the top section shows the sum of the values in the other two sections.

section corresponds, as stated above, to the sensitivity. The other two divisors have been adjusted to obtain a best fit between the sum of the values in the two bottom sections and the value in the third over all the examples. Where there is activity in both lower sections, this sum is shown by the broken line in the top section for comparison. It will be more convenient to use the units without the divisors for reference purposes as is done above.

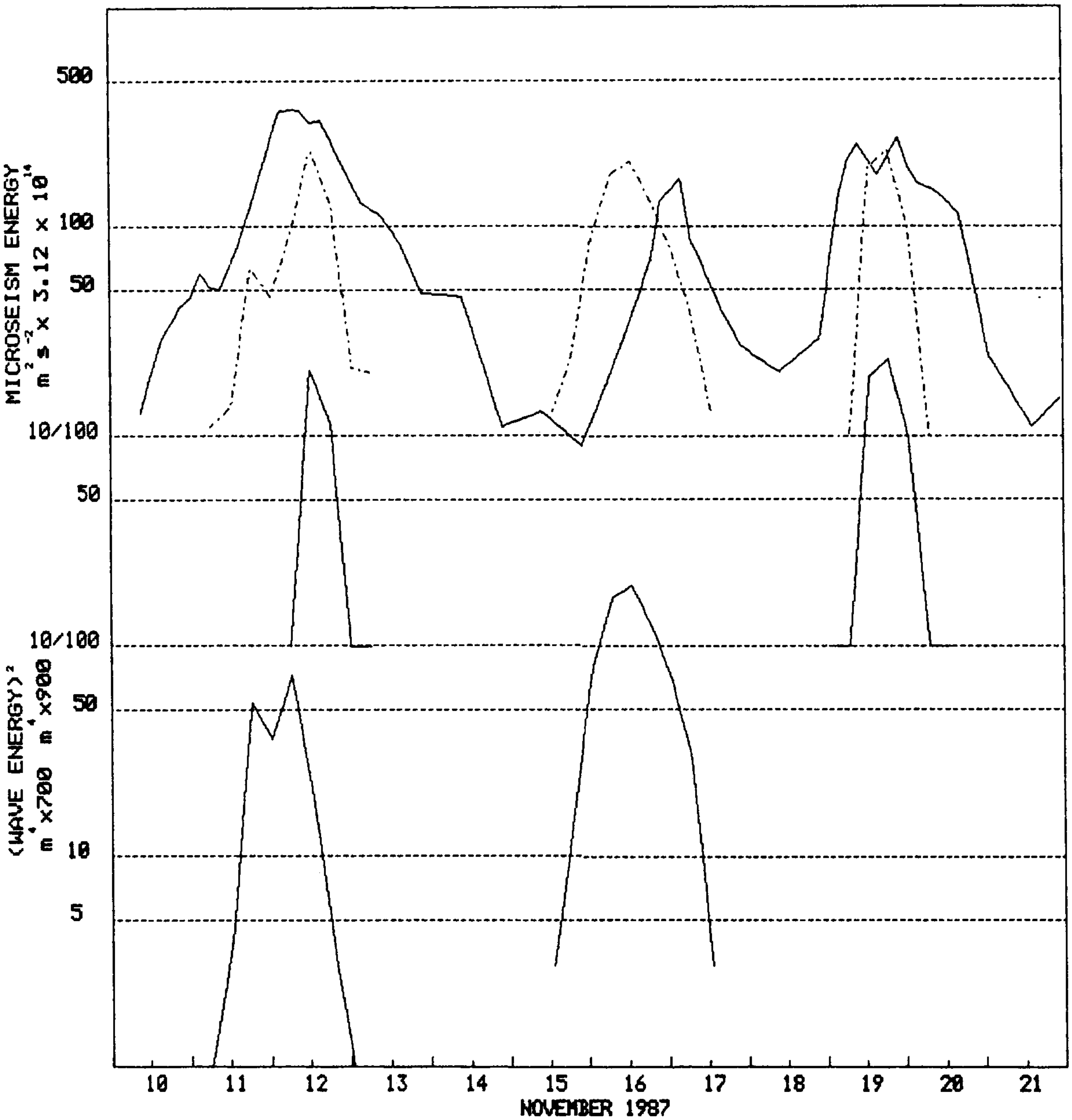

Fig. 4. Comparison of microseism energies and the square of hindcast wave energies for November 1987. (Microseism energies for 7 s and above. Wave values for 14 s and above.) The broken line in the top section shows the sum of the values in the other two sections.

3.1. Storms of September–November 1987

The first storm that satisfied the required conditions occurred on 28–30 September (Fig. 3). Although the storm was not so marked as were most of the succeeding ones, it aroused special interest as it was the first onset of any appreciable activity after a long period of inactivity. It was, in any case, very interesting as the intensity increased very rapidly and then diminished just as rapidly so that the intensity–time curve had a very sharp shape.

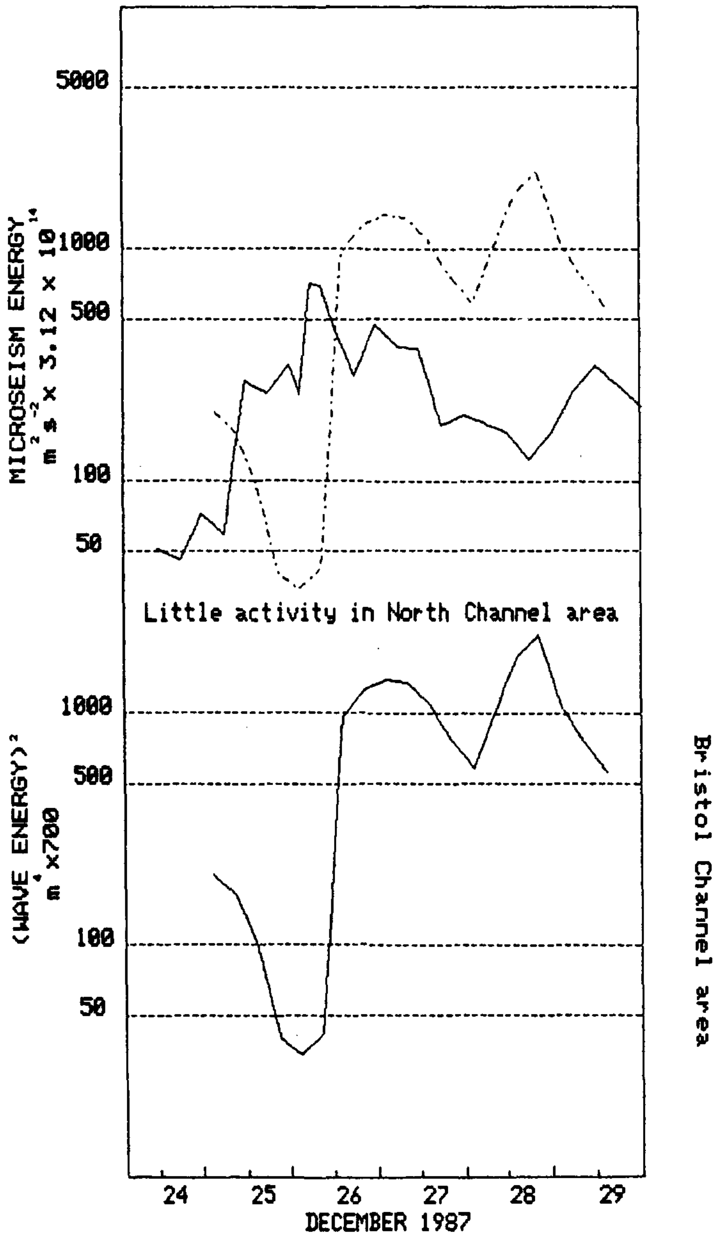

Fig. 5. Comparison of microseism energies and the square of hindcast wave energies for December 1987. (Microseism energies for 7 s and above. Wave values for 14 s and above.)

It was decided to compare the observations of this and succeeding storms with hindcasted sea-wave energy values for various regions of the N.Atlantic. The hindcasting model already described was used. For the September case there was a very intense storm starting at about 1000 km south of Iceland and moving northwards. The hindcasting model gave waves up to 30 m peak-to-trough height off the coast of Iceland. It was originally thought that these gave rise to the microseisms observed. The coast of Iceland is, however, in a divergent zone (Figs. 1 and 2), and the peak wave activity preceded that of the microseisms by 12–24 h. The storm, however, gave rise to very high waves on the west coast of Scotland, which although not so high as those off Iceland,

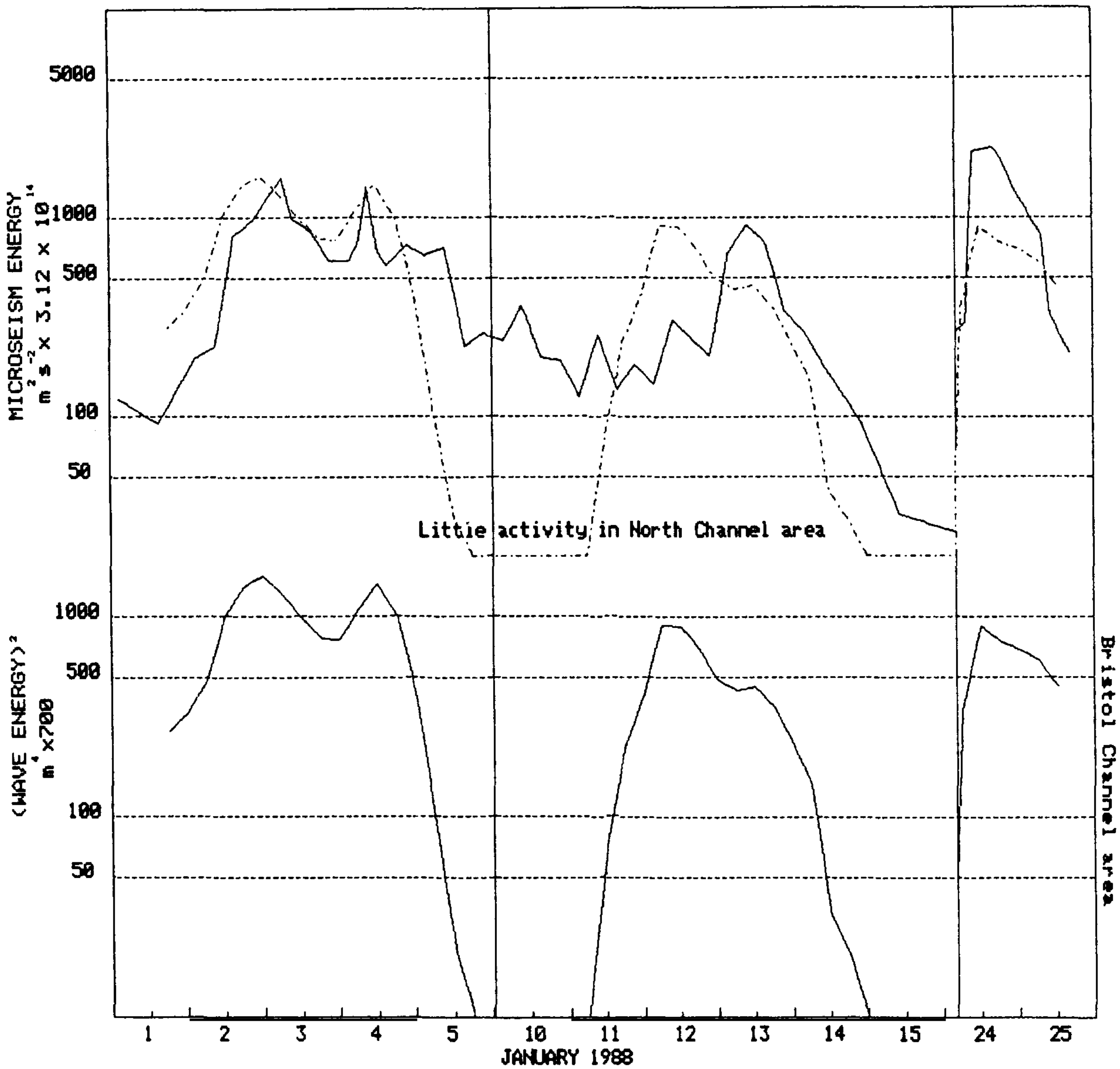

Fig. 6. Comparison of microseism energies and the square of hindcast wave energies for January 1988. (Microseism energies for 7 s and above. Wave values for 14 s and above.)

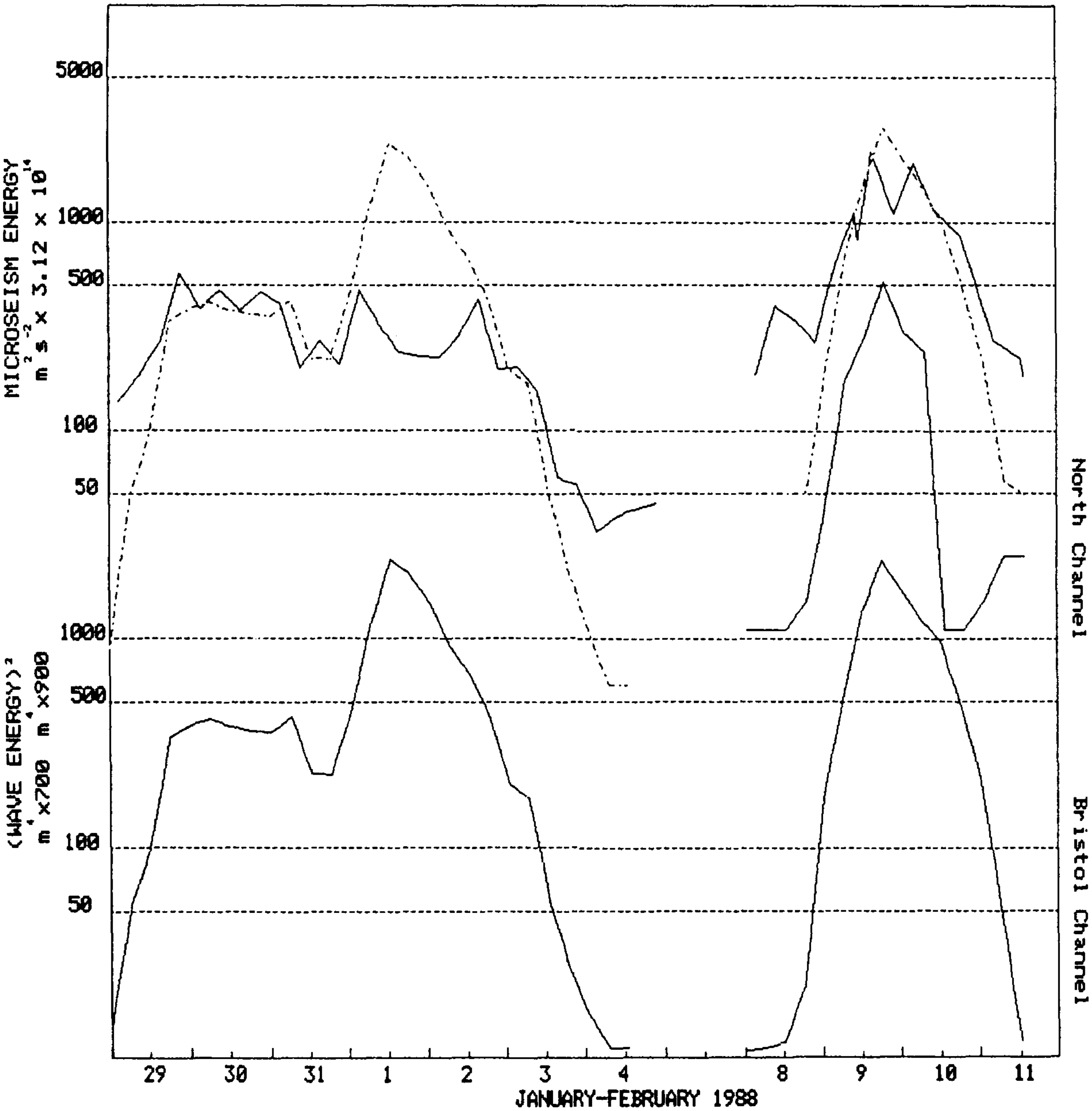

Fig. 7. Comparison of microseism energies and the square of hindcast wave energies for January–February 1988. (Microseism energies for 7 s and above. Wave values for 14 s and above.) The broken line in the top section shows the sum of the values in the other two sections.

would be more effective because of the shorter distance to the recording station.

The microseisms are clearly secondary ones as the periods are about half the wave periods and, according to Longuet-Higgins (1950), could only be generated by the simultaneous transmission of wave energy in two opposite directions. In the deep ocean this could happen, for instance, when waves formed by the winds in the northern sector of a storm meet others produced in the southern sector of another storm whose bearing to the first has a westerly component. Moreover, the winds and the fetch in both storms would have to be strong enough to produce appreciable 14 s waves and the area of wave interference would have to be in a non-diverging zone. During the time covered by this investigation these conditions were not met. The other possible cause of such wave interference is reflection off coasts. The western coasts of both Ireland and Scotland are very indented and any incident waves would, in general, be reflected in all directions and there would be only

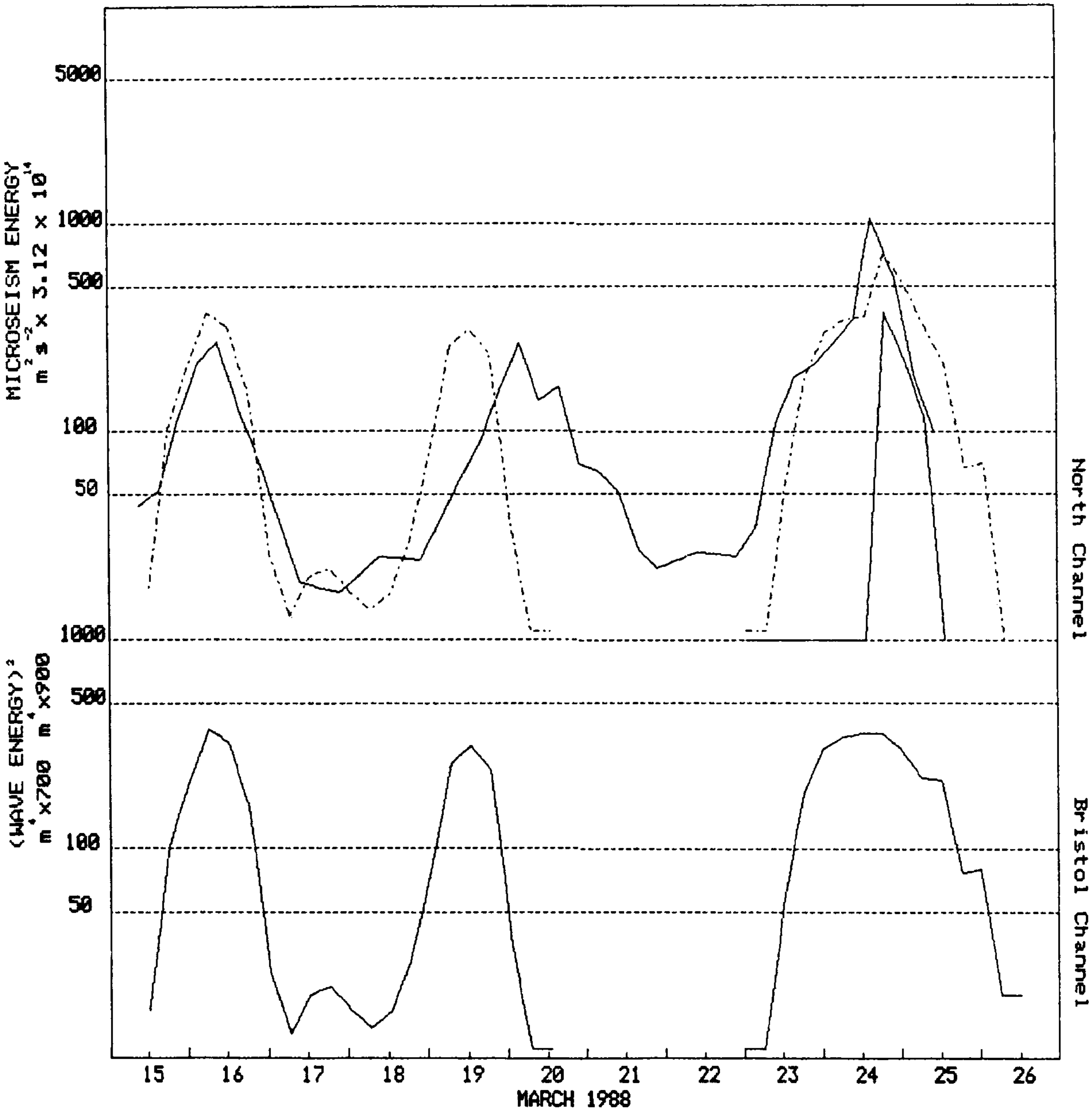

Fig. 8. Comparison of microseism energies and the square of hindcast wave energies for March 1988. (Microseism energies for 7 s and above. Wave values for 14 s and above.) The broken line in the top section shows the sum of the values in the other two sections.

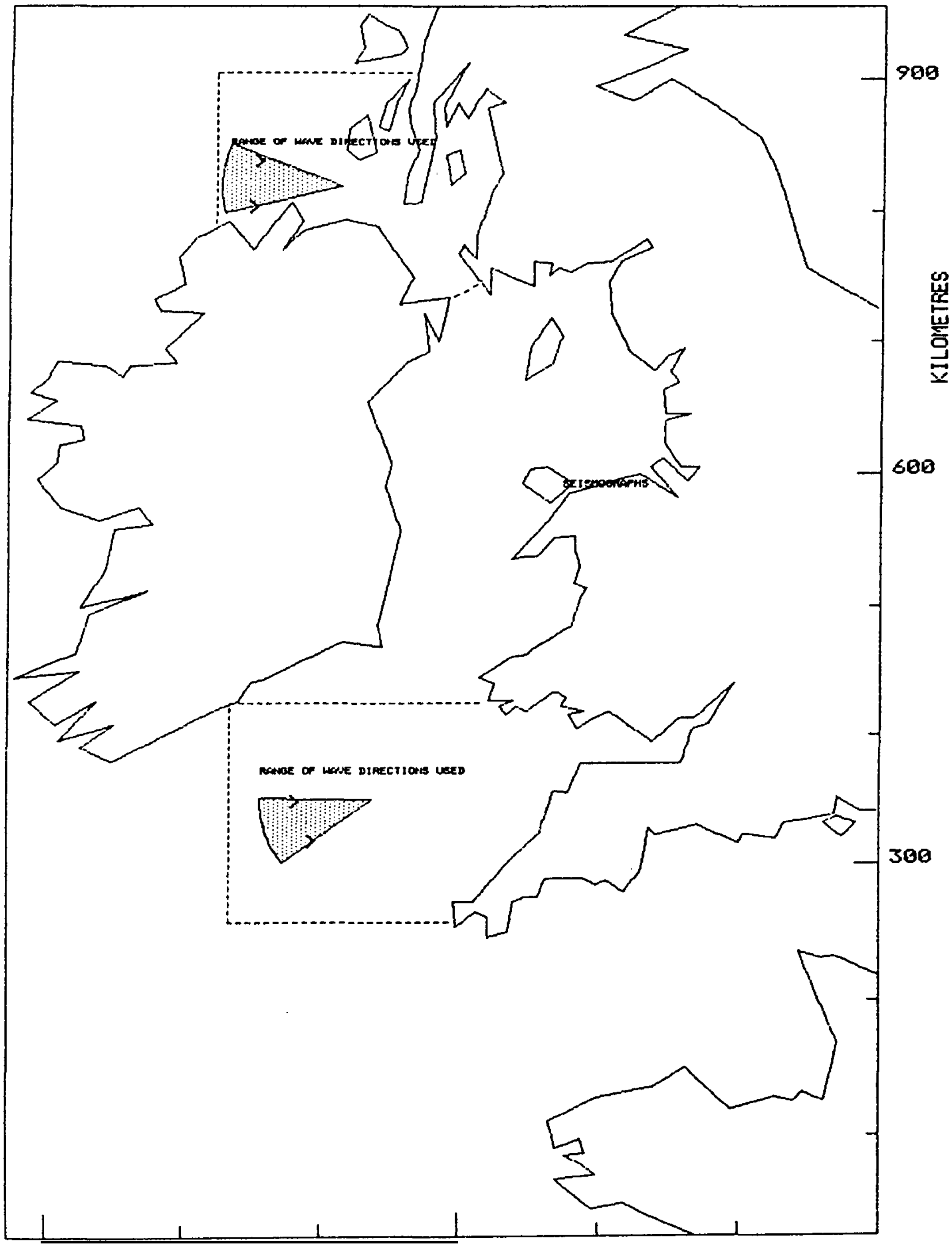

Fig. 9. Map of the western part of the British Isles showing the areas and the directions used for wave hindcasting.

a small proportion reflected back in the opposite direction to that of the main direction of incidence. A look at the map of the British Isles does, however, suggest two possible areas of reflection. One would be the passage north of Ireland to the south of the island of Islay leading to the west coast of Kintyre and then south to the North Channel. Kintyre has a relatively straight coast which faces west. This area will, for convenience, be called the North Channel. Another possible reflection area is the Bristol Channel. Although the coast is not so smooth, the general shape is that of a funnel with a closed end. Taylor (1921) has compared the effect of this area on long waves, such as those of the tide, to that of a horn on sound waves. For long waves there would be considerable reflection but this would be considerably reduced for the shorter waves we are dealing with. There is, additionally in our case, the loss of energy due to breaking. Even so there may be enough energy sent back to produce microseisms.

Figure 9 shows a map of the western part of the British Isles and the two areas are indicated bound by the coastlines and dotted lines. The sea wave energies are hindcasted for these areas. The values obtained apply to the open sea outside and no account has been taken of the effect of bottom friction and sea wave refraction inside the channels. In the case of the North Channel, the wave characteristics were calculated over two direction intervals of 22.5° from 67.5° to 112.5° and for the Bristol Channel for two direction intervals extending from 90° to 135°. These directions are measured in an anti-clockwise sense from the north and are indicated in Fig. 9. The wave energies were found for period intervals of 13.5–15.5, 15.5–17.5, and 17.5–19.5 s. These values were squared and added together. The results are shown in Figs. 3–6 and multiplied by 900 for the North Channel results and by 700 for the Bristol Channel, as described above. The hindcasts were limited to these two channels but this does not, of course, exclude the possibility of reflection from other coastal areas contributing to the activity, and it would be more realistic to regard these two areas as monitors.

The hindcast values and microseism observations agree reasonably well and are particularly good for the September case. The intensities match better, on the whole, than the times of maximum activity where the variation is between 0 and 12 h before and after. Seven incidents are considered. For four of these the energy came from the North Channel, for two it came from the Bristol Channel and one was a joint effort.

During the period covered by this part of the investigation there occurred the devastating storm of 16 October which caused great havoc in Southern England. The storm, however, was not very interesting from the point of view of microseisms. The effect of this storm reached its peak on the 16 October at 09.00 h when the energy above 7 s was 26 units and the total variance came to 1300 units with a dominating period of 4.5 s. The low value for the high period energy could be explained by the storm crossing the Atlantic and only intensifying when it reached the Bay of Biscay, which is in a divergence zone. The storm then travelled largely over land, crossing the English Channel and on to the North Sea. There were no high winds in the North Channel area and those in the Bristol Channel were not so high as those in areas further to the south-east.

3.2. Storms of December 1987–March 1988

The same techniques were used for the later storms but these were much more intensive, reaching about 700 units in December and over 2000 in January compared with the previous highest value of 400. Most of this activity came from storms approaching from the south-west and the reflecting area had to be the Bristol Channel as there was negligible activity in the North Channel during December and January. The results are shown in Figs. 5–8. The December example, shown in Fig. 5, has two distinct peaks on 25–26th and 28–29th. The correspondence is not very good in these cases, the time for the occurrence of peak activity varying by as much as 24 h in both instances and the energy at the second peak appears to be too low compared with the sea wave activity. The microseism activity was even higher in January but the agreement with the wave hindcasts is better. For the first two peaks during 2–4 January, the waves agree reasonably well for both onset times and intensities but for the second incident on 13–14 January, although the intensities match reasonably well, the times for the peak activity differ by 21 h. Agreement with both time and intensity is reasonable again for the third example on 24–25 January. In Fig. 7, there is very good agreement for 8–11 February, but it is not so good for the period 29 January–4 February. In Fig. 8, the agreement is again very good for all the examples in March.

While it is not the main object of this investigation, it was thought that the large coastal waves

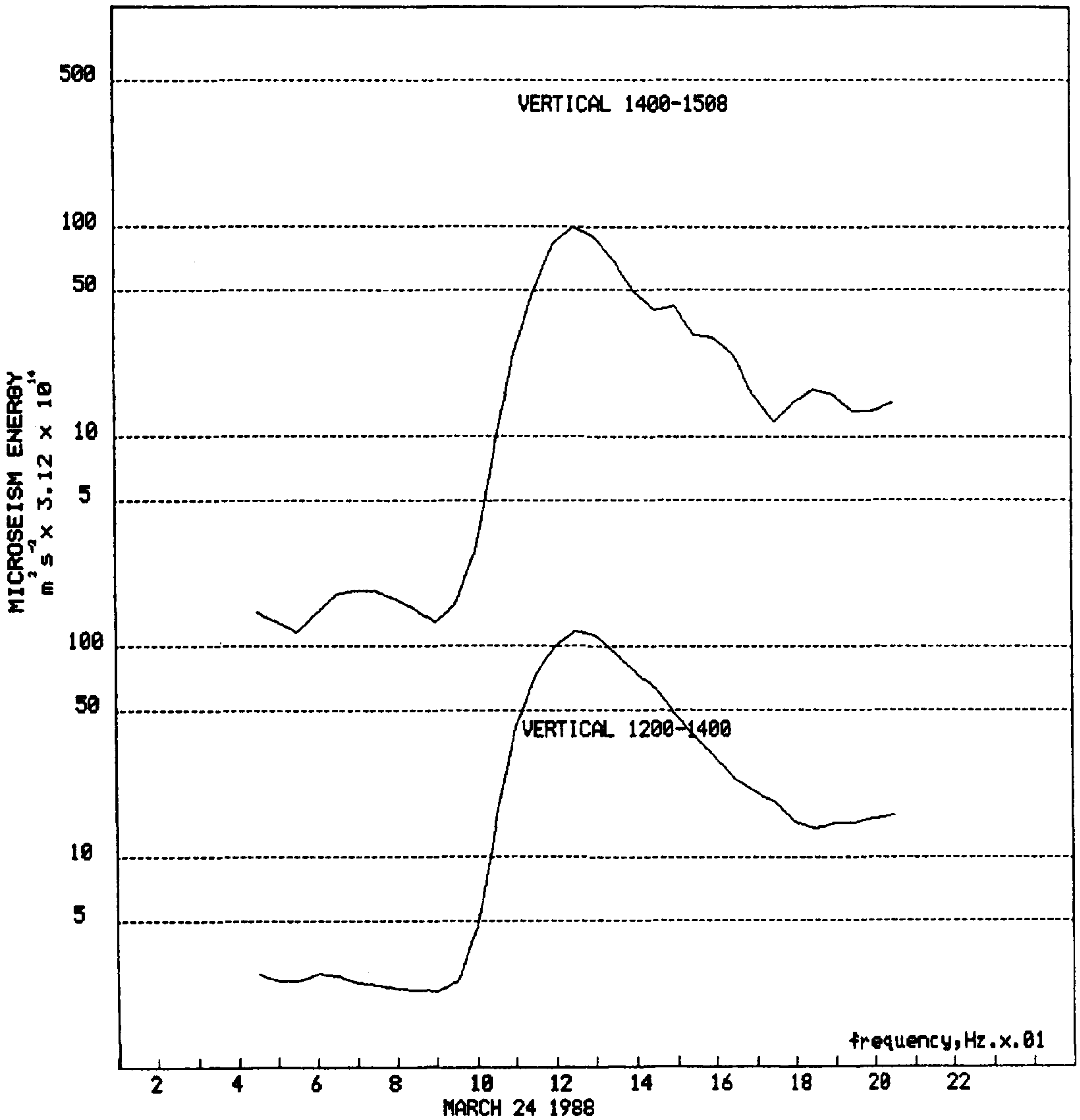

Fig. 10. Comparison of vertical spectra at different times for March 24, 1988.

during this period should produce significantly large primary microseisms. A previous investigation (Darbyshire and Okeke, 1969) had shown that the value for the primary energy was about two orders of magnitude below that of the secondary. In view of the success of Båth and Kulhánek (1987) and Gordeev (1990), it was decided to reinvestigate the matter. In the event the results confirmed the previous finding. A typical example is shown in Fig. 10.

4. Evaluation of results and comparison with theory

As already indicated, better agreement was found by multiplying the North Channel results by 900 and those for the Bristol Channel by 700 and then comparing these values directly with microseism energies in digitizer units. The corresponding results are plotted on a logarithmic scale in Fig. 11. A logarithmic scale is preferred as there

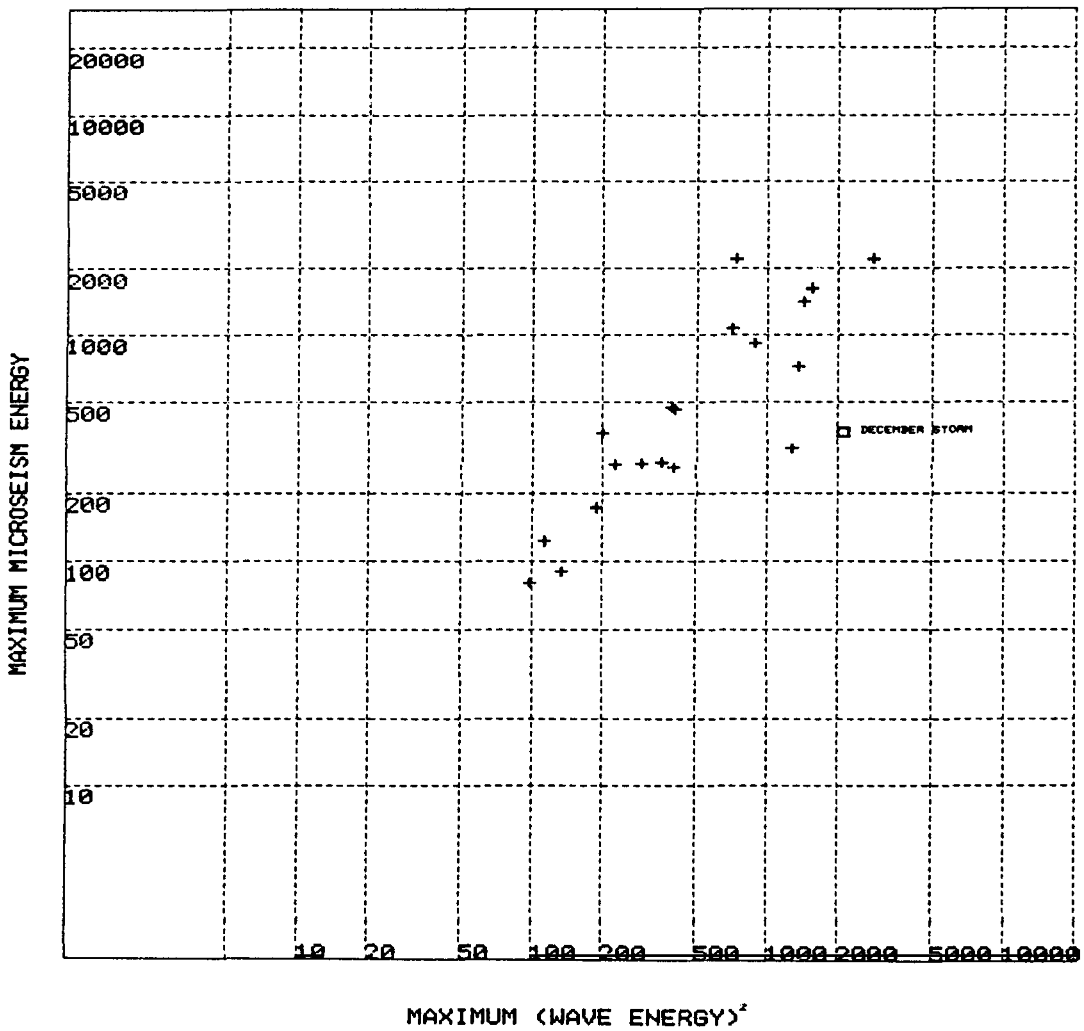

Fig. 11. Comparison of wave and microseism peak values.

TABLE 1

Errors in the prediction of the occurrence of peak times

	Difference between wave and microseism peak times
September 28–29	0
October 6–7	−3
October 8–9	−12
October 25–26	+9
November 12–13	+6
November 16–17	−12
November 19–20	−3
December 25–26	+21
December 28–29	−15
January 2–2	−3
January 13–14	−21
January 24–25	−3
January 29	0
January 30	0
February 1–2	−6
February 8–11	0
March 15–16	0
March 19–21	−15
March 24–25	0

Ignoring signs, the table indicates that 11 examples come within 0–3 hours, 3 examples within 6–9 h, 4 examples within 12–15 h and 2 examples within 18–21 h.

is a wide range of intensity from 100 to 2000 and there is a better indication of the relative error. The scatter is thus shown to be roughly the same for the whole range except for the second peak in December which is shown by a different marker.

The corresponding times for the peaks vary from −24 to +24 h, a negative value indicating that the waves preceded the microseisms. These times are shown in Table 1.

It will be appreciated that the wave energy values have been hindcasted from wind values and directions derived from weather charts published by the British Meteorological Office. The wave prediction model is liable to some error due to inaccuracies in the wind parameters and this is the more pronounced as only part of the spectral energy is hindcast and these values are then squared. Nevertheless the accuracy of the results compares favourably with that obtained in comparing hindcast wave characteristics with those of observed waves (see Elliott, 1987).

The results indicate that microseisms recorded at Menai Bridge can nearly always be related to the wave activity either in the Bristol Channel or the North Channel or both. This result would be useful not only for predicting the onset of microseisms at Menai Bridge but also for estimating the wave conditions in these two areas and indeed the whole of the western approaches from the observed microseism results.

From the work of Longuet-Higgins (1950), one can derive a formula for microseism generation from coastal reflection. It turns out to be

$$\frac{\delta^2}{2} = \frac{1.02 \times 10^{-15} (\Lambda/r)\, \Sigma\left(a_1^4 R^2/4\right)}{\mathrm{d}T \cdot \mathrm{d}\theta}$$

where $\delta^2/2$ is the variance of the microseism variation (square metres), $a_1^2/2$ is the variance of the wave variation (square metres), R is the reflection coefficient, Λ is the reflecting area (square metres), r is the distance of the reflecting area from the recording station (metres), $\mathrm{d}\theta$ is the direction range used for the calculation (radians) and $\mathrm{d}T$ is the period range used (seconds).

For the Bristol Channel case, $\Lambda = 1.04 \times 10^{10}$ m^2, $r = 2 \times 10^5$ m, $\mathrm{d}T = 2$ s, $\mathrm{d}\theta = 22.5° = 0.39$ rad. These values give

$$\frac{\delta^2}{2} = \frac{6.55 \times 10^{-11} \mathrm{R}^2 \Sigma a_1^4}{4} \tag{1}$$

For a comparison with the observations, one has to use the absolute sensitivity of the vertical seismograph. Thus

$$1 \text{ unit} = 0.0032 \times 10^{-12} \text{ m}^2 \text{ s}^{-2}$$

At a radian frequency of 1 (i.e. 2π s period), the vertical displacement is numerically equal to the vertical velocity. This is approximately the case for the microseisms considered here. With this sensitivity, the third highest peak-to-trough displacement during 24–25 January was 15 μ (velocity 15 μ s^{-1}), integrated over all the spectrum.

For the Bristol Channel, using this sensitivity, one obtains

$$\frac{\delta^2}{2} = \frac{2.24 \times 10^{-12}\, \Sigma a_1^4}{4} \tag{2}$$

This has to be compared with the theoretical result in eqn. (1). This gives $R^2 = 0.034$ and thus $R = 0.185$.

For the North Channel we replace 700 by 900 in deriving eqn. (2) and r is taken to be 2.8×10^5 m with Λ the same value as before. Then R comes out to be 0.25.

The reflecting areas are taken from the narrowest point to where the channel meets the open sea and so the areas used are less than the areas indicated in Fig. 8, which were used for wave hindcasting.

The value of the reflection coefficient seems to be too large to be physically realisable and this must be due, as mentioned above, to the effect of other sea areas such as the coasts off Ireland and Scandinavia contributing to the microseism activity. The microseism activity is not enhanced by the nature of the underlying rock as this is of Pre-Cambrian origin and is noted for its hardness.

5. Conclusions

The results show that a quantitative relationship can be established between the spectra of microseisms of 7–10 s period recorded at Menai Bridge and the squares of the components of wave spectra of 14–20 s period normal to the coast at two areas, one close to the Bristol Channel and one close to the North Channel. The relationship is reasonably good, bearing in mind the inherent errors in the data used for hindcasting.

From the theory, the relationship would require coastal reflection coefficients of about 0.15–0.25, which are higher than one would expect for these areas. It must be inferred therefore that they monitor similar wave activity in other areas.

The results also confirm a previous finding that primary microseism energy for these coastal areas is about 1/100 that of the secondary microseisms.

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