Symposium on Microseisms

Held at Arden House
Harriman, N. Y.
4-6 September 1952

SPONSORED BY THE OFFICE OF NAVAL RESEARCH, AND THE GEOPHYSICAL RESEARCH DIRECTORATE OF THE U. S. AIR FORCE

NATIONAL ACADEMY OF SCIENCES—
NATIONAL RESEARCH COUNCIL

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WASHINGTON, D. C.
DECEMBER, 1953

FOREWORD

Interest in the subject of microseisms has been growing in recent years because of their possible use in storm detection and location. Although there is no doubt that microseisms of certain periods are related to frontal and storm activity in some way, there is no general agreement as to the nature of the relationship. When microseisms are observed in connection with atmospheric activity, the coupling medium seems to be a water body. There are many and interesting explanations of the coupling mechanism and the means by which microseisms are generated. No single explanation is wholly satisfactory and perhaps none can be more satisfactory without the accumulation of more extensive and more refined data.

This situation in the field of investigation of microseims is perhaps not different from that in other fields of naturally occurring geophysical phenomena. However, because microseisms are the subject of such intensive research by a small but enthusiastic group of investigators and also because the phenomena probably have a potential application other than storm detection, it was considered that a symposium on the subject would provide a worthwhile opportunity for bringing together the existing observations in this field, for appraising their significance, and for stimulating further studies through the give and take of discussion.

The Office of Naval Research in joint effort with the Geophysical Research Division of the U. S. Air Force initiated arrangements for such a meeting. The symposium which was held at Arden House, Harriman, New York, on 4, 5, and 6 September, 1952, was organized by Dr. R. C. Gibbs of the National Research Council with the advice and assistance of an *ad hoc* group of interested scientists, chief among whom in thought and effort was Dr. Perry Byerly of the University of California at Berkeley who served also as moderator of the symposium. The National Research Council joins with the military research agencies, who provided the support for this symposium, in extending special thanks to Dr. Byerly for his outstanding contribution to the organization of the program and its direction at the symposium. The symposium brought together many of the outstanding and currently active investigators on microseisms from both this country and abroad. Those attending were:

John N. Adkins	Office of Naval Research, Washington, D. C.
Markus Bath	Meteorological Institute, Uppsala, Sweden
Perry Byerly	University of California, Berkeley
Joseph Caldwell	Army Beach Erosion Board, Washington, D. C.
Dean S. Carder	U. S. Coast and Geodetic Survey, Washington, D. C.
Frank Crowley	Air Force Cambridge Research Center, Mass.
G. E. R. Deacon	Natl. Institute of Oceanography, Teddington, England
Jacob E. Dinger	Naval Research Laboratory, Washington, D. C.
William L. Donn	Brooklyn College, Brooklyn, N. Y.
H. P. Gauvin	Air Force Cambridge Research Center, Mass.
R. C. Gibbs	National Research Council, Washington, D. C.
Marion H. Gilmore	U. S. Naval Air Station, Miami, Florida
Beno Gutenberg	Seismological Laboratory, Pasadena, Calif.
Norman A. Haskell	Air Force Cambridge Research Center, Mass.
J. Hughes	Office of Naval Research, Washington, D. C.
Columbus O'D. Iselin	Woods Hole Oceanographic Institution, Mass.
W. S. Jardetzky	Lamont Geological Observatory, Palisades, N. Y.
Gordon G. Lill	Office of Naval Research, Washington, D. C.
M. S. Longuet-Higgins	Trinity College, Cambridge, England
John Joseph Lynch, S. J.	Fordham University, New York, N. Y.
James B. Macelwane, S. J.	St. Louis University, St. Louis, Mo.
Ben S. Melton	Air Force Headquarters, Washington, D. C.
J. E. Oliver	Lamont Geological Observatory, Palisades, N. Y.
James A. Peoples	Air Force Cambridge Research Center, Mass.

Frank Press	Lamont Geological Observatory, Palisades, N. Y.
J. Emilio Ramirez, S. J.	Estacion Sismologica, Bogota, Colombia
Carl F. Romney	Geotechnical Corporation, Troy, N. Y.
J. G. Scholte	Meteorologisch Instituut, The Netherlands
Florence W. van Straten	Navy Department, Washington, D. C.
James T. Wilson	University of Michigan, Ann Arbor, Michigan
J. Lamar Worzel	Lamont Geological Observatory, Palisades, N. Y.

The papers given at the symposium are included in this volume. They stimulated much interesting discussion and speculation, some of which is also included in these proceedings. The discussions did not end at the conclusion of each day's meeting but continued after the dinner hour and far into the night. Since September, 1952, many interesting papers on microseisms have appeared in the literature. The sponsors of the symposium would like to feel that, at least in a small way, the discussions at the symposium were responsible for the continued emphasis on microseism research which is evident in the many recent scientific papers and reports on the subject.

GORDON G. LILL,
Head, Geophysics Branch,
Office of Naval Research

EDITOR'S PREFACE

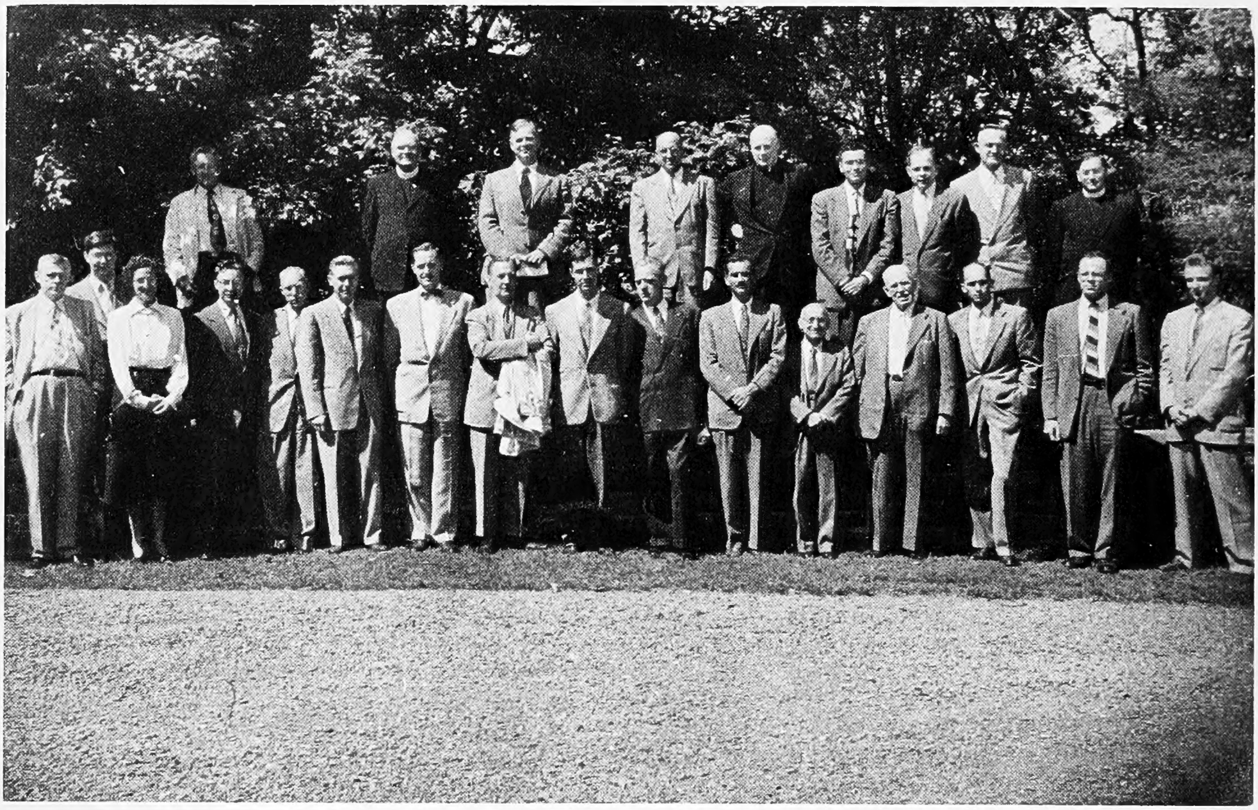
At the time of the Symposium the undersigned were given the task of editing for publication the presented papers, formal discussions, and comments from the floor. The latter, of course, posed the most serious problems, and we elected to handle them in the following manner: Each of us took rather complete notes and at the end of each half-day session these were used to select those remarks which we felt should be given in the exact words of the speaker and they were then asked to write out or dictate their remarks. Short questions, answers, or statements which seemed completely clear have been taken from our notes. For a number of these, however, the content was checked with the originator during informal discussions.

We wish to thank all of the participants for the splendid cooperation given us in compiling the informal discussions and we hope that in the process of editing and compiling we have not done injustice to any of their statements.

The papers and the formal discussions presented after each of them, have been edited as little as was consistent with the problems of printing, referencing, etc. Some consideration was given to rationalizing the various notations but this seemed unnecessary and unwise. Although we went over all of the papers jointly, most of the editing for the first half was done by Dr. Press and most of the editing for the second half by Dr. Wilson. The manuscripts came to us in good order and for this we wish to thank the authors.

Finally we wish to express our appreciation to Dr. R. C. Gibbs who has handled many of the problems and details that would normally have fallen to us.

JAMES T. WILSON
FRANK PRESS



Attendants at SYMPOSIUM ON MICROSEISMS held at Arden House, Harriman, New York, September 4-6, 1952

(front, left to right)	
Dr. Perry Byerly	University of California, Berkeley, Calif.
Dr. Florence Van Straten	Navy Department, Washington, D. C.
Mr. William Donn	Brooklyn College, Brooklyn, N. Y.
Dr. J. E. Dinger	Naval Research Laboratory, Washington, D. C.
Dr. James T. Wilson	University of Michigan, Ann Arbor, Michigan
Dr. John N. Adkins	Office of Naval Research, Washington, D. C.
Dr. W. S. Jardetzky	Lamont Geological Observatory, Palisades, N. Y.
Dr. M. S. Longuet-Higgins	Trinity College, Cambridge, England
Dr. J. G. Scholte	Meterologisch Instituut, The Netherlands
Dr. Norman A. Haskell	Air Force Cambridge Research Center, Mass.
Dr. Beno Gutenberg	Seismological Laboratory, Pasadena, Calif.
Dr. R. C. Gibbs	National Research Ccuncil, Washington, D. C.
Dr. Carl F. Romney	Geotechnical Corporation, Troy, N. Y.
Dr. Jack E. Oliver	Lamont Geological Laboratory, Palisades, N. Y.
Mr. Frank Crowley	Air Force Cambridge Research Center, Mass.
(back row, left to right)	
Dr. Marcus Bath	Meteorological Institute, Uppsala, Sweden
Dr. Ben S. Melton	Air Force Headquarters, Washington, D. C.
Rev. James B. Macelwane, S. J.	St. Louis University, St. Louis, Mo.
Dr. G. E. R. Deacon	Natl. Institute of Oceanography, Teddington, England
Dr. Marion H. Gilmore	U. S. Naval Air Station, Miami, Florida
Dr. John J. Lynch, S. J.	Fordham University, New York, N. Y.
Dr. Frank Press	Lamont Geological Observatory, Palisades, N. Y.
Dr. J. A. Peoples	Air Force Cambridge Research Center, Mass.
Dr. Dean S. Carder	U. S. Coast and Geodetic Survey, Washington, D. C.
Dr. J. E. Ramirez, S. J.	Estacion Sismologica, Bogota, Colombia

CONTENTS

MODERATOR'S COMMENTS

Perry Byerly

University of California at Berkeley

The idea that microseisms in the range of period from about 4 to 8 seconds are best correlated with marine phenomena has been widely agreed upon by seismologists for a long time. It is also generally agreed that they are surface waves—not body waves.

The theory that they are caused by surf breaking on rocky coasts was advocated by Weichert's school. Good correlations were established between high surf on the Norwegian coast and microseisms in northern Europe (and even Asia). However, the correlation was not so good with southern European stations at equal distance. This weakness was met by the assumption of geologic barriers between northern and southern Europe—called sometimes "deep seated faults." That heavy surf must impart some energy to the earth is unquestioned; that such energy would produce earth waves as nearly regular as microseisms seems unlikely considering the irregularities of coasts and of surf.

A parallel theory, advocated strongly by Banerji and Cherzi two decades ago, gave as the source of these microseisms some phenomenon accompanying storms far at sea. A number of particular cases were cited where the correlation seemed clear enough. The great objection then to this theory was physico-mathematical. Internal pressures due to water waves in deep water die off too rapidly. It was physically impossible for energy in the air to be transmitted through the ocean to its bottom.

The idea that microseisms must be Rayleigh waves is an old one. Many efforts have been made to get the direction of approach by analyzing components on this assumption. Then the tripartite method of getting direction of approach, free of the Rayleigh wave assumption, was applied simultaneously in America and Europe. This method as first applied suggested strongly that the center of storms at sea was the source of microseismic waves. However, an exhaustive pursuit of the tripartite method showed: 1) not always was the direction of approach that of the deep sea storm center, and 2) not at all tripartite set-ups did microseisms rise equally for storms at a given distance. Refuge was again taken in the assumption of geologic barriers. Their duty is to shield when microseisms are not observed accompanying a storm, and to reflect or refract when the computed direction of approach does not point to the storm center.

Although the surf theory has gone out of date there have remained those who connect their microseisms with cold fronts passing over the coasts and with storms only when they reach shallows near the coast. The surf is, however, disavowed as an intermediary.

For all theories to date one seems to have little difficulty in pointing out exceptions.

Only recently has the transfer of energy from the atmosphere over the deep ocean, through the water, to the earth become theoretically possible. If the ocean waves are standing waves then second order terms become effective. It appears that for such ocean waves of reasonable amplitudes the amplitudes in microseisms may be explained if the area covered by the water waves is reasonably large. The period of the microseisms should be half that of the ocean swell and has been so observed in England.

It is imperative that pressures on the ocean bottom at large depths be measured in some detail. We need to know whether or not standing waves under a storm are as common as microseisms. Currently without such knowledge the theory that microseisms originate under the deep sea stands hedged and inviolate. If there is no increase of microseisms with a storm, then there were perforce no standing waves—if they increase at some stations and not at others, the latter were protected by a barrier. If they appear to approach from the wrong direction at a station, then they were refracted or reflected at a barrier on their way from storm to station.

When two ideas persist for as long as the above (1) correlation of microseisms with deep sea phenomenon (2) correlation of microseisms with coastal phenomenon, one not violently interested may suspect there is something to both of them. Miss van Straten's paper in this symposium is admirable in bringing this out.

The purpose of a symposium such as this is primarily to broaden the minds of the members—not to offer each an opportunity to convert the others. As one member remarked, after many years observing microseisms at one station one may learn pretty well with what to correlate them. But this does not mean that he would be equally successful in another geographic locality. It would even appear that microseisms in Europe and America are not so comparable as we might expect.

It is to be hoped that each member of the conference will go home to review his own data with new possibilities in mind.

University of California at Berkeley
Berkeley, California

SKETCH OF THE HISTORY OF MICROSEISMOLOGY

J. B. Macelwane, S.J.
St. Louis University

John Milne, who is known as the father of modern seismology, published a paper [1883-84] in which he made the following statement. "The father of microseismical research seems to have been Father Timoteo Bertelli of Florence. In 1870 Father Bertelli suspended a pendulum in a cellar which he observed with a telescope . . . In 1873 Bertelli by means of microscopes fixed in several azimuths made 5,500 observations on free pendulums. He also made observations on reflections from the surface of mecury." Bertelli [1875] would seem to deserve the title given him by Milne because he appears to have been the first to undertake systematic studies of microseisms and because his publications moved so wide a circle of investigators to undertake research on microseisms and because he gave this name to the phenomenon. Actually the early observations of Bertelli referred to by Milne extended over the three years, 1869 to 1872. By 1874 daily observations were made at five stations in Italy and by 1884 at thirty.

Bertelli was, of course, not the first to observe that the surface of the earth is in a state of more or less continuous agitation. Astronomers and geodesists using a pool of mercury as a reference level found the surface of the mercury rarely quiet. George H. and Horace Darwin [1881], who had set up elaborate apparatus at Cambridge, England, to observe lunar tides with a magnification of 50,000 times, found such incessant ground vibrations that the experiment had to be abandoned. However, they seem to have made no attempt to study the vibrations as such.

John Milne in Japan interested himself very early in the observation of microseisms. In a paper read before the Seismological Society of Japan, Milne [1881] described a series of experiments he had made to determine the characteristics of microseisms, including the use, in February, 1880, of rotating mirrors with a magnification of approximately 250 times. He concluded: "From these results it would at first sight appear that the ground in Tokyo is almost constantly in a state of tremor." Two years later in a paper read before the Seismological Society of Japan Milne [1883] described the observation of earth vibrations in Italy, France and England and concluded: "Like observations have been made in Japan and it does not seem improbable that after farther experiments have been carried out we shall be brought to the conclusion that the surface of the whole globe is affected by similar microseismical disturbances." Milne noted the succession of intervals of comparative quiet followed by periods of hours or days of large amplitude disturbance, and he introduced the term "microseismic storm" to describe the latter. He presented a tabulation [Milne 1887] of a long series of observations of the north-south and east-west components of microseisms together with earthquakes, barometric heights, wind velocities and their gradients. He said: "In conclusion, so far as my observations have gone in Japan, it appears that the majority of earth tremors are movements produced by the action of the wind upon the surface of the earth and that these may often be propagated to distant places where wind disturbances have not occurred."

In Germany during the years 1892-1894 E. von Rebeur-Paschwitz was engaged in the observation of earth tides by means of horizontal pendulums. In the report [1895a] on his observations at Strasbourg a section [1895b] entitled "Die Mikroseismische Bewegung" was devoted to his observations on microseisms. These observations were continued at Strasbourg by Ehlert [1898].

With the beginning of the Twentieth Century, interest in microseisms had become general. In Japan F. Omori [1901] summed up the results of his observations in these words. "The chief characteristics of these movements, as observed in Tokyo, are the following:—

1. Pulsatory oscillations occur more frequently in winter than in summer.

2. Pulsatory oscillations continue generally for several days, there being no dependence of the frequency on the time of day.

3. The average period remains generally constant for several hours, not depending much on the amplitude.

4. The average period varies but little, the least value being 3.4 s. and the greatest value 8.0 s.

5. The direction of motion changes constantly, and each horizontal component shows a series of alternations of maximum and minimum groups; . . . . the average period being 6.8 s. . . . . "

Subsequently Omori [1903] wrote: "It thus seems probable that the pulsatory oscillations are essentially composed of two series of vibrations, whose periods are respectively about 4 sec. and 8 sec.; large pulsatory movements which are caused by very deep cyclones having generally the 8 seconds period."

Ten years later Omori [1913] reported a comparison of the microseisms recorded at Tokyo with those recorded at Hitotsubashi about two kilometers distant in which he found it impossible to identify individual vibrations on the two sets of records. This negative result, which may have been due to instrumental deficiencies probably led Honda [ ] and other later Japanese investigators to adopt the view that microseisms are stationary waves set up at the locality of the observing station.

At the meetings of the International Seismological conferences and later of the International Seismological Association, microseisms were of interest from the first and a standing committee for the study of microseisms was soon appointed. It was at the second International conference held at Strasbourg in 1903 that Wiechert [1905] proposed his well-known surf theory. Laska [1902] on the other hand correlated maximum amplitudes of microseisms with steepest barometric gradients on the basis of the records at Lemberg. Klotz [1909] presented a report at the Zermatt meeting of the International Seismological Association in 1908 in which he said among other things: "5. A well-marked Low sweeping up the Atlantic Coast from Florida to Newfoundland is almost always accompanied by marked microseisms. 6. Microseisms are but slightly, if at all, influenced by the movements of Lows across the continent."

It must be remembered that in nearly all of these early investigations attention was focussed on the band of microseismic frequencies which lies between one-fourth and one-eighth herz or those of still lower frequency because these microseisms were so prominent on the records of the seismographs then in use. The study of microseisms of higher frequency became possible much later with the introduction of more suitable types of instrumentation.

Gutenberg [1910] published his doctoral dissertation at Goettingen in which he presented extensive data which he interpreted as supporting Wiechert's surf theory and he correlated the microseisms of four to ten seconds in Germany with surf on the southern part of the west coast of Norway. This conclusion he supported vigorously through the succeeding years [Gutenberg 1912, 1921, 1924, 1927, 1928, 1931, 1936]. More recently he has modified and broadened his views in accord with his excellent later researches.

In addition to the surf theory, three other main theories or groups of theories have been advanced to explain microseisms. These are first, theories of local origin, meteorological or geological, at or near the recording station, secondly, theories of thermometric or barometric gradients travelling over continental areas; and thirdly, theories connected with storms or storm waves at sea.

The last named theory casually proposed by Bertelli [1878] and by Omori [1903] and formally reported by Klotz [1909] as a result of his observations at Ottawa, and specifically formulated and defended by Gherzi [1923, 1924, 1926a, 1926b, 1928, 1930, 1937], by Banerji [1929, 1930, 1935] and by Zanon [1936, 1938] in the nineteen twenties and thirties, has come to occupy the center of the modern microseismic stage.

Most of these studies were made by temporal correlation, as were a number of more recent investigations. At the second meeting of the Eastern Section of the Seismological Society of America held in Ottawa, Canada, a committee was appointed to correlate and map microseismic amplitudes recorded at all seismographic stations in the United States and Canada. A more ambitious program of correlation has been inaugurated by the Association of Seismology of the International Geodetic and Geophysical Union in 1952 involving simultaneous observations of microseisms in the whole world.

However, many seismologists have felt that temporal correlation of amplitudes and periods from place to place is not sufficient to distinguish between possible sources and have sought to determine the bearing of the origin through measurements of the direction of propagation of the microseismic waves. Some have attempted to do this by vector methods, using the separately recorded components on the assumption that microseisms are Rayleigh waves. Others beginning with Omori [1913] and Hecker [1915] attempted to find the direction independently of any assumption by means of the time interval between arrivals at closely spaced recording stations but failed because of inadequate instrumentation. Shaw [1922] in the years 1918 to 1922 obtained some evidence by this method that microseisms arrived at West Bromwich, England, from the northwest. Krug [1937] at Goettingen in 1936-1937 used three individually timed portable seismographs at the corners of an isoceles triangle with limited success. But the following year 1937 Trommsdorff [1939] and Ramirez [1940], independently using the simultaneously timed tripartite station method, found that the bearings determined from arrival times indicated the position of a low pressure area over the sea.

The Ramirez method was applied to the detection and tracking of hurricanes and typhoons by the United States Naval Aerological Service under Gilmore in 1944 [Gilmore 1946] and the following years. Anomalies in the indicated bearings, laid in part to refraction, have prevented the operational use of the method and have inspired critical investigations by Donn, VanStraaten, Kammer and Dinger, and others which still leaves the problem of the generating mechanism unsolved. Imbo [1931] then director of the seismological station at Catania, Sicily, published in 1930 a comparison of the periods of microseisms and of waves in the Mediterranean Sea and showed that the sea wave periods were twice as long. A similar result was obtained in 1947 by Deacon [1947] and in 1950 by Darbyshire [1950]. This relationship inspired a theoretical investigation by Longuet-Higgins [1950] which indicated that the second order terms in the equations of the pressure field produced by standing waves at sea integrated over a sufficiently large area could account for microseisms of half the period of the standing waves.

An example of the complexities involved in the problem of determining the cause or causes of microseisms of the types under discussion was brought out in the discussions at the "Study Week on Microseisms" which met at Rome in November, 1951, under the auspices of the Pontifical Academy of Sciences. While the "group" or "beat" form of the storm microseisms was a characteristic taken for granted by the participants from the Western Atlantic and Western Pacific stations it seemed to be less familiar to the European participants at their stations. Caloi in Rome presented arguments and observational data in favor of marine barometric gradients as a source.

Coming now to other bands of frequencies, research is still in its infancy on microseisms of two to three seconds period yet certain facts have been ascertained concerning them. In the case of "group" microseisms of four to eight seconds period, Klotz, Gherzi, Ramirez, Donn and others observed that the amplitudes rise rapidly and take on their characteristic form when a low pressure area leaves the continent and enters the ocean and that the amplitudes fall rapidly and the waves lose their regular form when the storm leaves the water and enters the land. This seems not to be the case with the microseisms of two to three seconds period. At Corpus Christi, Texas, they were found by Jennemann to arrive from the north at the tripartite station operated there by the United States Navy Aerological Service, and hence must have originated and been propagated on the North American continent. Father Lynch of Fordham, on contract with the Office of Naval Research, found that in southern New York state they arrive from the west and in North Carolina from the northwest.

Still less is known about the microseisms of frequency two to three which are widely observed in the records of open time-scale, short period seismographs. Research on those microseisms is in progress at Saint Louis on contract with the Office of Naval Research but has not progressed far enough to warrant any conclusions.

Geophysical prospectors are familiar with microseisms of still higher frequency which form unwelcome background noise in their operations. But, as far as the writer is aware, no systematic study has been published concerning them.

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Discussion

B. GUTENBERG

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The paper by Father Macelwane gives an excellent summary. However, the reported statement by European seismologists, that beats are not observed in European microseisms, does not correspond to the facts and is probably caused by an incorrect translation of the expression "beats" by scientists from Central Europe. "Schwebungen" (beats) are discussed in several European and especially German publications on microseisms (see e.g. the author's Handbuch der Geophysik, vol. 4, p. 282). They are also frequently found in microseisms recorded at Pasadena and other stations near the Pacific coast. There is no evidence that beats or groups in the regular type of microseisms with periods of 4-10 seconds are restricted to certain areas.

A few words may be added concerning the history of the division of microseisms into different types. The first detailed description of such types was given by Hecker [1906]. He distinguished four kinds of microseisms depending on whether the period was less than 4, about 7, about 30 seconds, or of the order of 1 minute. The first type is usually caused by local disturbances. Gutenberg [1910, 1912] suggested the division of microseisms into two major groups: a) microseisms caused by local effects and b) by distant sources of energy. Group a) included microseisms from traffic and industry with periods of less than 2 seconds, from local wind storms and from local surf. Group b) consisted of microseisms from ocean waves (surf) with periods from 3-10 seconds, of microseisms with periods of about ½ minute (from distant wind storms?—this type may be spurious and these "microseisms" may have been caused by air currents in the instrument vault), and of microseisms with periods of 1 or more minutes during periods of frost near the station. Later, additional types were reported [Gutenberg and Andrews 1951, Gutenberg 1951]. It is very important to recognize the type which is recorded in a given instance. Several types from apparently different causes have about the same periods, but differ in their appearance.

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Discussion

JAMES TINLEY WILSON

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Father Macelwane has given an excellent summary of the early history of the study of microseisms and of the steps which have led to our present state of knowledge. I do not feel it necessary to comment in detail on Father Macelwane's remarks, rather I would prefer to present some further comments. Most, if not all, of this information is already known to Father Macelwane and to the rest of you but it constitutes more of the data that we must keep in mind when considering the subject.

The fact that the study of microseisms is not a closed book is evidenced by this conference. If further evidence be needed, we have but to remember that there is not complete agreement among seismologists as to either the nature or the origin of microseisms, although they have been subject to study for some seventy-five years. Certain aspects of the subject are, of course, fairly clear. For example, Omori's observations in 1901 quoted by Father Macelwane would not look too out of place in a modern publication. A seismologist of the present day might study his seismograms (and the literature) for some time and not do much better.

The problem is such that we cannot afford to neglect any possibilities and must as far as possible consider all of the observations. Microseisms being almost always with us, we usually have more of the latter than we know what to do with.

The items I am going to mention come for the most part under the headings of "nature" or "origin." As Father Macelwane has done, I will limit myself primarily to those microseisms which have periods in the range of three to twelve seconds.

Although a case has been made on both observational and theoretical grounds for microseisms being stationary waves of some sort of "free vibration," the present concensus of opinion seems to be that they are traveling waves. The usual assumption has been that they are surface waves and more specifically of the Rayleigh type. I cannot help but feel that the latter assumption is based almost solely on the fact that they have a vertical component. The suggestion has been made, of course, that they are a mixture of Love and Rayleigh waves. In this connection it might be pointed out that attempts to obtain the direction of approach by comparing the phase of horizontal and vertical components on the assumption that the microseisms are Rayleigh waves has not led to as good results as the well known tripartite method. Various attempts have been made to compare the observed periods and amplitudes with those expected from Rayleigh waves in certain types of crustal structures. The results have usually been tantalizing but not conclusive.

While on the subject of the nature of microseisms, mention might be made of factors also related to origin. As noted before, it has been a common observation that microseisms are larger in the winter than in the summer and, in at least a general way, I think it can be said that they are larger in coastal regions than in the continental interiors. Further, all seismogram borrowing seismologists know that there are certain stations, for example Perth, where the microseisms are a constant nuisance. This variation with season and geography seems to have fathered some of the theories of microseismic origin. Wiechert's surf theory and the oceanic storm theories of Gherzi and

Banerji might be cases in point. I do not pass judgment on these theories here, but merely wish to indicate that there is a proximity factor in their historical development.

As Father Macelwane has pointed out, there are three general theories for the origin of microseisms (1) local meterological or geological conditions, (2) meterological gradients over continental areas, and (3) meterological conditions at sea. There seems to be almost universal agreement that the source is meteorological.

Two problems have been considered at length in this connection; first, what meteorological conditions supply the energy to the ground, and secondly, how is the energy transferred. The source of the energy is usually sought in "active" meteorological situations such as fronts or low pressure areas. Water bodies are frequently considered as a coupling medium to pass the energy into the ground. As mentioned by Father Macelwane, a long list of seismologists and meteorologists have formulated theories or portions of theories, but no one of them seems to explain all of the microseisms all of the time. Some of them fool some of the microseisms some of the time, and one might even say that some of them fool some of the microseisms *all* of the time.

I think a fact that is sometimes forgotten in trying to assess the various theories is that all of them must, of mathematical necessity, deal with rather idealized cases that may fit a given piece of geography fairly well, but will fail rather badly to match in other parts of the world. In this connection, I think it is an historical fact, and so worth mentioning in this historical discussion, that after any seismologist has tended the same station for a few decades he begins to know his own microseisms fairly well and, if he has become interested in them, he is likely to be able to relate his microseisms rather consistently to certain weather conditions, but he still may not be able to formulate a theory that will stand up for all the other stations in the world.

In studying both the nature and origin of microseisms, seismologists have made considerable use of large masses of data. Some of this has been done out of necessity in an attempt to extract useful microseism data from seismograms run for the routine recording of earthquakes. In other cases it has been in an attempt to obtain correlations between microseismic activity and weather conditions. To me, a very interesting transition has taken place in the past ten or fifteen years with more emphasis now on the study of individual microseismic storms and, in many cases, with the aid of seismographs specifically designed and operated for recording them. To some this might seem like a tree-to-tree examination before we have seen the forest, but having been unable for so long to get a clear picture of the forest, it may well be the proper method.

Discussion from the Floor

Bath. Dr. Gutenberg emphasized the parallel behavior of microseisms in northern Europe. In a comparison of microseisms at Uppsala, Bergen, and Copenhagen, this result was confirmed as far as the broad outlines are concerned, but there were significant deviations in detail. These could be explained by the hypothesis of an origin along a line source, but were not in accord with the hypothesis of a point source.

TRIPARTITE STATIONS AND DIRECTION OF APPROACH OF MICROSEISMS

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Seismographs set at the corners of a triangle were used as far back as 1884 by Milne and Japanese seismologists in Tokyo for the determination of the direction of approach of certain earthquake waves with various and generally unfavorable results [Imamura, 1902].

This three station distribution, generally consisting of 3 vaults, one at each corner of a triangle, each equipped with seismographs and known today as a tripartite station, began some 45 years ago to be used in an attempt to determine the velocity and direction of microseisms.

A microseismic wave originating from some distant source and traveling along the earth's surface would reach one of the corners of the triangle first and the other two corners subsequently. By measuring the time interval between the arrival of this microseismic wave at the 3 vaults the direction of travel and the velocity of the wave can be found.

The first trial (at least between two stations) was probably made by Omori in Japan in February, 1908, between the station of Hongo and Hitotsubashi, the mutual distance being 2.29 kms. but he "found it impossible to identify the individual vibrations at the two places" [Omori, 1909, 1913].

Hecker [1915] made a trial at Strasbourg in 1915. He used the NS and EW component of the observatory and placed a NS component first 0.58 km. due north of the central station, and later located it 2.4 km. northeast of the observatory.

In the first case, both stations receiving the time signals from the same clock, the waves at the instrument located 0.58 km. south, arrived now earlier and now later than at the other station (in one instance there was a maximum change of 1.0 second, during an interval of 1¼ minutes). In the second case, at a distance of 2.4 km. and with two different clocks marking the time signals, there were greater differences in the arrival times. The experiment was discontinued and "only the observations of one day were the ones that could be used and even on this particular day microseisms were not so strong.

Shaw [1922] reported at the Rome meeting of the first conference of the section of Seismology of the International Union of Geodesy and Geophysics upon his work in connection with microseisms and stated that at West Bromwich in 1918 during some experiments with Milne-Shaw seismographs situated 20 meters apart and in different building he noticed that the recorded microseisms were identical. During the spring 1919, and 1920, he demonstrated that at a distance of 3 km. each microseismic wave was still similar on each record. In the early part of 1921 an attempt was made to compare three stations about 16 km. apart. At this distance the waves were quite different and it was impossible to identify them for comparison. In the first months of the year 1922 two stations were arranged 4 km. apart and on different directions to the stations used in 1919 and 1920. At this distance the waves were again identified without difficulty. In the 1919 and 1920 experiments the differences in time of reception at the two stations were 0.83 seconds. The method gave some evidence that microseisms came from a north-westerly direction to west Bromwich.

From January to March 1927, Nasu and Kishinouye temporarily set three horizontal pendulum seismographs near the Seismological Institute of the University of Tokyo to study the phase relation of microseisms at different places. The three stations, with that of the Institute called B, made a set of four, located at the following distances in meters:

	A	B	C
B	520	. . .	920
C	920	430	. . .
D	1090	610	520

These investigators concluded, " . . . the observations are not sufficient to yield definite results, for they were obtained from records of NS component only." Furthermore, "it was very hard to find corresponding minute marks on records. So the comparison of records of four stations was drawn only in several cases . . . The variations of amplitude like beats may be due to oscillations of different period of a . . . land-block . . . bounded by vertical planes of discontinuity" [Kishinouye, 1935].

Krug [1937] undertook during March, May, September and October, 1936, and in January, 1937, the determination of the velocity of propagation and direction of microseisms at Göttingen. He used four portable horizontal seismographs of large magnification (over 6,000) for two movable stations in each of which he had a NS and an EW component. As a third station he used the Wiechert pendulum of the Göttingen Geophysical Institute (mass 1.200 kg.) with a mechanical magnification of 140 but increased by means of an optical system to 4,000. The period of the pendulum was about 11 seconds. The stations were located in the form of an isosceles triangle, with the two equal sides about 1,4000 meters. It is not clear from the article how the time signals optically marked on the records were received at each station, but it seems that no direct line or radio signals with automatic registration were used. He says: "Taglich wurde zweimal während des Naueners Zeitzeichens um 1 und um 13 Uhr M. E. Z. je 5 minuten gemessen und alle Zeitsignale als Gleichzeitigkeitsmarken optisch sufgezeichnet."

For waves of 4 to 8 seconds Krug found a velocity of 1100-200 meters per second which seems much too low. "Ob dieser unerwartet niedrige Wert zur Ausbreitung der Energie oder zur Ausbreitung einer bestimmten phase einer kombinierten Welle gehört, konnte nicht entschieden werden." An average of 80 per cent of all the readings gave a direction N 63° E ± 20°. He also found some correlation between the barometric depression on the Norwegian Coast and the intensity of movement on the geological conditions of the place of observation.

In 1937 Rev. James B. Macelwane, S.J., suggested to the author, then a graduate student at Saint Louis University, Saint Louis, Mo., an experimental determination of propagation and origin of group microseisms by means of a simultaneously timed tripartite station. This station array consisted of four Macelwane-Sprengnether seismographs: 2 EW components, one under the Saint Louis University Gymnasium, and one 6.4 kms. almost due West at Washington University. 2 NS components, one under the Saint Louis University Gymnasium and one 6.3 kms. almost due South, at Maryville College. Each component was recording identical time signals sent over leased wire every few seconds. The accurate and identical timing system, the instrumental homogeneity, and its special design for recording microseisms of periods between 3 and 9 seconds were characteristics of this new tripartite station.

The results were very satisfactory in demonstrating beyond doubt that microseismic waves are traveling and not stationary waves, that their direction of propagation can be measured, that the determination of the direction of arrival at Saint Louis of these waves in all observed cases indicated that they came from tropical cyclones over the ocean; and that the bearing followed exactly the movements of the low pressure center and not the location of surf on the rocky coasts.

The equations used for calculation of the direction of propagation of the microseisms were particular solutions suited to the case of a right triangle. Macelwane and Gilmore introduced equations valid for any tripartite station and the latter shortened the distances between the stations to about 600 meters using an isosceles triangle as the general shape of the tripartite stations for tracking hurricanes.

Simultaneously with the work at Saint Louis University, another tripartite station was being established by F. Trommdorff, at the University of Göttingen in Germany, leading to similar results.

According to Macelwane [1946] Trommdorff was not sure of the meaning of his results. Macelwane states, "Likewise the direction of propagation determined from the arrival times indicates the position of the low pressure area. This does not tell us whether it is the storm low itself or the surf caused by it on the coast which is the cause of the microseisms."

As a result of the experimental investigation at Saint Louis University, the Naval Aerological Service under the Guidance of Captain H. T. Orville, U. S. N., became interested in microseismic research because it presented the "possibility of saving lives, money and property by the ability of seismographs to determine the presence of embryonic tropical storm before there are indications of the geneses of those storms by other means and successfully track these storms without risking lives and property by weather reconnaissance near the eye of an active storm" (Naval Report, 1947a).

Accordingly the first naval tripartite station was installed at the Guantanamo Bay, Cuba. In September 1944, this was reported in complete operation and a few months later encouraging reports were given out with "some substantiation of successful tracking of the seasons hurricanes." To direct the establishment of this station and to act as officer in charge of the project, Commander M. H. Gilmore, U. S. NR, was appointed, an officer with many years of experience in geophysics and seismology.

A new improvement was made at the Guantanamo tripartite station which was later generalized to other stations. It consisted in connecting the instruments of the outlying vaults with the main vault by a lead shielded cable. Thus the recording of the three instruments could be made accurately on the tripple drum at the main vault, instead of recording at each vault, with a resultant saving of considerable time.

In December 1944 it was decided that additional tripartite stations should be established, one at the Naval Air Station of Richmond, Florida, (later moved to Opalocka) and one at the Naval Operating Base of Roosevelt Roads, Puerto Rico. The two new stations were in complete operating condition by August 1945. Thus a new technique was introduced as a check by triangulation on the origin of microseismic waves by means of the three stations of the tripartite type. The bearings and crossbearings obtained in 1945 with these stations seemed convincing, and the results were published in detail by Gilmore, Macelwane and others.

In 1946 a tripartite station was built at Corpus Christi, Texas, (First report came out 17 August), and a single station at Antigua (was ready for operation in October). In 1947 a single station was built at Trinidad and a tripartite station at Swan Island (February). At present tripartite or single stations include Bermuda, Whiting Field Jacksonville and Cherry Point. A tripartite station was established in the Pacific area, on Guam in the summer 1947, and single stations at Okinawa and at other points. The station installed at Corpus Christi on loose sand was discontinued due to the unsuitable ground foundation. Trinidad and Antigua were also discontinued; Roosevelt Roads and Guantanamo Bay are now single stations.

Finally in 1949 a special tripartite station was installed on the grounds of the Florissant seismograph station, for the purpose of studying the nature and origin of the 0.2-0.5 second period microseisms by means of special capacity seismographs developed at Saint Louis University under Dr. Joseph Volk and Dr. Florence Robertson. One corner of the triangle was established at the Florissant seismic vault, the second corner was approximately 600 feet due northeast and the third corner was about 800 feet due northwest of the Florissant vault.

Regarding the direction of approach of microseisms to a tripartite station, this was the view point of Naval Aerology in 1947: "The direction given by a tripartite station seems to be accurate frequently within 10 or 20 degrees, but sometimes greater deviations are to be expected due to geological idiosyncrasies, error of the observers and the fact that the source of the microseisms is not necessarily in the center of the hurricane and even may be an extended area with a different starting point of the longest wave for each station at given moment" (Navy report 1947, b). It was also remarked that for a particular storm "all the bearings obtained are through some portion of the storm. Some of them perhaps lead the storm center, while most of them lag behind."

It has been my experience also that in a particular storm the bearings obtained may differ widely, but averaging several readings leads to a truer indication of the direction of approach.

According to various authors this variation of the intervals of arrival of waves at the corners of a tripartite station may be accounted for as due to the possibility that the energy source of microseisms waves may not be at a point but that it is rather a wide source varying its maximum around the center of the storm, or even that it is due to interference from various other sources such as a new system of microseismic disturbances coming from a different direction.

More recently Gilmore has found evidence to state that microseisms "may not always be propagated outward through the earth's crust from the center of the storm in straight lines because of refraction and reflection," and that in order to track storms with microseismic cross bearings from tripartite stations accurately, charts are needed showing all refraction around each station. Hence, a new method has been developed by the Navy Microseismic Research, called the micro-ratio technique of storm tracking, which according to its author Gilmore, may permit very accurate tracking of storms that are far from land.

Summing up, the tripartite station system, either in the form of a tripartite station for determining the bearing of the source or in the form of a set or sets of tripartite stations for locating and following the origin of the source of microseisms has been used and still can be used successfully, provided the stations have a suitable ground foundation, a proper distance between themselves, and as much as possible, instrumental homogeneity, accurate and identical time system, and simultaneous recording on a triple drum.

The direction of approach of microseisms can be rather accurately determined by each tripartite station by averaging of several readings. The crossbearings from tripartite stations can locate energy sources and track them continuously from hour to hour and detect them days before they can be detected by any other method.

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Discussion

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The tripartite station is a useful tool for studying the direction of approach of microseisms. However, in any discussion of a tripartite station it is essential to point out the limitations which must be taken into consideration if one is to obtain the greatest usefulness from the tripartite network.

First, one must recognize the tolerance imposed on the accuracy of computed bearings when taking into account the maximum accuracy of the measurements obtained from a given set of tripartite instruments. The size and shape of the tripartite network, the speed of the paper, and the sharpness of the trace all affect the maximum accuracy that can be achieved, assuming that well-formed microseisms are being propagated across the tripartite network. In fact, the accuracy varies with the direction of approach to a given network. A triangle having one large oblique angle will have a considerably greater accuracy if the microseisms approach along a direction parallel to the long side than will be the case if the approach is at right angles to this long side. An equilateral triangle will come the nearest to giving equal accuracies in all directions. Let us take one typical network to illustrate the instrumentation errors one might encounter. Assume an equilateral triangle having sides of 4,000 feet, a chart recording speed of 150.00 cm/min, and a microseismic wave traveling across the network at 8,000 ft/sec in a direction which bisects one of the angles. If one can superimpose the traces with an accuracy of ± 1 mm, the maximum errors that can enter the bearing computation will be ± 11°; a spread of 22°. This example I believe approaches the ultimate in instrumental accuracy; in practice the errors may be considerably larger.

In view of the fact that in the computation of a bearing one is in effect measuring the relative phase differences between the three recordings, it is highly essential that the three seismographs do not introduce any phase shifts into the record; or at least that the three seismographs introduce identical phase shifts. This factor demands special attention if any component of the system, such as the pendelum or galvanometer, has a natural period in the range of the periods of microseisms being recorded. Strict attention must be paid to proper damping of all such components. So far as phase shift is concerned a seismometer working into an electronic amplifier which in turn actuates the recording mechanism is to be preferred over a seismometer working directly into a recording galvanometer. In the former case there is no reaction of recording element on the seismometer to complicate phase relations.

A serious limitation of the tripartite station arises from the very nature of the microseisms. This limitation has been pointed out in the literature by a number of writers, among them being Trommsdorff [1939], Bungers [1939], Leet [1949], Donn & Blaik [1952], and Kammer & Dinger [1951]. This limitation arises from the observation that, in general, microseisms crossing a tripartite station do not consist of a single coherent wave train but rather are the composite of several wave trains which may differ in direction, period, and wave type. One obtains evidence that the microseisms do not consist of a coherent wave when the separation of the three seismometers is large (several miles), for in this case it is often difficult to identify the corresponding portions of the three records. The lack of coherency is also illustrated by Leet [1949] in a five-minute sample record made by a three-component seismograph. This sample record shows a mixture of Love and Rayleigh waves. Leet suggests using a three-component registration at each corner of the triangle so that the type of wave motion in a given interval of the record can be determined. A complete record of this sort will possibly permit one to select wisely the portion of the record to be used for bearing computation.

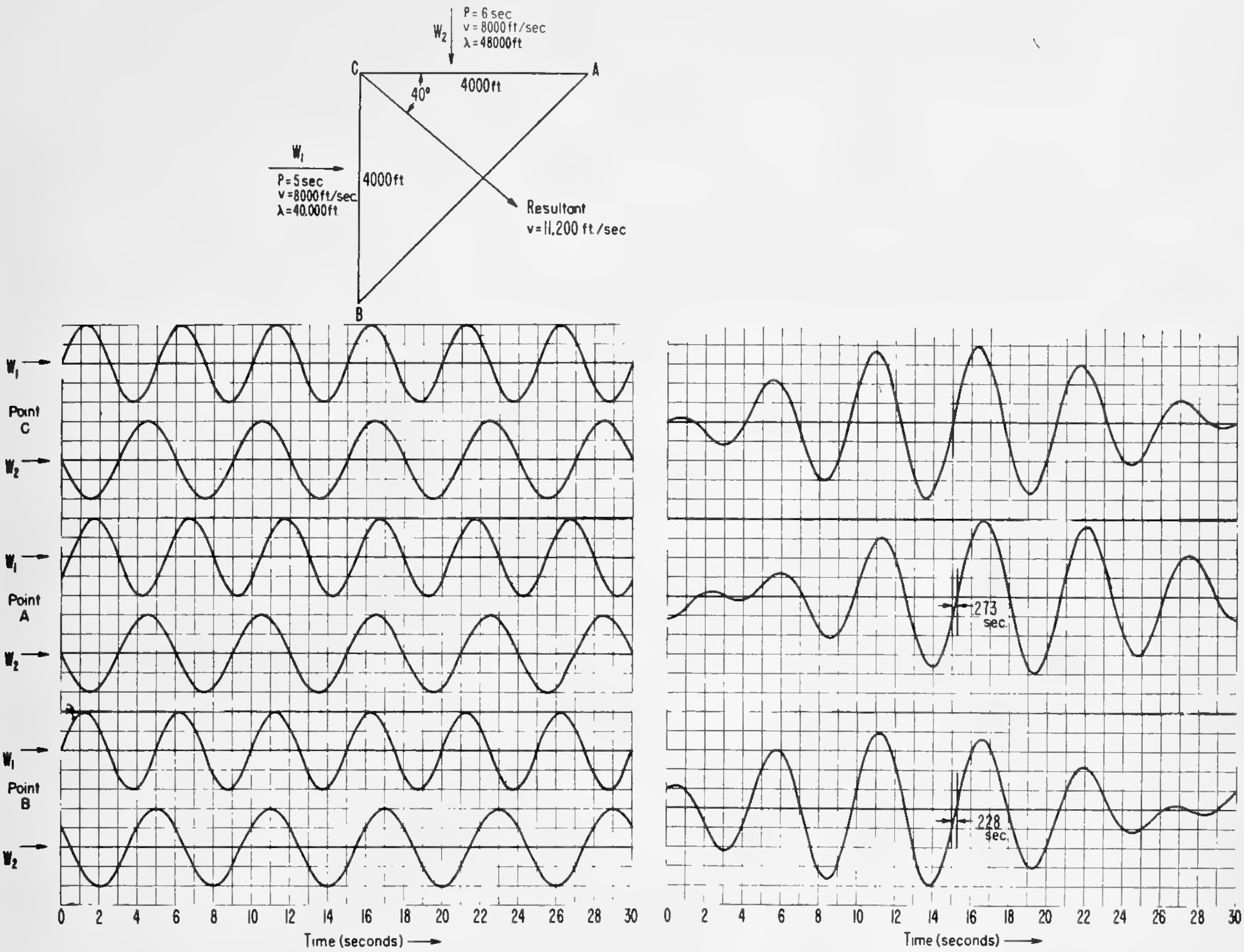


Figure 1. Illustration of two wave trains crossing a station at right angle

Perhaps the most serious factor which introduces incoherency in the recorded microseisms is the possibility of two or more wave trains simultaneously crossing the station in different directions. In this case the trains will add together in different phases at each seismometer with the net result that a false direction and velocity will be computed. Figure 1 illustrates a very simple case of two wave trains crossing a station at right angles. For this example the apparent velocity is 11,200 ft/sec as compared to 8,000 ft/sec for the component waves; and the direction is intermediate between the direction of the component waves. Figure 2 shows a graph of the ratio of apparent velocity to the real velocity as a function of the angle between the direction of propagation of two similar wave trains.

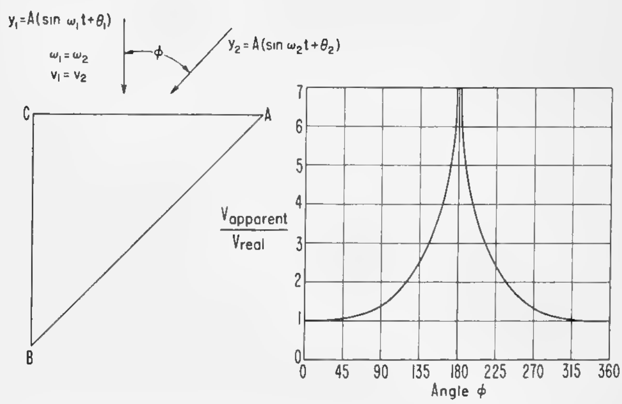


Figure 2. Ratio of apparent velocity to real velocity as a junction of angle between direction of propagation of two similar wave trains.

This indication that the apparent velocity of the recorded microseisms is increased if two or more wave trains simultaneously cross a network has suggested a method of selecting the bearings computed from a tripartite station. This method is reported by Kammer and Dinger [1951] and later studied by Donn and Blaik [1952]. Figure 3 shows a plot on a polar graph of a typical series of bearings obtained by the Naval Research Laboratory tripartite station on 21 November 1950. The distance of each dot from the origin is a

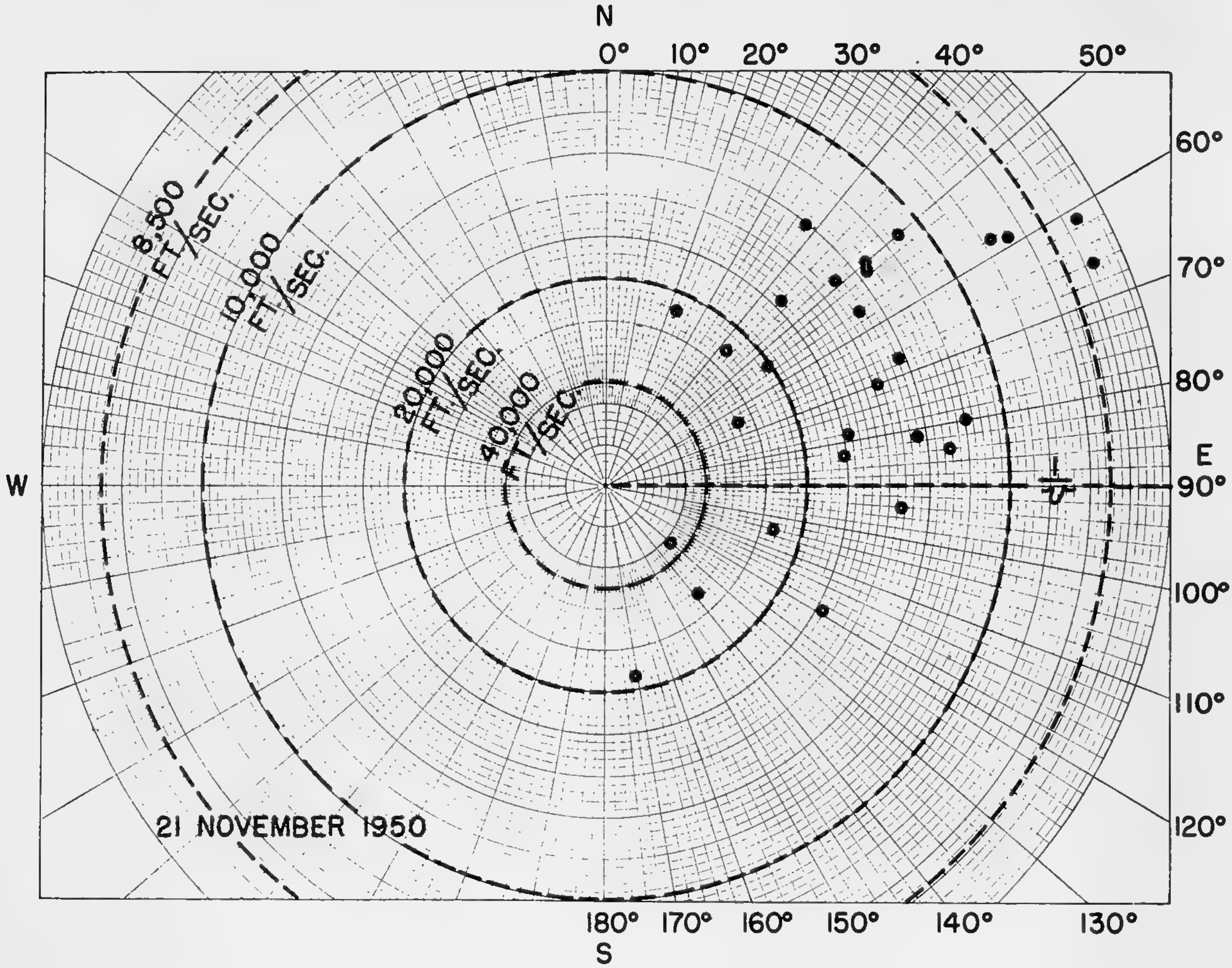
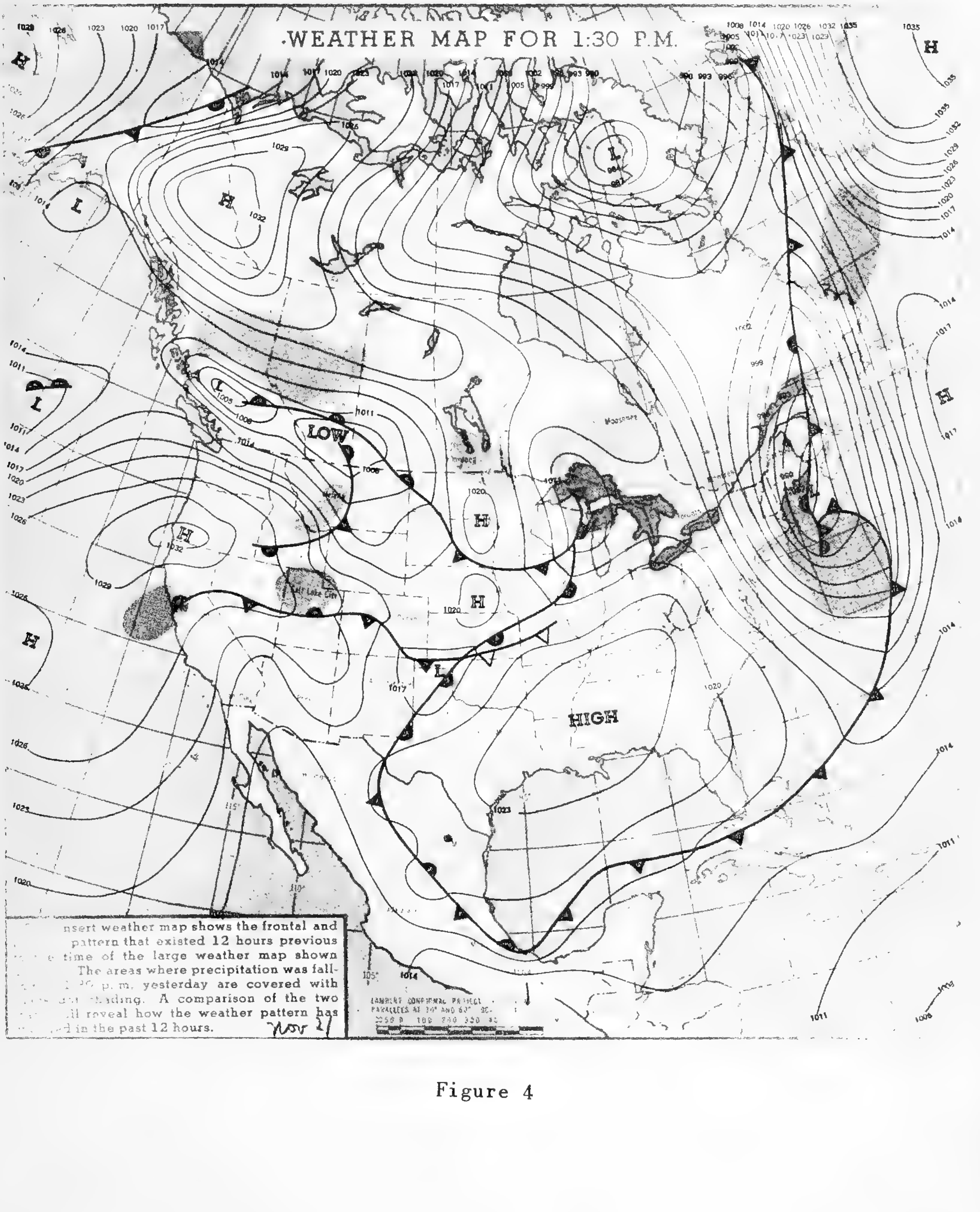


Figure 3. Typical bearings and velocities observed on 21 November 1950

measure of the reciprocal of the apparent velocity and the orientation of the point with respect to the origin gives the computed direction for the bearing. A wide scatter in direction is apparent when all points are considered, however, the spread in direction is narrowed, in this case to about 12 degrees, if only those bearings having an apparent velocity of 10,000 ft/sec or less are considered. Figure 4 shows the weather map existing at the time the bearings of Figure 3 were taken. The interpretation placed on these and similar results for various storms is, that at those instances when the computed velocity is a reasonable value, the recorded microseisms consists of a coherent wave train coming from a single source.

The question of refraction and reflection was raised in the preceding paper as a limitation on the usefulness of the tripartite station. Donn and Blaik [1952] have also referred to refraction as a possible source of error in pointing to the area of generation of microseisms. It may well be that refraction is important, but it is believed that the existence of refraction will be very difficult to identify as long as so much uncertainty exists in knowing where the true area of generation really is. The assumption that refraction is the reason why tripartite bearings do not point to the center of a hurricane or low-pressure area does not seem justified until it is proved that these centers are the area of generation. However, the use of earthquake records to study refraction of seismic waves is a valid approach since in this instance the location of the source is well known.

To summarize, it can be stated that the tripartite station has definite limitations. To obtain the greatest accuracy from a tripartite station one must (1) use the most advanced technique in instrumentation and (2) some method must be applied which selects the portions of the records to be used to compute the bearings so as to insure the use of the most nearly coherent wave trains that exist during a given microseismic storm.


WEATHER MAP FOR 1:30 P.M.
LOW
HIGH
Salt Lake City
Moosonee
nsert weather map shows the frontal and
pattern that existed 12 hours previous
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The areas where precipitation was fall-
p. m. yesterday are covered with
A comparison of the two
reveal how the weather pattern has
in the past 12 hours.
Nov 21
LAMBERT CONFORMAL PROJECTION

Figure 4

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Discussion

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Father Ramirez has given an excellent 65-year history on tripartite stations used in the study of earth motions connected with earthquakes and microseisms. He has described the systems used and mentioned some of the results obtained by investigators in many countries, including a few comments on his own experiments at St. Louis University in 1939. His summary of the view point of Naval Aerology in 1947 and again in 1952 is essentially correct.

Before one is able to discuss adequately the reliability of bearings and cross-bearings from microseismic storms it is first necessary to show where they originate. Father Ramirez dismissed this important point in these words, "The results were very satisfactory in demonstrating beyond doubt that microseismic waves are traveling and not stationary waves, that their direction of propagation can be measured, that the determination of the direction of arrival at St. Louis of these waves in all observed cases indicated that they come from tropical cyclones over the ocean, and that the bearing followed exactly the movements of the low pressure center and not the location of surf on the rocky coasts." In spite of these statements there are still a few doubts concerning the actual source of regular typhoon-hurricane microseisms with periods of 3.5 to 6.0 seconds. The pounding of large ocean swells from a storm at sea upon a land mass or a continental shelf cannot be the direct cause of large storm microseisms unless an abundance of observational data are disregarded. Microseisms have been repeatedly recorded several days before the energy front from newly formed storm swells could reach a land mass on which the seismograph was located. Therefore, in order to establish again the fact that this type of storm microseism is generated when the energy from a tropical storm is transmitted by some coupling mechanism directly to the ocean floor around the area of the storm, the following observational data are submitted:

1. Microseisms travel approximately 100 miles per minute while the energy front of storm produced swells seldom exceed 18 miles per hour. In other words microseisms could travel 740 miles in seven minutes or less but it would take at least 40 hours for newly formed ocean swells to travel the same distance. The data presented in Figure 7 of the following paper shows three typhoons passing over an area of the Pacific that is almost equal distance from Guam, Okinawa and Manila, or approximately 740 miles from each station. A critical analysis of the data will show that each storm quickly intensified into a typhoon with greater wind force and that the microseisms, in all three cases, immediately registered a sharp increase in amplitude at Guam, Okinawa and Manila. It is physically impossible for swells to have had anything to do with the simultaneous increase in microseisms because there would have been a delay of at least 40 hours for the intensified swell crest to reach the nearest land mass. Nor could the sudden intensification, which caused the increased microseismic activity, have occurred two or three days earlier for, had such been the case, the swells would have reached Guam several days before reaching the other two stations. The microseismic amplitude curves, Figure 1, show no such delay in either of the three storms. The simultaneous arrival of the larger microseisms at the three stations can be explained only by the theory that energy from severe tropical storms is transmitted to the bottom of the ocean where it immediately generates microseisms that are propagated outward in all directions at a speed slightly greater than 100 miles per minute. It will be noted, also, that the three typhoons were going away from Guam and approaching the other two stations. It is common knowledge that a swell traveling ahead of a storm will attain great height and period, while those traveling in the opposite direction never become prominent. But the per cent of increase in microseisms at Guam was as large or larger than at the other two stations.

2. Another argument against the surf theory of generation of microseisms is that the amplitude of microseisms recorded at Guam, Swan Island and Bermuda, islands surrounded for great distances by uniformally deep water, show no correlation with the state of the sea surrounding them. Heavy swells may pound

against a steep coast within one hundred feet of a seismograph or approach a continental shelf without causing the least increase in hurricane type microseisms. At other times each of the above stations have recorded microseisms more than ten times normal amplitude while the state of the sea around the islands ranged from a dead calm to swells of only one to three feet. Many other investigators also have reported a complete lack of correlation between surf and microseismic amplitude.

3. There is still another obvious reason why storm microseisms used in tracking severe tropical disturbances could not be caused by ocean swells striking a coast. The energy front of ocean swells, racing ahead of similar storms at any average velocity should travel equal distances in water of uniform depth, regardless of the direction from which they come. In other words, if storm swells generate storm microseisms, a seismograph located on an island surrounded by several hundred miles of uniformly deep water should start recording large microseisms as soon as the hurricane is a fixed distance from the station regardless of the direction of approach. If this were true lines drawn around such a station, for example Bermuda, showing the location of a 90 knot hurricane when microseisms are first recorded should be nearly circular. But this is far from true. The Bermuda seismograph has recorded many large microseismic storms when hurricanes of similar intensity moved directly toward the station from the east, the south, or the west and these data were used to prepare the chart shown in Figure 6 of the following paper. The microseismic range for hurricanes located south or west of the station is double the range to the east. It is therefore obvious that swell activity reaching the Bermuda coast cannot account for the generation of such storm microseisms because the limiting distance is not the same in all directions. The only theory that agrees with all the available data is that the microseisms are generated in the ocean bed in the vicinity of the disturbance.

The microseisms generated in the vicinity of a storm are transmitted outward in all directions through the earth's crust according to established laws of physics pertaining to wave motion in an elastic medium. Wave motion through a perfectly homogeneous substance is transmitted in straight lines from the source and decreases in amplitude, in direct proportion to the square of the area covered. However, such conditions of complete homogenity exist over very limited areas of the earth. Thus, it is only natural to expect that most microseisms generated by storms do not arrive at a station from the exact direction of the storms. "Microseismic Barriers," major discontinuities in the earth's crust or gradual changes in the density and elasticity of a portion of the earth's crust, are very logical phenomena that appear to reflect, to refract, and to absorb microseisms.

The following conclusions are therefore apparent: (a) The true origin of storm microseisms appears to be in the area of strong winds of a hurricane or typhoon. (b) It is possible to calculate accurately the direction microseisms are traveling when they pass over a tripartite station. If such microseisms from tropical storms were always propagated outward in concentric circles it would be very simple to determine their exact origin and the location of the storm by means of cross bearings from two or more tripartite stations. However, microseisms, as explained above, do not always travel in straight lines, and this results in large errors unless maps of the characteristic refraction patterns are constructed around each tripartite station. Until such charts are made and the proper corrections applied microseismic bearings will continue to look like those in Figure 1.

Discussion from the Floor

Melton. I would like to comment at this point that many of our observations constitute a statistical problem, and it should be valuable to examine some of our observations in that light. This term "coherence" which we have been using is much discussed in the literature, and, in particular, the subject of cross correlation as we move these two seismographs apart. It is obvious that if we place them on the same pier their correlation should be unity or 100 per cent, provided the instruments are operating properly. We should like to see the curve of this correlation factor as the separation is increased to the order of distances we have been discussing.

Byerly. (Commenting on tripartite measurements) It makes a difference how you set these seismographs down, too. You can set them down side by side and they won't show the same thing at all if you don't set them down properly.

Melton. Yes. But this business of properly planting a seismograph is a controllable thing. Working in the marshes of Louisiana which literally float, I have observed that when reflection instruments were simply set down in the mud, some of the reflections came out lying on their backs. However, the proper way to plant such instruments is on about 20 feet of pipe pushed firmly into the marsh, and if one plants them this way the reflections come out properly.

(*Byerly* asked if there were any cases of microseisms and no storms. *Gilmore* replied that there are a few cases. *Peoples* asked if the reverse was true in any case. *Gilmore's* reply was yes, but then it turns out the storm has been over-rated in intensity. *Melton* asked about the reliability of the intensity of the storms. *Van Straten* replied that as the storms slow down they usually intensify. *Macelwane*

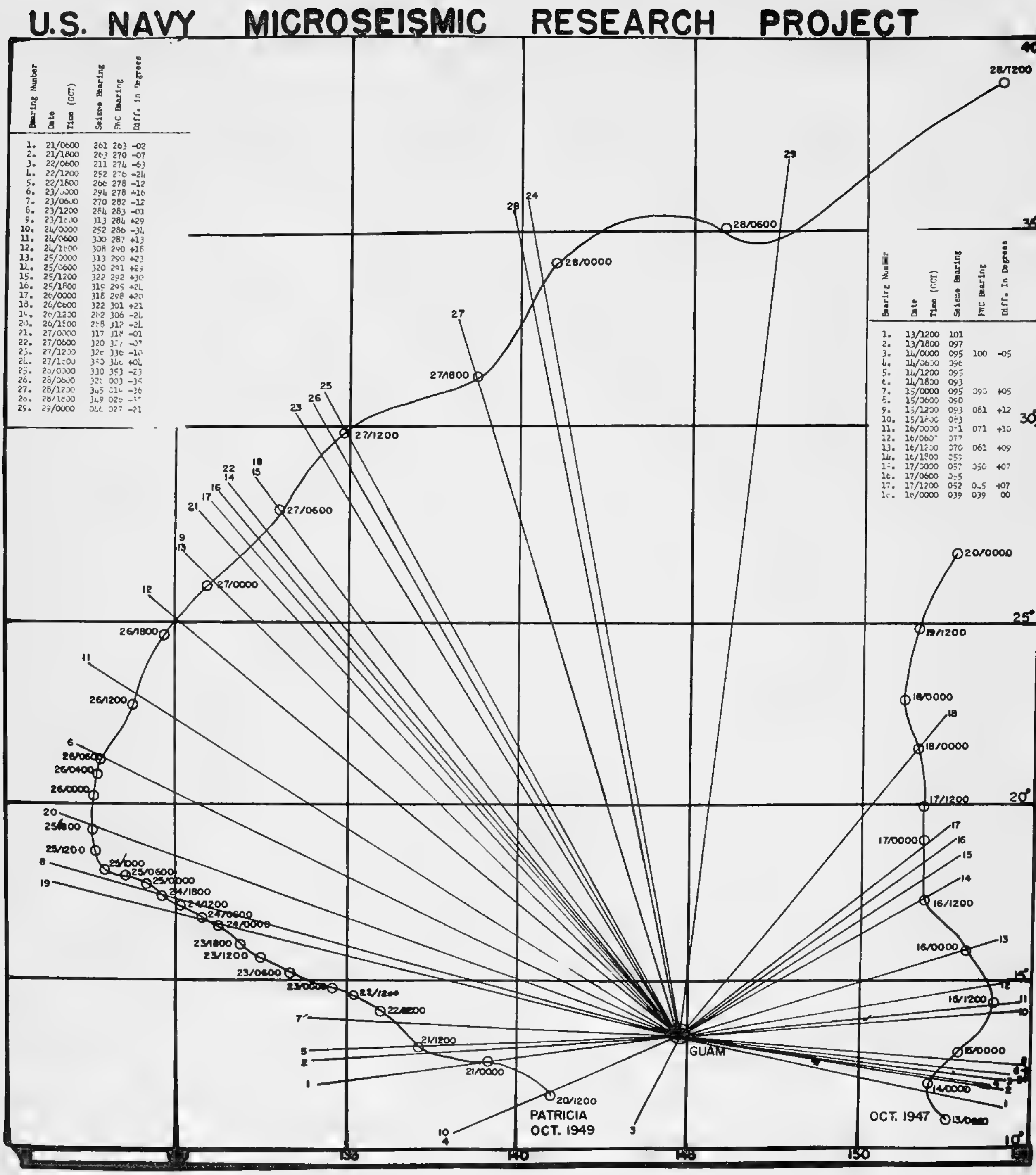

Figure 1. Microseism bearings from Guam

asked for the difference between tropical and extra-tropical storms. *Van Straten* replied that there are not the sharp fronts in the tropical storms. *Deacon* raised the point that it is the energy in the storm that is important. Bath asked if the waves differed greatly. *Deacon* said no.)

Deacon mentioned attempts recently made by Mr. Darbyshire in his laboratory to find the direction of the source of microseisms by comparing phases and amplitudes of the east-west and north-south components using recordings from the Galitzin seismographs at Kew made available to him by the Superintendent of the

Observatory. The comparisons were made for selected, simple, meteorological situations when there was little doubt of the actual source of the microseisms. The experiments were most disappointing and the main conclusion was that the ground movements at Kew were so complicated, presumably by the interference of waves which have suffered multiple refractions and reflections, that comparison of the two horizontal components gave little information about the direction of the source. During the experiments Fourier analyses were made of simultaneous recordings of waves and the three microseism components. The most striking features of these analyses was the striking similarity of the period spectra of the north-south, east-west, and vertical components. Though it was fairly certain that the direction of the storm changed over a time of 30 hours from west through northwest to north, the relative amplitudes of the east-west and north-south records and spectra remained the same.

The horizontal components always had the same range of periods as the verticals; this might indicate that there was no appreciable movement due to Love waves which would be likely to widen the period range of the horizontal components.

The amplitudes of the vertical movements were roughly twice those of the horizontal components.

The use of tripartite stations is based on the assumption that the microseisms approximate to a regular, simple wave system, with the wave crests travelling as straight lines. The fact that tripartite stations in some places, and at some times, give excellent results shows that the microseisms sometimes do travel as simple waves, but experience in Great Britain indicates that such behavior is exceptional there.

AMPLITUDE DISTRIBUTION OF STORM MICROSEISMS

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Abstract—The work of the Microseismic Research Project has been carefully reanalyzed in an effort to determine as many facts as possible concerning the nature of microseisms and their possible operational value in detecting and tracking severe tropical storms. From these data there is no longer any doubt that dominant group microseisms are generated by various types of meteorological disturbances over the oceans. It is still impossible to track storms by means of cross bearings from tripartite stations. A new method has been developed by the Navy Microseismic Research Project that permits accurate tracking of storms that are far from land. This micro-ratio technique of storm tracking is entirely independent of microseismic bearings from tripartite stations, or geology of the earth's crust, or theories concerning microseism generation; it is dependent only upon the amplitude of microseisms actually recorded from a storm at sea. It is possible, by the use of special microseismic charts, to detect, to track, and to determine changes in the intensity of a storm when it is within range of three or more microseismic stations.

The Problem of Recording Storm Microseisms—The Naval Aerological Service initiated the Microseismic Research Project in 1943 with one major objective, which was to determine if severe tropical storms could be detected and tracked by recording changes in the amplitude and period of microseisms. Rapid progress is now being made in solving this problem. The many new data obtained over a period of eight years by recording microseisms generated by several hundred tropical storms in the Pacific and Caribbean aid in verifying certain theories concerning the origin and method of propagation of storm microseisms. However, this paper will present only the details of the new Micro-Ratio technique of storm tracking and give facts and figures showing the degree of accuracy obtained by these methods in forecasting tropical storm movements.

The Microseismic Research Project uses Sprengnether type, horizontal component, electromagnetic seismometers with natural periods of approximately 7.0 seconds. Both the seismometer and the galvanometer are critically damped and of exactly the same period. In order to standardize the work at each station the magnification is held rigidly to 5,000 for ground motion with periods between 3 and 5 seconds, and each seismometer is orientated in a N-S direction. A *trace up* on each record represents an east movement of the ground. There are now 24 such instruments in operation at three single and six tripartite stations in the southeastern United States, the Caribbean, the North Atlantic and the Western Pacific. The data derived from records made on properly calibrated seismometers are trustworthy in all respects, as in Figure 1, which shows similar microseisms recorded at Bermuda on three different days from instruments one-half mile apart. This high degree of standardization is maintained at each station and is the same year after year.

There are many classes of microseisms recorded on seismographs but the type generally called "Group Microseisms," Figure 1, is the only class discussed in this paper. Macelwane [1951] says: "These microseisms are regular in wave form and appear in a succession of groups of a few large waves, each with short intervals of slight motion between the groups. They do not appear at all times but in discreet sequences which may last for a period of hours or days, building up to a maximum and dying down again. Such a sequence has come to be known as a microseismic storm." There is no longer any doubt that these storm microseisms are generated by various types of meteorological disturbances, but there is yet no complete agreement on the exact manner in which energy derived from the storm is transferred from the storm to the ground. However, it may be pointed out here that the newly developed technique of using the amplitude of microseisms to detect and track tropical storms is valid regardless of the method of generating storm microseisms.

Group microseisms are always recorded at each Navy seismograph station as soon as a generating source, such as a hurricane or cold front, comes within range of a station. Since this has happened hundreds of times in the past eight years at one or more of the microseismic stations it is now possible to describe certain outstanding characteristics of storm microseisms. These facts are all the more noteworthy because they also direct attention to the important problem of "How microseisms

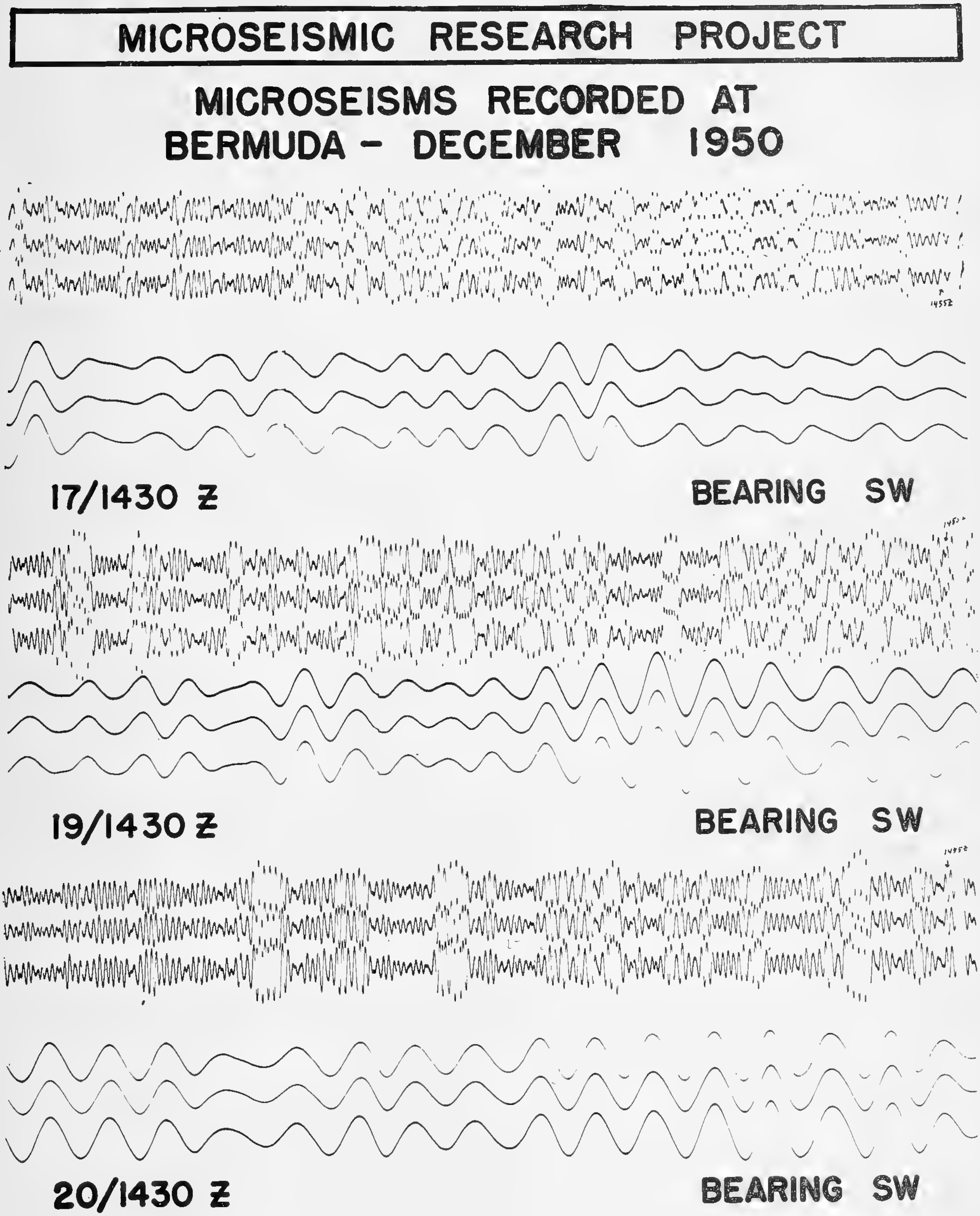

Figure 1. Group microeisms recorded at Bermuda

are generated," and throw considerable light on its solution.

Hurricane Microseisms—The first method attempted was to determine if tropical storms could be located within operational accuracy by microseismic cross-bearings from two or more tripartite stations. It is possible, with microseisms from a tripartite station such as shown in Figure 1, to calculate accurately the direction they are traveling when passing over a station. If microseisms from tropical storms were always propagated outward in concentric circles it would be very simple to determine their exact origin and the location of the storm by means of cross-bearing from two or more tripartite stations. But accumulated data show that microseisms do not always travel in straight lines, often resulting in large errors in tracking tropical storms, Figure 2. The tripartite cross-bearings proved unsatisfactory because the bearings pointed to the area of a storm only when it was traversing regions that lay in specific directions from a microseismic station, such as south and west of Guantanamo Bay, Cuba. It was therefore necessary to seek a new and different technique for solving the problem: one that would be completely independent of the method of generation and propagation of microseisms. Facts derived from microseismic records over a period of many years led the way to the development of a new and important technique for detecting and forecasting the movement and intensity of hurricanes and typhoons. This new method consists only of using observed microseismic data with little or no regard to the physical processes involved in their generation. The amplitude and micro-ratio charts were first constructed in 1950 and recent results indicate that detection and tracking of severe tropical disturbances is well within the necessary operational accuracy.

The tracks of six hurricanes are drawn in Figure 3, and along each track are listed the corresponding amplitude of recorded microseisms in mm on top of line and the intensity of the storm in knots on the bottom. The storm intensities and corresponding microseismic amplitudes are very consistent throughout the map, especially at points where the tracks cross

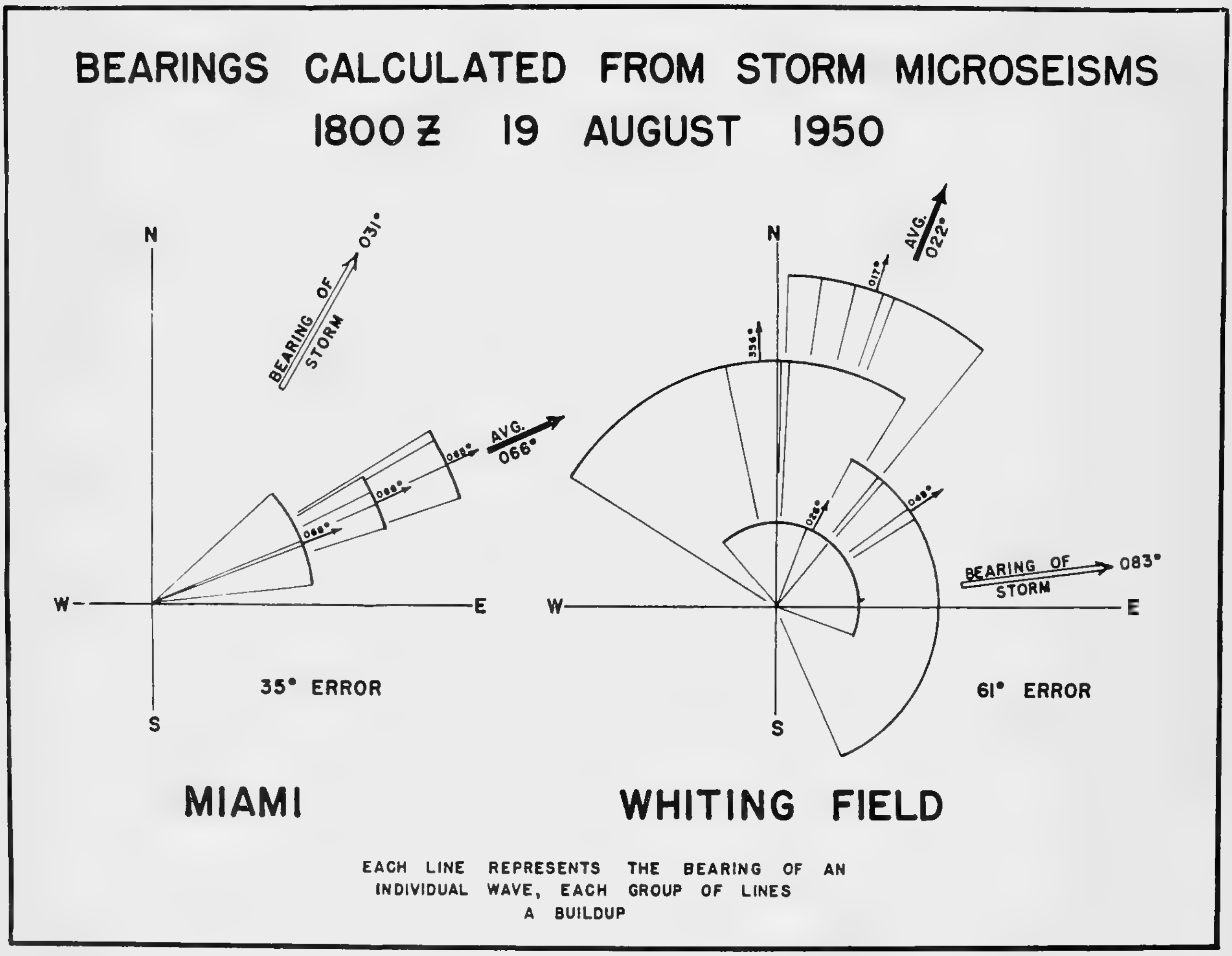

Figure 2. Microseismic directions obtained at Tripartite Station

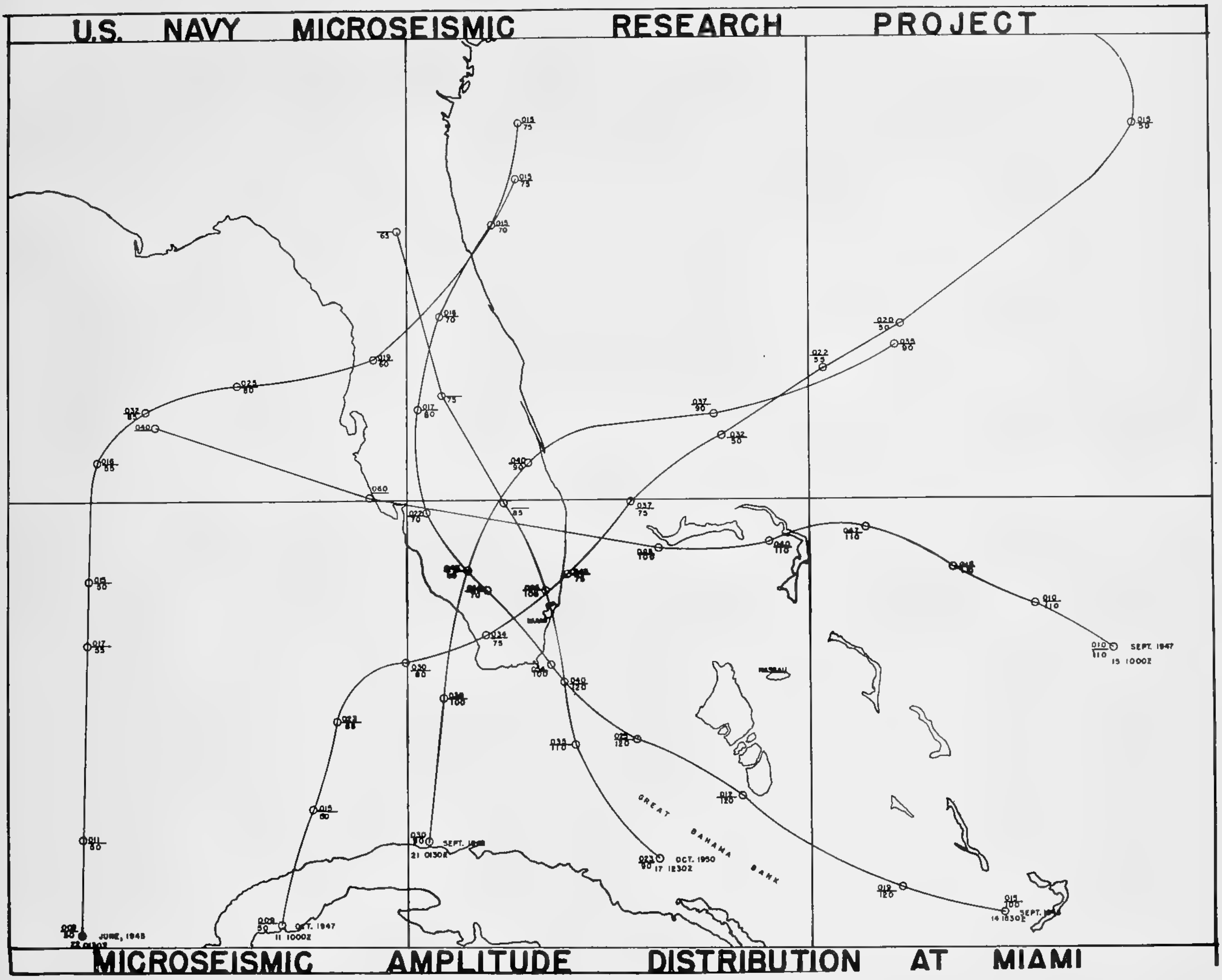

Figure 3. Hurricane tracks around Miami

each other and for the beginning of first increase in microseisms. From such data it is possible to draw lines of equal microseismic amplitude for a storm of any intensity around a station where sufficient storm microseisms have been recorded. The lines of equal amplitude around Miami for a 90 knot storm are shown in Figure 4. It will be noted that these lines cross Florida and that some storms continued with 90 knot winds. Considerable pertinent information concerning a storm can be obtained from such amplitude charts. It is an early warning for any station because it is impossible for a 90 knot storm to exist inside the 10 mm line on seismometers standardized by the Microseismic Research Project and not cause the ground to move or shake sufficiently to make microseisms of 10 mm in amplitude. This, therefore, makes the seismograph a one hundred per cent detector of severe tropical storms, regardless of all other types of information. Storms of greater than 90 knot winds will register the 10 mm amplitude of microseisms a greater distance from the station so that such lines can then be drawn for storms of any intensity. After a storm comes within range of Miami by crossing the 10 mm line, its intensity can be determined by plotting the position on the 70, 90, 110, or 130 knot amplitude chart. When a particular amplitude chart agrees with the location of the storm and the amplitude of the microseisms, then the intensity on that chart is very close to the actual intensity of the storm. Ten knots, more or less in the intensity of a severe storm will make little difference in its destructiveness to life and property.

The amplitude lines drawn in Figure 4 are valid for storms moving in any direction in relation to Miami. This suggests that the source of microseisms cannot be along the coast or continental shelf because, were this true, a hurricane approaching Florida would undoubtedly cause larger microseisms than a storm leaving Florida. The lines around Miami are very similar to those around all other stations in that none are concentric with the station in the center. They tend to run close together on one or two sides of a station and are more

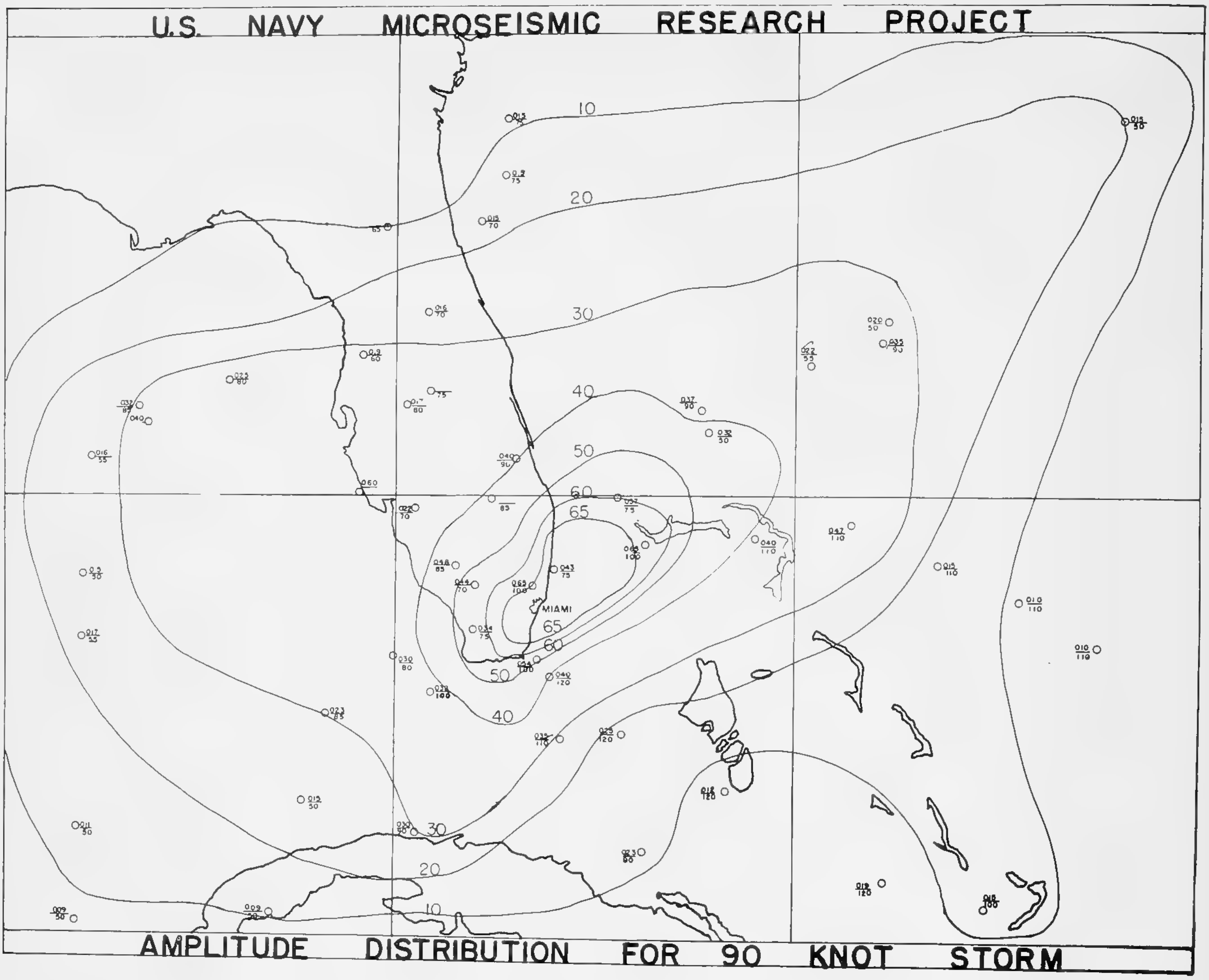

Figure 4. Amplitude chart for 90 knot storms in range of Miami

widely spaced on the other sides. Each storm passing a station inside the 10 mm line develops the same characteristic type of storm microseisms and will fit one of the amplitude charts, depending only on the intensity of the storm and its distance from the station. Similar storms, in passing over the same area as a previous storm, will generate point by point the amplitude of microseisms as shown on the chart. Microseisms recorded simultaneously at four stations located between 190 and 740 miles from the storm center produce storm microseisms of a uniform character, regardless of the direction of the storm from the station, Figure 5. Any similar storm in the same location will duplicate these microseisms at the four stations, provided the instruments are maintained in the finest of calibration and at a standard magnification.

Use of Amplitude Charts—The Fleet Weather Centrals at Guam and Miami direct aircraft to fly reconnaissance into and around tropical storms, when they are far from land. Some reconnaissance planes are now equipped with radar for tracking storms at night, but because of various interferences these radar reports are not always reliable. Valuable time is consumed in alerting the planes, flying to the storm location and transmitting reports. Occasionally a storm far from an airfield cannot be tracked, other than by extrapolation, because the planes have been grounded for various reasons: chiefly, mechanical trouble. Ship reports are often just as valuable, when available, but once the storm is located ships are warned to stay clear of the storm area and consequently give little additional information.

The Fleet Weather Centrals at Guam and Miami have been using the data from microseismic records for several years as a helpful aid in forecasting the intensity and movement of tropical storms. There are many specific cases in which microseisms have given valuable information that could be obtained in no other way.

The type of pertinent storm information furnished by the amplitude chart is adequately

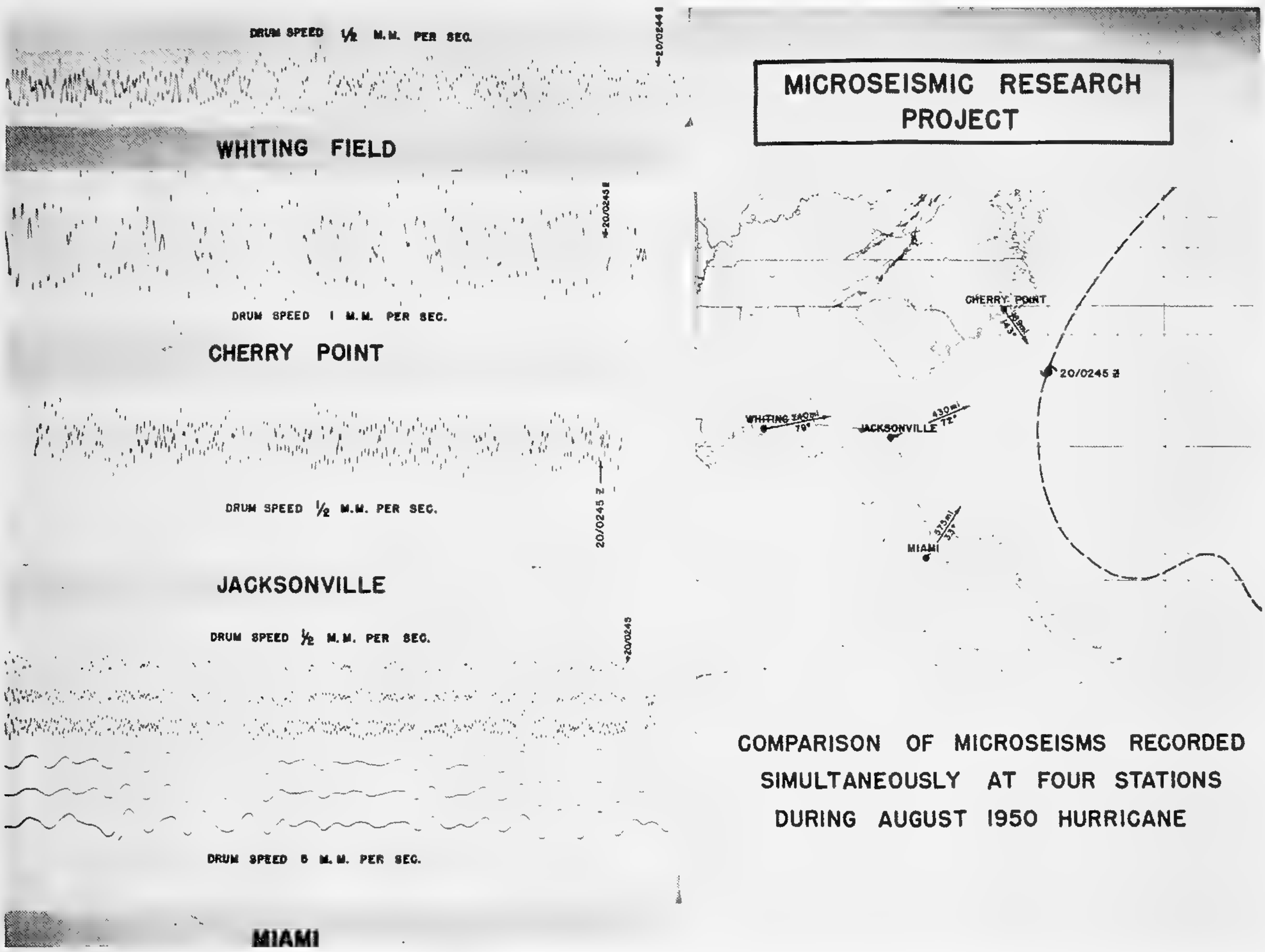

Figure 5. Group microseisms simultaneously recorded at four stations

illustrated in Figure 6 for a 1950 storm. Neither aircraft nor ship reports were available for a period of 42 hours from 7 September 1230 GCT. The forecasting agencies were left with no alternative but to rely entirely on microseismic data from Bermuda to make forecasts of the movement and intensity of the hurricane. Because of a large high pressure cell building up to the north, upper right of Figure 6, a continued northward movement was unlikely. An amplitude of 30 mm was recorded at 1200 GCT, 7 September, lower right of Figure 6, at which time the storm was reported by aircraft to have 100 knots of wind and its position as reported was located between the 35 and 40 mm lines on the 90 knot amplitude chart. That position and the 100 knot wind agreed with the amplitude chart because the 100 knot storm was located just outside the 30 mm line of a 90 knot chart. A maximum amplitude of 44 mm was reached at 0300 GCT on 8 September and, within an hour, it decreased 6 mm. The storm could not move north because of the high cell; it could not move west without causing an increase in microseisms at Cherry Point. The only forecast possible to make, from microseismic data, was that the storm was drifting slightly northward and slowly filling. This forecast was fully verified at 0630 GCT on 9 September, when the storm was located by a radar plane and found to be between the 45 and 50 mm lines, with only 75 knots of wind. The microseisms were only 37 mm in amplitude. The storm then moved westward, causing Bermuda amplitude to drop rapidly, while that at Cherry Point started increasing. It might be pointed out here that a 100 knot storm was within 600 miles of the coast for three days, yet no increase in microseisms were recorded at Cherry Point nor Jacksonville, despite the fact that storm-produced swells had ample time to reach the coast.

The tracks of three typhoons passing through the area between Guam, Manila and Okinawa are plotted on map, Figure 7, together with the corresponding amplitudes of microseisms recorded at the three stations. Typhoon Rita developed as a strong tropical depression late in 1948, west of Guam and continued westward. The Guam, Okinawa, and Manila amplitude chart, lower left of Figure 7, shows a very slow increase in microseisms until the 1200 GCT report of 6 November. At that time the storm quickly developed into a typhoon with winds increasing from 55 to 100 knots in the next 24 hours. A simultaneous sharp increase in microseisms occurred at all three stations but was greater at Okinawa because the storm was nearer that station. This increase occurred at all stations long before storm swells from the intensified winds around the storm could possibly build up and travel the great distances to the three stations.

Typhoon Patricia in October 1949 was a small tropical storm with only 35 knots of wind when it passed Guam on the 20th. This storm also rapidly developed into a typhoon with 90 knot winds, starting about 2400 GCT on 23 October. The microseisms increased rapidly at all three stations when the storm was about 740 miles from each station. The maximum increase was greater and a day later at Okinawa because the storm was approaching that station and crossing increasingly larger lines of equal microseismic amplitude. Here again it was physically impossible for the intensified swells to reach the three land masses and account for the rapid increase of microseisms on 24 October.

Almost one month later typhoon Allyn passed south of Guam on 17 November and the seismograph registered microseisms of 240 mm in double trace amplitude. The microseisms fell rapidly to 83 mm as the storm moved westward and the winds decreased to 80 knots by 1200 GCT on 20 November. This typhoon suddenly intensified when it was located by aircraft at a midway point about 740 miles from Guam, Okinawa, and Manila. The simultaneous sharp increase in microseismic amplitude at the three stations, lower right of Figure 7, again occurred before storm generated swells could build up after the storm intensified and travel the necessary 740 miles.

The rapid intensification of the three typhoons was identical with the sharp increase of microseismic amplitude at each station. It may be pointed out here that these three typhoons generated very large microseisms at three stations 740 miles away in three different directions; yet a very similar storm was within 600 miles of Cherry Point for three days without causing the slightest increase in amplitude. Each storm was moving away from Guam when the microseisms suddenly increased in amplitude; i.e. the storms were

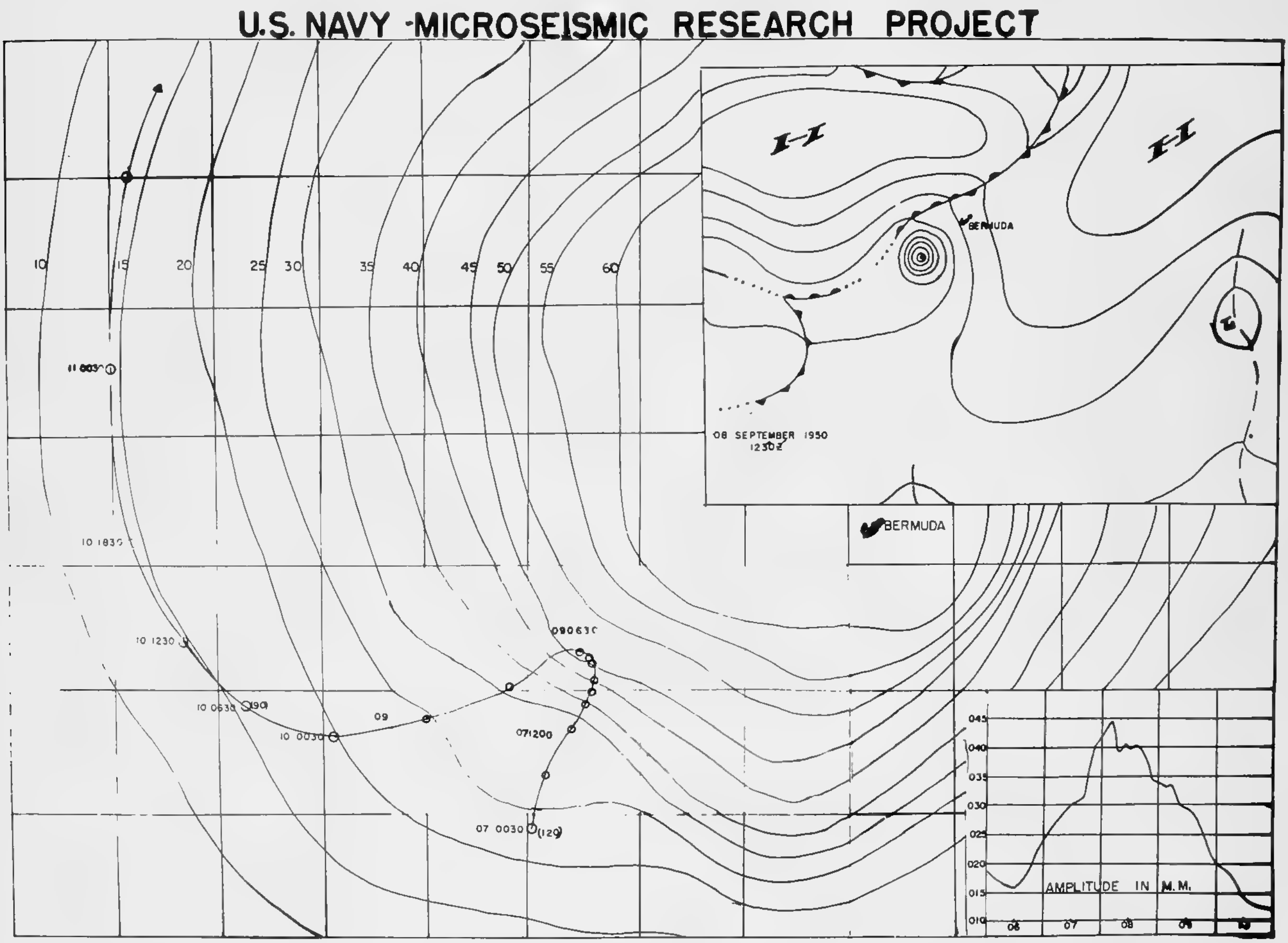

Figure 6. Forecasting movement of storm by microseismic amplitude at Bermuda

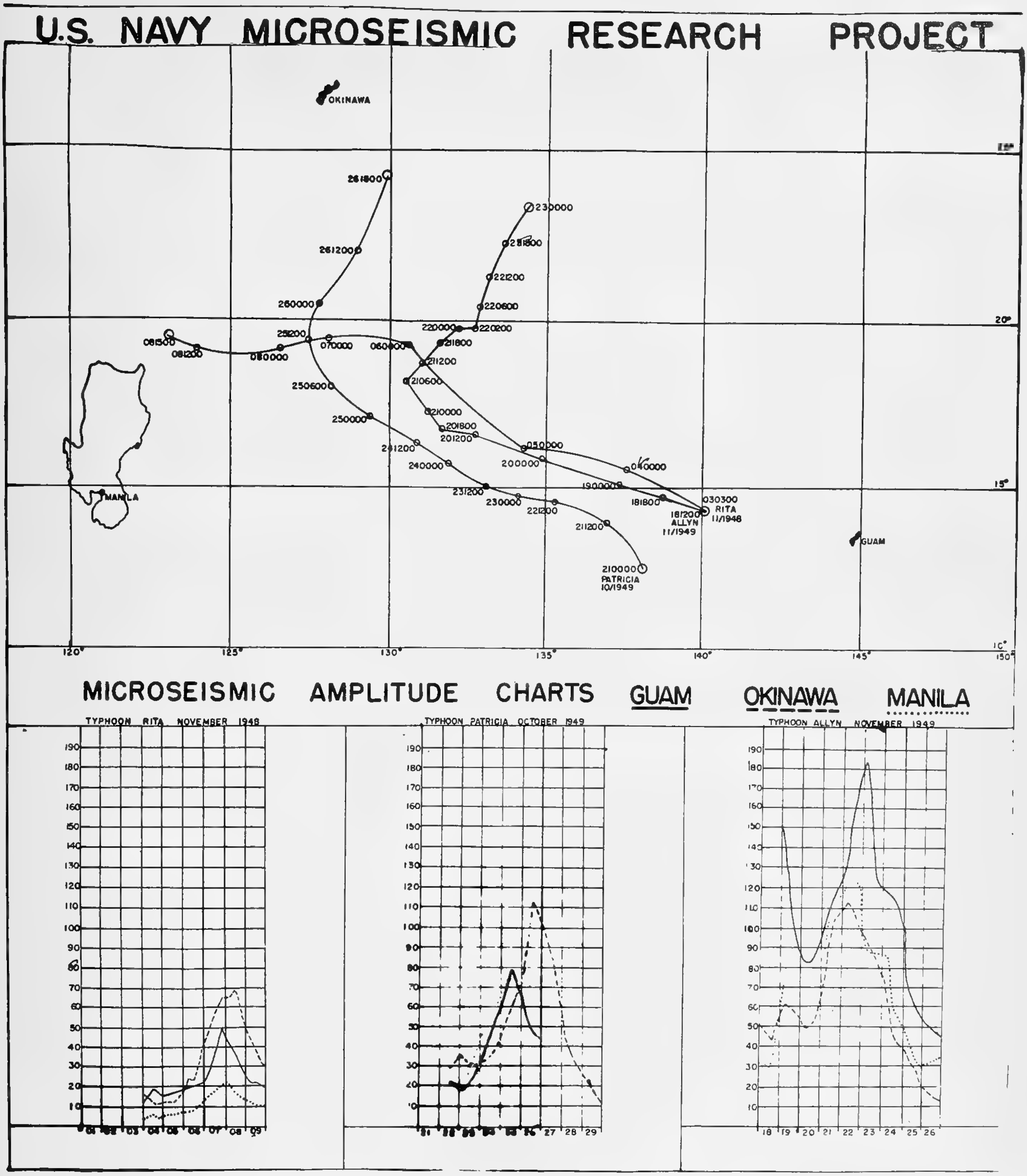

Figure 7. Forecasting movement of storm by microseismic amplitude at Guam, Okinawa and Manila

crossing decreasing lines of equal amplitude, and the microseisms should have been decreasing if the intensity of the storm had remained constant. The fact that the observed microseisms were too large at each station to fit the proper amplitude chart gave every evidence that the storms were increasing in intensity.

Microseismic stations can detect tropical storms long before their existence can be determined except by direct observation by planes or ships. This priority of detection is sometimes difficult for the Carribbean stations because of the presence of many islands with weather facilities between the possible source of storms and a recording seismograph. Nevertheless it has been accomplished. Figure 8 shows an example of early storm detection. The left side of the picture shows a portion of the Caribbean just south of the west end of Cuba, where the synoptic situation was not conclusive as to the existence of a hurricane. On the basis of an abrupt increase in microseismic amplitude at Swan Island, a weather plane was sent to investigate the area and the situation shown in the right side of the picture was found to exist. This storm developed rapidly and the intensification was immediately registered by the Swan Island seismograph before any other sort of warning, such as increasing swells or winds, reached Cuba, Swan Island, or Yucatan. Property located seismographs would always give similar advance warnings. In other words, it would be impossible for a storm to develop into a dangerous hurricane or for a fully developed hurricane to approach a seismograph station without giving sufficient warning to permit the carrying out of all necessary precautions. The seismograph at Guam often detects typhoons long before they are otherwise known. Moreover, a microseismic station in an area which is homogeneous with that over which the storms are traveling, can observe immediately, from an increase or decrease in the amplitude of the microseisms, any changes in the intensity of the storms being tracked. Tropical storm How in October 1951 was first determined to be a hurricane by the rapid increase of microseisms during the night at Miami and Jacksonville. Even when the synoptic reports indicated that the storm was apparently filling, the seismographs at Jacksonville and Cherry Point showed that it continued to be attended by 90 knot winds. This feature of storm detection, that of giving a good estimate of storm intensity, is an especially valuable aid in the forecasting of hurricanes, especially at night and at other times when there are no direct observation by planes and when no ship reports are available.

The Micro-Ratio Charts—The microseismic amplitude charts are primarily used for detection of storms, and after the storms are located, to

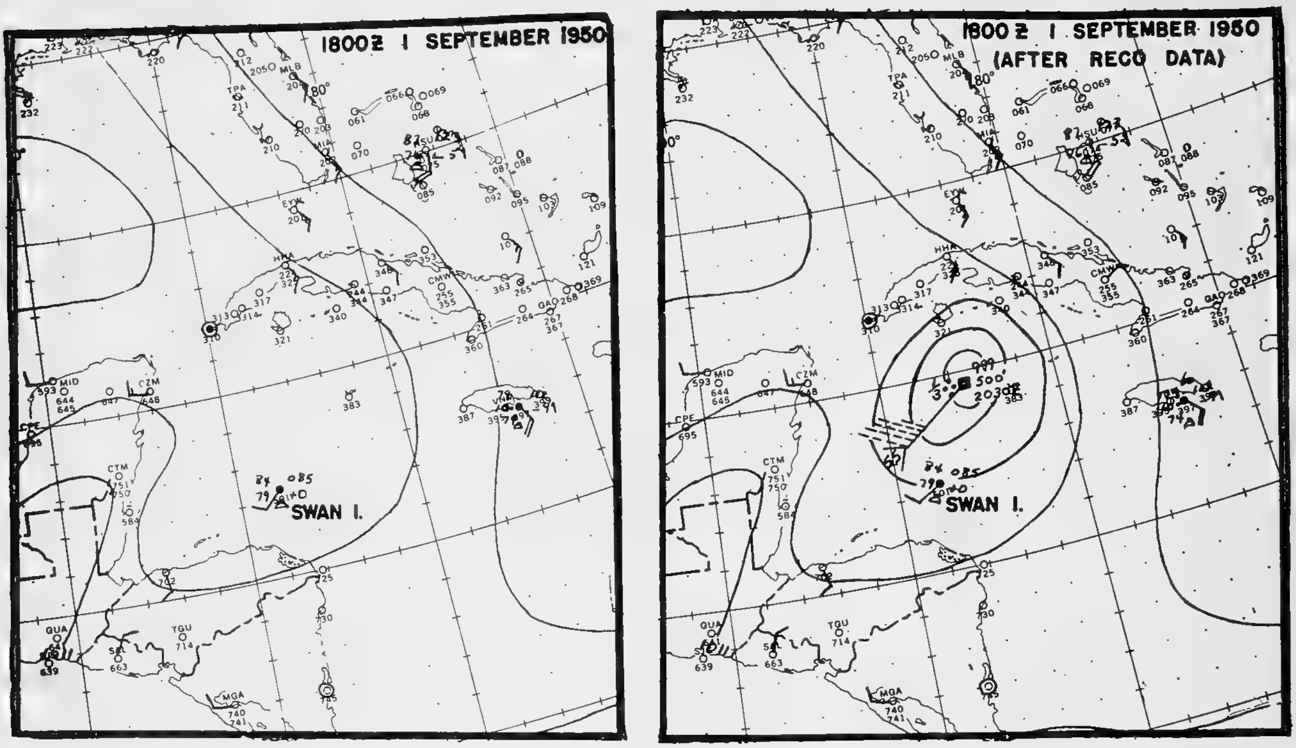

Figure 8. Early detection of hurricane by microseismic amplitude increase at Swan Island

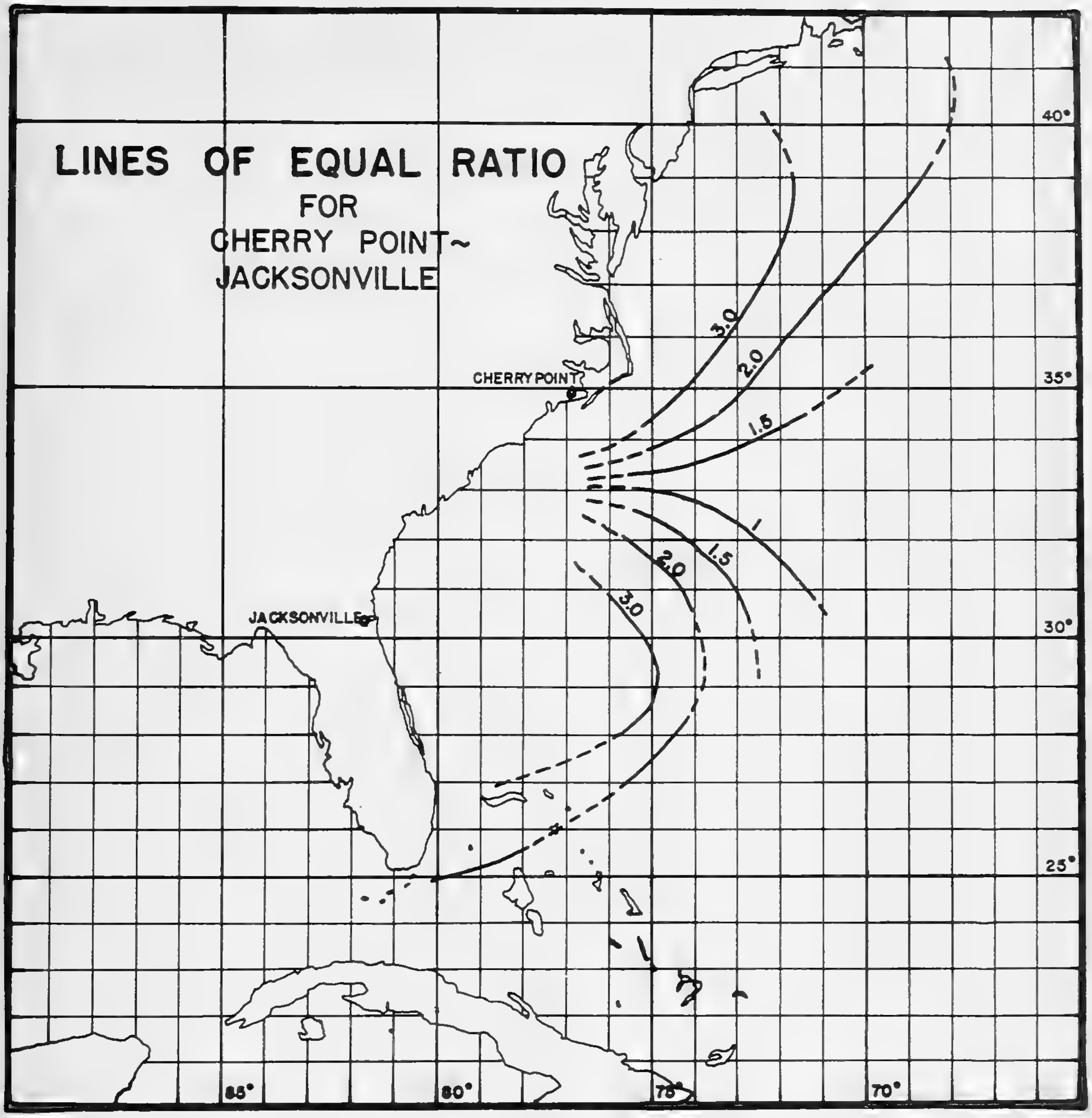

Figure 9. Micro-ratio chart between Jacksonville and Cherry Point

find their intensity. Amplitude charts, therefore, are very important in storm forecasting. The micro-ratio technique mentioned briefly before, is still another step forward in the tracking of severe storms. This technique can be used independently of the amplitude charts, but the two together can give all the information necessary for detecting and tracking of a storm and for finding its intensity.

The technique involved in the construction of micro-ratio charts is based upon the observed fact that a storm at any specific place will cause the amplitude of the microseisms at two recording stations to be in a definite ratio to each other regardless of the intensity of the storm, the geologic formation through which the microseisms pass, or the distance of the storm from the station. It is important to note that this technique does not presuppose anything concerning the method of generation or of propagation of microseisms, since one hurricane at any particular place will generate and transmit microseisms in the same manner as any other hurricane in the same place. For example, when the first storm of 1950 was located at 39°N and 70°W, the ratio of the microseismic amplitudes between Cherry Point and Jacksonville was 2.5. The fourth storm of the year, one of less intensity, passed over

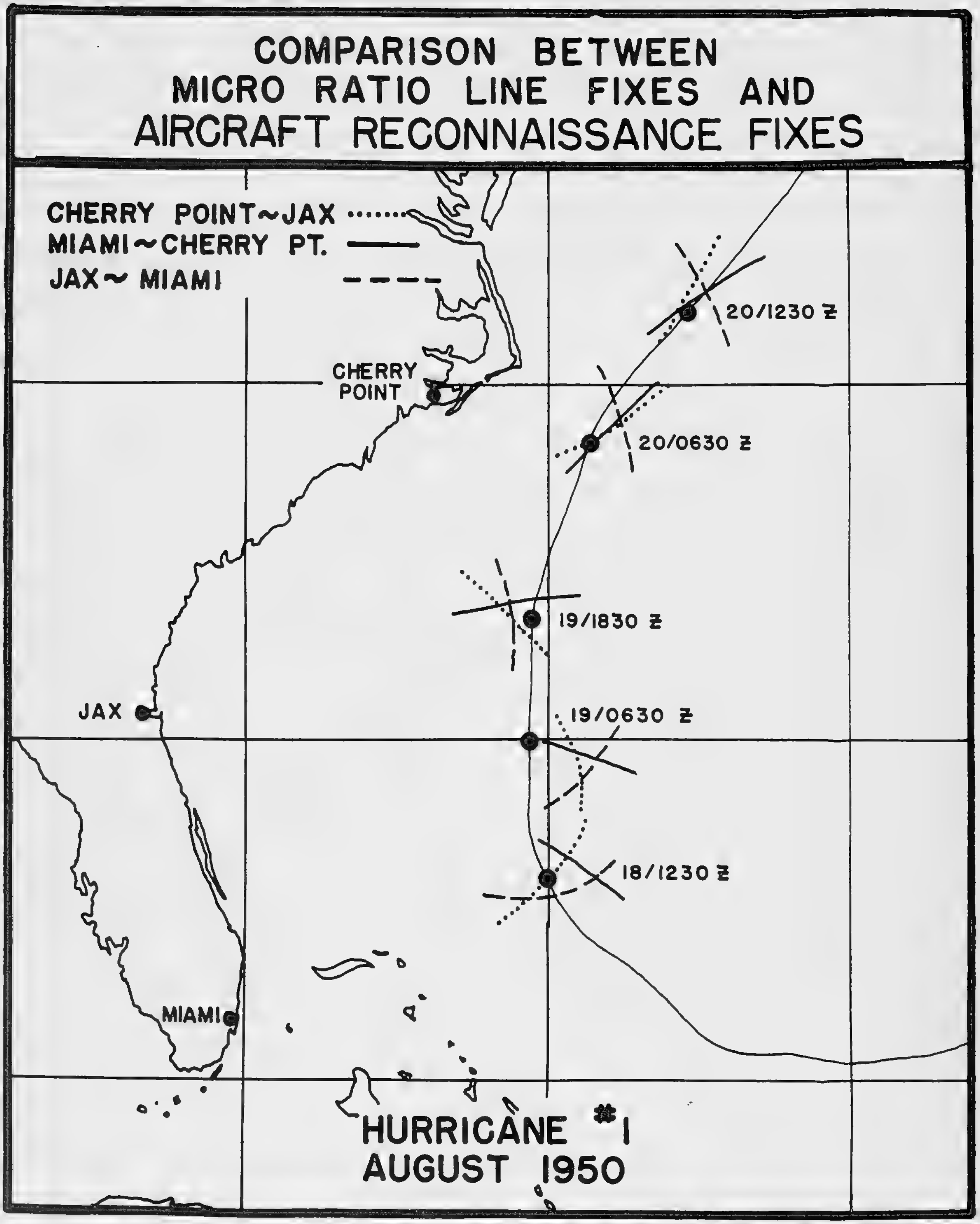

Figure 10. Hurricane tracked by micro-ratio technique in 1950

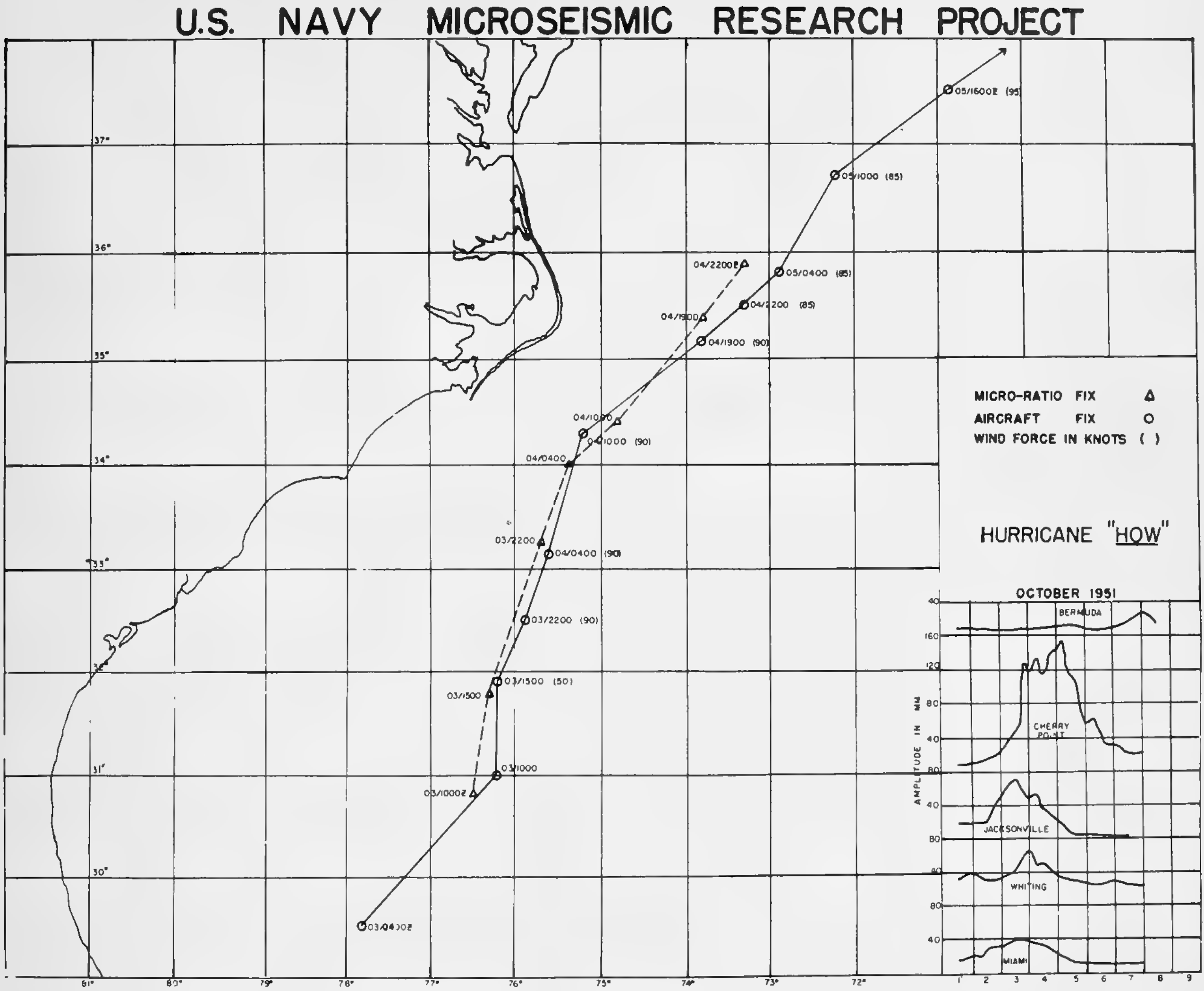

Figure 11. Hurricane tracked by micro-ratio technique in 1951

the same point and smaller microseisms were recorded at each station, but the ratio of their microseisms was again 2.5. This fact has been observed at all other points for which reliable data are available.

A chart utilizing such data can be made by plotting on a map lines through all points of equal ratio, Figure 9. Thus any storm generating microseisms that are recorded simultaneously at the two stations must be located along the ratio line on a chart corresponding to the ratio of the amplitude between the recorded microseisms at the two stations. Similarly, a ratio between either station and a third station can also be computed and the intersection of the two ratio lines will fix the position of the storm, Figure 10. Naturally, when other stations record microseisms generated by the same storm, the additional computed ratios will greatly improve the accuracy of the fix.

This method should locate and track a storm in any area when three or more microsemic stations record vibrations solely from the storm, regardless of the storm's intensity, position, its size or its method of generating microseisms. The final accuracy of any micro-ratio fix will depend principally upon the amount and reliability of accumulated storm data used in making the ratio charts between each pair of stations and the number of stations recording the storm simultaneously. An example of the accuracy that has been obtained is demonstrated in Figure 10 for hurricane ABLE in August 1950. The five positions plotted along the storm track were taken from the Fleet Weather Central's final history of the storm. The micro-ratio fixes corresponding to the actual storm positions are shown as the intersection of the three ratio lines. This was the first attempt to use the micro-ratio charts and resulted in positions not far from the true location of the storm.

The results of the second attempt is shown in Figure 11. This shows the storm track, plotted from advisories issued by the Fleet Weather Central, and the track plotted from micro-ratio technique. The micro-ratio fixes were, for the first time, released along with the Weather Central advisories. The two tracks are almost identical. Hurricanes ABLE and BAKER in August and September 1952 gave additional proof that the new technique of storm tracking is a very valuable tool for the forecaster of tropical storms. Accurate information as to the location and intensity of these storms were derived from microseismic charts.

The validity of such amplitude and micro-ratio charts is naturally dependent upon the accuracy of the data used in determining the intensity of past storms. Wind force is difficult to estimate, especially the higher velocities, and is often as much as 10 knots in error, even in some aircraft reconnaissance reports. If it were always correct, it would still be only a rough gauge of the relationship between the total energy of the storm and the amplitude of the microseisms produced at each station. Also there is the stipulation that no other major source of microseismic generation, except the hurricane, be influencing the station. This, of course, can be assured only by the placement of additional single microseismic stations throughout the entire hurricane belt in such numbers that stations influenced by other sources would not have to be depended upon. These data indicate that, when a proper network of microseismic stations is available, it will then be possible to detect and track any and all tropical storms.

Conclusions —— The data presented in this report clarify and solve to some extent some of the problems involved in microseismic storm detection and show what is necessary for the future microseismic forecasting of tropical storms. The primary problem of detecting and tracking tropical storms with microseismic data appears to be near solution with the newly developed micro-ratio technique. Even a few secondary problems concerning the actual method by which energy is transmitted from a storm to the ocean floor and then propagated in the form of microseisms may be near solution. There are three important items concerning the fundamental problem of complete microseismic storm detection to which attention is invited.

A. Early Tropical Storm Detection.

From all the information obtained in the research there is no longer any doubt that the seismograph, when properly located in relation to a tropical storm, can detect the storm when it is over water. The effect of apparent irregularities in the earth's crust that tend to impede microseismic transmission can be greatly minimized, if not altogether eliminated by placing seismograph stations on each side of all known "microseismic barriers." Early storm detection is obviously a valuable aid to any hurricane warning system and it is very probable that the present operational network. A sufficiently close network of such stations would, in addition to giving early warning of a storm, greatly aid the weather centrals in scheduling flights into suspicious areas.

B. Tracking Tropical Storms.

The newly developed micro-ratio technique for tracking tropical storms is based entirely upon an empirical use of microseisms and for that reason is not influenced in any manner by changes in geologic formations between the storm and the observing station nor by the size or intensity of the storm. The accuracy obtained in tracking past hurricanes, if maintained with future storms, could fulfill operational needs. These hurricanes were the first storms passing within range of three or more microseismic stations since the new technique was developed. If this high degree of accuracy can be obtained again with this same group of stations and with new stations in other areas, it would, without doubt, be a long step towards the answer to the original problem in tropical storm tracking.

C. Detecting Changes in the Intensity of a Storm.

It is just as important to know whether or not a hurricane is intensifying, when approaching a populated area, as it is to know its approximate position. For this reason alone the amplitude charts are of prime importance in hurricane forecasting. The amplitude charts give far more accurate information on storms than merely moving them along at a constant forward speed, when weather reports are not available. This is especially true at night when no reconnaissance can be made. It is obvious that information from such charts will become more valuable as the charts are improved by increasingly accurate data from future storms and as the number of effective stations is increased.

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Discussion

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The technique that was just presented by Mr. Gilmore on the use of empirical amplitude relationships of storm microseisms as a method of tracking tropical storms is highly interesting and shows promise. However, a large amount of observational data is needed before its usefulness can be verified. The method, to be reliable, needs precision calibration of the instruments, as was outlined by Mr. Gilmore; and in addition, data from storms traveling in different directions over different though intersecting paths must be checked against one another, so that if a storm is over any given point in the ocean the amplitude ratios recorded at any two stations used in the net must be the same regardless of the direction of approach of the storm to that point. Presumably this condition has been met by the Microseismic Research Project and if the method has been proven by independent observers, there is little room for adverse argument.

Although Mr. Gilmore favors the idea that storm microseisms which are recorded in force on land are generated beneath the eye of the storm, he nevertheless reports that the amplitude relationships at the several stations will hold regardless of how the microseisms are generated. I concur with Mr. Gilmore only up to a certain point.

If the microseisms are generated on the ocean bottom beneath the storm, then barring extraneous sources, the amplitude ratios at two stations should be about the same.

Further if two storms approach a given spot in the ocean from the same direction and the microseisms are caused as a result of the fetch reaching the nearest shore, then amplitude relationships should also hold. But if the

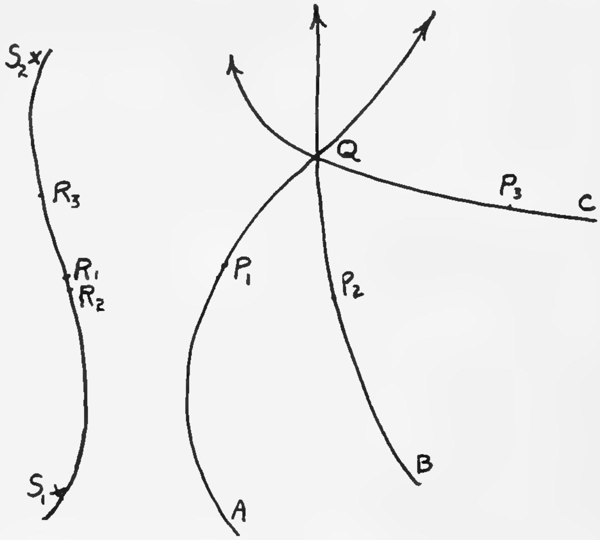

Figure 1. Paths of 3 hurricanes; A, B, C, passing over point Q producing microseisms recorded at stations S_1 and S_2.

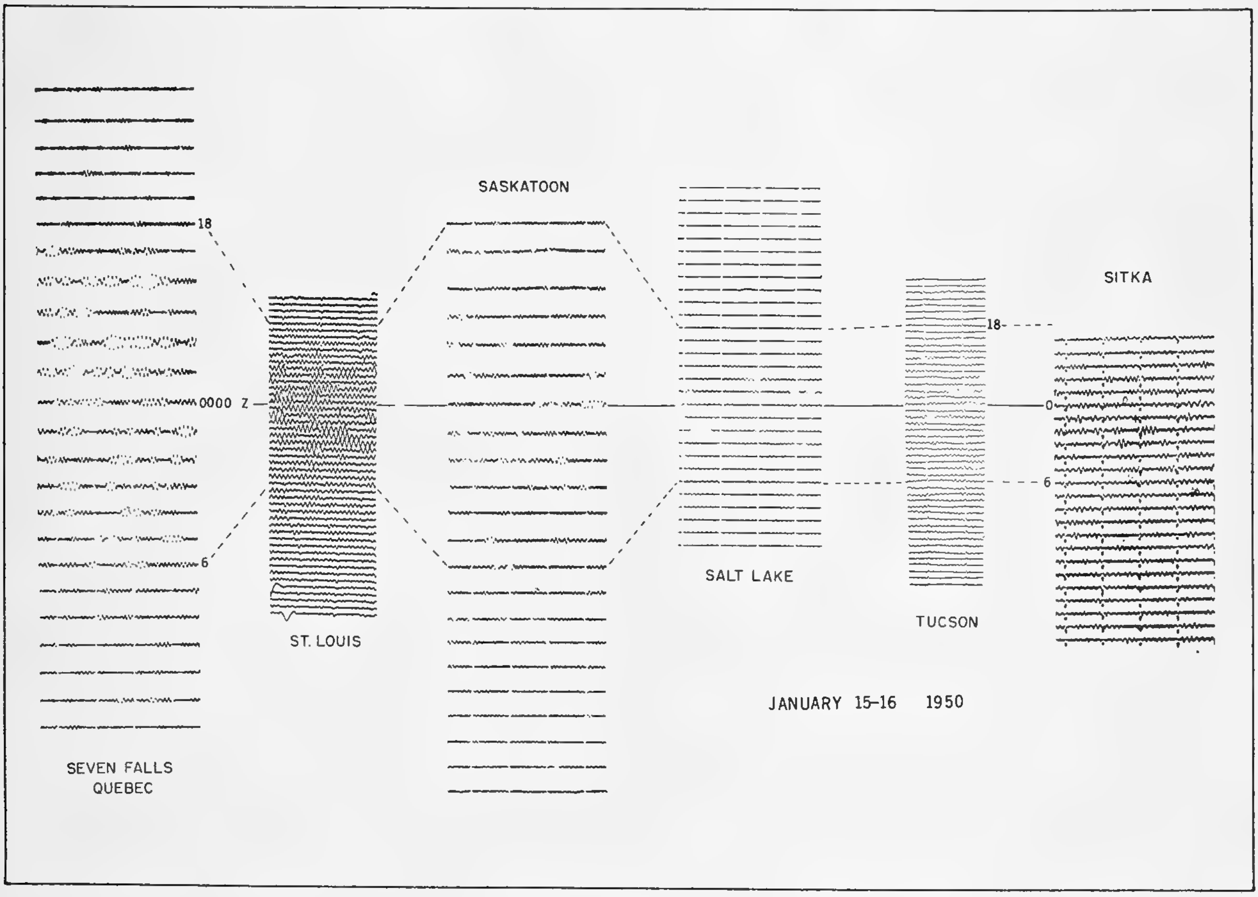

Figure 2. Storm microseisms from an east coast source recorded at representative North American stations.

storm approaches the point from different directions, unless it is far at sea, amplitude ratios as a means of locating the storm center would hardly be expected to be reliable because in the two cases swells generating the microseisms would generally reach the shore at different points. Consider for example three hurricanes, A, B, and C, Figure 1, which in the course of their history had passed over point Q. Suppose that microseisms were generated near the shore as a result of the fetch from the hurricanes reaching the shore at R_1, R_2, and R_3, leaving the storm areas at P_1, P_2, and P_3, the storm in the interim having moved to Q. In the illustration R_1 and R_2 are close together and would be expected to yield nearly the same microseismic ratios as stations S_1 and S_2, but in case C, R_3 is closer to S_2 and the amplitude ratio S_2/S_1 would be relatively larger, although other things equal, the microseisms would be much smaller.

Past and future verification of the amplitude ratio technique of every circumstance may possibly verify the contention that hurricane microseisms are generated under the ocean beneath the storm area. But I have strong evidence that this condition does not generally apply to extra-tropical storms in the western half of the Northern Hemisphere. I will now outline the case history of 5 cyclonic storms of the past three winter seasons and will show that microseisms generated in connection with these storms are not by virtue of the position of the storm over the ocean, but more probably by virtue of strong winds on the shore-directed limb of the storm while the storm may have been as far as several hundred miles at sea. The microseisms, however, may appear on land as much as 48 hours later after the swells generated by these winds have had time to reach the shore or meet oppositely moving swells generated by an earlier or later storm. The storm in the meantime may have moved inland or far at sea, or it may have become dissipated.

Source of data. In this study the data are from long or intermediate period seismographs operated by the U. S. Coast and Geodetic Survey and cooperative institutions. Data from cases 1 and 3 are also from the stations of the Canadian network; the Berkeley and Pasadena nets, and from individual stations at St. Louis, Cleveland, and Bermuda. Most of the Canadian and Coast and Geodetic Survey data are from direct-recording seismographs and are therefore non-selective to the periods under study.

Source location and transmission of microseisms. We will first make two basic assumptions. Figure 2 are shown cross-sections of records from representative stations arranged in order from east to west across North America. The solid lines connecting the records represent the same hour. The records show a microseismic buildup beginning, peaking, and ending about the same respective times. On most records, the normal, background period was 4 or 5 seconds before and after the buildup. The period on all of the records was 6.0 sec. At the beginning and 6.5 to 6.8 sec. during the peak of the buildup. It will be assumed therefore that these microseisms were generated in the same general area; and since the amplitudes from east to west were progressively from relatively large eastward to relatively small westward, it will be assumed that this area was on or off the east coast.

It follows that microseisms were transmitted from the east coast to the west coast with low attenuation and with no material increase in period.

The records shown here are representative of all North American stations, parts of California excepted. The history at Pasadena paralleled that of other North American stations. Berkeley and Ukiah, on the other hand, although they recorded 6 + sec. periods among others as part of the normal background, recorded no increase in amplitude nor a concentration of 6 + sec. periods during the time of the buildup at other stations. Reno and Fresno have short-period instruments. Reno recorded definite 6 + sec. periods during the storm, but not before or afterward. Fresno definitely recorded a 6 + sec. period on the vertical and indefinitely on the horizontal components.

It follows that the Rocky Mountain system transmits 6 + sec. microseisms from east to west, although not as well as the eastern lowlands perhaps. Furthermore, it follows that the Sierra Nevada, or more likely the Central Valley of California because of its deep sedimentary rocks, may possibly be an effective barrier to these microseisms. During this time there was only slight evidence of a 6 + sec. period in very faint waves during lulls in local microseismic activity at Bermuda where the dominant periods were 4-5 sec. There was no evidence of a 6 + sec. period at San Juan. Bermuda is closer to any east coast area than Saskatoon, yet 6 + sec. microseisms at Saskatoon were quite strong. The North Atlantic, therefore, apparently absorbs 6 + sec. microseisms.

Figure 5 shows an arrangement similar to that of Figure 2 except that amplitudes are relatively much larger on the west coast, more specifically on the northwest coast, than toward the east. The parallelism in amplitude buildup shown here was characteristic of all North American stations except Pasadena and perhaps Ukiah and Tucson. Before and after the amplitude buildup the period was somewhat less than 7 sec. at most places. During the peak the period was about 8 sec. at Sitka and 8.5 sec. elsewhere including Victoria, Ukiah, and Tucson, but not including Pasadena. There was no evidence of an 8 + sec. period at Honolulu at this time.

Again it may be assumed that the microseisms during the peak activity at all stations where 8 + sec. periods were dominant had a common source area, and that in this case the area was near the west coast, more specifically nearer Sitka than Victoria, for instance.

This series of records indicates that these microseisms may undergo a slight increase in period during the first few hundred kilometers of their journey, but thereafter they continue across the continent, including the western mountain systems, with no material change in period. Since Berkeley records microseisms

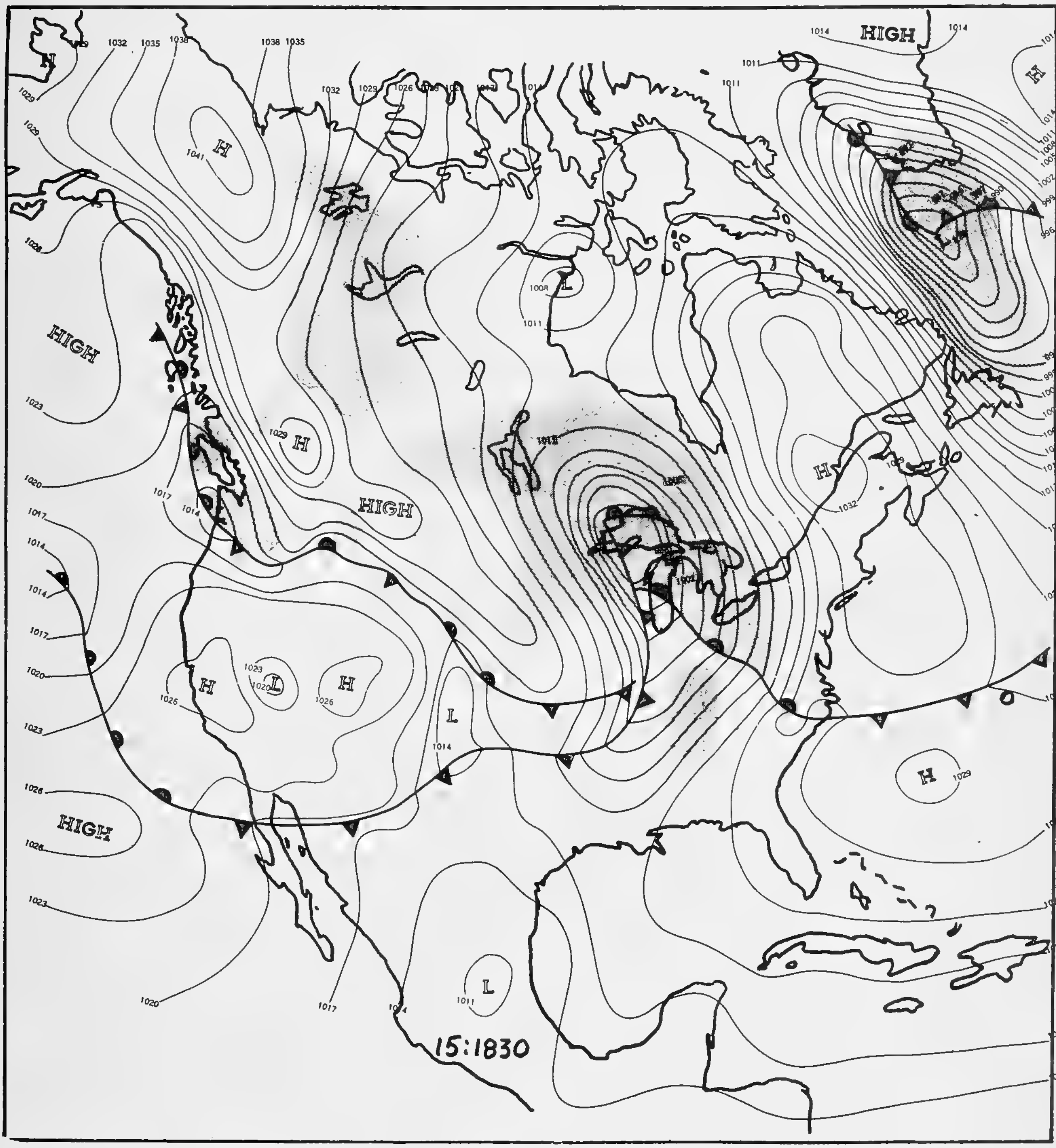

Figure 3. Weather chart of North America for 15 January 1950 at 18h 30m GCT showing a storm moving off the coast of labrador which is believed responsible for the generation of the microseisms shown in figure 2, courtesy of U. S. Weather Bureau.

from this source and Pasadena and Honolulu do not in this case, there is additional support for the idea advanced earlier than the Central Valley of California and the ocean basin, in this case the North Pacific, absorbs these microseisms.

It should be noted that conclusions reached in this section imply only that the North American continent, with one or two exceptions, is a good transmitter of microseisms and that the adjoining ocean basins are not; and except at the very first, the period does not increase with distance. No inference as to source or source mechanics was made except that the location was in a general area near one coast or the other.

The relationship of these and other selected microseisms to weather conditions will now be presented.

Case I. On January 15, 1950, a violent

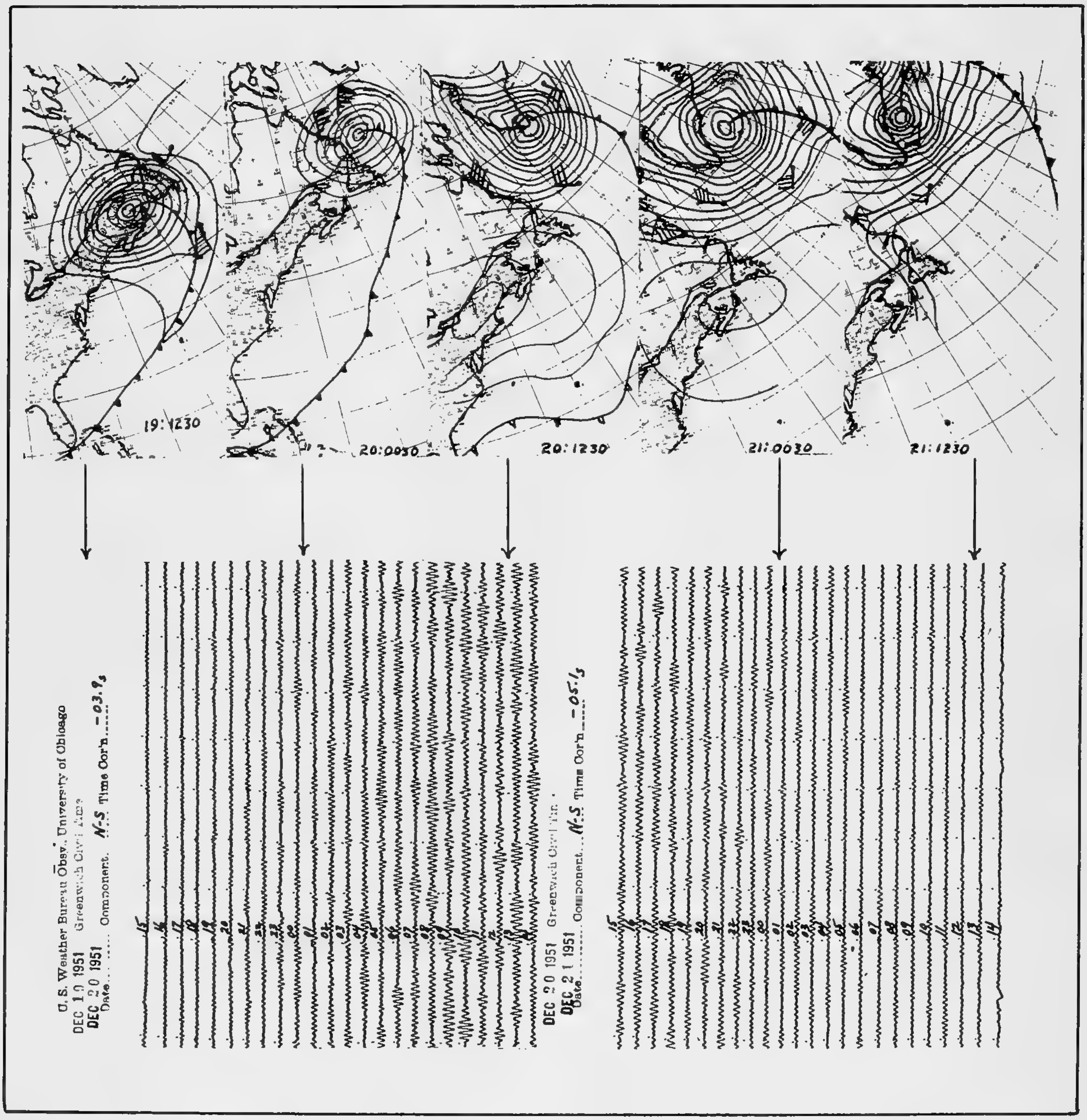

Figure 4. Weather charts 19-21 December 1950 showing a storm moving off the coast of northeast North America and microseisms recorded at Chicago during this time. In this and other illustrations, the flags show wind velocity in 10 knot intervals and the numerals represent days, hours and minutes. Isobar interval is 5 millibars.

storm moved off the coast of Labrador. This storm was undoubtedly the indirect cause of the microseisms shown in Figure 2, some of which were received as far away as Fairbanks, Alaska, but not at Bermuda, which is much closer. Normal microseisms east of the Rocky Mountains had periods of 5 sec. or less. The 6 sec. microseisms first began to appear at mid-continent stations about an hour before the time shown in Figure 3 and about 8 hours after the center of the storm had moved oceanward from Labrador, at which time storm-generated swells on the northwest limb had time to approach the Labrador coast. The microseisms died out about 15 hours later after the storm had moved oceanward off Greenland.

Case II. The source of heavy microseisms shown in Figure 4 is undoubtedly a violent storm accompanied by 75-knot winds on its northwest limb, moving oceanward off Newfoundland. An amplitude buildup accompanied by 6-6½ sec. periods began at Chicago a few hours after the storm center had left Newfoundland. Heavy microseisms continued until the storm center had moved off the east shore of Greenland and after the winds off Labrador had shifted from onshore to offshore. The storm continued violent off Greenland long after the continental microseisms had died down. During this time 6+ sec. microseisms were also recorded at Sitka along with 4½ sec. periods from a local source.

Case III. Figure 6 shows two possible sources, one in Davis Strait and the other moving inland from the Pacific off Alaska. The Davis Strait source perhaps was responsible for much of the 6± sec. normal background over the continent.

While the Pacific source was moving offshore along the Alaska Peninsula, local 7 sec. microseisms were generated, reaching Fairbanks and Sitka but not in force inland. Then after the storm center had moved inland, 8-8½ sec. microseisms began to appear at all continental stations as shown in Figure 5.

This was a rapidly moving storm. If it were responsible for the 8+ sec. microseisms that were recorded on a continent-wide basis, of which in my mind there is no doubt, the storm moved ahead of the mechanism that actually produced the microseisms.

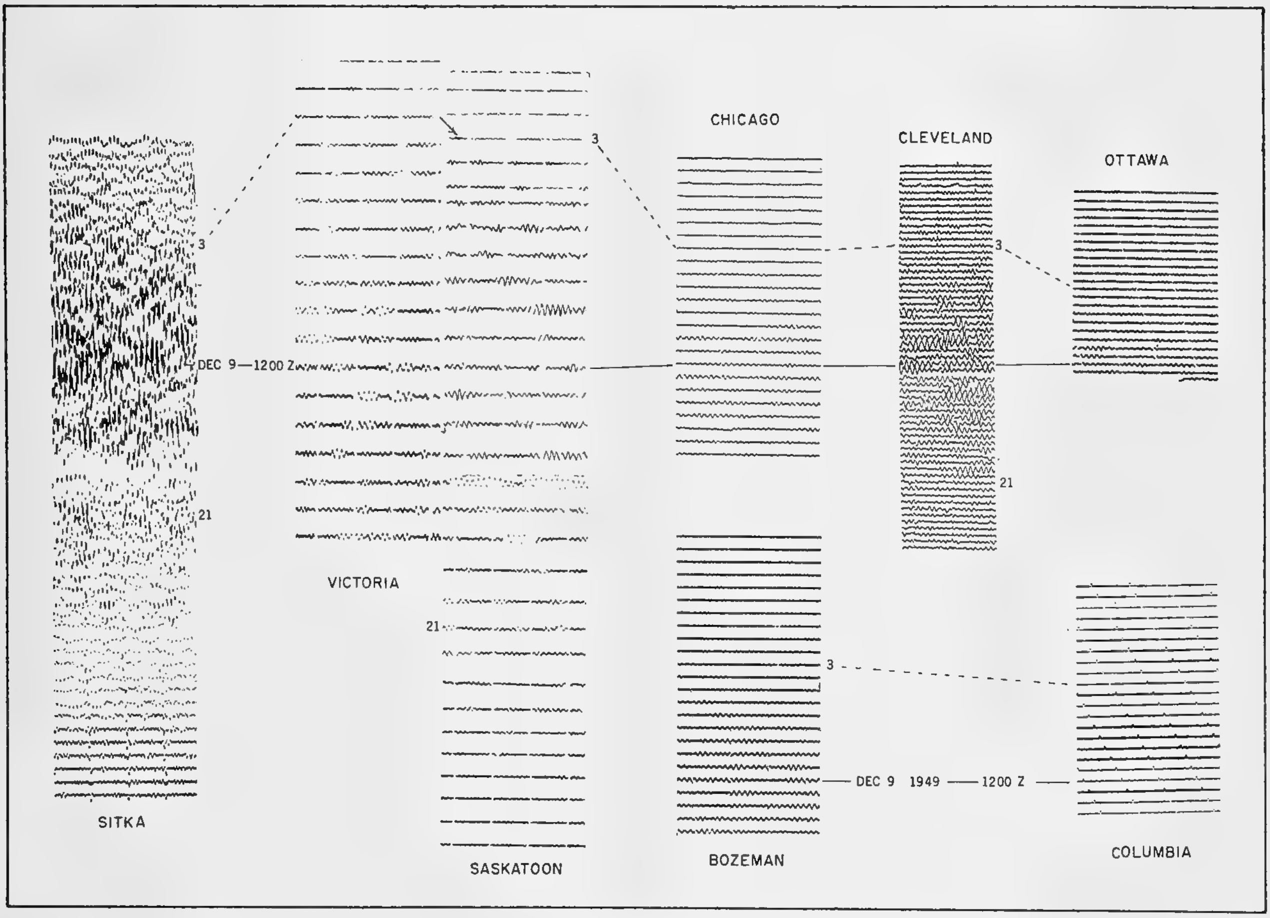

Figure 5. Storm microseism from a west coast source recorded at representative North American stations.

Case IV. Figure 7 shows (a) a violent storm, the center a few hundred miles off the North Pacific coast, and (b) a storm moving northeast across the Great Lakes, then along the St. Lawrence Valley. Two Greenland lows appeared earlier but dissipated and are of no importance to us. The Great Lakes low moved north across eastern Labrador into the Davis Strait. The Pacific storm moved very slowly toward the coast. While it was most violent on late January 20, continental microseisms were normal. While it was still violent, but abating slightly, 7 sec. microseisms appeared on a continent-wide basis at a time before the eastern storm had a chance to produce them. While the Pacific storm having abated somewhat was still off the coast the center just a hundred or so miles from Sitka, and also at the time the eastern low was in Davis Strait when it should have been producing the observed microseisms if it were the generating medium, continent-wide microseisms, including those at Sitka, had returned to normal.

Case V. At first glance this seems an indefinite source, except that heavy 8½-9 sec. microseisms appeared suddenly at Sitka and College, Alaska. At nearly the same time 8½-9 sec. microseisms with almost parallel amplitude buildups appeared also at Chicago and Columbia. The source must have been on the west coast because of the large amplitudes at Sitka and College.

In Figure 8 the contours represent the elevation of the 1000 mb. surface in intervals of 200 ft. This is equivalent to about 7 mb. where the intervals on earlier illustrations were 3 or 5 mb. The larger map shows a fairly heavy storm over the Aleutian Islands but which abated before it reached the mainland. But the front from this storm reached the entire Alaskan coast almost simultaneously 36 hrs. later and about 4 hrs. before the 8½+ sec. microseisms began to appear.

This storm, in my opinion, supports the

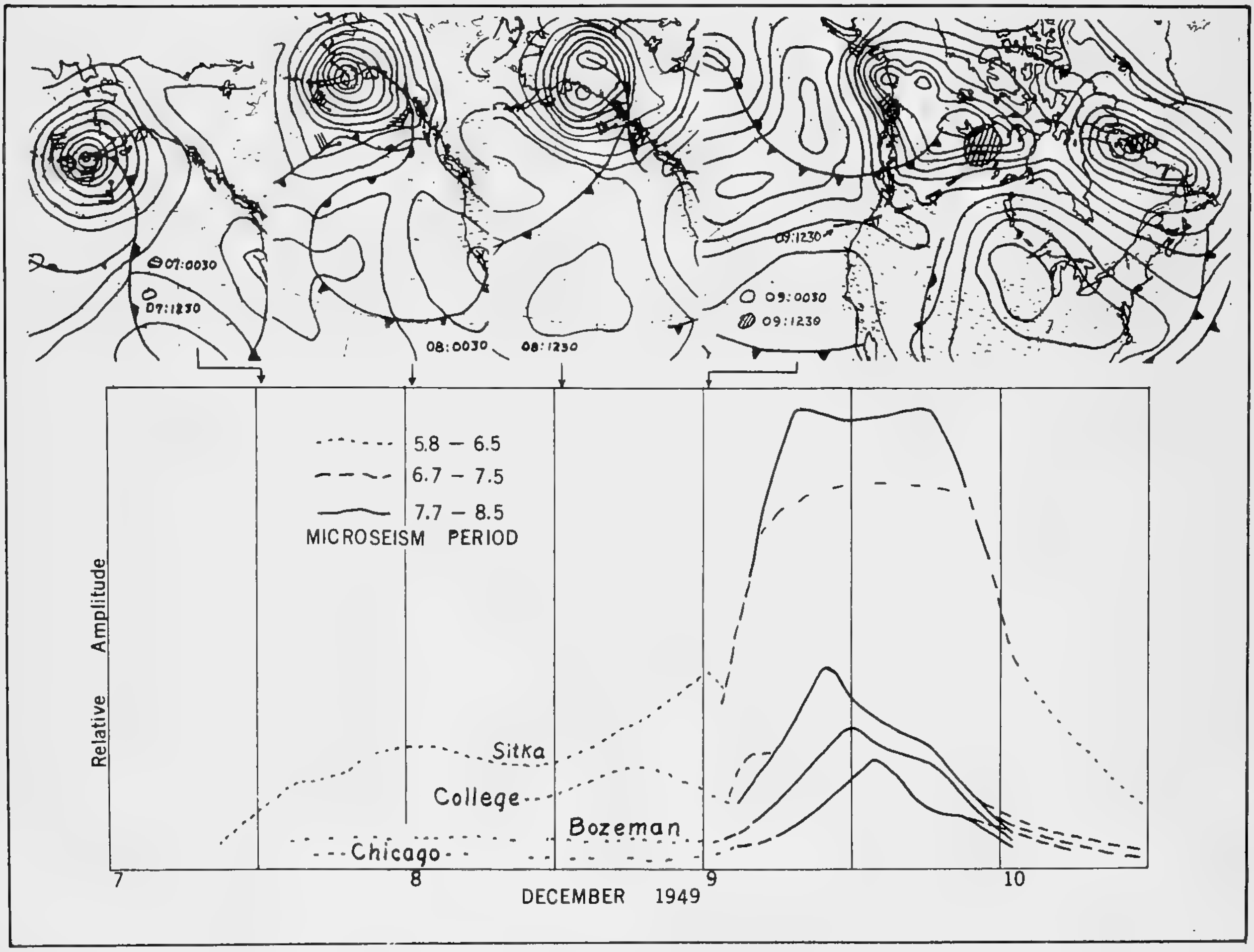

Figure 6. Weather charts 7-9 December 1949 showing a storm moving inland from the coast of Alaska, and a time-period-amplitude graph of microseisms recorded at certain North American stations. This storm is believed to have been responsible for the microseisms shown in figure 5, but the microseisms appeared after the storm had moved inland. Interval 5 millibars.

ocean swell viewpoint of microseism generation. There seems to be no other source of these microseisms except possibly the cold front moving off the Atlantic coast which produced 4 or 5 sec. microseisms that were also recorded at Chicago.

In the five cases presented here the microseisms under discussion were more nearly associated either with the high winds on the shore-directed limb of cyclonic storms hours or days before the microseisms appeared, or with the front from the storm sweeping inland or oceanward. Certainly there is no direct association between the microseisms and the storm center, or even the area in the storm about the center associated with a given wind velocity. In one case presented here the storm area had dissipated before the appearance of the microseisms.

The ocean swell idea of microseismic generation has the best support. If these five cases and also some of the examples of hurricanes presented by Gilmore are examined more critically, microseismic generation could very conceivably result in part from swells driven before winds on one limb of the storm meeting swells from oppositely directed winds of an earlier storm, or even the same storm if it has small area and is moving fast enough in accordance with the idea advanced by Deacon.

Other case histories of microseisms which were recorded during the past three seasons could have been used instead of those which were presented here, and would have told the same story. At no time could near-the-continent-generated microseisms be identified in force at Honolulu, Bermuda, or San Juan, indicating that the North Pacific or North Atlantic basins are effective absorbing media. On the continent, absorbtion across the western mountains is probably greater than across the central plains area, but the mountains certainly do not cut off the longer period microseisms. The Sierra Nevada, or more probably the Central Valley of California, are possible exceptions.

If the data presented in this paper are correctly interpreted and if the conditions outlined are general, it is inconceivable that strong microseisms on the continent can have their source far at sea.

This is supported by Gilmore's paper. On page — he states that two hurricanes 600 miles offshore send microseisms to Bermuda (probably from its Bermuda directed NE limb) but not to the nearest continental station 600 miles

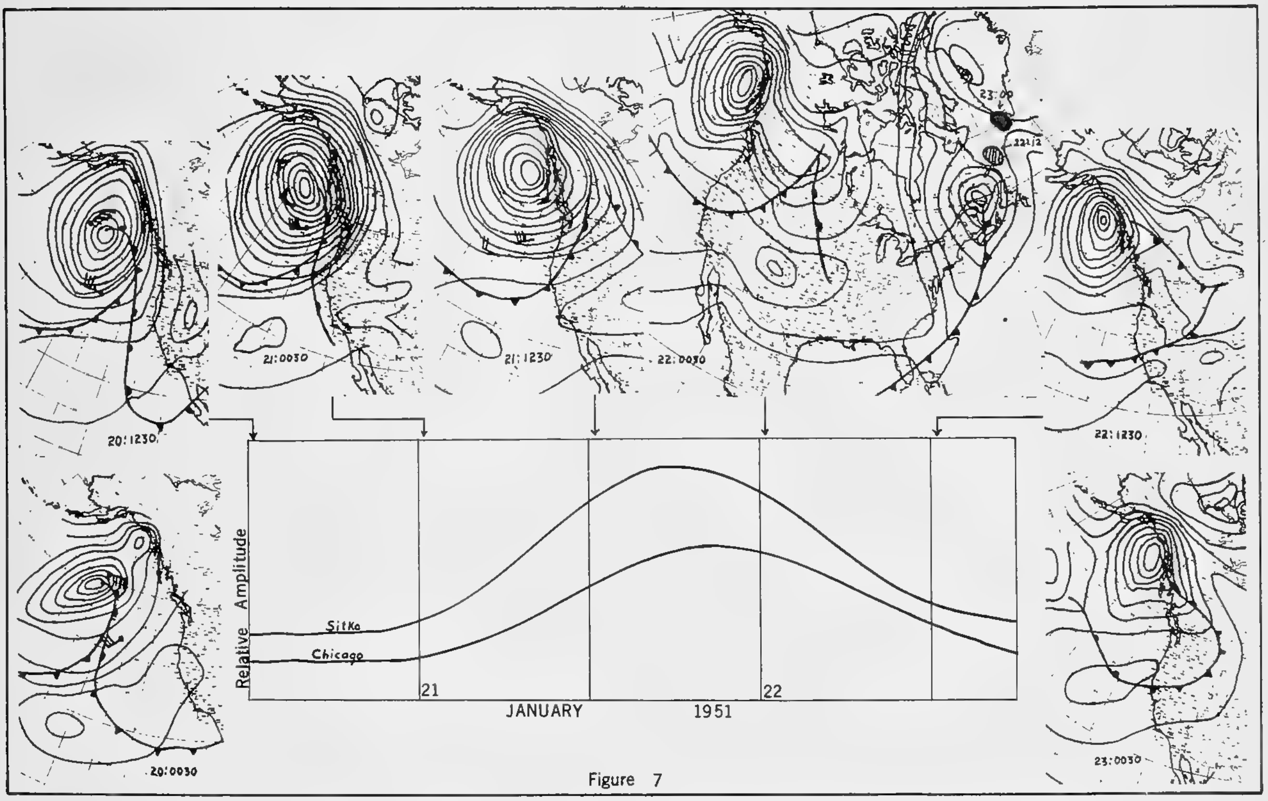

Figure 7. Weather charts 20-23 January 1951 showing a violent storm off the west coast of North America and time-amplitude graph of 7 sec. microseisms recorded at Sitka and Chicago during the same interval.

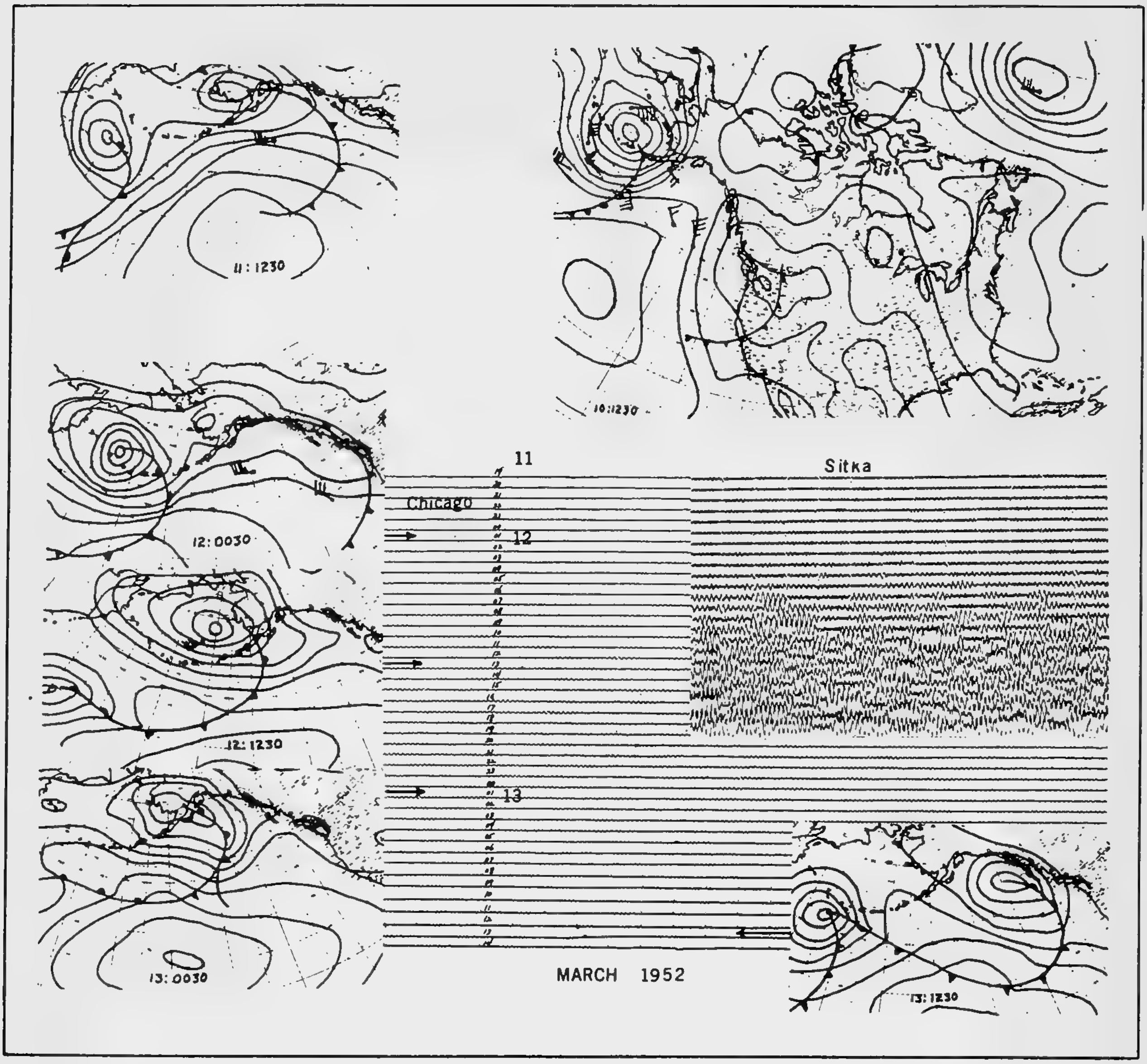

Figure 8. Microseisms recorded at Sitka and Chicago 11-13 March 1952 and weather charts 10-13 March 1952 showing successive northeast Pacific storms. Isobar interval about 7 millibars. The storm centering over the Aleutian Islands at 12:30 10 March, shown upper right, is believed responsible for the microseisms as illustrated, rather than the storm shown at left center which followed two days later. Note that the first storm had abated by the time the microseisms had reached their peak.

away. Yet two late August 1952 hurricanes on the coast of the Carolinas or southward sent microseisms as far as 1700 miles inland to Bozeman, Mont.

Referring to Figure 5 of Gilmore's paper the amplitudes at Whiting Field are about the same as at Jacksonville which is much closer to the storm center; and about twice the amplitude at Miami which is about the same distance as Whiting. The slightly greater ocean path to Jacksonville about offsets the additional land path to Whiting, but the somewhat greater ocean path to Miami is not nearly enough to compensate for the longer land path to Whiting.

Discussion

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Abstract—Although the application of empirical amplitude data to the problems of location and intensity of hurricanes eliminates some important sources of error inherent in tripartite operations there are still several factors which are at present adverse to operational success. These factors are considered here.

A new method of amplitude study is presented which employs resonant seismographs. These instruments permit discrimination among microseisms having different period bands and presumably different areas of generation.

Studies of simultaneous microseism and ocean wave and weather data together with local sensitive atmospheric pressure instruments suggest that atmospheric impulses resulting from the single or combined effects of turbulence, gustiness or pressure oscillations are the ultimate source of microseisms. The observations given all negate ocean waves or swell as being significant in the generation of the type of microseisms under discussion.

Introduction—A study of the intensity and variations of microseism amplitudes affords much information applicable to the practical use, and the problem of origin of microseisms. The use of such data has been presented in detail by Gilmore but the theoretical significance bearing on microseism origin, although stressed, was not supported strongly by quantitative or correlative data. The discussion given here is hence organized in three parts: (1) an evaluation of the use of amplitude data in describing the location and intensities of hurricanes or violent storms; (2) the introduction of new instrumentation and resulting data for the study of microseism amplitudes; and (3) a discussion of the mechanisms of origin of microseisms based on empirical amplitude data.

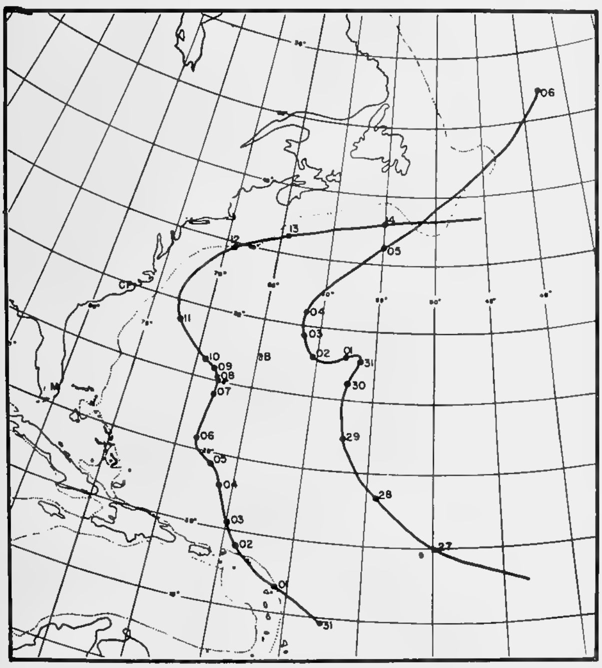

Figure 1. Paths of the hurricanes of August 27-Sept. 6, and August 31-Sept. 14, 1950.

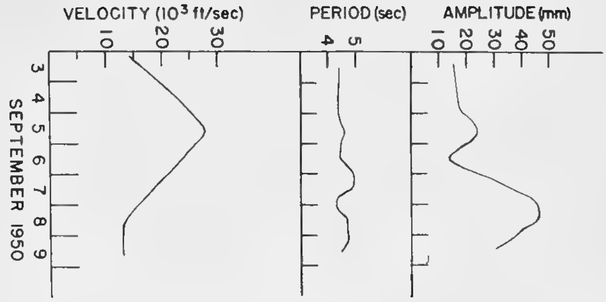

Figure 2. Bermuda microseism data for Sept. 3-9, 1950.

Part I—Amplitude Data Applied to the Position and Intensity of Storms—The amplitude distribution and micro-ratio methods of Gilmore seem to have definite, although limited success for the two cases given. It is noted that storm fixes by his method have an accuracy which varies from about 35 to 300 nautical miles, with the average being higher than, but closer to the lower value. Since this is strictly an empirical procedure the question now is whether continued success can be obtained or whether the cases shown are ideal for the procedure. Even these cases show "fixes" over relatively small portions of the hurricanes' paths. Hurricane No. 1, 1950 was plotted on weather charts as early as August 13 but the first fix is given on August 18. The procedure could not be applied to Hurricane "How," 1951, during the early and more critical part of its course. Prolonged experience in microseisms study suggests that a number of the difficulties that beset tripartite operation still have an adverse effect on the application of amplitude distribution, together some that are not involved in tripartite stud. However, the micro-ratio technique eliminates the effects of refraction, which has been a major obstacle in tripartite work. It also eliminates any effect of possible short-crested microseisms, the presence of which would introduce serious tripartite errors. A discussion of the sources of error that should be considered in connection with the further application of the micro-ratio method is given below, together with suggestions for minimizing or removing them when possible.

1. The presence of two storms within range of the recording seismographs will render the micro-ratio method subject to considerable ambiguity. This is a very real problem. During the 1950 hurricane season in the

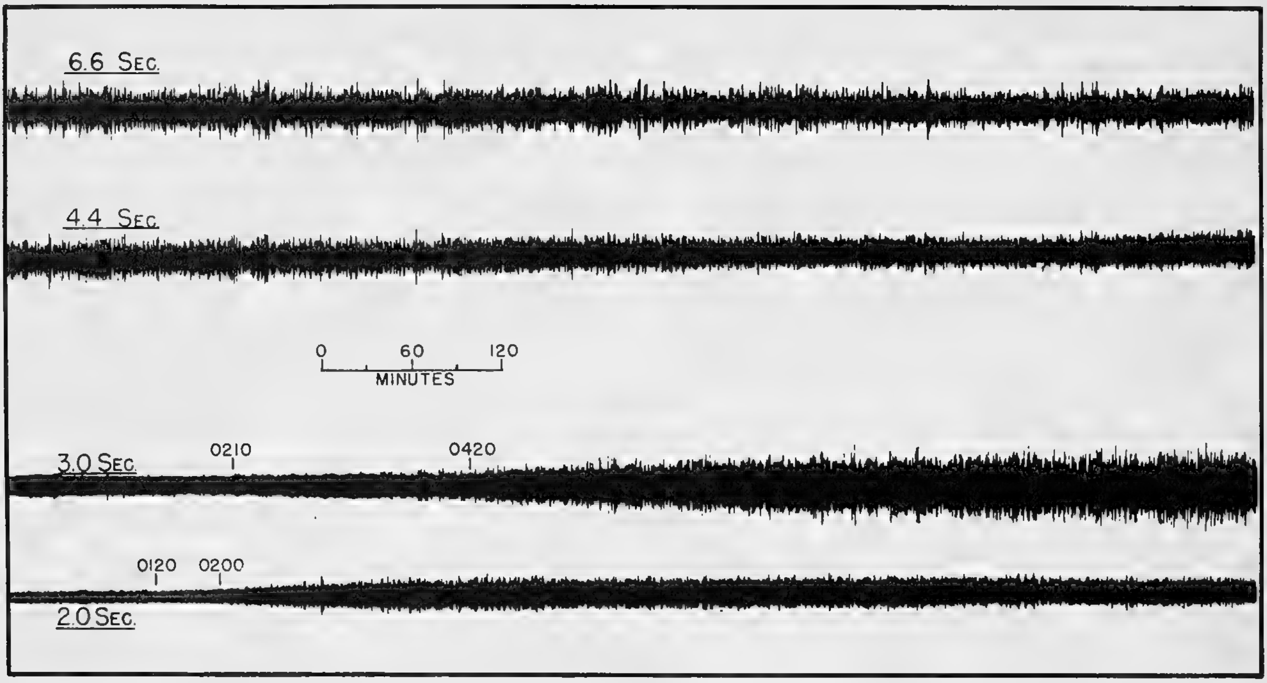

Figure 3. Palisades resonant seismograph traces for May 5, 1952. Time shown is GCT.

western North Atlantic Ocean two and occasionally three hurricanes were present simultaneously on 25 days and were in a position to cause ambiguity at most of the hurricane tracking stations. This does not include cases when strong low-pressure areas could also have produced ambiguous results.

In this connection it may be of interest to consider one of the cases given by Gilmore, namely the hurricane of September 3 to 9, 1950. It should be noted that for a portion of this interval and preceding the time of Gilmore's study, two hurricanes existed in opposition to Bermuda. One was to the south and approaching, and the other to the north, and receding, as shown in Figure 1. Figure 2 shows a first amplitude maximum (24mm) corresponding exactly to the time of velocity maximum (for the waves across the net) and showed the same trend. It should be noted that the value of 30,000 ft/sec was the average of the measured velocities across the Bermuda tripartite station for the minority of the waves present. Most of the waves recorded at each element of the net appeared to arrive at exactly the same time indicating infinite velocities or a standing wave pattern from oppositely moving wave trains. By selection of waves with measurable time differences, azimuths pointing toward both storms were obtained. The amplitudes during this interval (September 4-6) seem to be clearly the result of two hurricanes. The later, larger amplitude build-up was produced by the close approach of the storm from the south. Certainly estimates of intensity as well as position would be affected by the presence of two or more storms. If simultaneous hurricanes are well-separated, some stations in the detection net would respond uniquely to one storm, but a prior knowledge of their existence and position would be necessary before the data would be useful. Another possible solution is given in the second part of this paper.

2. Relatively local high wind areas can introduce considerable ambiguity in the application of amplitude distribution studies. Assuming that the microseisms of local wind and hurricane origin can be successfully distinguished (which is not always possible), the effect of microseisms of local origin must be removed before amplitude ratios are determined. This has frequently proved a matter of some difficulty and much ambiguity in the writer's study of hurricane microseisms. Interpretation of the hurricane intensity would be similarly subject to error.

3. The size and shape of the generating storm can also effect estimates of their position and intensity. Microseism amplitudes have been described as being a function of both area and intensity of an atmospheric disturbance [Donn, 1952a]. The strongest microseisms observed by the writer were produced by a large area cyclone with 35 to 40 knot winds while closer but smaller hurricanes with 100 knot winds generated much smaller microseisms at the same station. It is believed that the intense wind areas of hurricanes vary sufficiently in size so that estimates of velocity based only on microseism amplitude can be unreliable. Elongated storms should result in

poorer fixes than those with circular shapes, especially since the wind velocity may have considerable variation in different sectors of a moving hurricane.

4. It has been shown [Donn 1951] that amplitudes are a function of the spectrum of the microseism periods and the tuning of the receiving instruments. Much ambiguity could be removed in the study and application of amplitude distribution by the use of instruments sharply tuned to the periods shown by hurricane microseisms for a given region. This would eliminate much local high wind microseisms usually of short-period, and might distinguish between microseisms from simultaneous hurricanes providing they are of different periods. This will be referred to again in Part II.

5. The use of horizontal-component seismographs in making amplitude distribution and ratio studies introduces possible sources of error. The equal-amplitude lines around a station will tend to be elongated in the direction of pendulum movement, as found by Gilmore. This has also been shown in studies of amplitudes from horizontal components oriented at right angles, and is consistent with the Rayleigh-wave idea of microseisms. The comparison of amplitudes among several stations would be more uniform if vertical components were used. It has further been shown [Lee 1935] that amplitudes from vertical components are less affected by local geologic differences than those from horizontal components.

6. It is believed that the amplitude distribution and micro-ratio procedures can operate best when microseism amplitudes are much above background since this decreases the error in making comparisons and ratios with other stations. This may explain the difficulties in applying the micro-ratio procedure to the first five days of Hurricane No. 1, 1950, referred to earlier. Since the tripartite method requires only the recognition of the first sinusoidal waves of a new microseism storm, it may still be more applicable for early location of hurricanes. Both procedures can probably be refined beyond their present stages and may be valuable supplementary tools.

In view of the factors just considered it appears that the converse of the observation that a hurricane of a given intensity will at a given position generate microseisms of a definite amplitude at a particular station, is not always true. However, this is the basis for the application of amplitude distribution data to the determination of the position and intensity of a hurricane.

Part II—Amplitude Studies by Means of Resonant Seismometers—It is considered an empirical fact at least for east-coast stations, that the microseisms period-spectrum varies

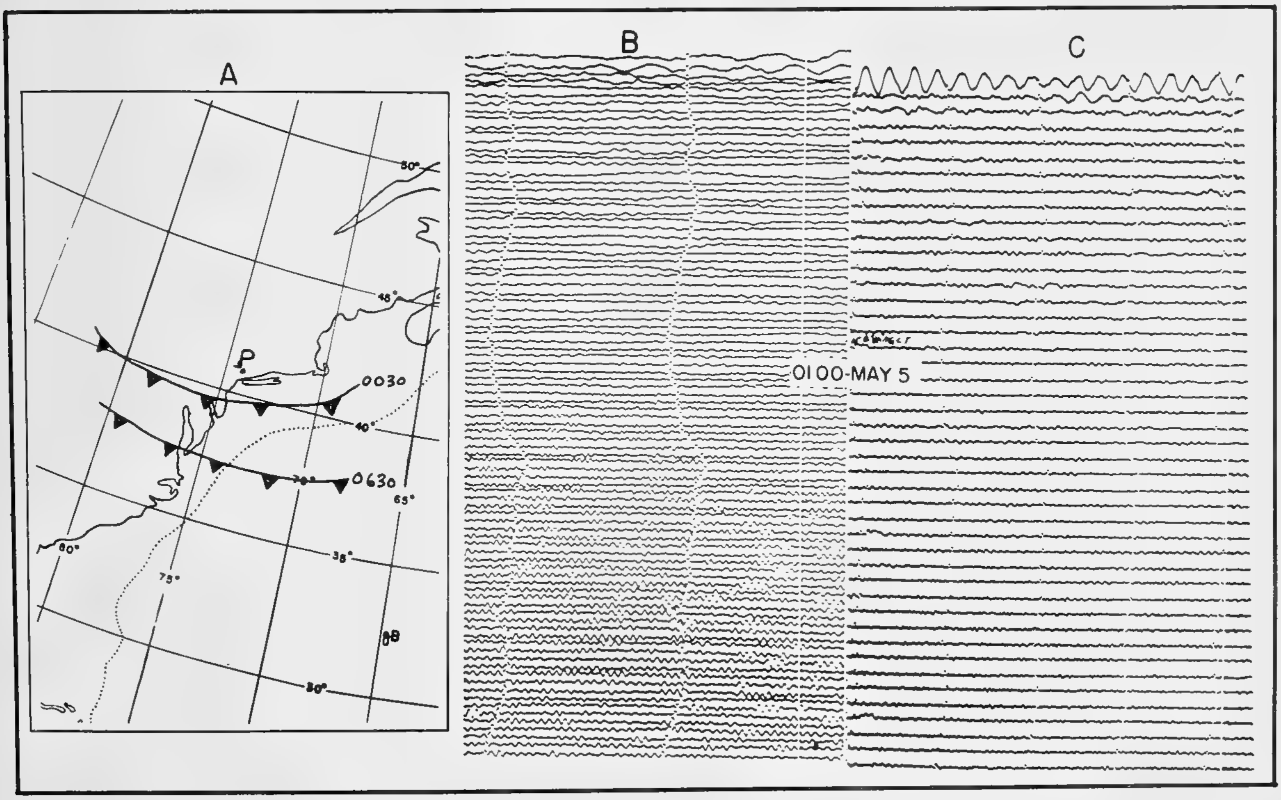

Figure 4. Cold front positions (A) and Benioff (B) and Columbia (C) verticals for May 5, 1952.

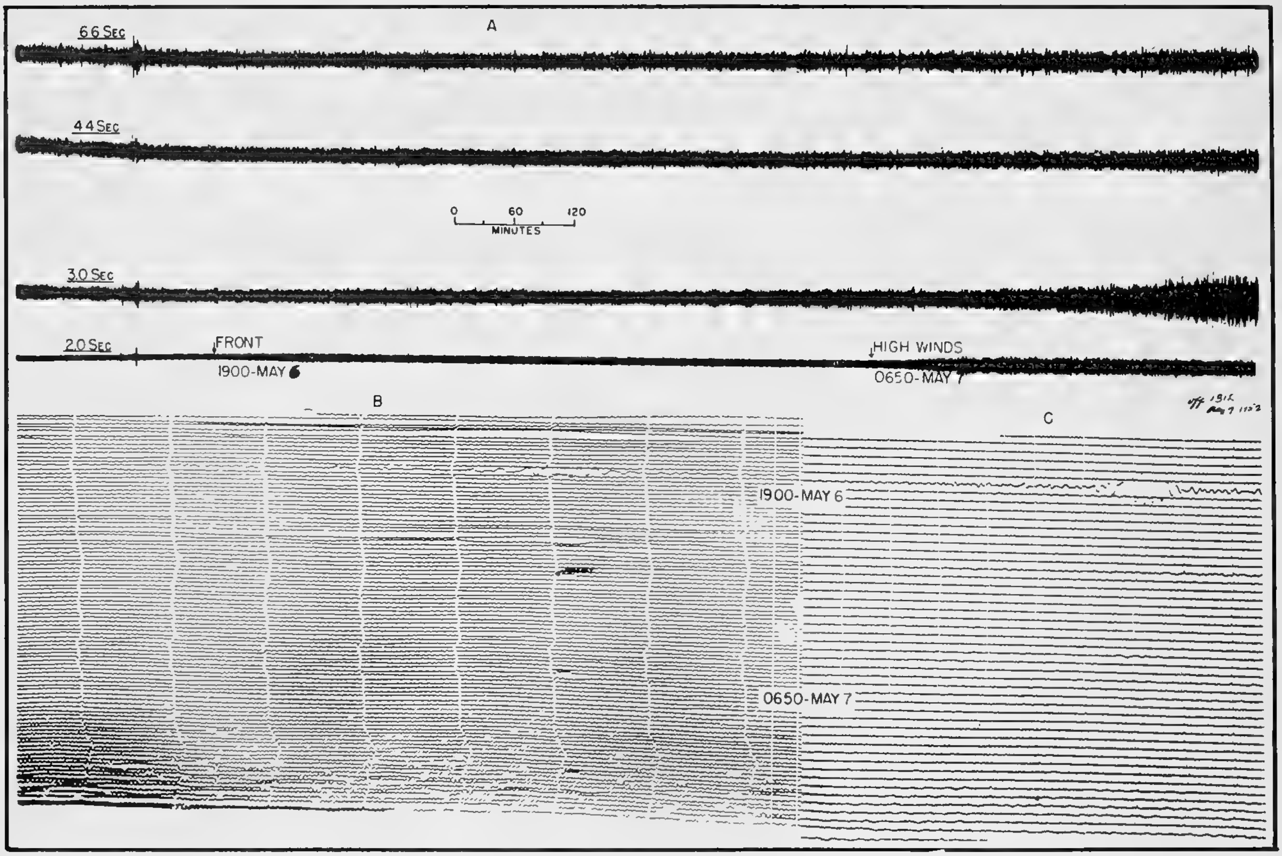

Figure 5. Comparison of Palisades resonant seismograph traces (A) and Benioff and Columbia vertical records (B and C) for May 6-7, 1952.

in width and position with the position of, and environment beneath meteorologic disturbances responsible for the microseism storms. In order to select microseisms from a particular environment a series of sharply-tuned vertical seismometers were designed at Lamont Geological Observatory and recently put into operation. Although this study is in a preliminary stage and a later report will give a full description of the instruments it seems worth describing some of the data and results here.

During most of the time of operation the resonant periods were set at 2, 3, 4.4 and 6.4 seconds, and 15 second Leeds and Northrop galvanometers are used. Although not yet calibrated the instruments are considered to have "very sharp" tuning, which seems supported by a comparison of records. Also by comparison with other instruments at the Lamont Geological Observatory at Palisades, New York, the magnifications have been estimated for the 2-second seismograph at about 35,000, and for the 3-, 4.4-, and 6.4-second seismographs at about 50,000. The seismometers are undamped with the pendulums having a logarithmic decrement of about ¼. In order to attempt rapid study of the onset times of microseism storms, and the envelope of amplitudes without the laborious measurements heretofore used, these instruments record on drums making one rotation in 25 hours. Although a time scale is given on the illustrations to follow, since no time marks were made on the early records, automatic time marks are being imposed on current records.

A few cases are given below which illustrate the nature of the data and results obtained so far, and compare them with other records made at Palisades.

Case 1. May 5, 1952—Figure 3 shows a portion of the records made (from bottom to top) by seismometers whose free periods are indicated on each trace. A 1.5- second instrument has just been added to this array. On the 2-second trace a detectable amplitude increase occurs about 0120, with a slightly stronger rise at 0200. The 3-second trace shows a detectable rise at 0210 followed by a stronger one about 0430. It can be noted that the 2-second trace is returning to background level while the 3-second trace is still high. During this interval, the upper-long-period traces are at normal background level which appears higher than background for the shorter-period instruments. It is evident that real amplitude variations of several minutes occur, some of which show sharp peaks that may correspond to the maximum wave or wave groups common

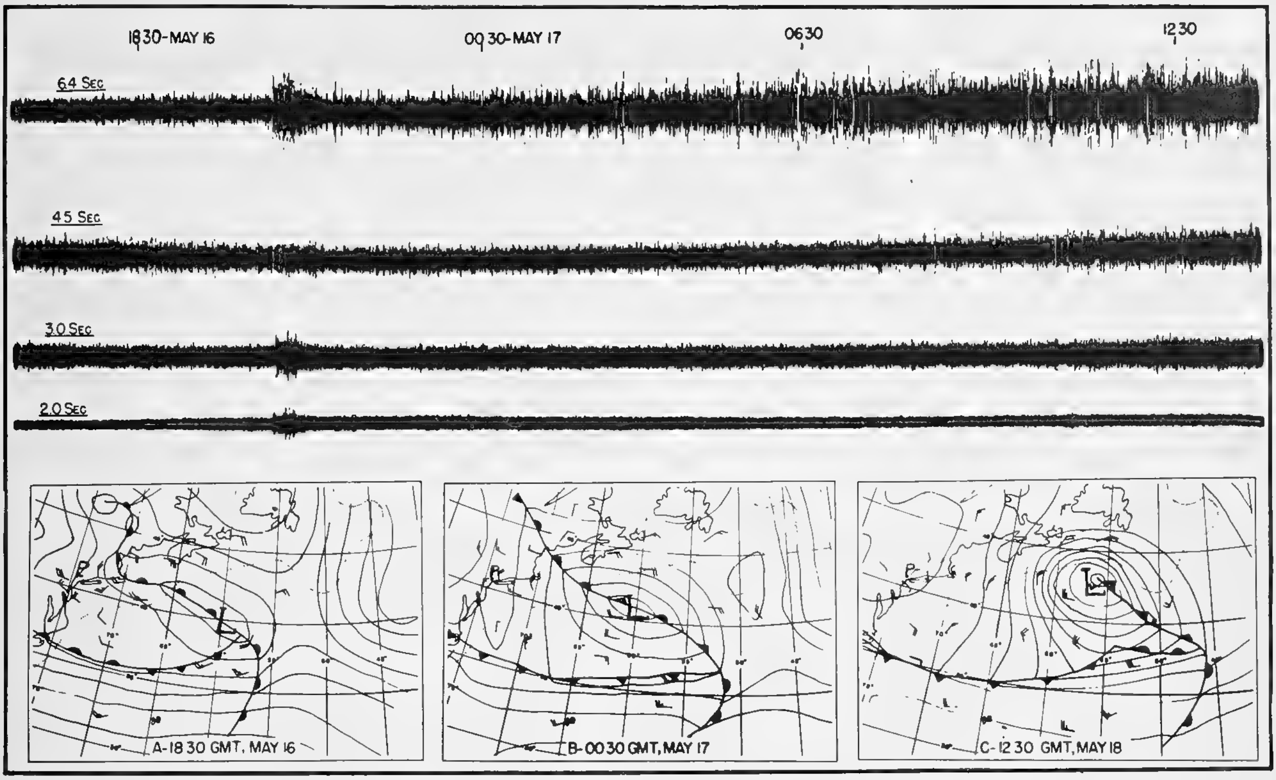

Figure 6. Palisades resonant seismograph traces for May 16-17, 1952 and charts of the associated marine cyclone development.

on conventional records. This will be investigated by running the instruments simultaneously at both speeds. It is significant that these peaks cannot be identified from trace to trace, indicating the sharpness of the tuning. This will be given more attention below.

Figure 4 shows the records of the Benioff vertical (B) and the Columbia vertical, (C), both at Palisades together with a chart (A) showing the 0030 and 0630 positions of the cold front associated with the microseism storm. (The dot below "P" on the chart marks the position of the Palisades station; the dotted line off the coast represents the 1,000 fathom depth contour.) The earliest detectable activity on the Benioff, between 0100 and 0200, shows microseisms of 1.4 to 1.6 seconds, with 2-second microseisms being recorded between 0200 and 0300. This suggests that the early activity on the 2-second trace may be a response to less than 2-second microseisms. It is hoped that the 1.5-second instrument will show greater magnification at this level and even earlier response to fronts. Although the Palisades vertical shows some discernible activity of short-period at this time, it is of no value for studying this situation.

Case 2. May 6-7, 1952—On Figure 5 A, the 2-second trace shows a gentle amplitude increase about 1900 GMT, on May 6. This corresponds within one to two hours with the offshore passage of a cold front in the neighborhood of the station. The other traces are at background at the time. A more prominent amplitude increase on the 2-second trace occurs at 0650 on May 7. This corresponds almost exactly with the time of abrupt increase of winds from the NW to force 4 and 5 over local, shallow waters near the station. The 3-second trace shows a later increase corresponding to the spreading of these winds over a more extensive water area. Although the Benioff (Figure 5 B) shows activity corresponding in time to the latter event, no evidence of the less intense earlier frontal micro-

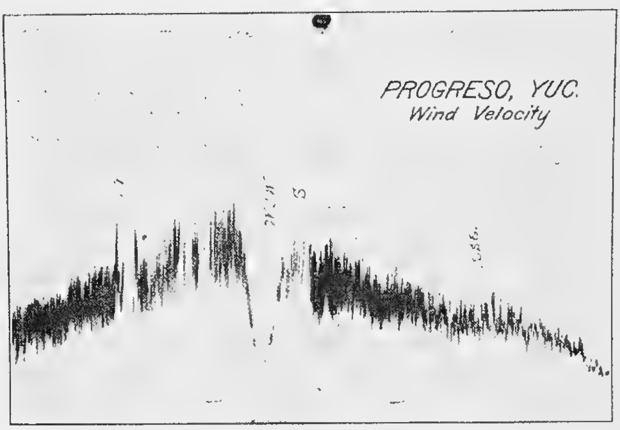

Figure 7. Wind velocity recording for Yucatan showing gustiness during the passage of a storm.

seisms is shown. The Columbia vertical shows only weak activity at the time of the latter event. The quake at 1740 is shown well on all traces.

Case 3. May 16-17, 1952- On Figure 6 the 2-second trace shows an amplitude increase about 1800, May 16, with the other traces showing gentle increases following the quake at 2100. The 4.5- and 6.4- second traces show very prominent activity during the first 12 hours of May 17, while the 2-second trace has returned nearly to background level. Weather chart A, (1830, May 16) Figure 6 indicates a

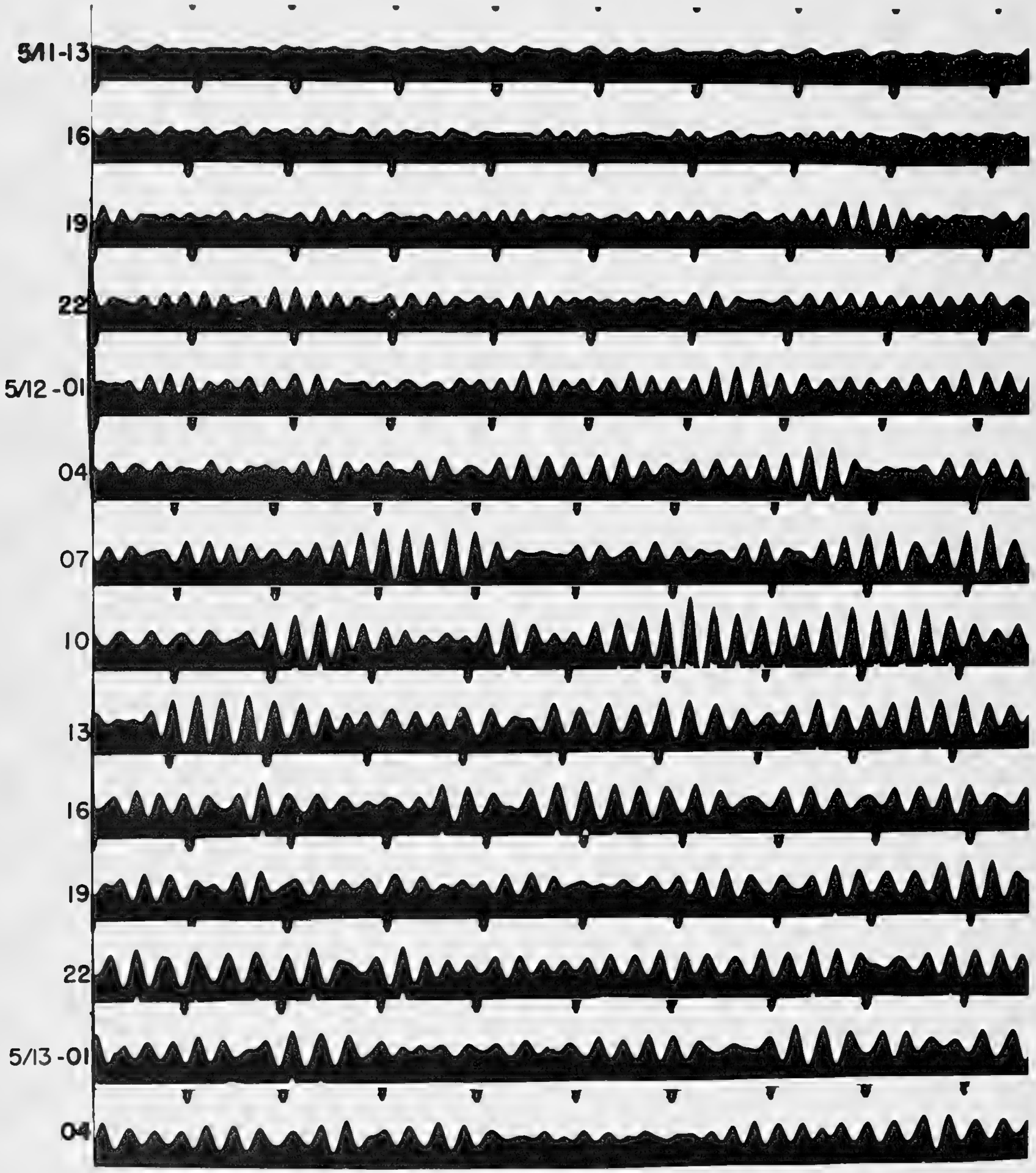

Figure 8A. Wave-gauge records of May 11-13, 1946 made near Woods Hole, Massachusetts.

low pressure area developing over near ocean waters with relatively high winds in the cool air over local shallow waters near the station. These winds appear responsible for the initial 2-second microseism increase. The rapid intensification of the "low" from 0030 to 1230, May 17, shown on charts "B" and "C", respectively occurring over deeper and more distant waters appears to be responsible for the generation of the long-period microseisms. This is consistent with all previous studies of similar situations. Wind velocities decreased over shallow waters as the "low" intensified and with this decrease occurred the decrease in 2-second activity. The conventional records show irregular long-period microseisms during this interval.

From the three cases given here, the resonant instruments seem capable of distinguishing between microseisms generated in different environments. The amplitude variations of several minutes duration, referred to earlier, are even more prominent on the records of case 3. Since these amplitude groups cannot be traced as simultaneous events among the different traces they suggest an origin in different parts of the generating area, and on the basis of this and data given in earlier studies, water depth is considered to be the dominating factor. If these groups were recorded with the same pattern simultaneously at different stations having equivalent instruments, it would tend to prove the relation of these groups to conditions in the generating areas. It may be of interest to compare these records with a wind-velocity recording, Figure 7, made during the passage of a storm. The wind record has the same time scale and shows similar amplitude variations. No conclusions are drawn from the obvious similarities, for doubtless, ocean wave records would have a similar appearance on this scale. The groups in all of these phenomena are related to conditions of origin and propagation, and their complete explanations are equally difficult.

In view of the response of the resonant seismometers as given in the study so far, their use in hurricane tracking seems to have definite advantages.

Part III—Amplitude Studies Applied To Problem Of Microseism Origin. Gilmore refers to the commencement of microseism storms with the primary generating hurricanes at a distance so great as to preclude the simultaneous arrival of swell in local waters. It is believed

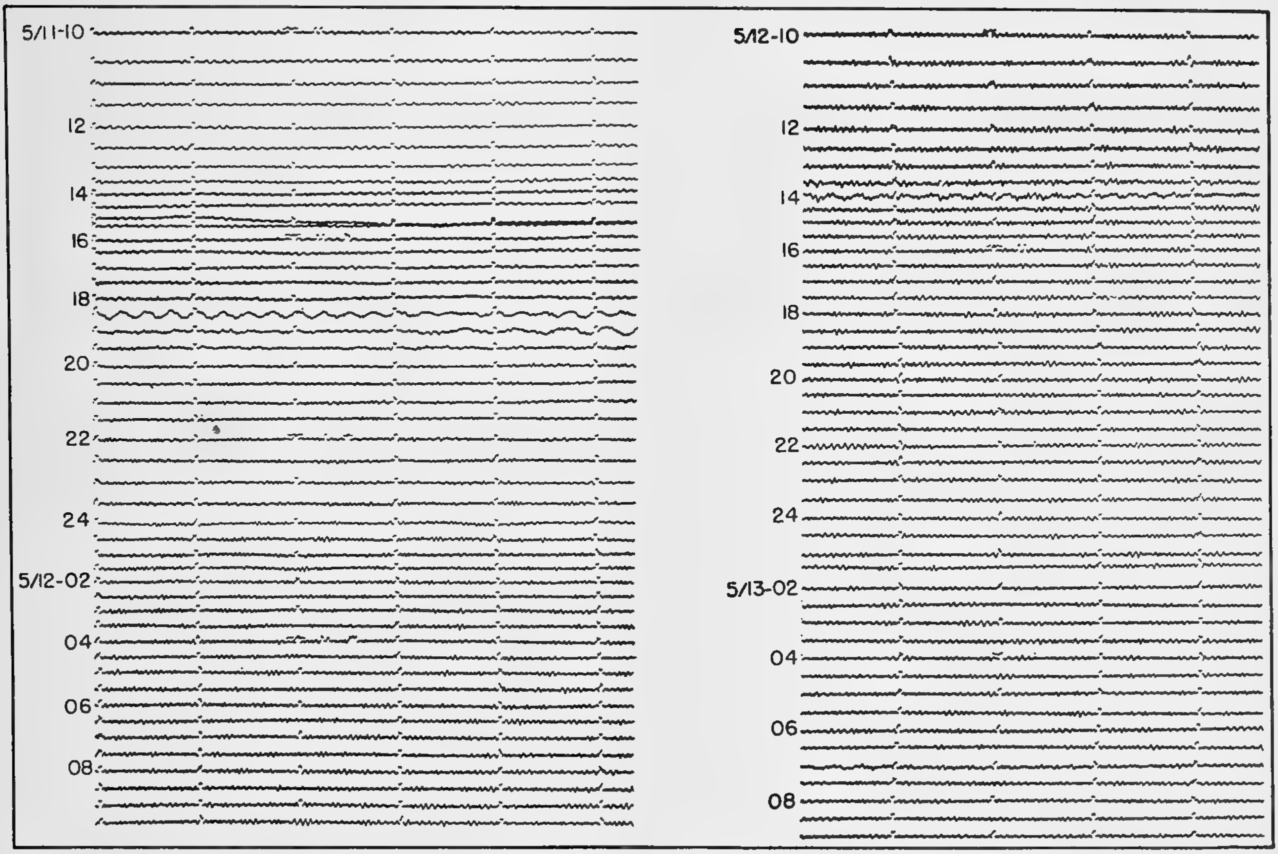

Figure 8B. Weston microseisms of May 11-13, 1946, recorded by long-period vertical.

that the significance of such amplitude studies in connection with the problem of microseism origin is high enough to warrant additional stress.

Many cases of simultaneous ocean wave and microseism recordings have been studied in connection with related weather conditions. These cases include all significant combinations of these interrelated conditions; namely, cases when (1) atmospheric storms were very prominent, (2) ocean waves were very prominent without significant atmospheric disturbances, (3) microseisms were especially high regardless of the first two conditions, and (4) ocean conditions were calm with prominent microseismic and atmospheric disturbances occurring. By this means it seemed possible to test empirically many of the proposed mechanisms of origin and determine whether any one could uniquely account for the variety of observed microseisms.

A recently published case (Donn 1952 b) gives data for the hurricane of September 13 to 16, 1946 as recorded at southeastern New England. Although prominent swell and strong microseisms were associated with the hurricane a study of microseism and wave recordings with simultaneous weather data eliminated the possibility of local high waves and swell in shallow water anywhere along the coast from being the microseism-exciting mechanism. A comparison of amplitudes with hurricane position related the microseism origin to conditions within the hurricane. Microseism period showed a trend opposite to that of recorded swell.

The two cases shown in Figures 8, A and B, and 9, A and B, which are taken from a recent report (Donn 1952 c) show ocean bottom pressure records of high magnitude near Woods Hole, Massachusetts, and the simultaneous long-period vertical microseism records made at Weston Observatory some 50 miles to the north. The ocean waves were generated by protracted onshore southerly winds of 4 to 6. Bottom pressure variations of 1/12 to 1/20 atmospheres were recorded yet the simultaneous microseism records are essentially at background level, showing none of the intense storm-type microseisms under discussion. However, a close examination of the microseism records does show low-amplitude, short-period

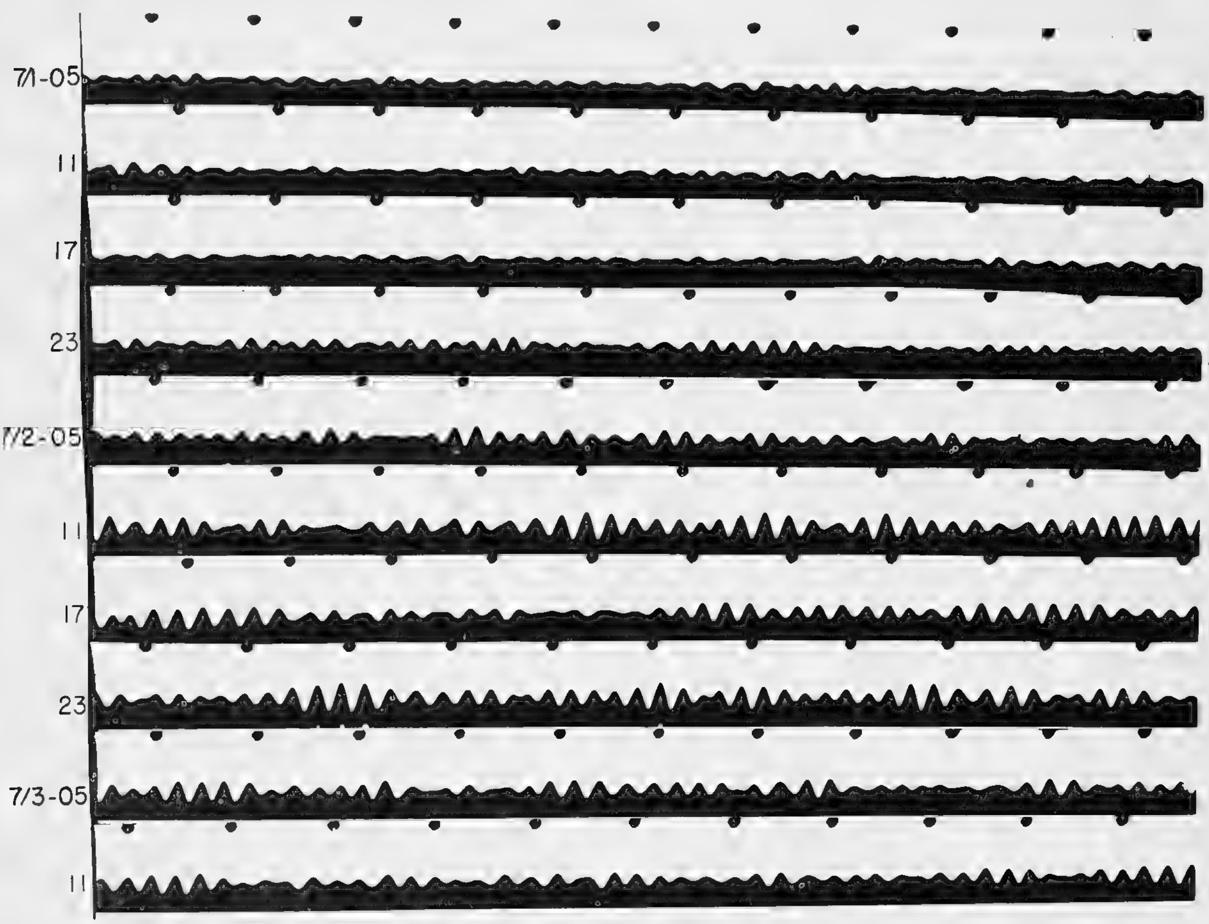

Figure 9A. Woods Hole wave records of July 1-3, 1946.

microseisms which match very well the times of beginning, maximum and termination of the ocean waves. It is suggested that these are the microseisms generated by the relatively strong pressure variations or impulses resulting either from the waves in shallow water or the surf action on the shore.

This should be contrasted with the very intense microseisms that develop almost as soon as a cold front passes seaward from land, as shown in Figure 10. In this case too, strong, onshore winds preceded the cold front which must have generated high waves. It may be safely assumed that in all the above cases, high ocean waves were general along the New England coast. If microseisms in this area are ever generated from standing waves resulting from reflection and interference of ocean waves, they should have been prominent in these and other similar cases. It may be argued that the shallow-water area of the strong progressive waves was too small for effective microseism generation through any shallow-water effect. But fronts and post-frontal disturbances produce microseisms when over as small or smaller a water area in the same locality. Any weak second order effect from interference between onshore waves and waves that may be set up by the offshore-moving front would be limited to the same small area. Since there would be a stiking difference between the periods of the two sets of waves and since an angle of greater than 10° occurs between ocean waves from the south and fronts from the west, the development of such an effect seems even less likely.

It has been argued that the bottom in the vicinity just described (off New England) is for some reason not conducive to microseism generation as an explanation of the negligible effect of strong ocean waves. But at other times the same bottom seems quite conducive to microseism generation by fronts or offshore cold winds of such a limited extent that no other area could possibly have been involved.

Two additional cases will be given now in which the sea surface off northern New Jersey (and presumably southern New England) was calm preceding and during the generation of strong microseism storms.

Case 1. Analyzed wave records of the Beach Erosion Board made from a bottom pressure gauge off Long Branch, New Jersey, give sea conditions as "calm" from February 12 to 15, 1952, with surface heights of 0.4 to 1.0 ft. and 9 to 10 seconds period for February 11.

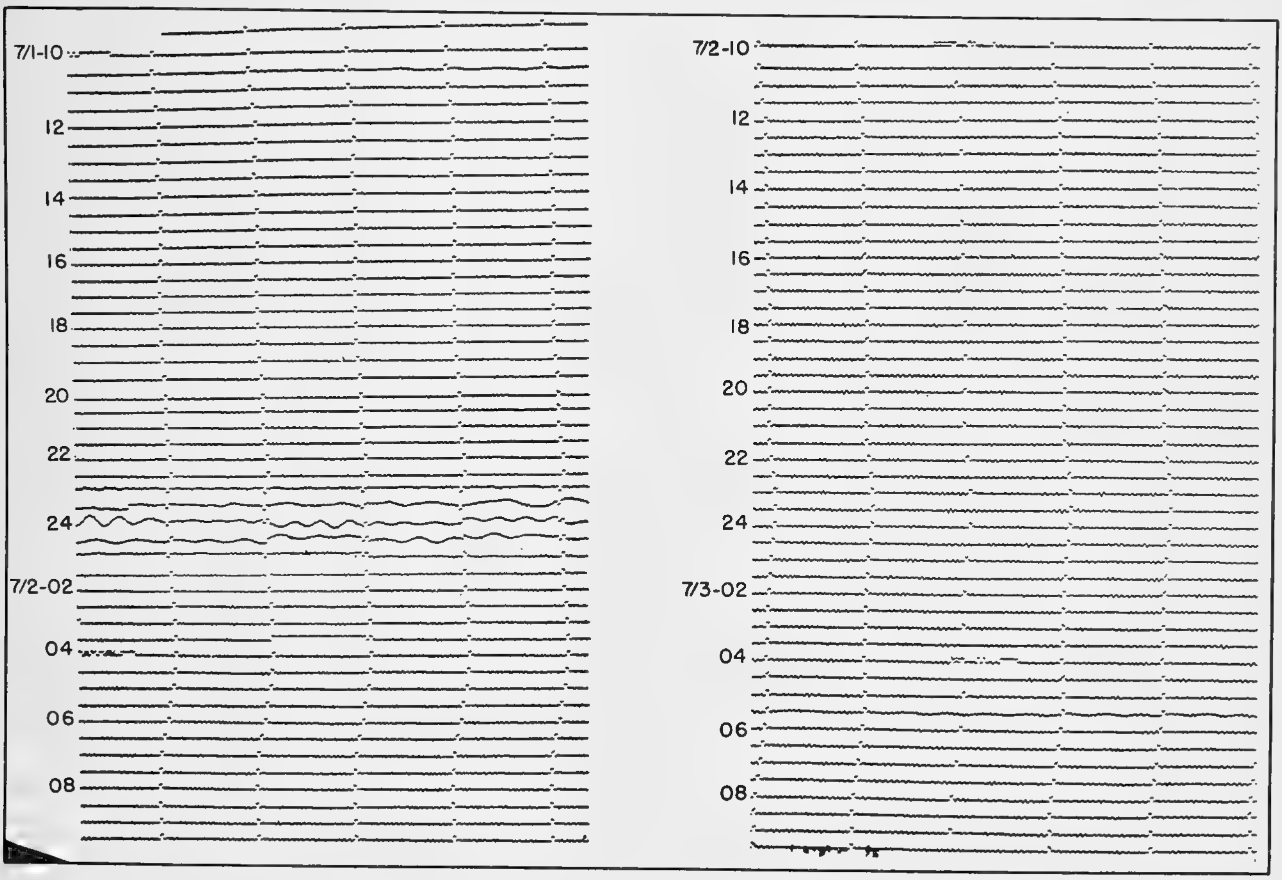

Figure 9B. Weston microseisms of July 1-3, 1946, recorded by long-period vertical.

Figure 11 E is a simplified surface weather chart for 1830, February 11, and shows a cold front which passed seaward over the coast some 3 to 4 hours earlier. The heavy broken line shows the frontal positions 6 hours earlier, at 1230. The dot below the "P" marks the Palisades station of the Lamont Geological Observatory where the records A to D of Fig. 11 were made. "A" is from the seismically compensated microbarograph (6); "B" is a recording from the hot-wire microphone; "C" is from a vertical component seismograph uncompensated for pressure changes and "D" is from the Columbia vertical component electronic seismograph.

The microseisms on "D" can be detected earliest at about the time of the 1830 weather chart and have a period of 2.5 seconds. Earlier studies (Donn 1951) with shorter-period instruments have detected microseism onset when the front was much closer to the coast. Both the sensitive microbarograph and the hot-wire microphone show pressure changes of different, but relatively short-period, beginning about 1500 to 1600, when the front had probably just passed the coast. This would place the Palisades instruments in the cooler turbulent zone just to the rear of the front. The existence of this zone has been confirmed by recent findings of Roschke (1952). Since Palisades is close to the coast these atmospheric conditions must have passed seaward very soon thereafter, at just about the time of commencement of the microseism storm. Much more severe atmospheric disturbances commenced at Palisades at 0200 probably in the main mass of cold air following an unmarked secondary cold front (which is actually shown on earlier weather maps). The microseisms on "D" show marked and continued increase as the turbulent cold air passed seaward and continued for the following day, with calm sea conditions prevailing. It is interesting to note that seismogram "C", which is uncompensated for pressure changes shows simultaneous microseisms and somewhat longer-period pressure changes shown by long-period, rather irregular waves which can be

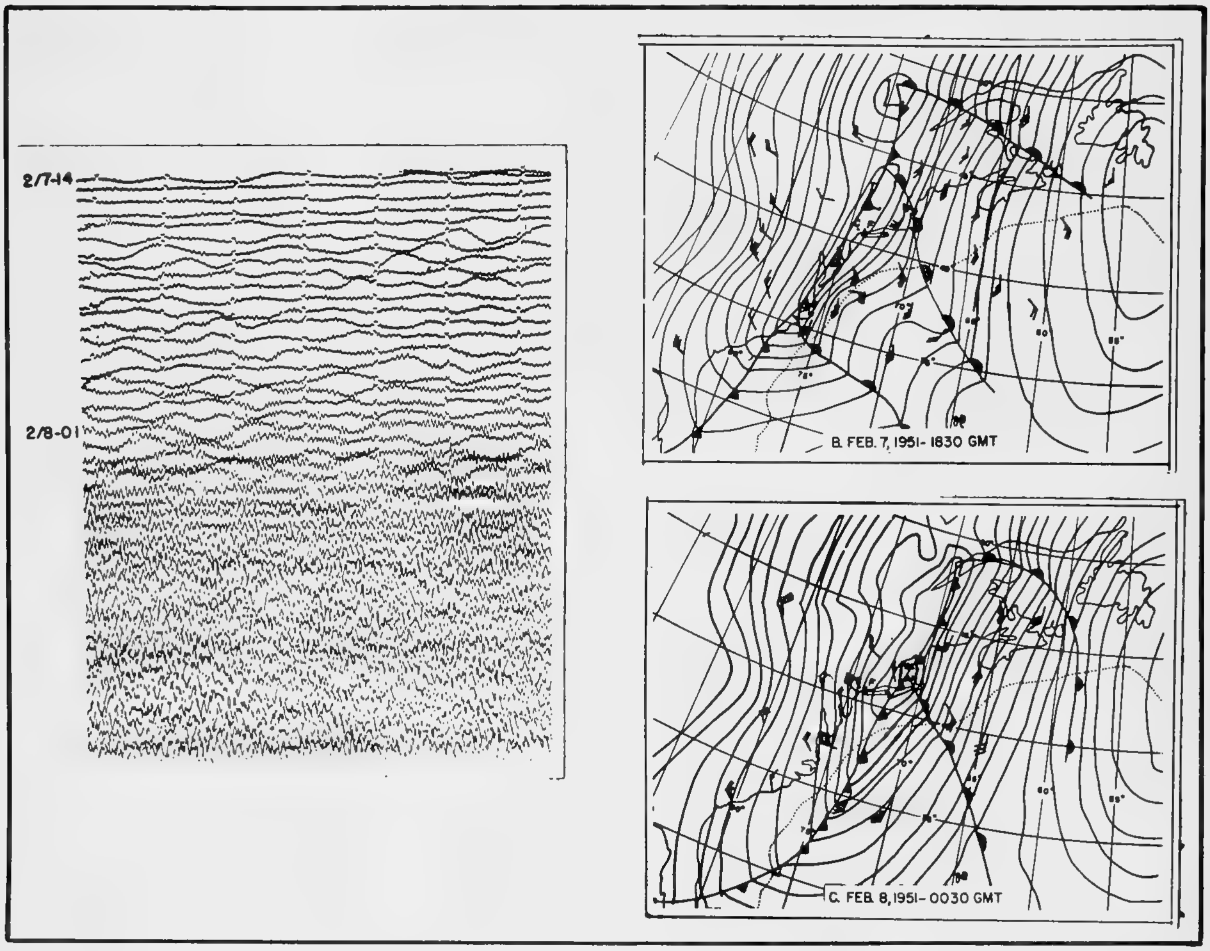

Figure 10. Portion of Weston long-period vertical recorded for February 7-8, 1951, and charts of related marine weather conditions.

matched with those on record "A". In addition to pointing to a direct meteorologic origin for the microseisms, this case also suggests that bearings or azimuths computed by any of the procedures in use should not be expected to give a direction to the center of an extra-tropical cyclone, but rather to the center of the most turbulent zone or the cold air mass.

Case 2. Calm seas existed in the same area from March 4 at 2000 to March 6 at 1200 according to wave records of the Beach Erosion Board. Figure 12 D is a simplified surface weather chart for 0030, March 5, 1952 (Palisades is shown by the dot to the right of "P"). A vigorous cold front is seen approaching Palisades. The sensitive microbarograph and hot-wire microphone records show short-period pressure fluctuations prior to and during frontal passage at Palisades, culminating in very severe variations. The latter, between 0200 and 0300, March 5, was attended by heavy showers and high winds which backed from NE to NW establishing frontal passage. During this time no short-period microseisms occurred, although long-period microseisms are present which can be correlated with a more distant storm present for some time prior to this record. However between 0400 and 0500 on March 5 the beginning of a new microseism storm was recorded showing fairly regular 2.5 second microseisms. This would just allow time for the atmospheric pressure disturbances recorded at Palisades to travel to local offshore waters. It would appear that the pressure disturbances recorded on the upper halfs of records "A" and "B" generated

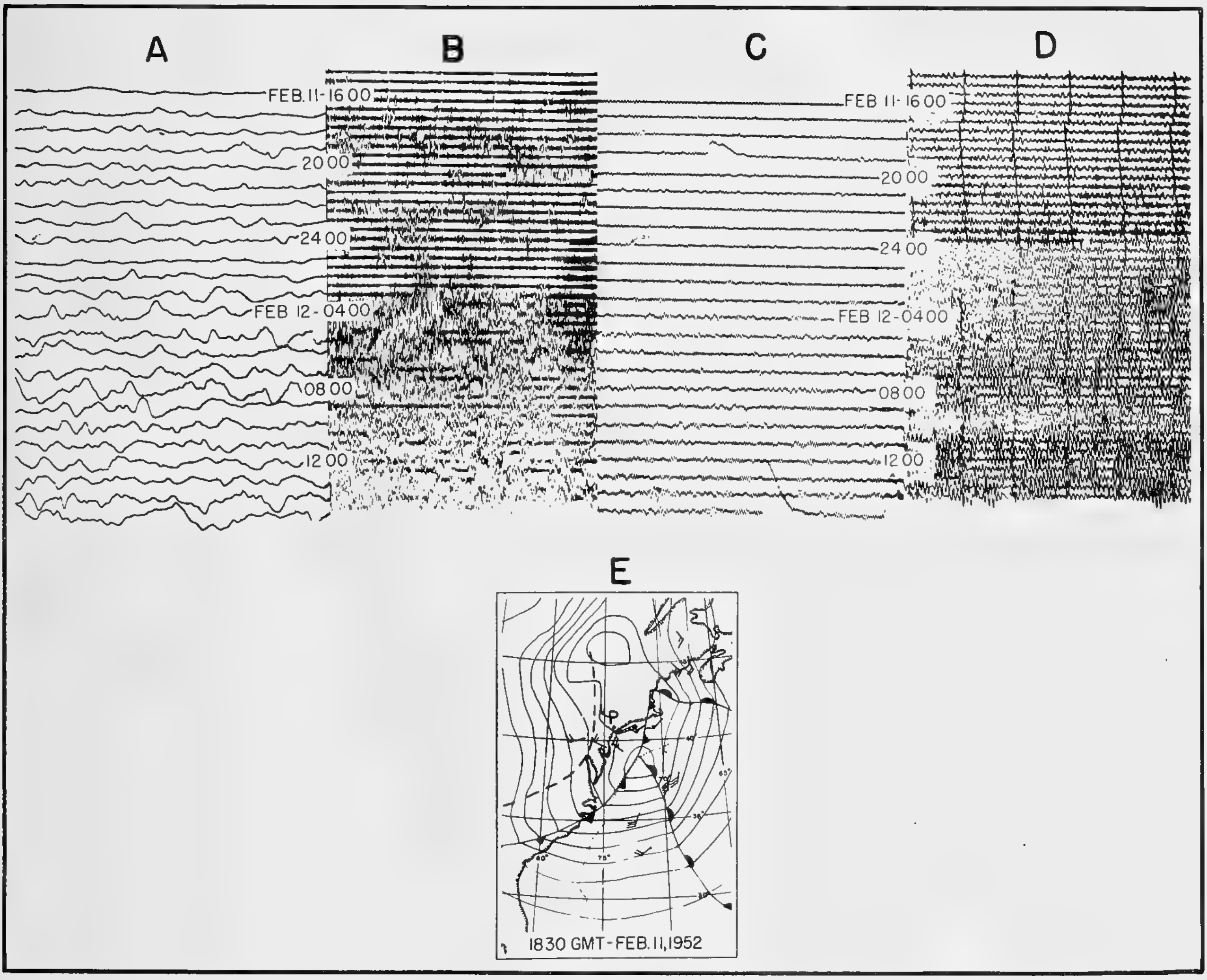

Figure 11. Atmospheric pressure, and microseisms at Palisades for February 11-12, 1952 together with related weather data. A - record of seismically compensated microbarograph (T_o= 10.5, T_g=76); B - record of hot-wire microphone; C - record from vertical seismograph, drum speed=15mm per minute (T_o=10; T_g=75); D - record from Columbia vertical component electronic seismograph (T_o=12), drum speed=30mm per minute.

the microseisms shown on the middle and lower parts of record "C".

This is consistent with earlier findings of microseisms associated with fronts and cold air masses (Donn 1951, Jones 1949). It may be interesting to compare these quotations from the studies of Donn in December 1951 and February 1952 with that of Roschke in June 1952: Donn — "Microseism intensity may be maintained at a high level by fresh to strong winds (in the cold air) which may follow a cold front. It seems significant that winds of similar strength in the warm air preceding a cold front have no noticeable effect in the production of microseism storms. This suggests the effect of gustiness or turbulence as being of special significance in microseism origin." "It is suggested that pulsations or oscillations in the air striking the water resulting from instability or turbulence in the cold air are coupled to the sea surface by some mechanism." Roschke — "It is shown that high-velocity flows of cold air are much more efficient mechanisms for producing extended intervals of maximum-amplitude micro-oscillations in the air than corresponding warm air flows."

Lee (1934) observed that although microseism storms can always be associated with some atmospheric disturbance, the intensity of the microseisms varied despite similar conditions of pressure gradient and winds within the atmospheric disturbance. This effect is more explainable on the basis of pressure oscillations in the air which will depend on factors of temperature, density, stability, etc. in addition to pressure gradient and wind force. The following observations given by Roschke are again applicable: "Extended intervals of maximum-amplitude micro-oscillations occur concurrently with the combination of a tight horizontal surface pressure gradient and a very cold polar air mass; however, the occurrence of either a tight pressure gradient or a particular air mass does not, of itself, signify a particular characteristic

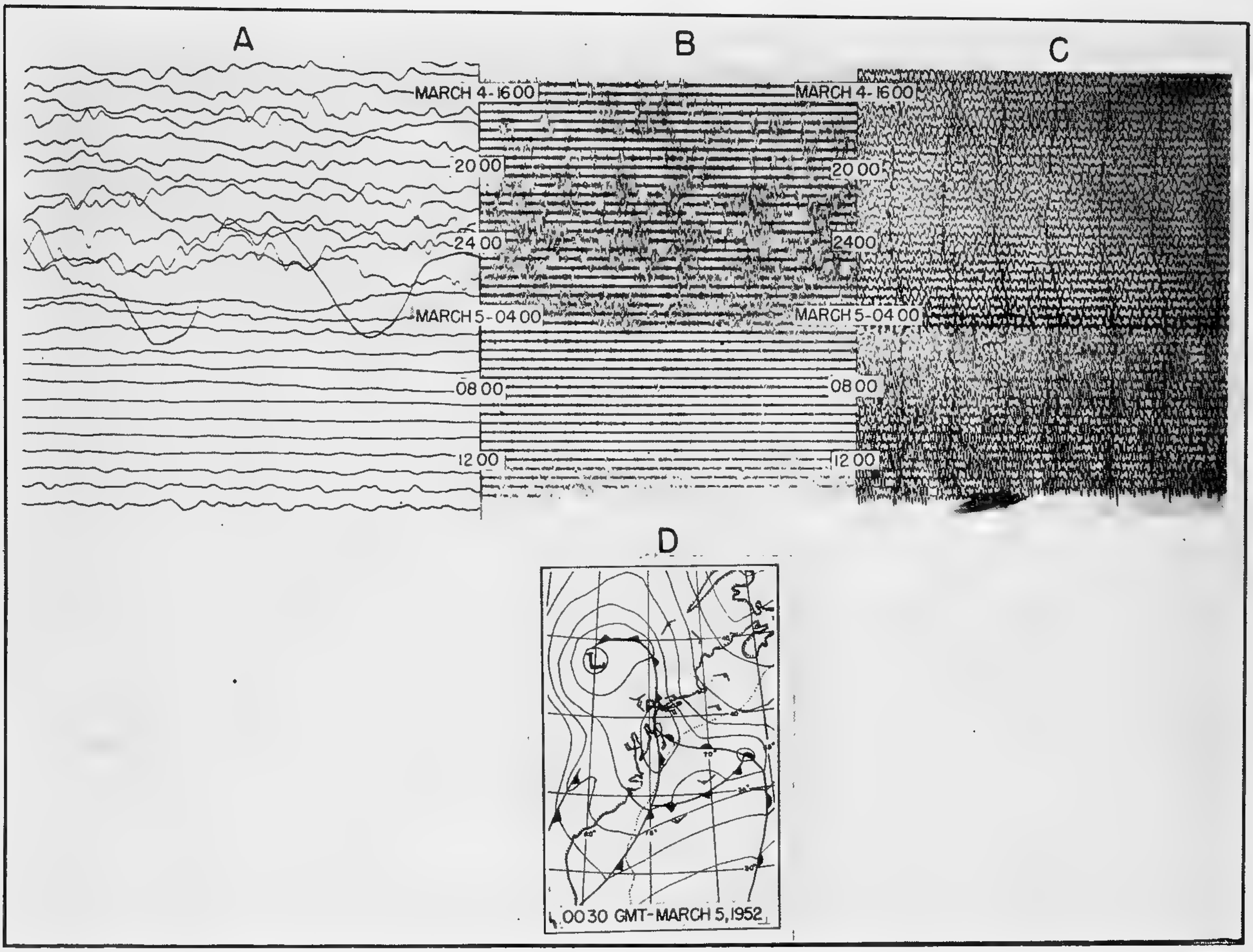

Figure 12. Palisades pressure and microseism data for March 4-5, 1952: A - record from microbarograph; B - record from hot-wire microphone, C - microseism record from Columbia vertical, D - weather data.

microbarographic activity."

Microseism storms of the short-period, irregular type shown here have been shown (Donn 1952 a) to grade continuously into regular long-period microseisms as the generating atmospheric disturbance moves to distant and deeper waters, with the microseism period appearing to be more a function of water depth than distance. This is shown on simultaneous recordings with Palisades short- and long-period seismographs in Figure 13, with the generating cyclone shown in Figure 14. The difference in the type of microseisms seems to be one of the position and environment of the generating disturbance. Microseism period is shown to increase continuously until the generating cyclone reached deepest ocean water. The reverse effect is expected when fronts or storms approach coastal stations from shoaling seas.

Conclusions—1. The amplitude distribution and micro-ratio technique for locating hurricanes and estimating their intensities is a strictly empirical procedure. Only further application will determine its operational value. Although the serious effects of refraction and possibly of short-crested microseism waves are eliminated by this new procedure there are still a number of serious adverse factors which would at present prevent application to all hurricanes and to early positions of many if not most hurricanes.

2. The use of resonant seismographs seems to permit the study of a narrow micro-

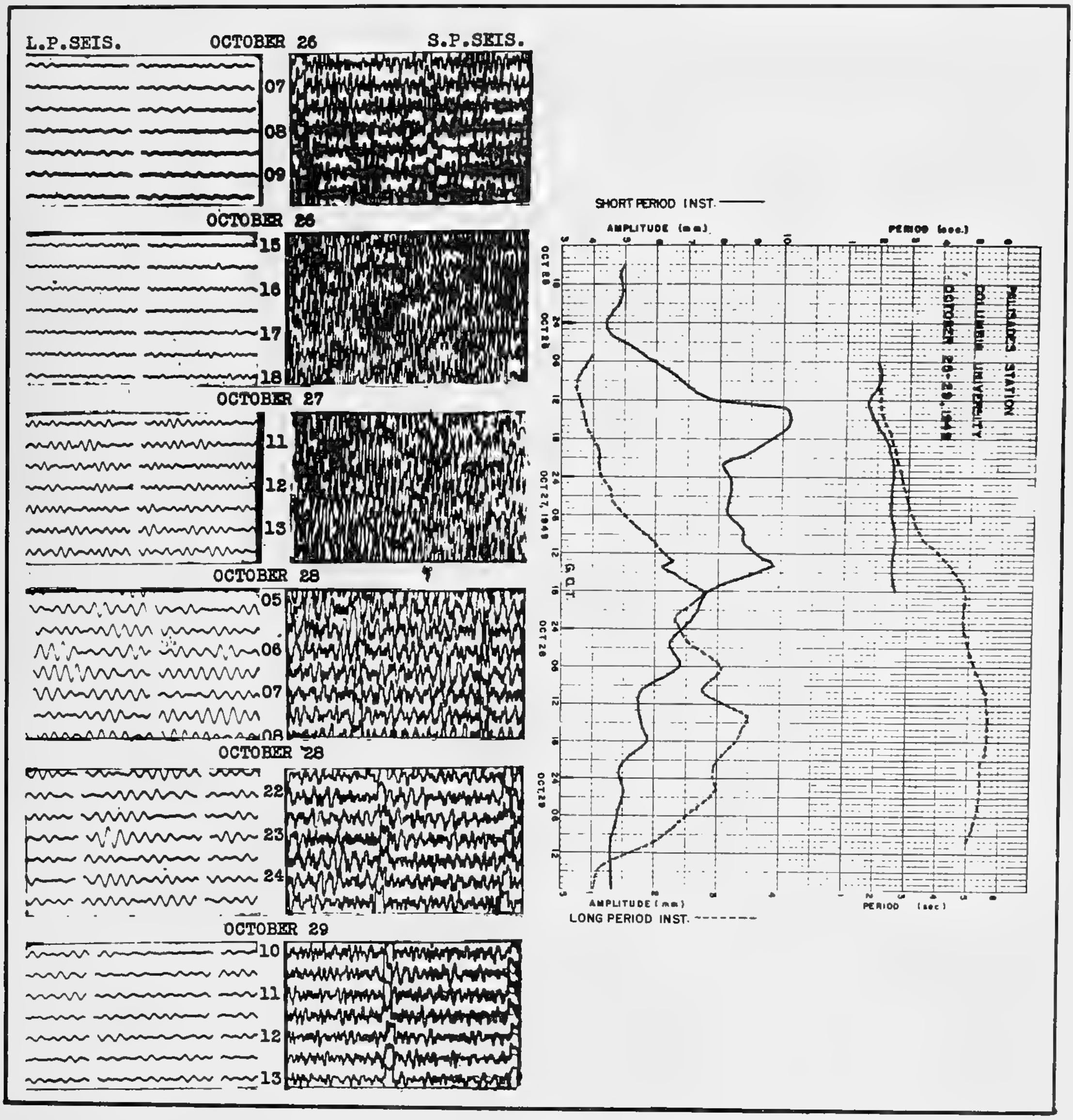

Figure 13. Comparison of microseism data from long- and short-period Palisades instruments for October 26-29, 1949 showing shift of microseism energy to longer period with the retreat of the cyclone shown in Figure 14.

seism-period band, and permits discrimination between microseisms from different generating areas, or different parts of one generating area. Further use of this new method of microseism study seems promising.

3. Empirical studies of the intensity and times of beginning and termination of amplitude changes during microseism storms recorded at east-coastal stations together with simultaneous weather and ocean wave data for east-coastal waters permit discrimination among suggested mechanisms of microseism origin. The only unique method of origin seems to lie in excitation within the area of an atmospheric disturbance and by direct coupling of the energy of some impulsive air disturbances to the sea surface. At lower than hurricane wind velocities, cold air is a much more efficient microseism source than warm air. All of the observations made on the east coast negate ocean waves or swell (whether progressive or standing) as being transitional in the generation of microseisms from an original energy source in the air. This appears to conflict with observations and interpretations made on the western coasts of Europe and North America. Since these coasts are the targets for both storms and their associated ocean waves which generally travel westward in the latitudes of microseism study a clear possibility of ambiguity exists. Even here it has been interpreted that ocean waves produced by the cold sectors of storms are especially efficient in microseism generation. On the east coast, where storms move offshore, it is possible to distinguish between these factors as shown in this paper.

Acknowledgments—The study and instrumentation involved in this research was supported by Contract N6-onr-27133 and Contract AF19 (122)441 between Columbia University and the Office of Naval Research and the Geophysical Research Division of the Air Force Cambridge Research Center, respectively. Weather data was supplied by the United States Weather Bureau Office at La Guardia Field.

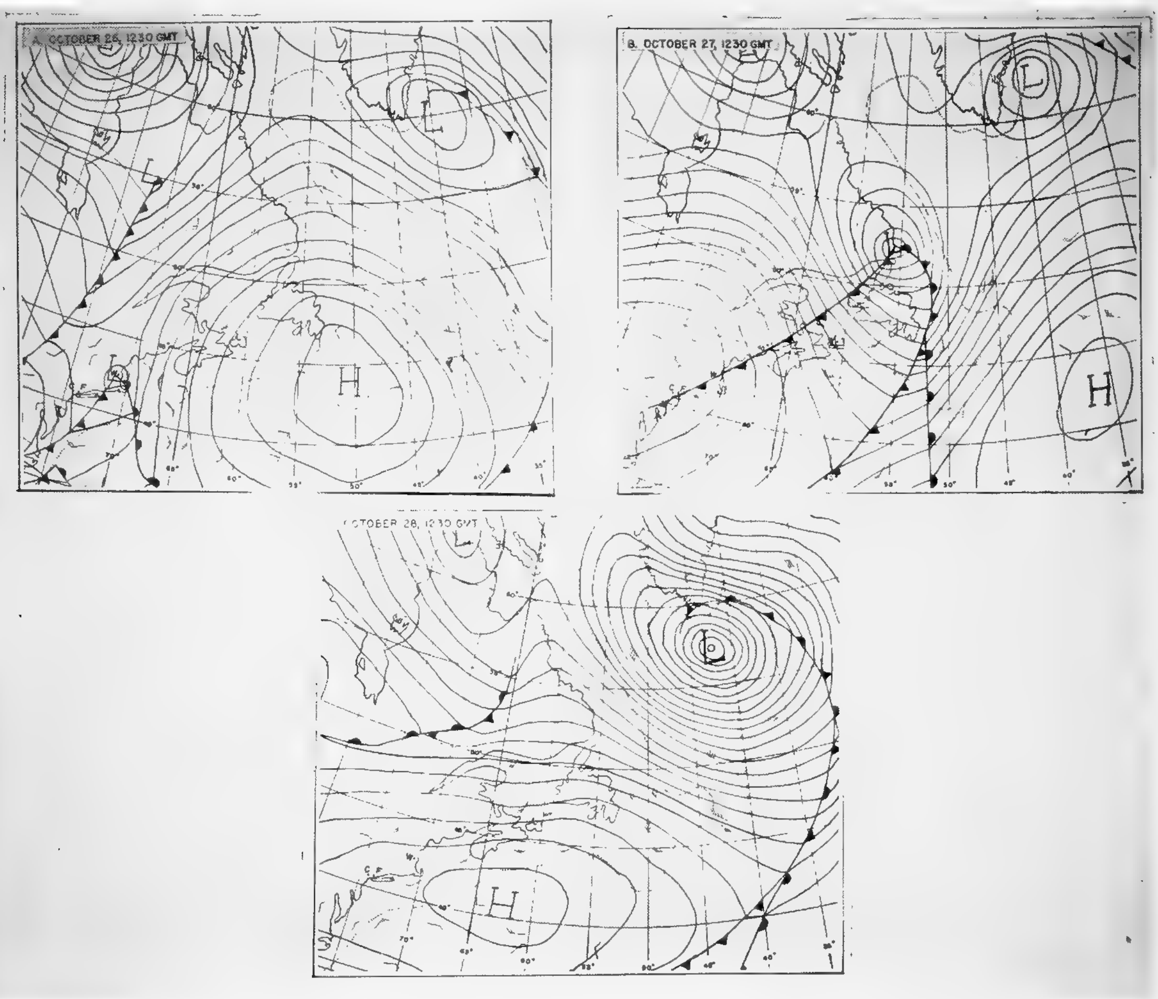

Figure 14. Marine weather charts showing development and positions of the cyclone related to the microseism storm shown in Figure 13.

Both the Woods Hole Oceanographic Institution and the Beach Erosion Board of the Corps of Engineers, U. S. Army, generously made available a large supply of wave records and wave record analyses.

Records from U. S. Navy tripartite stations were obtained from the Microseismic Research Project at the Hurricane Weather Central in Miami through the generous cooperation of M. A. Gilmore. The writer is deeply grateful to all the institutions and individuals who have so cooperated.

Discussion from the Floor

Byerly asked about the fact that the isointensity lines were drawn directly across the Florida peninsula. In private discussion later *Gilmore* answered the question and said that while such had not been observed, he thought it was because storms with 90 knot winds did not occur over the peninsula, but he thought that microseisms would be generated if they did. *Press* asked about the periods, and *Gilmore* replied that the periods are generally the same for the same location. *Gilmore.* Storms of different intensities in the same area produce different microseisms at the same station. The larger storms will always produce the larger microseisms. In some areas a 90-knot storm will not produce microseisms at a particular station, whereas a storm of larger intensity will produce larger than normal microseisms at that station.

(*Peoples* pointed out that the micro-ratio lines might well indicate the geology of the region. *Dinger* asked if *Gilmore's* method had been applied to hurricane "Easy" of 1951. *Gilmore* replied no. *Ramirez* asked how accurately the wind velocities are known.) *Gilmore.* It is very difficult for the forecaster to establish the true velocity of hurricane winds. The best that he can do is to get an average velocity of the wind which may be, and often is, as much as ten or more knots above or below his estimate.

(*Melton* asked if the tripartite stations are still running and do they always show the same errors. *Gilmore* replied yes, and probably no. *Båth* asked if *Gilmore* had plotted amplitude-period ratio lines. *Gilmore* replied, in only a few cases.) *van Straten.* For some time, I have been concerned about the terms by which tropical storms are described. One hears discussion and comparison of "90-knot storms" or "120-knot storms." Actually, the magnitude of the wind at the center of the storm is only one factor in the description of a storm. The area under the influence of strong winds seems another significant factor.

In order to determine what factors might be related to microseismic generation, I requested that Fleet Weather Centrals Guam and Miami plot three factors concerning a storm against time: (1) center wind speed, (2) area enclosed by the highest closed isobar, (3) area enclosed by the 50-knot isovel.

The initial reports indicate that the maxima in microseismic amplitude correspond closely to the area enclosed by the 50-knot isovel. This correspondence is much greater than that attempting to relate center wind speed with microseismic amplitude.

Press. As Dr. van Straten showed this morning, microseisms are affected not by conditions at the very center of the storm, but by conditions over an area enclosed by a given isovel. For this reason, statements concerning the presence or absence of the microseisms as a function of the position of the center of the storm are rather dangerous, especially for cyclones extending over areas as large as those considered by Dr. Carder. If one adds to this the very sharp effect of barriers, one might interpret Dr. Carder's results differently. It may be possible to come up with a different interpretation.

(*Press* also pointed out that earthquakes indicate a barrier off the California coast. *Gutenberg* pointed out that a hurricane off the lower California coast had given large microseisms at Pasadena and Tucson, but none at Santa Clara and Berkeley. *Deacon* pointed out that we do not know very much about where the actual wave interference may take place. *Longuet-Higgins* pointed out with a small origin there was more attenuation close to the source.)

Longuet-Higgins. One cause contributing to the apparently rapid attenuation with distance of "hurricane" microseisms may be mentioned. If the microseisms are surface waves spreading out horizontally from the generating area, their amplitude can be expected to decrease like $r^{-1/2}$, where r is the distance from the center (viscous dissipation and structural barriers being disregarded). Microseisms originating in a small generating area, such as may be associated with a hurricane, and recorded near the center, would decrease rapidly with r; but microseisms from a large generating area, such as an extra-topical cyclone, and recorded at greater distances, wouid fall off less rapidly.

MICROSEISMIC PERIOD SPECTRA AND RELATED PROBLEMS IN THE SCANDINAVIAN AREA

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Meteorological Institute at Uppsala

I. Introduction—The problem of microseismic periods is intimately connected with the microseismic problem as a whole. We are here concerned with microseisms in the general period range of 3-8 sec. Notable theories to explain their origin have been proposed on one hand by British workers on the subject (see *Longuet-Higgins*, 1950), on the other hand by *Press* and *Ewing* (1948). Both theories are capable of explaining the observed microseismic periods, but in completely different ways. Therefore the only fact that observed and theoretical periods coincide cannot be taken as a stronger support of one theory than of the other. Other facts indicate that the observations in Scandinavia are better explained by the former theory than by the latter. A further discussion of this matter is given by the present author in the paper (1951 b).

In this paper we will study some characteristics of the microseismic periods in the Scandinavian area, leaving aside the question how the periods originate. By periods we usually mean the periods corresponding to the maximum amplitudes. By constructing frequency curves of all periods existing at a certain time (period spectra) we will be able to study also the period corresponding to the frequency maximum as well as the mean period. It is a well-known fact that the periods increase with distance. But it is still an open question if this is due to a real lengthening of the period of each wave as they propagate or if it is due to a more rapid extinction of the shorter periods, whereas the period of each individual wave is constant. Another fact, which will also be studied, is the tendency of periods to vary in unison with the amplitudes. The fact that the periods of microseisms are functions of several variables (such as distance and intensity of the source), which usually vary simultaneously, requires great care in period studies in individual cases.

The periods of microseisms have earlier been studied from various points of view by the present author. See (1949), pp. 8-9 (frequencies of periods for different months), pp. 23-24 (annual variation of periods), pp. 26-28 (diurnal variation of periods), pp. 42-44 (beats), pp. 60-66 (relation between amplitude and period, and mean periods in different situations), p. 109 (comparison between periods on N-S and E-W components), pp. 119-142 (period studies in individual cases); see also (1951 a), pp. 371-374 (comparison between periods at Bergen and Uppsala). In the latter paper (p. 374) also a hyperbola method for locating the source by means of the periods was indicated. This method will be studied below in the light of the new results.

II. Materials and Methods Used—Period spectra have been constructed for four different situations (Oct. 7, 1947, Oct. 28, 1947, Jan. 14, 1949, and March 23, 1949, at 07^h M.E.T. in all cases) from the records of the N-S and E-W components of the Wiechert seismographs at Bergen, Copenhagen, and Uppsala, and the Mainka seismograph at Helsinki (all with mechanical recording). In each case every individual period within ± 15 minutes of 07^h M.E.T. was measured. Due to this concentration of the measurements the period spectra correspond in all cases to well defined weather situations; the changes of the weather situations taking place during the interval of 30 minutes are of no consequence. The number of observations is given in every case in Table 2 below. In the mean every frequency curve is based on more than 200 observations. Various sources of error will now be considered.

1. A certain period spectrum is generated at the source. The seismographs do not generally reproduce this spectrum unchanged but act as filters due to their different response to different periods. In comparing the records of different seismographs due account must be taken of this fact. Table 1 gives the seismograph constants in our cases, and Figure 1 gives a few representative curves of the dynamic magnification V. As we are not so much concerned with the magnification itself as with its variation with the period T (especially within the range of the microseismic periods), the curves in Figure 1 have been displaced so that they all pass through the point $V = 200$ for $T = 5$ sec in order to facilitate their comparison. The circumstance that Copenhagen usually has the shortest free period and Helsinki in all cases has the largest free period necessitates some discussion. The free periods are

usually all larger than the microseismic periods and there seems to be no influence from the different constants. This is obvious for the following reasons.

a. The periods corresponding to frequency maxima and amplitude maxima are not consistently larger at Helsinki, nor are they consistently lower at Copenhagen.

b. The upper limit of the period spectra is not larger at Helsinki than at the other stations.

c. The lower limit of the period spectra increases from Bergen to Uppsala in spite of equal free periods; it also increases from Bergen to Copenhagen in spite of generally lower free periods at Copenhagen than at Bergen.

d. Also the mean periods increase from Bergen to Copenhagen and from Bergen to Uppsala.

Therefore the conclusion seems to be justified that the different seismographs have the same response to the microsisms under consideration, and the spectra at the different stations are directly comparable.

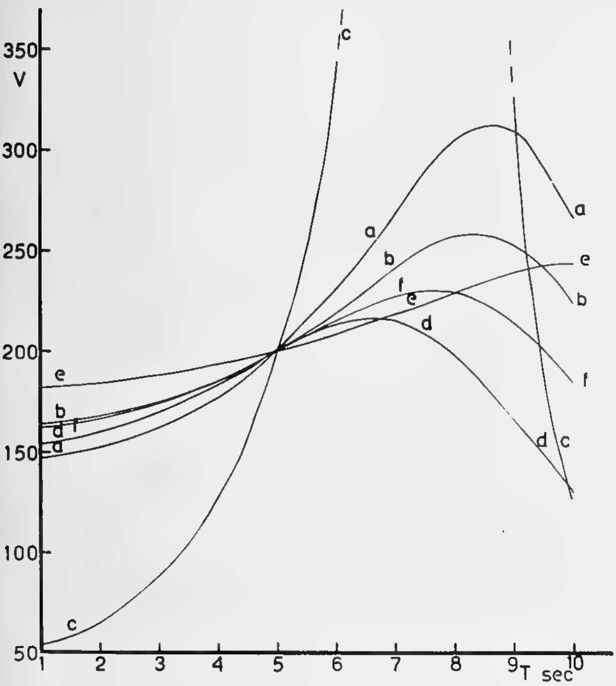

Figure 1. Dynamic magnification curves. Explanation:

Curve	T_0	V	E	Valid for
a	9.3	180	2.4	Bergen N
b	9.2	129	2.7	Bergen E
c	7.7	220	1.4	Bergen E
d	8.2	200	4.3	Copenhagen N, E
e	12	140	3.5	Helsinki N, E
f	9.2	189	3.9	Uppsala N, E

All curves have been displaced so as to pass through the point $T_0 = 5$ sec, $V = 200$.

2. The ground (from the source to the station) also acts as a filter. Bergen (gneiss), Uppsala (granite), and Helsinki (gneiss) are all within Fennoscandia, and no general differences exist. Copenhagen (chalk) is outside Fennoscandia, but owing to the relatively small part of the path, consisting of sediments, as well as owing to the uncertainties of any correction for their influence, the Copenhagen period spectra have also been used without modification.

3. An essential requirement is that the microseisms at the different stations compared have the same source. This has earlier been shown to be the case for Bergen and Uppsala by the author (1951 a) as well as for Copenhagen (1952). In the light of our present experience the statement is justified that the microseisms at Helsinki also have the same source as at the other three stations (for further discussion see below).

4. Only measurable periods have naturally been included. The smaller number of observations in a few cases depends on weak microseisms, i.e. fewer measurable periods. This procedure necessarily entails a certain selection, depending on the sensitivity of the seismograph. With regard to what has been said in 1. above as well as to the results (Table 2) this circumstance does not seem to be of any importance.

5. The drum speeds are 12mm/min at Copenhagen, 15 mm/min at Bergen and Uppsala, and 20 mm/min at Helsinki. The measurements were made to 0.1 mm and then converted into seconds and tenths of seconds. There was a clear tendency in all cases of obtaining frequency maxima at 1.0, 1.5, 2.0 mm (the measurements were made with a glass scale, divided into half millimeters). As they were certainly not real, smoothing has been made according to the formula $\frac{1}{4}$ (a + 2 b + c). The frequency curves thus obtained are certainly nearer to the truth than curves based directly on the original observations.

6. In a procedure like this where successive periods are measured, false periods may arise at points where one wave train gives place to another wave train. A too long false period may arise if a small quiet interval separates the two wave trains. Likewise too short false periods may occur where one wave train is replaced by another wave train without separation. However, this source of error is of no influence if due account is taken of it in the measurements; moreover, the microseisms measured are generally regular and continuous.

III. Discussion of the Results—For convenience in writing we introduce the following notations:

T_f = period, corresponding to frequency maximum,

T_m = mean period,

T_a = period, corresponding to amplitude maximum,

n = number of observations,

N = N-S component,

E = E-W component,

B = Bergen,

C = Copenhagen,

H = Helsinki,

U = Uppsala,

I = October 7, 1947, at 07^h M.E.T.,

II = October 28, 1947, 07^h M.E.T.,

III = January 14, 1949, 07^h M.E.T.,

IV = March 23, 1949, 07^h M.E.T.

The period spectra are given in Figs 2-5. Table 2 contains values of T_f, T_m, and T_a with standard errors for T_m and T_a and the number of observations in each case. T_f, T_m, and T_a have also been indicated in Figs. 2-5 by vertical lines, the shortest for T_f, the next longer for T_m, and the longest for T_a. The whole investigation is based upon 7083 individual period measurements on the records. The weather situations at 07^h M.E.T. are obvious from Fig. 6, copied from the official Swedish weather maps. The agreement with the official British weather maps is very good.

The aim in the following study has mainly been to establish general rules for the periods. The statistical significance of these rules has been investigated in every case, usually by application of Bernoulli's theorem. The deviations from the general rules which may occur in individual cases, require more detailed studies of the particular cases in order to be explained.

1. The shape of the frequency curves.

The frequency curves have in general only one pronounced maximum around which the curves are symmetrical. The microseisms may therefore be characterized as regular. Notable exceptions occur at Bergen, especially for N. This component has usually two maxima at Bergen. As the main source of the microseisms in Scandinavia lies along the Norwegian coast, we understand that the shape of a frequency curve depends on the position of the station in relation to this coast. If the position is such that the coast length takes up a large distance interval, from the station, as the case is for Bergen, a wider and less regular spectrum is obtained. In case IV the microseisms are less regular at all stations than in cases I-III.

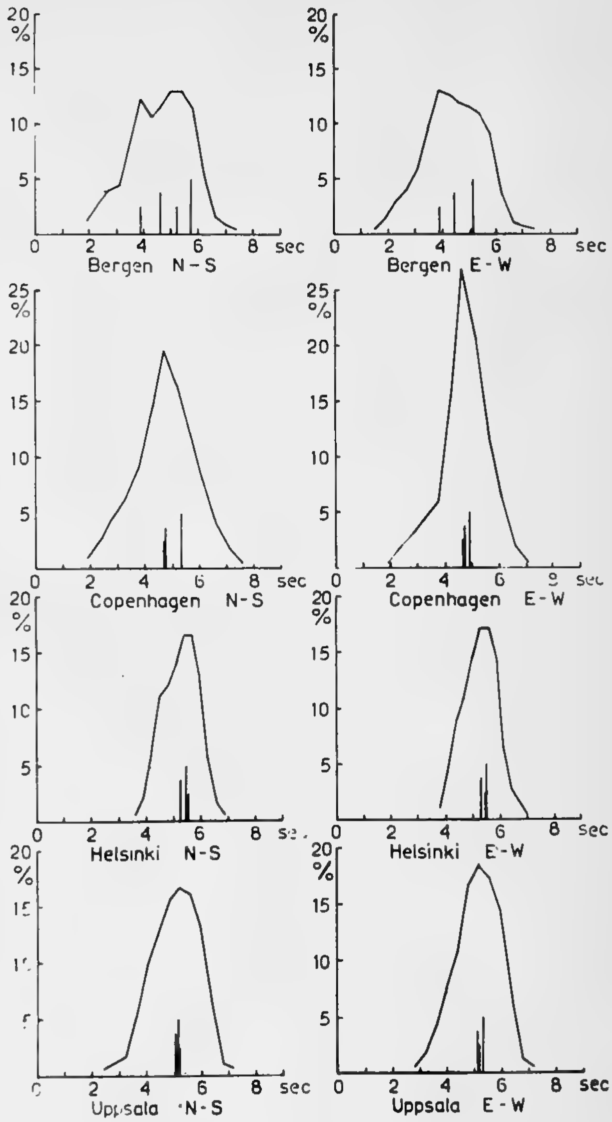

Figure 2. Period spectra on October 7, 1947, at 07^h M.E.T. The shortest vertical line indicates T_f, the next longer T_m, and the longest T_a.

2. Comparison of stations.

The mean periods (n = 8; calculated from Table 2) for both components and all situations are as follows:

Station	T_f sec	T_m sec	T_a sec
B	4.79	4.71	5.76
C	5.05	5.10	5.52
H	5.45	5.44	5.85
U	5.43	5.35	5.65

For T_f it seems to be an increase from B to H: B < C < U < H. The various differences, B < C, C < U, U < H, B < U, and C < H are, however, not statistically significant. But

Table I. Seismograph constants.

T_o = free period, V = static magnification, ξ = damping ratio, r = deviation due to friction.

Time	Station	N-S				E-W			
		T_o sec	V	ξ	r mm	T_o sec	V	ξ	r mm
Oct. 7 1947	Bergen	9.5	177	2.5	1.5	9.2	129	2.7	1.3
	Copenhagen	7.7	200	2.2	1.1	8.3	200	4.3	0.9
	Helsinki	12	140	3.5	-	12	140	3.5	-
	Uppsala	9.2	192	3.7	1.3	9.3	186	4.0	1.4
Oct. 28 1947	Bergen	9.5	177	2.5	1.5	9.2	129	2.7	1.3
	Copenhagen	8.4	190	6.0	0.6	8.4	200	2.9	0.8
	Helsinki	12	140	3.5	-	12	140	3.5	-
	Uppsala	9.2	192	3.7	1.3	9.3	186	4.0	1.4
Jan. 14 1949	Bergen	9.0	182	2.2	1.9	7.7	220	1.4	0.8
	Copenhagen	8.2	200	5	0.7	8.6	200	4.2	0.8
	Helsinki	12	140	3.5	-	12	140	3.5	-
	Uppsala	9.1	192	4.0	0.7	9.3	187	3.9	1.1
March 23 1949	Bergen	9.0	182	2.2	1.9	7.7	220	1.4	0.8
	Copenhagen	8.0	210	5	0.7	8.1	210	4.0	0.8
	Helsinki	12	140	3.5	-	12	140	3.5	-
	Uppsala	9.1	192	4.0	0.7	9.3	187	3.9	1.1

B < H is significant. Therefore we may conclude that there is no pronounced difference of T_f between the different stations except for a slight indication in the manner already mentioned. For T_m there is a very clear increase from B to H: B < C < U < H, very well substantiated by the individual cases. Every difference is significant, possibly with the exception for U < H. For T_a there is no very obvious difference between the different stations, except that H has generally larger values. The only difference which is no doubt significant is C < H. The relatively large values of T_a at B for our situations are remarkable. Earlier results (1951 a, p. 373) have indicated that in general T_a is less at B than at U. By means of the data used in (1951 a, p. 373) this was shown to be significant. By means of the corresponding data for Copenhagen (1952) it has been shown that for T_a B < C with a high degree of significance, whereas for the same data there was no difference of T_a between C and U.

The main reason for the sequence B < C < U < H, especially clear for T_m, is the stronger extinction of shorter periods (see below). The fact that C comes between B and U is probably explained by a relatively larger importance at C of the southwestern part of the Norwegian coast, from where the microseisms arrive at C before they arrive at U. At U and H the whole Norwegian coast is of about equal importance.

3. Comparison of components.

The mean periods (n = 16) for all stations and all situations are as follows:

Comp.	T_f sec	T_m sec	T_a sec
N	5.35	5.20	5.82
E	5.01	5.10	5.57

For all three periods E < N. This result may be considered significant only for T_a, but the tendency of E < N exists for all three periods. The result that E < N for T_a confirms my earlier results for Bergen and Uppsala (1951 a, pp. 372-373). It has a high degree of significance for Uppsala, but not for Bergen. It has also been proved for Copenhagen with a high degree of significance, using the observations corresponding to those at Bergen and Uppsala in (1951 a, p. 373). This is therefore a characteristic feature of Scandinavian microseisms. The most probable reason is a distance effect. The N component is most sensitive to actions at the more remote northern part of the Norwegian coast, whereas the E component reacts strongest to actions at the west coast (around B). This has been established beyond doubt from numerous cases at Uppsala.

An increase of the period T_a with increasing distance has been observed in several investigations. But the nature of the phenomenon is not clear: if it is mainly an actual increase of period with distance, or if it is only a more rapid extinction of the shorter periods. No decision seems to be possible only from a knowledge of the increase of T_a with distance as both effects produce the same result (for further discussion see 6, below).

4. Comparison of the different periods.

The mean periods (n = 32) for all stations, both components, and all situations are $T_f = 5.18$ sec, $T_m = 5.15$ sec, $T_a = 5.69$ sec. The result $T_m < T_a$ is valid in practically every case (exceptions are UIIE and HIIE) and has a very high degree of significance. The result $T_f < T_a$ is also highly significant, but there are a few more exceptions to this rule than to the rule $T_m < T_a$. This result is a reflection of the fact that the amplitudes corresponding to the shorter periods are relatively small. Furthermore, $2_f \simeq T_m$, i.e. no significant difference, corresponding to the generally symmetrical nature of the frequency curves.

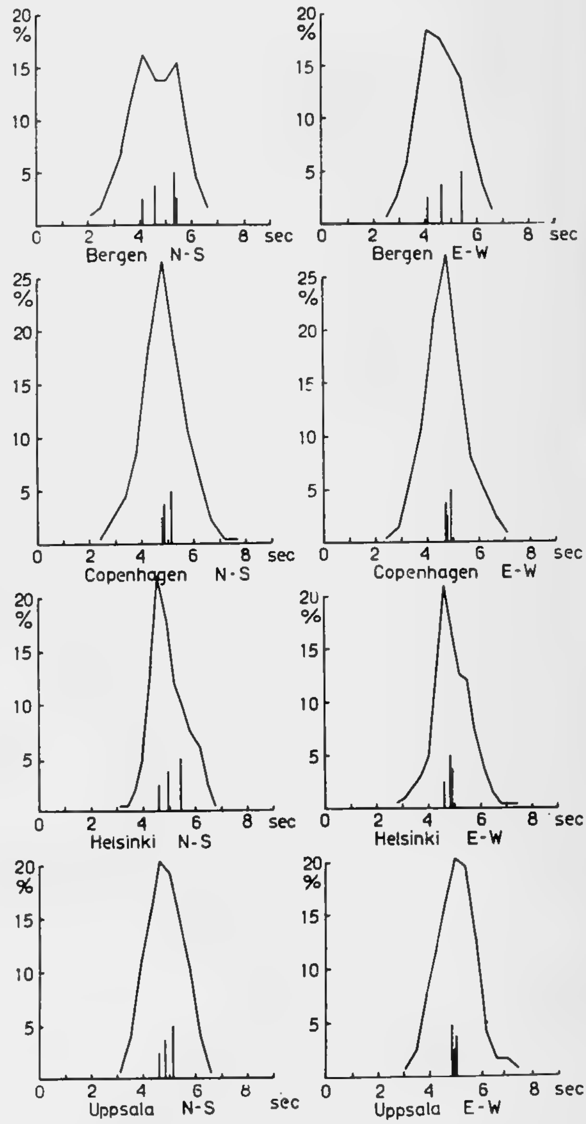

Figure 3. Period spectra on October 28, 1947, at 07^h M.E.T.

Table 2. Periods.

Time	Station	T_f sec N	T_f sec E	T_m sec N n	T_m sec N	T_m sec E n	T_m sec E	T_a sec N n	T_a sec N	T_a sec E n	T_a sec E
Oct. 7 1947 (I)	Bergen (B)	(3.9); 5.2	3.9	255	4.60±0.07	267	4.43±0.07	5	5.72±0.23	6	5.13±0.22
	Copenhagen(C)	4.7	4.7	221	4.75±0.07	201	4.76±0.06	8	5.33±0.16	6	4.95±0.11
	Helsinki (H)	5.55	5.45	180	5.27±0.05	181	5.30±0.05	10	5.46±0.08	6	5.50±0.10
	Uppsala (U)	5.2	5.2	175	5.04±0.07	168	5.13±0.06	6	5.13±0.16	6	5.33±0.13
Oct. 28 1947 (II)	Bergen (B)	4.1; (5.4)	4.1	117	4.59±0.09	158	4.63±0.07	5	5.32±0.08	8	5.40±0.08
	Copenhagen(C)	4.8	4.8	226	4.86±0.06	262	4.74±0.05	6	5.13±0.21	9	4.92±0.10
	Helsinki (H)	4.6	4.6	200	4.96±0.05	206	4.90±0.05	8	5.43±0.08	4	4.83±0.14
	Uppsala (U)	4.6	5.0	177	4.82±0.05	118	5.03±0.07	7	5.11±0.07	6	4.87±0.09
Jan. 14 1949 (III)	Bergen (B)	(4.0); 5.6	5.4	281	4.89±0.08	276	4.88±0.06	8	5.75±0.15	12	5.57±0.09
	Copenhagen(C)	4.7	4.7	275	5.07±0.07	242	4.97±0.07	9	5.79±0.06	8	5.30±0.21
	Helsinki (H)	5.7	6.0	190	5.61±0.06	240	5.59±0.06	9	5.87±0.15	9	5.97±0.06
	Uppsala (U)	5.9	5.5	223	5.50±0.06	216	5.30±0.07	7	5.79±0.11	6	5.77±0.09
March 23 1949 (IV)	Bergen (B)	(4.0); 6.0	4.0; (6.0)	233	5.07±0.11	235	4.58±0.09	8	6.85±0.09	10	6.32±0.13
	Copenhagen(C)	7.2	4.8; (6.2)	272	6.07±0.09	291	5.54±0.08	8	6.89±0.13	6	5.85±0.14
	Helsinki (H)	5.7	6.0	156	5.92±0.11	199	5.96±0.08	5	7.12±0.08	6	6.60±0.13
	Uppsala (U)	6.0	(4.0); 6.0	209	6.10±0.09	206	5.85±0.09	5	6.40±0.18	5	6.80±0.22

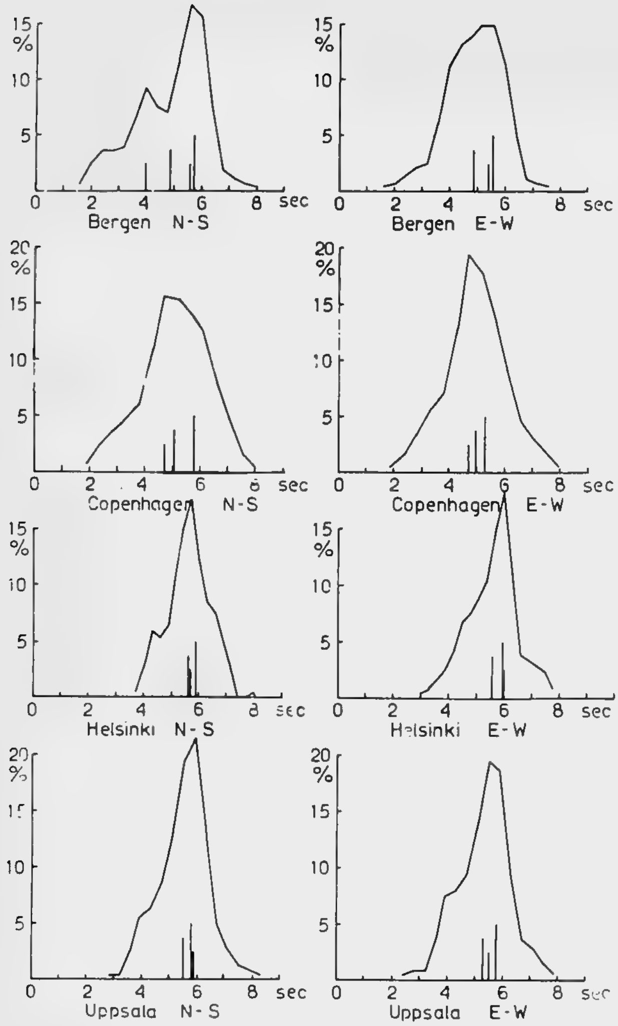

Figure 4. Period spectra on January 14, 1949, at 07^h M.E.T.

5. Comparison of situations.

The mean periods (n = 8) for all stations and both components are as follows:

Case	T_f sec	T_m sec	T_a sec
I	4.99	4.91	5.32
II	4.58	4.82	5.13
III	5.44	5.23	5.73
IV	5.71	5.64	6.60

These values indicate the sequence II < I < III < IV for all three periods. This sequence is in general well established from the individual cases. All differences II < I, I < III, III < IV are significant for T_a (BE is the only partial exception). The differences are also significant for T_m, possibly with exception for the difference II < I (exceptions occur for T_m for CN and BE). On the other hand, the differences for T_f are not significant, not even II < IV; nevertheless, the general trend is the same for T_f as for T_m and T_a.

The differences between the different situations can hardly be explained as only distance effects, e.g. in II both active coast and cyclone center are at the greatest distance, nevertheless the periods are shortest. On the other hand, the different intensities of the storms seem to afford an explanation. This is another instance of the parallel behaviour of amplitudes and periods. The maximum ground amplitudes, expressed in μ, are given in the following table.

Case	B		C		H		U		Mean
	N	E	N	E	N	E	N	E	
I	2.2	1.9	1.1	1.4	2.3	2.1	1.0	0.8	1.6
II	0.9	1.2	1.3	1.4	1.8	1.6	0.7	1.0	1.2
III	(6.2)	2.1	2.8	2.5	2.3	3.0	2.1	2.0	2.1 (2.9)
IV	1.8	0.6	3.8	2.7	2.2	2.0	2.5	2.0	2.2

If the anomalous amplitude BNIII is excluded, we get the amplitude sequence II < I < III < IV, i.e. the same as for the periods. Plotting the periods against the mean amplitudes we find that I-II-III forms a reasonable sequence, whereas the small amplitude difference between III and IV is not in good accord with the relatively large period differences. This could possibly be due to a distance effect, as in IV the more active part of the Norwegian coast is in the northern part. A numerical calculation of the rate of change of T_a with distance to center of the active coast gives approx. 2.10^{-3} sec/km, a value which lends further support to this idea (1949).

6. Upper and lower limits of the period spectra.

From an inspection of the period spectra (Figures 2-5) it is clear that the upper limit is remarkably constant from station to station in a given situation with no general variation, whereas the lower limit shows a very pronounced increase in the direction B-C-U-H. This increase occurs for both components in every case without exception. The increases of the lower limit are in the mean from B to C about 0.25 sec, from C to U about 0.7 sec, from U to H about 0.6 sec. The total increase from B to H is in the mean about 1.5 sec.

The results concerning the upper and lower limits of the period spectra strongly support the conclusion that the change of the spectrum with distance is due to a more rapid extinction of the shorter waves rather than to an actual increase of periods. It would naturally be valuable to extend such investigations to greater distances, wherever possible, provided the source of the microseisms is the same for the whole area investigated.

7. The hyperbola method.

A hyperbola method for locating the source of microseisms from the periods observed at a number of stations has been given in my paper (1951 a, p. 374). This method rests on the observation that the observed period T_a apparently increases with distance. The method may briefly be described as follows. For a mean apparent increase of T_a with distance Δ from the source $f = \frac{dT_a}{d\Delta}$ we get the following equations, combining three different stations two and two:

$$
\begin{aligned}
T_a'' - T_a' &= f\,(\Delta'' - \Delta')\\
T_a''' - T_a' &= f\,(\Delta''' - \Delta')\\
T_a''' - T_a'' &= f\,(\Delta''' - \Delta'')
\end{aligned}
$$

f is assumed to be the same in all three equations as a first approximation. This set of equations means that $(\Delta'' - \Delta') : (\Delta''' - \Delta'') : (\Delta''' - \Delta'')$ is given, geometrically defining the source. As f is not exactly known, the source can be located by trial and error, until all three hyperbolas intersect in a point, assuming a point source. When the source has been located, it is possible to calculate f. The method can only be used with success if the source of the microseisms is exactly the same for all stations for which the periods are used. For a common point source the method may be expected to lead to results. In our case, however, we have a line source, the length of which is comparable with and often larger than the mutual distance between the stations. Different parts of the Norwegian coast are of different importance to the different stations as already indicated above, i.e. the source is not exactly the same for all our stations. Therefore the hyperbola method may not be expected to lead to any useful results in these conditions.

Among further desirable investigations the following may be mentioned:

a. Extension of the investigation of periods by means of period spectra to greater distances.

b. Application of the hyperbola method to cases with a point source.

c. Correlation of microseismic periods with other phenomena, notably the periods of sea waves and swell. Investigations of the last-mentioned kind have been done by British investigators, but an extension to other localities is desirable.

Summary—Microseismic period spectra have been constructed for four different situations (I = Oct. 7, 1947, II = Oct. 28, 1947, III = Jan. 14, 1949, IV = March 23, 1949) for both components (N, E) at Bergen (B), Copenhagen (C), Helsinki (H), and Uppsala (U). The microseisms studied are usually regular and continuous in the general period range 3-8 sec. The following results have been obtained for the period T_f, corresponding to frequency maximum, for the mean period T_m, and for the period T_a, corresponding to amplitude maximum.

1. Especially T_m increases clearly from B to H: B < C < U < H. This is explained as a distance effect.

2. For all periods E < N; this is especially clear for T_a. It is also explained as a distance effect.

3. $T_m < T_a$ is valid practically without exception. Furthermore $T_f \simeq T_m$, corresponding to the generally symmetrical nature of the frequency curves.

4. For all periods a comparison of the situations shows that II < I < III < IV. This is explained as mainly due to different intensity of the microseismic storms.

5. A comparison of the upper and lower limits of the period spectra at the different stations clearly indicates that there is a greater extinction of the shorter waves, whereas there is no indication of an actual period increase.

6. The hyperbola method for locating the source can be expected to lead to useful results only when the source is exactly the same for all stations compared and preferably a point source.

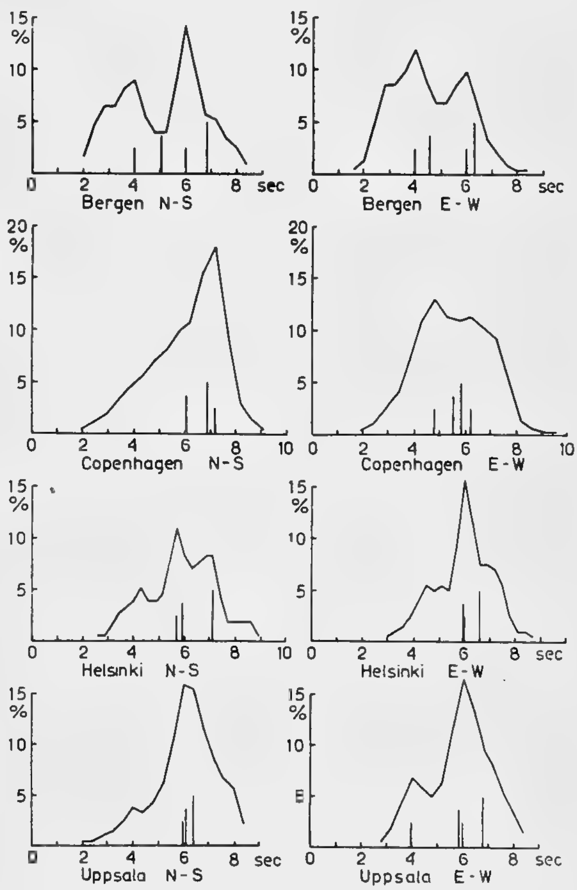

Figure 5. Period spectra on March 23, 1949, at 07^h M.E.T.

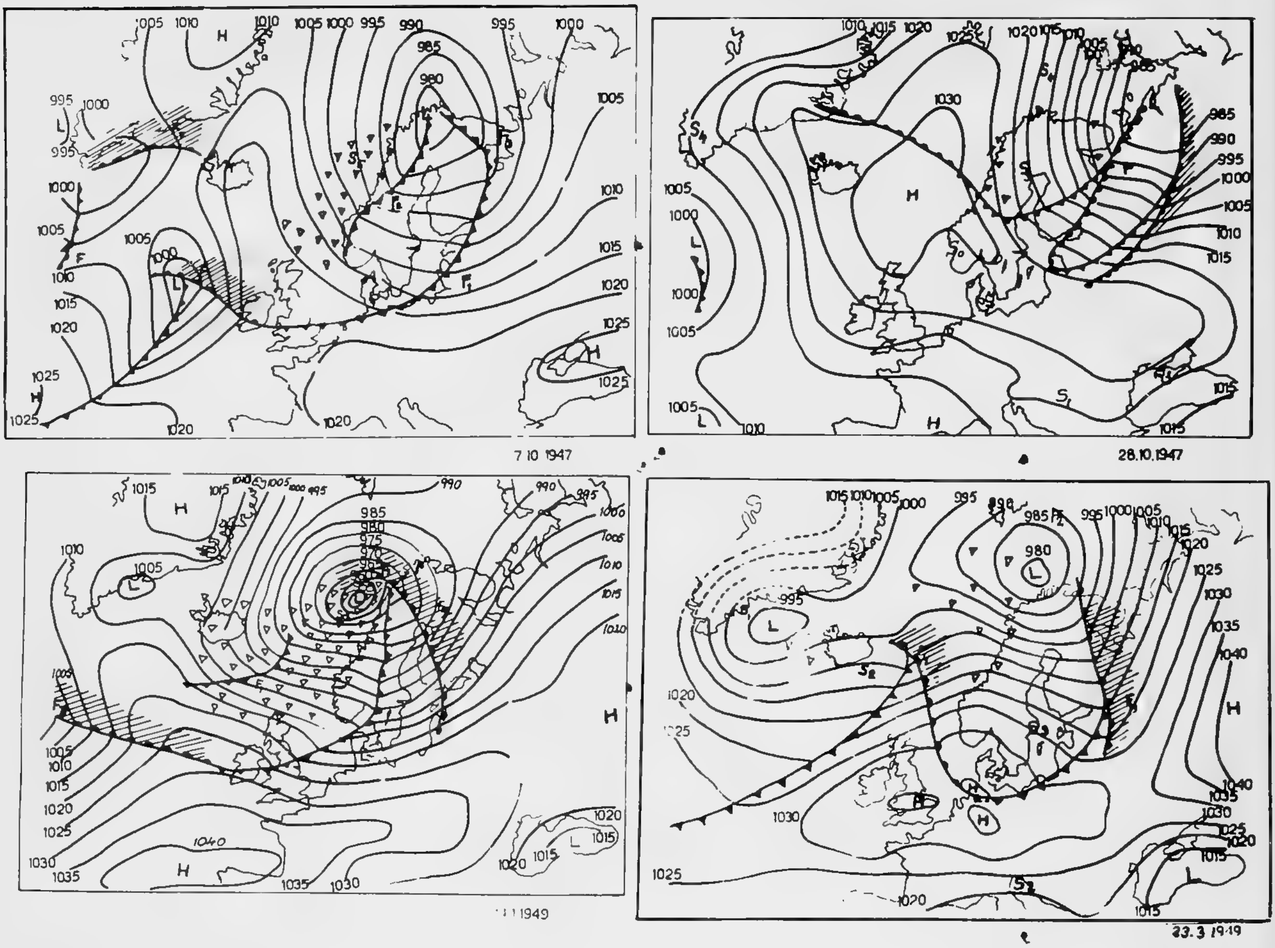

Figure 6. Weather situations at 07^h M.E.T. on October 7, 1947, October 28, 1947, January 14, 1949, and March 23, 1949.

Acknowledgments—This investigation has been made at the Seismological Laboratory of the Meteorological Institute, Uppsala. The Directors of the Seismological stations at Bergen, Copenhagen, and Helsinki have lent me their original records and given information on seismograph constants, etc. Miss I. L. Andersson, Uppsala, has assisted me in drawing the figures. My best thanks for valuable help are due to all the above-mentioned persons.

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Discussion

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In the brief space of a few pages Doctor Båth has assembled a surprisingly large volume of first hand observational data on the periods of microseisms recorded in the Scandinavian area. Anyone familiar with such measurements realizes what an expenditure of time and painstaking labor is involved. Seismologists the world over will be unanimous in their appreciation of the factual material thus made available.

A few similar studies have been published in the past for other regions. But one wonders what is the overall significance of a period spectrum which seems to take no account of superposition of wave trains simultaneously arriving from different directions. There would be less likelihood of inhomogeneity if only the periods accompanying the maximum amplitudes of well-formed groups are considered; but care is required even in this case to select only groups of regular waves. Neither from one microseismic storm to another nor within a group in the same storm has the writer found the period to increase consistently with amplitude.

Recently Gotch under the writer's direction made a study of the relation between amplitude and period in the most regular portions of wave groups in nineteen microseismic storms recorded by Galitzin-Wilip seismographs at Florissant in 1949, 1950 and 1951. The average period of the maxima for all nineteen storms was 6.66 seconds. The shortest period was 5 seconds and the longest was 8.8 seconds. In the four storms in which the average peak period was over eight seconds the amplitudes were only moderate; while in those storms in which the mean peak amplitudes were largest the corresponding periods were less than the overall average. The writer has not succeeded in finding a tenable storm to storm relationship between amplitudes and periods in the mid-continent area of North America. Within a given microseismic storm after it is fully developed the period tends to remain approximately constant.

The periods generally recorded at Florissant and Saint Louis are sensibly longer than those shown by Donn and by Gilmore in records of east coast stations. This might be and has been interpreted as a distance effect. However, the Florissant periods are approximately the same as those listed by Thompson for the Palmer Land station in Antarctica and yet the probable source fronts were often very near to that station and sometimes over relatively shallow waters.

Many seismologists will fail to see the cogency of the argument for a linear source at a given distance drawn from the variation of the plane of vibration of microseismic waves at single stations. Even the actual instantaneous directions of travel of well-formed groups of waves across a tripartite station vary through many degrees of arc so that an average of many observations is necessary to determine a bearing that is representative of the energy flow across the network. How then can it be shown from observations at single stations that the source is linear and not a real? And in view of the inconclusive relationship of period to distance how can the source be considered as certainly lying along a given coast line?

Measurements Made by Mr. Gotch on Florissant Seismograms

T = period; A = amplitude in mm = microns/600 approx.

Microseismic Storm	T	Amax	T	Amax	T	Amax	T	Amax	Average T (Seconds) of Maximum Amplitude
January 1, 1949									
V	....		....		....		....		
N-S	5.5	3.5	5.75	4.0	5.75	3.5	5.5	2.75	5.6
E-W	5.5	3.75	5.75	3.5	5.75	3.75	5.5	3.5	
January 15, 1949									
V	....		....		....		....		
N-S	7.25	2.75	7.0	2.5	6.5	2.75	6.75	2.75	6.9
E-W	7.0	4.25	7.25	3.25	6.5	3.25	6.75	3.5	
January 30, 1949									
V	5.25	1.25	5.75	1.0	....		5.25	1.75	
N-S	....		5.25	1.75	....		5.75	2.25	5.4
E-W	5.75	2.0	5.25	2.5	5.25	2.25	5.5	2.25	
February 8, 1949									
V	....		5.0	3.0	5.25	2.75	5.25	2.0	
N-S	5.25	2.25	....		....		5.5	2.0	5.4
E-W	5.5	2.75	5.75	3.0	5.75	2.75	5.5	2.75	
February 11, 1949									
V	6.25	3.75	6.5	3.0	6.25	3.25			
N-S	6.75	3.0	6.0	3.5	6.5	3.5			6.4
E-W	6.75	3.75	6.0	3.75	6.25	4.5			
February 18, 1949									
V	....		....		....				
N-S	6.25	2.25	6.0	2.75	6.0	2.5			6.1
E-W	6.0	3.5	6.25	2.75	6.0	2.5			

Microseismic Storm	T	Amax	T	Amax	T	Amax	T	Amax	Average T (Seconds) of Maximum Amplitude
March 1, 1949									
V	....		5.25	2.5	....				
N-S	....		....		....				5.0
E-W	4.75	2.0	5.25	2.25	4.75	2.25			
March 8, 1949									
V	8.75	2.75	8.75	2.5	8.25	2.5			
N-S	....		....		....				8.6
E-W	8.25	2.0	8.75	2.0	8.75	2.0			
March 26, 1949									
V	7.75	1.75	7.25	1.75	....		7.75	1.75	
N-S	....		....		7.25	2.75	7.75	2.25	7.6
E-W	7.75	2.5	7.75	2.25	7.5	2.25	7.5	3.0	
March 31, 1949									
V	....		....		....		....		...
N-S	....		....		....		....		8.5
E-W	8.75	4.0	8.0	4.0	8.5	4.5	8.75	3.75	
November 1, 1949									
V	4.75	3.75	5.75	3.25	....		....		
N-S	....		....		....		4.75	3.25	5.1
E-W	4.75	3.75	5.75	4.0	4.75	3.75	5.0	3.25	
November 9, 1949									
V	9.0	3.5	9.0	3.75	8.75	4.25			
N-S	....		....		8.75	4.25			8.8
E-W	8.5	3.5	9.0	3.0	8.75	2.75			
December 8, 1949									
V	8.25	4.75	8.25	4.5	....		....		
N-S	7.75	4.75	....		....		....		8.4
E-W	8.75	5.5	8.25	4.75	9.0	3.75	8.5	4.25	
December 29, 1949									
V	5.75	5.75	6.5	4.75	5.75	4.75			
N-S	....		6.5	3.75	6.25	3.75			6.2
E-W	6.5	4.75	6.25	3.75	....				
November 26, 1950									
V	....		....		....				
N-S	....		....		....				6.3
E-W	6.25	3.75	6.25	3.5	6.5	4.75			
November 28, 1951									
V	6.25	6.75	6.25	5.75	6.0	7.5	6.25	6.75	
N-S	....		6.25	4.25	6.5	4.0	....		6.2
E-W	....		6.0	4.5	6.0	5.0	6.25	5.5	
December 9, 1951									
V	6.5	3.5	6.5	3.75	6.75	3.0	6.5	4.0	
N-S									6.7
E-W	7.0	3.25	....		7.0	2.5	6.75	2.5	
December 17, 1951									
V	6.75	4.75	6.75	4.5	6.75	3.5			
N-S	7.25	3.0	7.0	3.5	7.25	3.0			6.9
E-W	7.0	3.25	6.75	2.75	7.0	2.75			
December 19, 1951									
V	6.25	9.75	6.5	10.25	6.25	10.75	6.5	11.75	
N-S	6.0	5.25	6.0	5.5	6.25	4.75	7.0	6.0	6.4
E-W	6.25	6.25	....		6.75	6.5	6.5	9.5	T. A. 6.66

Discussion

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Microseismic motions over a wide range of periods have been detected and reported in the literature; however, quantitative information on the ground amplitudes associated with the various periods is surprisingly difficult to find. Further, measurements describing the spectrum existing at a given locality and time are generally found to include only a narrow band of periods, usually of about one octave band width selected by the filter characteristics of conventional observatory seismographs. Most of the known information deals with the same "storm microseisms" discussed by Dr. Båth's paper, which is chiefly concerned with periods in the range 4-8 seconds. The spectra to be presented here cover a much wider band width, extending over nearly four octaves. Such data describe more completely the state of earth activity during a given meteorological situation, and in addition, variations in spectra due to location and time provide some insight into the problem of extinction of short period microseisms, which is studied by Dr. Båth.

During August of 1950, identical horizontal seismographs were operated at the Harvard College Observatory, Harvard, Massachusetts, and at the Rensselaer Polytechnic Institute's Pinewoods Observatory, near Troy, New York, through the cooperation of Dr. L. Don Leet of Harvard and Dr. Roland F. Beers of R. P. I. Seismometers were of the capacitance-bridge type, and ground motion was recorded on paper by means of a Brush Instrument Company penmotor after suitable amplification. Prior to installation, both seismographs were calibrated by means of a shaking-table at the Pinewoods Observatory, and a field calibration technique was developed to insure that changes in the instrumental constants would be known. The seismometers were operated with a free period of 1.5 seconds and with critical damping.

Both stations were in operation during the period from August 20 through August 23, at which time an intense hurricane was moving parallel to the Atlantic coast line between Cape Hatteras and a point south of Greenland. This storm produced a rapid rise in microseismic amplitudes at both stations, reaching a maximum on the early morning of the 21st, and decreasing to essentially the normal level by the 23rd. Measurements were made of the microseisms on August 21, near the time of the storm's least distance of about 200-350 miles from the stations, and again on August 23, when the storm was about 1700 miles away. The measurement technique consisted of finding amplitudes (peak to trough and associated periods for the largest nearly sinusoidal groups appearing on the records; special attention was directed toward finding groups with as widely different periods as possible. Seismogram amplitudes were reduced through the experimentally known steady-state response curves to ground motion in microns, and plotted as a function of period on logarithmic paper. Displayed in this manner, the points show a rather surprising regularity. Figures 1 and 2, for the "normal" day, when the storm had moved off to a great distance, show the observed spectra at Harvard and at Pinewoods. Within the range from 1.4 to 5.0 seconds, the peak microseismic ground motions were found to increase very nearly as the cube of the period. Important differences were observed, however. Note that while both stations showed ground motion for the long period (5.0 seconds at about 0.7-1.0 microns, the shorter (1.5 second) periods were noticeably smaller at Pinewoods, approximately 120 miles inland from Harvard.

A marked change in the spectrum at each

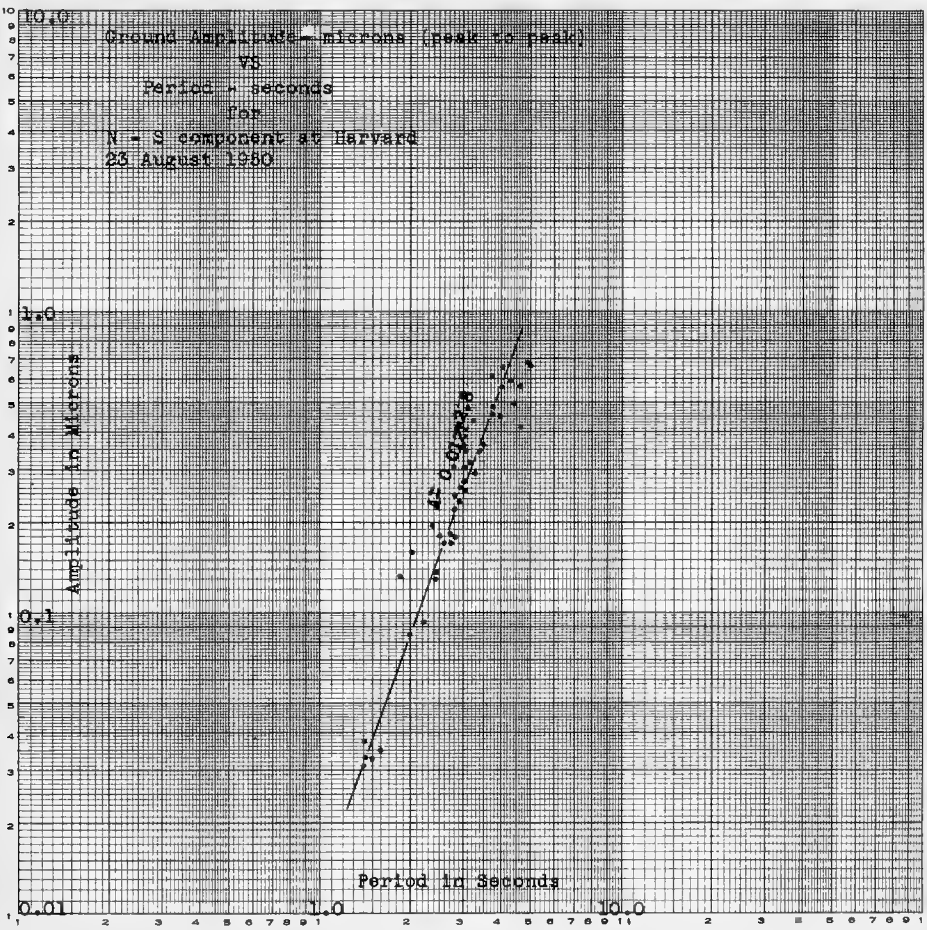

Figure 1. Microseismic Spectrum for Harvard, 23 August 1952. Normal day.

station was found on August 21, when the hurricane was at a minimum distance from the stations. As seen in Figures 3 and 4, not only did the general activity increase, but the shape of the spectra changed so that amplitudes were found to increase nearly as the fourth power of the period. A comparison of all of this data is shown in Figure 5 relative to the Harvard spectrum on 23 August 1950, here used as a standard spectrum. If all amplitudes are reduced to ratios of those at Harvard on the normal day, the following emerges. First, on a normal day the long period noise, presumed to originate in the Atlantic, is reduced by only a small amount in traveling the distance between Harvard and Pinewoods, while the short period noise (also presumed to originate in the ocean) is reduced by a factor of about 3:1. This is interpreted as being in agreement with Dr. Båth's conclusions, that there is a greater extinction rate for short period waves than for long period waves. Second, on the day of the hurricane's near approach, both stations showed a marked increase in amplitude, with the long period waves increasing in size much more than the shorter periods. It may be observed that Pinewoods, at roughly 50% greater distance from the storm than Harvard, showed an amplitude increase of 500% at 5 seconds, while Harvard showed an increase of 1,000% at 5 seconds; short period amplitudes about doubled at each station. This, incidentally, tends to verify the assumption that the short period (1.5 second) microseisms also originate in the ocean, as evidenced by the increase in amplitude associated with the hurricane.

Further knowledge of the microseismic spectrum in the vicinity of Troy, New York, was obtained on November 12, 1950. On that date, between the hours of 1100 to 1200 E.S.T., means of a vertical shaking table, so that the total response characteristic at any filter setting was determinable by simply adding the known filter response to that of the seismometer. Results are shown in Figure 6. The first line shows electrical noise in the system, which may be seen to be rather unimportant cominstrumental data were obtained showing microseismic amplitudes over a considerably wider frequency range. The instrumental setup consisted of a conventional vertical component electromagnetic seismometer whose output was amplified, played through a Krohn-Hite Ultra-Low Frequency Band-Pass Filter (model 330-

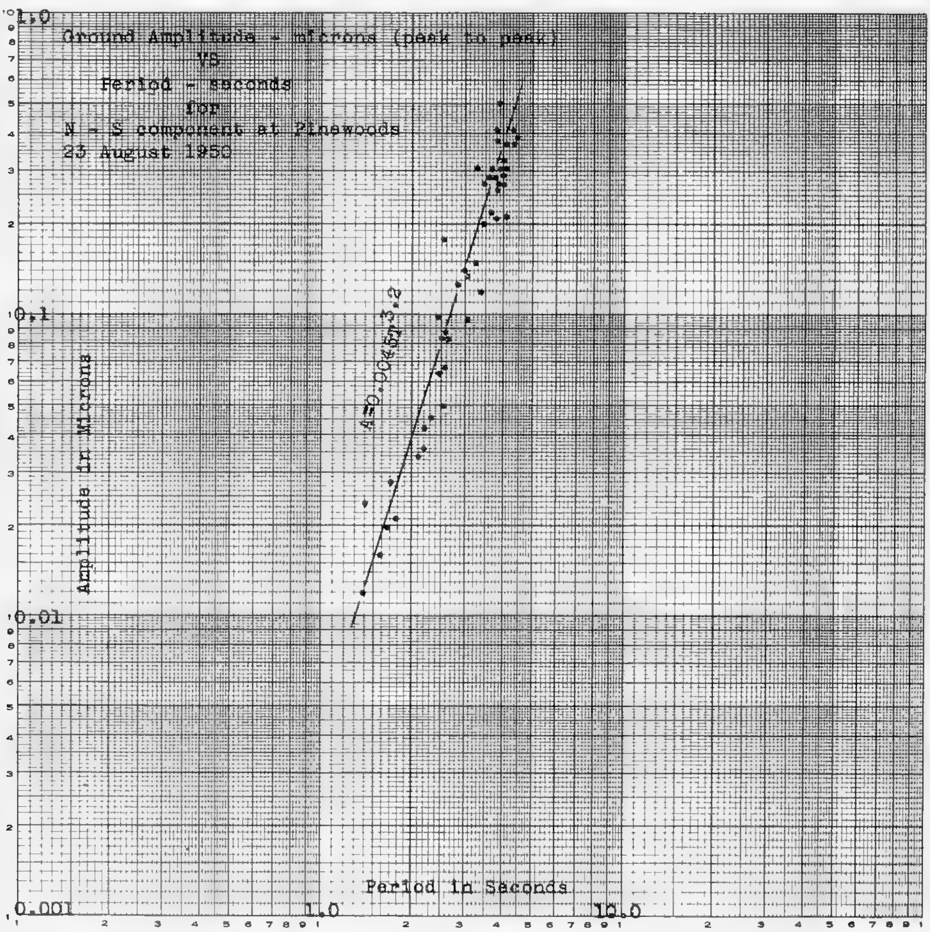

Figure 2. Microseismic Spectrum for Pinewoods, 23 August 1952. Normal day.

A), and recorded by means of a Brush recorder. The Krohn-Hite is a selective filter which will pass frequencies of an input signal, with no loss in gain, within any desired band between 2,000 cps and 0.02 cps (or 50 seconds period). High and low cut-off frequencies are independently adjustable. Outside of the pass band, the response falls off at 24 db per octave. Using this detector system, it was possible to examine the microseisms within a one octave region of the spectrum, and to vary the pass band so that it was possible to observe activity in several regions of the spectrum without interfering "noise" from other frequencies of ground noise. The seismometer was calibrated by pared to the recorded microseisms. Although the total instrumental magnification at successively larger periods decreases roughly as the cube of the period due to the response of the seismometer ($T_o \doteq 1.2$ seconds, damping critical) it was found necessary to further reduce the gain for longer periods; from this we infer that the ground amplitudes increase faster than the cube of the period. Calculated true ground amplitudes are shown on Figure 7, which shows an amplitude increase about proportional to the fourth power of the period in the range $0.5 \leq T \leq 5$ seconds. Indications are that the ground amplitude was maximum at 5.0 seconds, decreasing slowly for longer periods. Measurements on wave periods beyond 5 seconds are not shown, since there were evidences of instrumental instability at extremely long periods. It was estimated, however, that the waves with periods exceeding ten seconds had about 20% of the amplitude of the five second microseisms. Synoptic meteorological conditions on this date have not been studied.

No evidence for the existence of discrete microseismic "bands" at Harvard or Pinewoods was found during the investigation described here. In all cases, the sprectrum appeared to be continuous, or nearly so, with no wave period exhibiting amplitudes departing to any marked extent from the very regular rate of increase of amplitude with period. On the other hand, Macelwane and his colleaques have published examples showing narrow and sharply defined short period microseismic bands. It is tentatively suggested that the difference in these findings is attributable to regional geological differences; both Harvard and Pinewoods are situated on relatively homogeneous metamorphics, in contrast to the horizontal sedi-

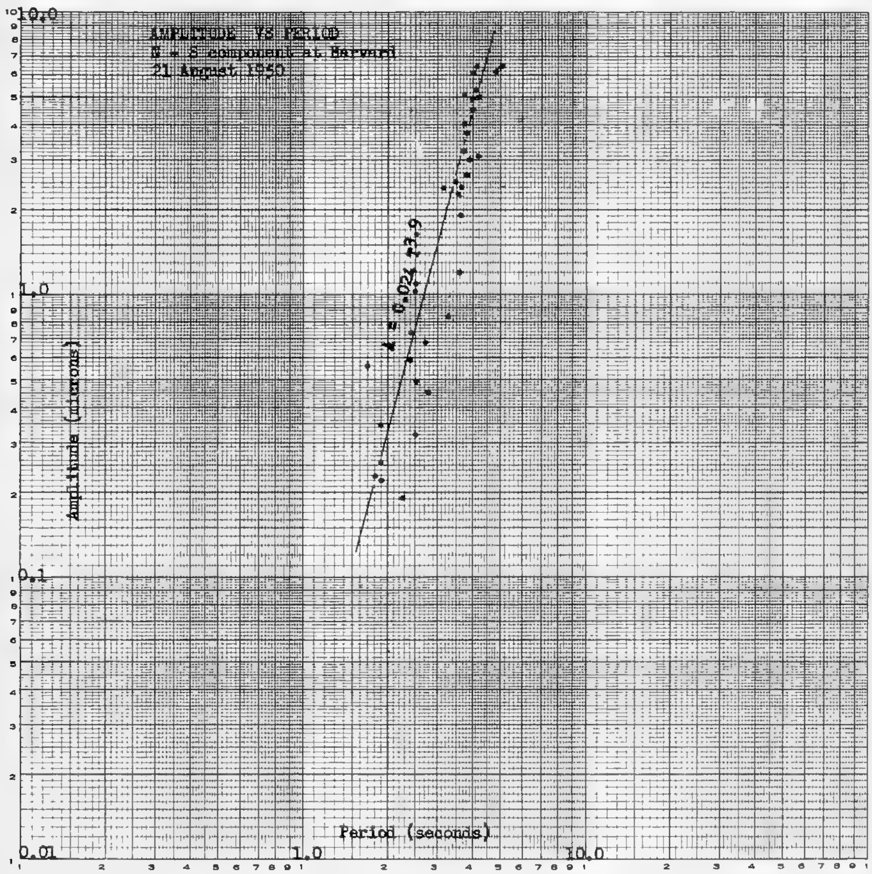

Figure 3. Microseismic Spectrum for Harvard, 21 August 1952. Hurricane at nearest approach to station.

mentary strata prevailing throughout the central United States.

Considering the great amplitude differences found for the different frequencies of motion, it is apparent that a "flat" seismograph response curve is useless for studying the microseismic spectrum over a wide frequency range. In such a case, the limiting factor is the dynamic range of the recorder, which will seldom exceed 40 db (100:1), whereas the phenomenon to be observed has an amplitude range of at least 60 db (1000:1). Under these conditions, analysis of the seismogram by any method will not reveal useful information on the short period components — such information being irretrievably lost in recording. Two approaches are possible in obtaining broad band microseismic coverage. One is a method used here, employing the selectivity characteristics of low frequency filters; or equivalently, using tuned detectors. The second approach is to use a seismograph response characteristic that is nearly the inverse of the microseismic spectrum. In this case, the product of ground motion and seismograph magnification will be nearly constant, and all ground frequency components will be registered equally well on the recorder. Fourier analysis or autocorrelation methods may then be used to obtain true ground amplitudes over a wide frequency range. It may be observed that the short period Benioff seismograph when critically damped has nearly the correct characteristics for the spectra discussed here, since the Benioff response decreases nearly as the cube of the period for periods longer than one half second. Mechanical seismographs, or seismographs with capacitance — bridge transducers will not have as great a microseismic band coverage, because their response decreases as only the square of the period on the long period side of the peak magnification.

Conclusions from the foregoing are that the microseismic spectrum at Pinewoods is essentially continuous in the range from ½ to 5 seconds, increasing at a rate about proportional to the third or fourth power of the period, depending to some extent on meteorological conditions over the adjacent regions on the North Atlantic Ocean; that the ocean is the source of at least part of the short period microseismic activity; and that the shorter period microseisms suffer a more rapid attenuation due to distance traveled than the longer periods. It

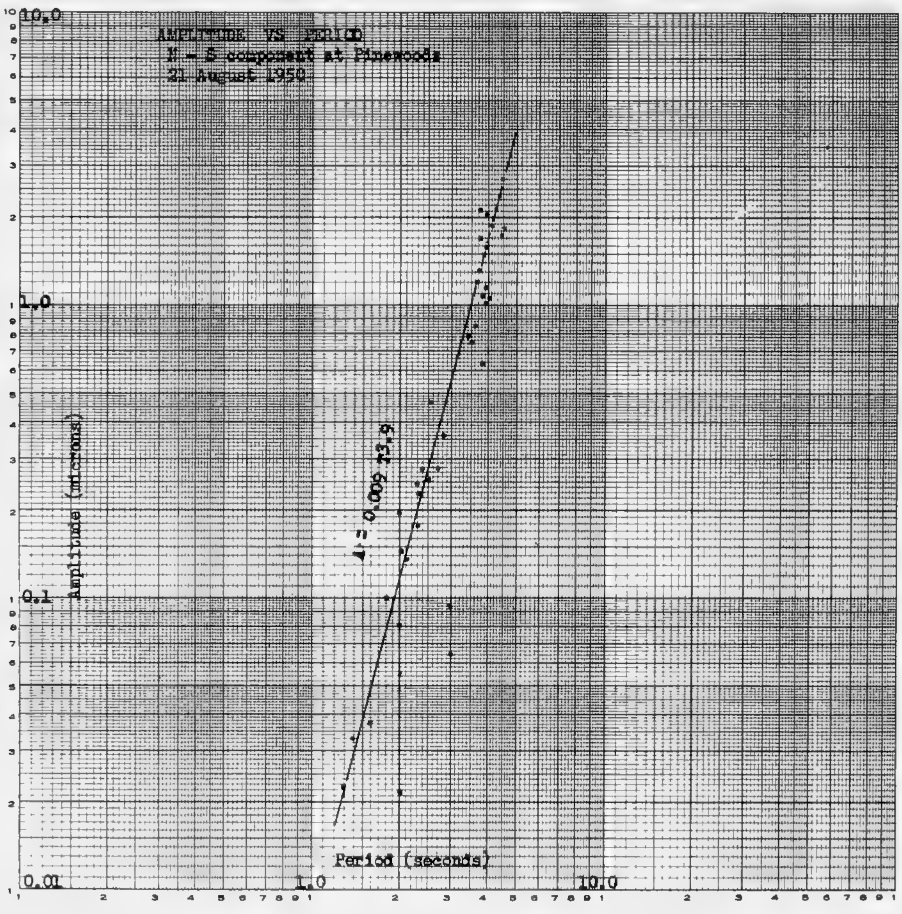

Figure 4. Microseismic Spectra for Pinewoods, 21 August 1952. Hurricane at nearest approach to station.

is also shown that broad band microseismic recording may be obtained on one seismograph, if the instrument and instrumental parameters are correctly selected.

Discussion from the Floor

(*Caldwell* asked how the periods for the period-amplitude graphs were selected. *Romney* replied, by looking for each period on the record and then measuring its amplitude. *Donn* inquired from Båth whether or not his fixed maxima of eight seconds and variable minumum corresponded with water depth. Båth replied that there was no connection apparent.)

(After *Father Macelwane's* paper, *Deacon* said that the lack of any relationship between microseism periods and amplitudes at Florrissant was not surprising if it was assumed that the microseisms were made by sea waves of twice their period. *Father Macelwane* had not examined short microseisms partly because they did not reach Florissant and partly because he had examined the most prominent groups. He was dealing with microseisms in the period range of 6½ to 8½ seconds. There is no relationship between the amplitude of waves of 13 to 17 seconds periods. Thirteen seconds would be the dominant wave period produced by a surface wind of 35 knots blowing for a long time, and there would be 17 seconds waves present in the complex wave-pattern produced by such a wind, but they would be much smaller than the 13 second waves. It is also possible that the 8½ second microseisms were made by 17 second swell which had a small amplitude because it had travelled a long distance over the ocean.)

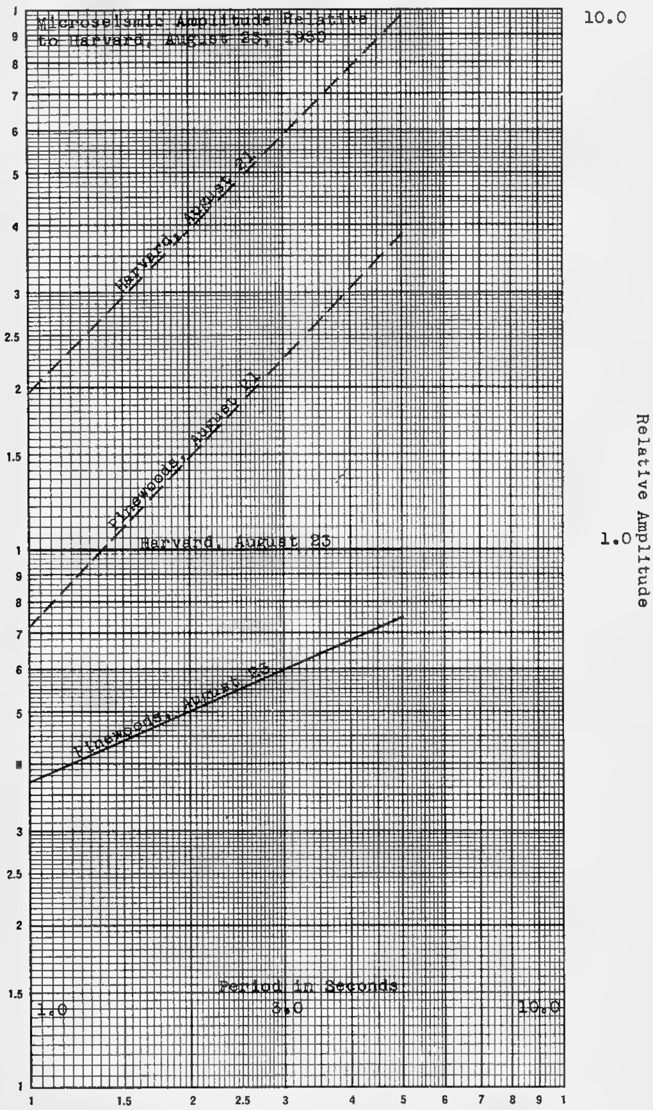

Figure 5. Ratios of Microseismic Spectra to Spectrum at Harvard on 23 August 1952.

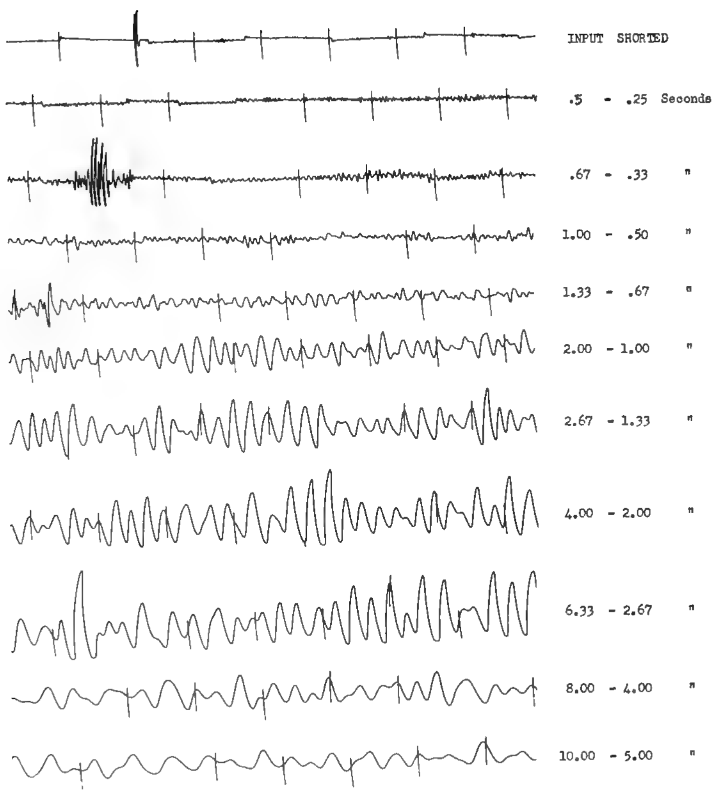

Figure 6. Microseismic Spectral Analysis by Krohn-Hite Filter. Figures to the right of each line give pass band of filter, but not of combined seismometer-filter system. Large disturbance on third line due to near approach at automobile. System gain reduced for longer periods.

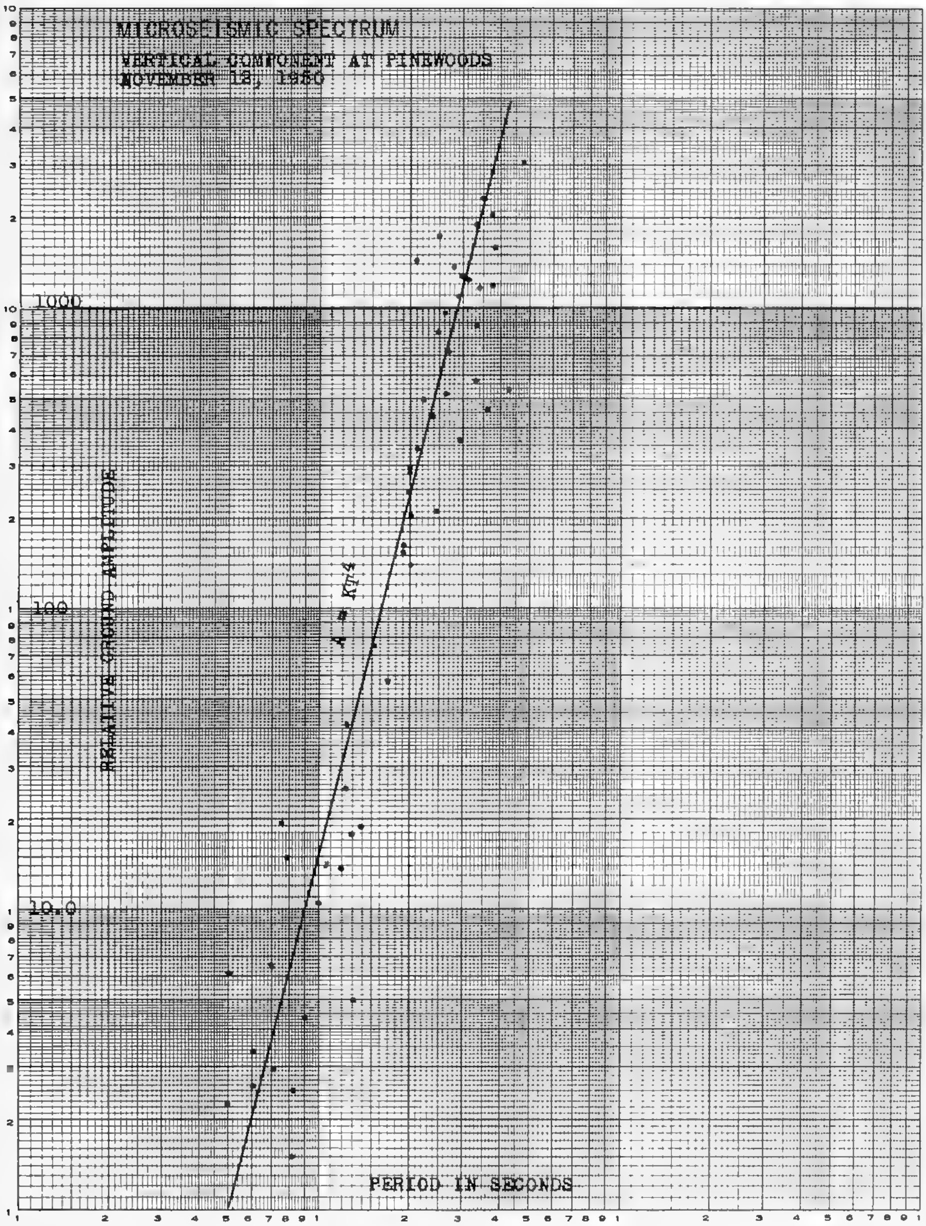

Figure 7. Microseismic Spectrum at Pinewoods from data of Figure 6. Only relative ground motion is shown.

CAN SEA WAVES CAUSE MICROSEISMS?

By M. S. Longuet-Higgins
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Abstract—This paper is an exposition of the "wave interference" theory of microseisms. Simple proofs are given of the existence, in water waves, of second-order pressure fluctuations which are not attenuated with depth. Such pressure fluctuations in sea waves may be sufficiently large to cause microseisms. The necessary conditions are the interference of opposite groups of waves, such as may occur in cyclones or by the reflection of waves from a coast.

Introduction—It has long been known that there is some connection between certain types of microseisms and deep atmospheric depressions over the ocean; and the similarity between microseisms and sea waves — their periodic character and the increase of their amplitude during a "storm" — naturally suggests some causal relation between them. But until recently there have seemed to be many difficulties, both theoretical and observational, to supposing that sea waves could, by direct action on the sea bed, be the cause of all these microseisms; for the latter have been recorded while the corresponding sea waves were still in deep water, whereas theory seemed to show that the pressure fluctuations associated with water waves were quite insufficient, at such depths, to produce any appreciable movement of the ground.

However, recent theoretical work in hydrodynamics has altered this situation: Miche (1944), in quite another connection, discovered the existence, in a standing wave, of second order pressure variations which are not attenuated with the depth; a much shorter demonstration of this result was given by Longuet-Higgins and Ursell (1948), and the result was extended by the present author (1950) to more general systems of waves. In the latter paper it was shown that such pressure variations may be quite sufficient, under certain circumstances, to produce the observed ground movement, the chief conditions required being the interference of waves of the same wavelength, but not necessarily of the same amplitude, travelling in opposite directions. This, then, may be called the "wave interference theory."

In the latter paper (which will be referred to as I) the results on which the theory depends were derived in a general and concise form, with detailed proofs. In view of the interest of the subject it seems desirable to clarify the main ideas behind the theory and to discuss further some of the more unexpected results. This will be attempted in the present paper, in which we shall rely as far as possible on physical reasoning, and refer where necessary to the former paper for rigorous proofs of the results quoted. We shall conclude with a brief historical review of the theory.

1. **The importance of the mean pressure**—Let us suppose that seismic waves are to be generated by some kind of oscillating pressure distribution acting on the surface of the earth or of the sea bed. If the period of the oscillation is T, and the corresponding wavelength of seismic waves is L, then the pressure distribution over an area whose diameter is small compared with L may be regarded as being applied at the same point, so far as the resulting disturbance is concerned; for the time-difference involved in applying any pressure at another point of the area would be small compared with T. Hence the resulting disturbance is of the same order of magnitude as if the mean pressure over the area were applied at the point. Now the wavelength of a seismic wave is many times that of a gravity-wave (sea wave) of the same period. It is therefore appropriate to consider the properties of the mean pressure, over a large number of wavelengths, in different kinds of gravity-wave. We shall first consider some very special but physically interesting cases, when the waves are perfectly periodic and the wave-train is infinite in length. It will be assumed for the moment that the water is incompressible.

2. **The progressive wave**—Consider any periodic, progressive disturbance which moves, unchanged in form, with velocity c (see figure 1). Let $\bar{p}$ (t) denote the mean pressure on a fixed horizontal plane (say the bottom) between two fixed points, A, B, separated by a wave length

We may show that $\bar{p}$ (t) is a constant. Let A and B denote the points, separated from A and B respectively by a distance ct. Then since the motion progresses with velocity c the mean pressure over A'B' at time t equals the mean pressure over A B at time O, i.e. $\bar{p}$ (0);

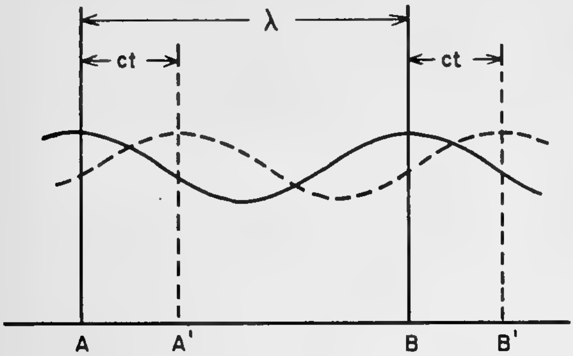

Figure 1. Positions of the profile of a progressive wave at two different times.

the total force on A'B' is $\lambda\ \bar{p}\ (O)$. But since the motion is periodic the force on A A' equals the force on B B'. Hence, by subtraction, the force on A B equals $\lambda\ \bar{p}\ (O)$; and the mean pressure on A B equals $\bar{p}\ (O)$ which is independent of the time. Thus there is no fluctuation in the mean pressure on the bottom over one wave-length, or over a whole number of wavelengths; in any interval containing more than N wavelengths the fluctuation in the mean pressure is less than $N^{-1}\ p_{max}$ where p_{max} is the maximum pressure in the interval. In other words, in a progressive wave the contributions to the disturbance from different parts of the sea bed tend to cancel one another out.

There is a second reason why progressive water waves may be expected to be relativly ineffective in producing seismic oscillations of the sea bed: not only the mean pressure fluctuation $\bar{p}$, but also the pressure fluctuation p at each point decreases very rapidly with depth and is very small below about one wavelength from the surface. This fact is closely connected with the vanishing of the mean pressure fluctuation; the motion below a certain horizontal plane can be regarded as being generated by the pressure fluctuations in that plane; and hence we should expect that the contributions to the motion from the pressure in different parts of the plane would tend to cancel one another out.

3. The standing wave—Consider now a standing wave, and let A and B be the points where two antinodal lines, a wavelength apart, meet the bottom (see figure 2). To a first approximation, a standing wave can be regarded as the sum of two progressive waves of equal wavelength and amplitude travelling in opposite directions. Therefore the mean pressure on the bottom between A B vanishes to a first approximation. However, the summation of the waves is not exact; if two progressive motions, each satisfying the boundary condition of constant pressure at the free surface, are added, (i.e. if the velocities at each point in space are added) there is no "free surface" in the resulting motion along which the pressure is always exactly constant; although if the elevations of the free surface are added in the usual way, the pressure is constant along this surface, to a first approximation. We should not expect the motions to be exactly superposable, on account of the non-linearity of the equations of motion.

It can be seen from the following simple argument that the mean pressure on the bottom, in a standing wave, must fluctuate. Consider the mass of water contained between the bottom, the free surface, and the two nodal planes shown in figure 2. Since there is no flow across the nodal planes, this mass consists always of the same particles; therefore the motion of the center of gravity of this mass is that due to the external forces alone which act upon it. Figure 2 shows the mass of water in four phases of the motion, separated by intervals of one quarter of a complete period. In the first and third phases the wave crests are fully formed, and in the second and fourth phases the surface is relatively flat (though never exactly flat; see Martin et al., 1952). When the crests are formed the centre of gravity of the mass is higher than when the surface is flat, since fluid has, on the whole, been transferred from below the mean surface level to above it. Thus the centre of gravity is raised and lowered twice in a complete cycle. But

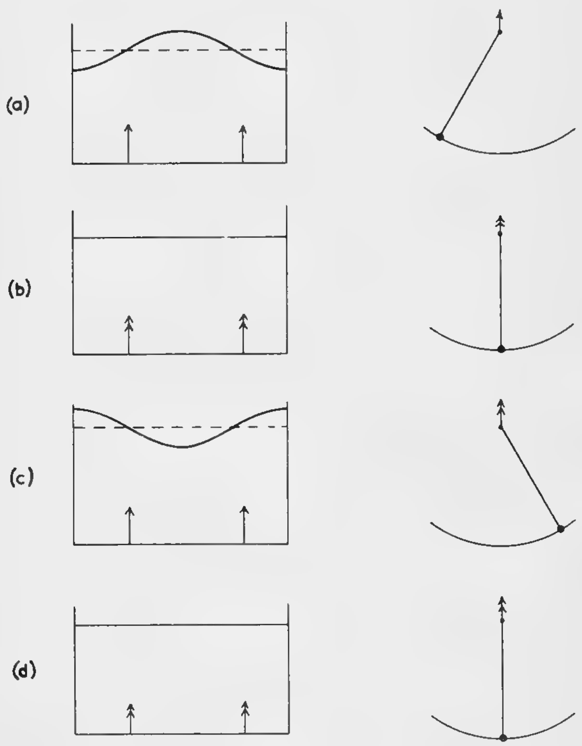

Figure 2. Comparison of a standing wave with a swinging pendulum, at four different phases of the motion separated by a quarter of a period.

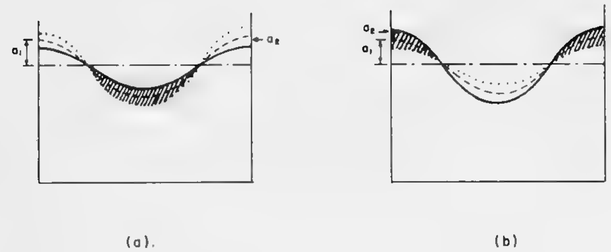

Figure 3. Two phases of the interference between two waves of equal length but different amplitudes a_1 and a_2 travelling in opposite directions. The profile of the first wave (dashed line) is reduced to rest by superposing on the system a velocity $-c$; the second wave appears to travel over the first with velocity $-2c$. The full line shows the final wave form.

the external forces acting on the mass are, first, that due to gravity, which is constant, (the total mass being constant); secondly the force from the atmosphere, which is also constant, since the pressure p_o at the free surface, if constant, will produce a constant downwards force λp_o; thirdly the forces across the vertical planes, which must have zero vertical component, the motion being symmetrical about these planes; and, lastly, the force on the bottom, which equals $\lambda\ \bar{p}$. Since all the other external forces besides $\lambda\ \bar{p}$ are constant it follows that $\bar{p}$ must fluctuate with the time. In figures 2(a) and 2(c) the mass of water above the mean level is proportional to the wave amplitude a; since it is raised through a distance of the order of a, the displacement of the centre of gravity, and hence the mean pressure fluctuation, is proportional to a^2.

An explicit expression for p can easily be derived. Let z denote the vertical coordinate of a particular element of fluid of mass m, so that z is a function of the time t and of, say, the position of the fluid element when $t = o$. If F denotes the vertical component of the external forces acting on the mass of water, we have, on summing the equations of motion for each element of fluid, and cancelling the internal forces:

$$F = \sum\left(m\,\frac{\partial^2 z}{\partial t^2}\right) = \frac{\partial^2}{\partial t^2}\left(\sum m\ z\right) \quad (1)$$

the summation being over all the particles. The expression in brackets on the right-hand side will be recognized as g^{-1} times the potential energy of the waves; in an incompressible fluid

$$\sum m\ z = \rho\int_0^{\lambda} \tfrac{1}{2}\ \zeta^2\ dx + \text{constant} \quad (2)$$

where x is a horizontal coordinate, p is the density, λ is the wavelength, and ζ (x, t) is the vertical displacement of the free surface. But by our previous remarks

$$F = \lambda\,(\bar{p} - p_o - \rho g\ h\ ), \quad (3)$$

where h is the mean depth of water. On equating (1) and (3) we find

$$\frac{\bar{p} - p_o}{\rho} - g\ h = \frac{\partial^2}{\partial t^2}\ \frac{1}{\lambda}\int_0^{\lambda} \tfrac{1}{2}\ \zeta^2\ dx. \quad (4)$$

Now for a standing wave

$$\zeta = a \cos kx \cos \sigma t \quad (5)$$

where $k = 2\ \pi/\lambda$ and $\sigma = 2\pi/\tau$ (τ being the wave period), and higher-order terms have been omitted. On substituting in (4) we find, after simplification,

$$\frac{\bar{p} - p_o}{\rho} - g\ h = -\tfrac{1}{2}\ a^2\ \sigma^2 \cos 2\sigma t \quad (6)$$

This shows that, to the second order, the mean pressure $\bar{p}$ fluctuates sinusoidally, with twice the frequency of the original wave, and with an amplitude proportional to the square of the wave amplitude. The pressure fluctuation is independent of the depth, for a given wave period, though of course the depth enters into the relation of the wave period to the wavelength, given by

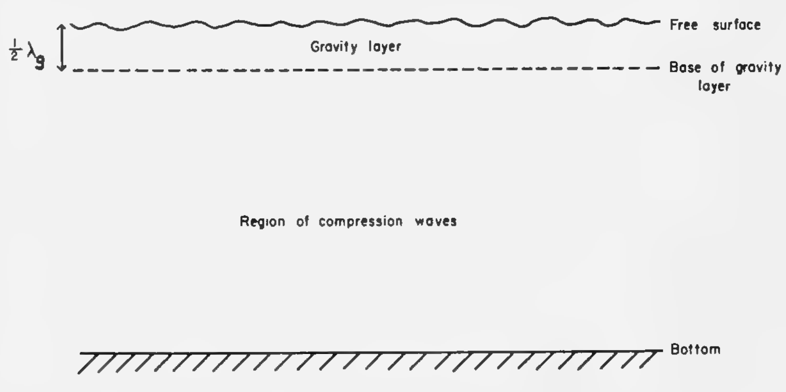

Figure 4. Waves in a heavy, compressible fluid.

$$\sigma^2 = g\ k \tanh\ k\ h \tag{7}$$

There is a close analogy with the motion of a pendulum (see figure 2). In a complete cycle the bob of the pendulum is raised and lowered twice, through a distance proportional to the square of the amplitude of swing, when this is small. The only forces acting on the pendulum are gravity, which is constant, and the reaction at the support. Hence there must be a second-order fluctuation in the vertical component of the reaction at the support. Furthermore the reaction will be least when the pendulum is at the top of the swing (the potential energy is greatest) and will be greatest when the pendulum is at the bottom of its swing (the potential energy is least).

It will be noticed that the above analytical proof does not necessarily involve the idea of the centre of gravity, whose vertical coordinate $\bar{z}$ is defined by

$$(\Sigma\, m)\ \bar{z} = \Sigma\, (m\ z)\ . \tag{8}$$

The theorem on the centre of gravity that was used previously is in fact usually derived from equation (1): but in the present proof we have appealed directly to the original equations of motion for the individual particles, without introducing $\bar{z}$.

4. Two progressive waves—The above proof can easily be extended to the more general case of two waves of equal period but unequal amplitude travelling in opposite directions. For, such a disturbance is exactly periodic in space. Thus we may consider a region one wavelength in extent, as for the standing wave. This will not always contain the same mass of

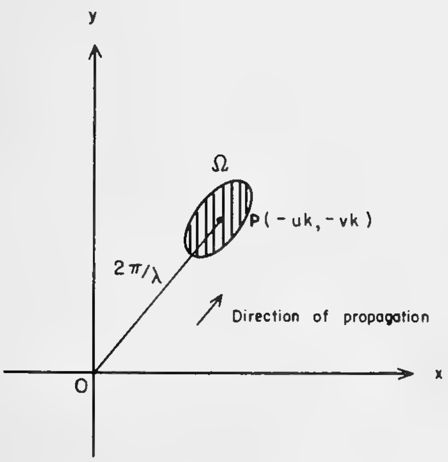

Figure 5. The spectrum representation of a wave group.

water; but, owing to the periodicity, the vertical reaction on the bottom due to the flow of water across one vertical boundary will be exactly cancelled by that due to the flow across the opposite boundary (see I Section 2.2); thus equation (4) is still exactly valid. The wave profile in this case is represented by

$$\zeta = a_1 \cos\ (kx - \sigma t) + a_2 \cos\ (kx + \sigma t) \tag{9}$$

and so

$$\frac{1}{\lambda} \int_0^{\lambda} \tfrac{1}{2}\ \zeta^2\ dx =$$

$$\tfrac{1}{4}\ (a_1^2 + a_2^2 + 2 a_1\ a_2 \cos\ 2\sigma t) \tag{10}$$

giving

$$\frac{\bar{p} - p_0}{\rho} - g\ h = - 2 a_1\ a_2\ \sigma^2 \cos\ 2\sigma t \tag{11}$$

The mean pressure fluctuation on the bottom is therefore proportional to the product of the two wave amplitudes a_1 and a_2. When these two are equal ($a_1 = a_2 = \frac{1}{2}$ a) we have the case of the standing wave, and when one is zero (a_1 = a; a_2 = 0) we have the case of the single progressive wave.

A physical explanation of this result may be given as follows. Suppose that one of the waves, say the wave of amplitude a_1, is reduced to rest by superposing on the whole system a velocity - c in the direction of x decreasing (this will not affect the pressure distribution on the bottom). The second wave will now travel over the first with a velocity −2c. The crests of the second wave will pass alternately the troughs and the crests of the first wave - each twice in a complete period. Figure 3 shows the two phases. One may pass from figure 3(a) to figure 3(b) by transferring a mass of fluid, proportional to a_2, from a trough to a crest of the original wave, i.e. through a vertical distance proportional to a_1 (the transferred mass does not of course consist of identically the same particles of water). The vertical displacement of the centre of gravity of the whole mass is therefor shifted by an amount proportional to $a_1 a_2$; and hence the fluctuation in $\bar{p}$ is also proportional to $a_1\ a_2$.

5. Attenuation of the particle motion—The fact that there is a pressure fluctuation on the bottom even in deep water does not, however, mean that there is movement at those depths. In fact it may be shown (Longuet-Higgins 1953) that in exactly space-periodic motion, whether in a simple progressive wave or a combination of such waves, the particle motion decreases exponentially with the depth, apart from a possible steady current. Now if the velocities at great depths are zero, or steady, it follows from the equations of motion that the pressure-gradient must be independent of the time. Thus if there is a pressure fluctuation it must be uniform in space, i.e. it must be applied equally at all points of the fluid. This indicates that below a certain depth, in a strictly space-periodic motion, the pressure fluctuations are uniform and equal to the fluctuation $\overline{p}(t)$ in the mean pressure on the bottom, which has been evaluated. The effect of the waves, at great depths, is then the same as would be produced by an oscillating pressure applied uniformly at the upper surface of the fluid—for example an oscillation of the atmospheric pressure. Alternately one may imagine a rigid plane or raft to be floating on the surface of the water and completely covering it, and the pressure to be applied to this plane by means of a weight attached to a spring and oscillating in a vertical direction.

6. An experimental verification—The above results were verified experimentally (Cooper and Longuet-Higgins 1951) in the following way. Waves were generated at one end of a wave tank and allowed to travel towards the far end, where they were dissipated on a sloping beach. The pressure beneath the waves was detected by means of a hydrophone and was recorded continuously. On starting the motion from rest, no appreciable pressure fluctuations were recorded until the wave-front, travelling with approximately the group-velocity of the waves, passed over the hydrophone. The pressure fluctuations then built up quickly to a constant amplitude, and had a period equal to that of the waves. The amplitude agreed well with the first-order theory; it diminished exponentially with depth, and was negligible below about half a wavelength.

A vertical barrier was then placed in the wave tank, between the hydrophone and the beach, which reflected the waves back over the hydrophone. As soon as the reflected wave

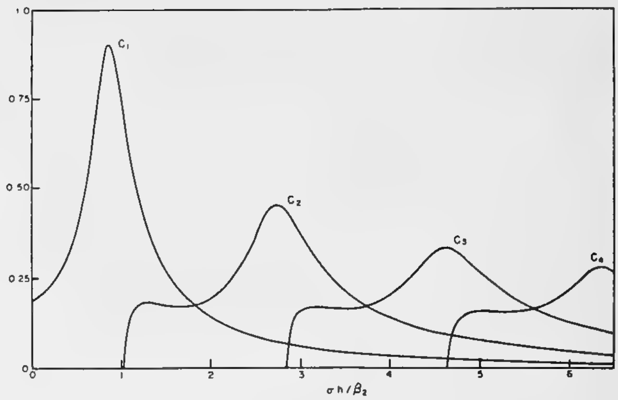

Figure 7. Graph of C_1, C_2, C_3 and C_4 as function of $\sigma h/\beta_2$, showing the relative amplitude of the vertical displacement of the "sea bed" in the first four modes.

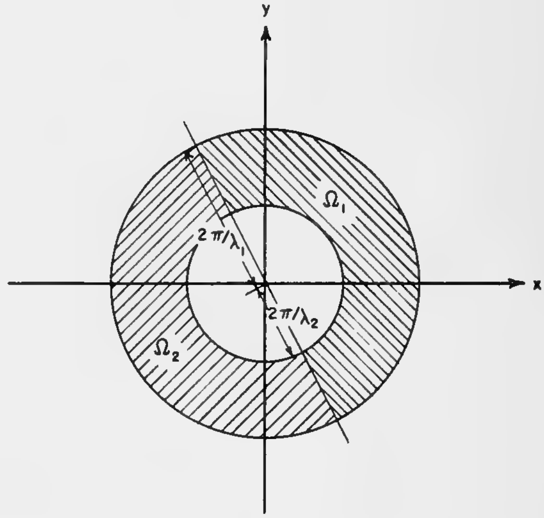

Figure 8. The form of the wave spectrum in a circular storm.

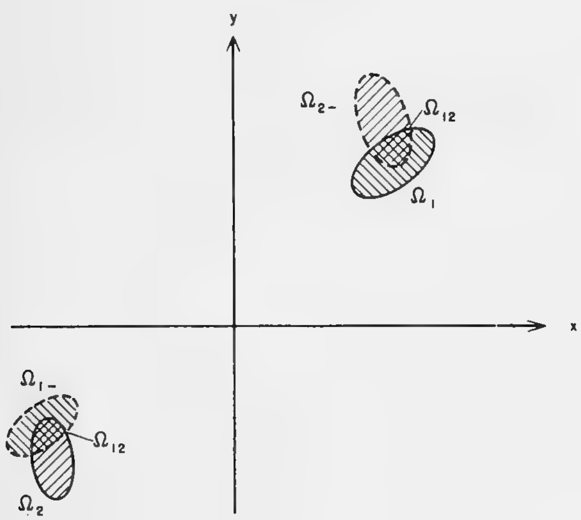

Figure 6. The regions of interference of two groups of waves in the spectrum.

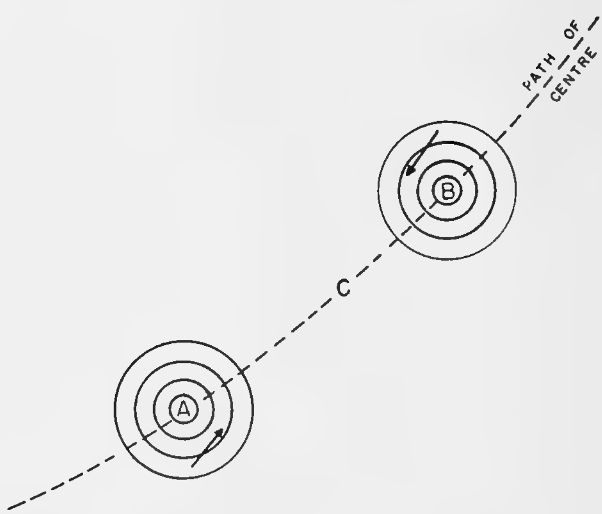

Figure 9. Wave interference caused by moving cyclonic depression.

front arrived over the hydrophone the appearance of the pressure record was changed. At moderate depths there were not only first-order pressure fluctuations from the incident and the reflected wave, but also considerable second-order pressure fluctuations, of twice the fundamental frequency. At greater depths the first-order pressure fluctuations become negligible and only the pressure fluctuations of double the frequency remained. The amplitude of these was in good agreement with equation (6). When the barrier was removed, and the rear end of the reflected wave train had passed the hydrophone, the second-order pressure fluctuations rapidly died out.

Interference between waves of unequal amplitude was obtained by placing in the tank a vertical barrier extending only to a certain depth below the free surface, which allowed the waves to be partly reflected and partly transmitted. The coefficient of reflection from such a barrier is known theoretically for different ratios of the depth of the barrier to the wavelength of the waves, and it was verified that the amplitude of the second-order pressure fluctuations was proportional to the amplitude of the reflected wave. Indeed this property seems to provide a convenient method of actually measuring the coefficient of reflection from different types of obstacles or from plane beaches.

Since standing waves produce only second-order pressure fluctuations below moderate depths one would expect that, if pressure fluctuations were induced deep in the water, standing waves of half the frequency would be produced at the surface. An experiment of this kind was in fact performed by Faraday (1831); (see Section 13 of the present paper) who produced standing waves, of half the fundamental frequency, by means of a vibrating lath inserted in a basin of water. Faraday remarked that the general result was little influenced by the depth of water: "I have seen the water in a barrow, and that on the head of an upright cask in a brewer's van passing over stones, exhibit these elevations." (1831, footnote to p. 334). The present author has observed a similar phenomenon on board ship: a pool of water on deck, when excited by the vibration of the ship's engines, sometimes shows a standing-wave pattern whose amplitude gradually builds up to a maximum, and then collapses; the process is repeated indefinitely.

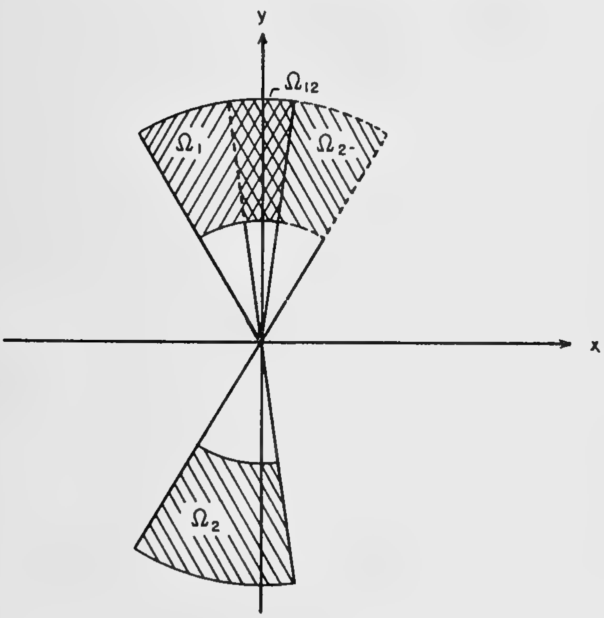

Figure 10. The spectrum representation of incident and reflected wave-groups.

7. Standing waves in a compressible fluid— The water has so far been assumed to be incompressible, and we have seen that in this case the pressure fluctuations below about half a wavelength from the surface occur simultaneously at all points of the fluid. But this can only be true if the least time taken for a disturbance to be propagated to the bottom and back is small compared with the period of the waves. In the deep oceans, where the speed of sound is about 1.4 km/sec and the depth may be of the order of several kilometers, this time may be several seconds. Thus the compressibility of the water must be considered.

The first-order theory of waves in a heavy, compressible fluid (in which all squares and products of the displacements are neglected) indicates that water waves of a few seconds' period fall into two classes (Whipple and Lee 1935). On the one hand there are waves approximating very nearly to ordinary surface waves in an incompressible fluid, in which the particle displacement decreases exponentially downwards, to first order; these may be called gravity-waves. On the other hand there are long waves controlled chiefly by the compressibility of the medium and hardly attenuated at all with depth; these may be called compression-waves; their velocity is nearly the velocity of sound in water. The wavelengths of a gravity-wave and a compression wave will be denoted by λ_g and λ_c respectively. For waves of period 10 sec. λ_g/λ_c is of the order of 10^{-2}.

However, the pressure variations which are of interest to us at present are of second order. To investigate the effect of the compressibility, therefore, a complete example, namely a motion which in the first approximation is a standing gravity-wave, has been worked out in full to a second approximation (I Section 4). The result is as follows.

Near the free surface, that is within a distance small compared with λ_c, the waves are unaffected by the compressibility of the water

—as one might expect, since a disturbance could be propagated almost instantaneously through this layer. At a distance of about ½ λ_g from the free surface the first-order pressure variations are much attenuated, and the second-order pressure variations are practically those given by the incompressible theory (equation [6]). Below this level the displacements are comparatively small, but, instead of the uniform, unattenuated pressure fluctuations in the incompressible fluid, there is now a compression wave, whose planes of equal phase are horizontal: the pressure field in this wave is given by

$$\frac{p - p_0}{\rho} - g\,z = -\tfrac{1}{2}\,a^2\,\sigma^2\,\frac{\cos 2\sigma(z-h)/c'}{\cos 2\sigma h/c'}\cos 2\sigma t \qquad (12)$$

very nearly, where z is the vertical coordinate measured downwards from the mean surface level, and c' is the velocity of sound in water. This wave can be regarded as being generated by the unattenuated pressure variation (6). There is a resonance, or "organ-pipe," effect: when $\cos 2\sigma h/c'$ vanishes, the pressure on the bottom $(z = h)$ becomes infinite. This happens when

$$2\sigma h/c' = (n + \tfrac{1}{2})\pi \qquad (13)$$

that is, when the depth is $(\tfrac{1}{2} n + \tfrac{1}{4})$ times the length of the compression wave. In general, however, the displacements in the compression wave are small, being only of the order of a^2/λ_c; the displacement of the centre of gravity of the layer at the surface of thickness ½ λ_g is of the order of a^2/λ_g. This explains why the compressibility of the fluid below has little effect on the pressure fluctuations at the base of the surface layer.

We have then the following picture (see figure 4): there is a surface-layer, of depth about ½ λ_g, in which the compressibility of the water is, in general, unimportant: this may be called the "gravity-layer." Below this layer there exist only second-order compression waves, generated by the gravity-waves in the surface layer, and of twice their frequency.

8. **Application to sea waves**—So far we have considered only the very special cases of perfectly periodic and two-dimensional waves. Such waves cannot be expected to occur in the ocean, although the sea surface usually shows a certain degree of periodicity. We shall now consider how the sea surface is to be described in this more general case.

It can be shown (See I Section 3.2) that any free motion of the sea surface can be expressed as a Fourier integral:

$$\zeta(x, y, t) = R\int_{-\infty}^{\infty}\int_{-\infty}^{\infty} A(u,v)\,e^{i(ukx + vky + \sigma t)}\,du\,dv \qquad (14)$$

where (x, y) are horizontal coordinates, k is a constant and σ is a function of (u, v):

$$\sigma^2 = g\,k\,(u^2 + v^2)^{\frac{1}{2}}\tanh\,(u^2 + v^2)^{\frac{1}{2}}\,k\,h \qquad (15)$$

A (u, v) is in general complex, and R denotes the real part. The expression under the integral sign represents a long-crested wave with crests parallel to the line

$$u\,x + v\,y = 0 \qquad (16)$$

and of wavelength λ given by

$$\lambda = \frac{2\pi}{(u^2 + v^2)^{\frac{1}{2}}\,k} \qquad (17)$$

If the point P, = (- uk, - vk) is plotted in the (x, y) plane (see figure 5) the direction of the vector O P is the direction of propagation of the wave-component and the length of O P equals 2π divided by the wavelength. Points on a circle centre O correspond to wave components of the same wavelength; diametrically opposite points correspond to waves of the same length but travelling in opposite directions. When the energy is mainly grouped about one wavelength and direction, the complex amplitude A(u, v) will be appreciably large only in a certain range of values of (u, v), say Ω, as in figure 5. The narrower this region, the more regular will be the appearance of the waves.

The spectrum A(u, v) of the waves is determined uniquely by the motion of the free surface, at a particular instant, over the whole plane (see I, Section 3.2). Since we shall want to consider the wave motion in only a certain part of the plane, say a square S of side 2R, it is convenient to define a motion ζ' which, at any time, has the same value as ζ inside S but is zero outside. Let A' be the spectrum function of ζ', so that

$$\zeta' = R\int_{-\infty}^{\infty}\int_{-\infty}^{\infty} A'(u,v)\,e^{i(ukx + vky + \sigma t)}\,du\,dv \qquad (18)$$

A' is very closely related to A; if k is chosen so that

$$k = \pi / R \quad (19)$$

and if R is large compared with the wavelengths associated with most energy in the spectrum then (see I Section 3.3)

$$A'(u,v) = \int_{\infty}^{\infty}\int_{\infty}^{\infty} A(u_1,v_1) \frac{\sin (u-u_1)\pi}{(u-u_1)\pi} \frac{\sin (v-v_1)\pi}{(v-v_1)\pi} e^{-i(\sigma-\sigma_1)} du_1 dv_1 \quad (20)$$

where $\sigma_1 \equiv \sigma(u_1, v_1)$. In other words A' is the weighted average of values of A over neighboring wavelengths and directions. Since u and v are proportional to the number of wavelengths intercepted by the x— and y—axis in S, a "neighboring" wave component is one which has nearly the same number of wavelengths, in each direction, in S. A' gives a "blurred" picture of A; but the larger the side of the square, the less is the blurring. The region Ω' in the (u, v)—plane which corresponds to the blurred spectrum will be almost the same as the region Ω corresponding to the original spectrum. A' also varies slowly with the time—the waves in S change gradually—but this rate of change is slow compared with the rate of change of the wave profile, or compared with $\sigma A'$.

The energy of the waves is given very simply in terms of the spectrum function A'; in fact, if a denotes the amplitude of the single long-crested wave which has the same mean energy inside S,

$$a^2 = \int_{-\infty}^{\infty}\int_{-\infty}^{\infty} A'A'^{\star} \; du\, dv \quad (21)$$

where a star denotes the conjugate complex function (I equation [189]). a may be called the equivalent wave amplitude of the motion.

9. General conditions for fluctuations in the mean pressure—We shall evaluate the mean pressure p at the base of the gravity-layer, i.e. at a distance of about $\frac{1}{2}\lambda_g$ below the free surface, over a square of side $2R$. (Here λ_g refers to the mean wavelength of the predominant components in the spectrum.) Consider first the two-dimensional case. The mass of water contained between the surfaces $z = \zeta$ and $z = \frac{1}{2}\lambda_g$ and the planes $x = \pm R$ no longer consists of the same particles of water; but it is possible to extend the analysis of Section 3 so as to take account of the motion across the boundaries (see I Section 2.2). Provided that the horizontal extent $2R$ of the interval is large compared with λ_g the effect of the flow across the vertical boundaries can be neglected (I Section 3.1). Further, since the motion decreases rapidly with depth the effect of flow across the horizontal plane $z = \frac{1}{2}\lambda_g$ is small. The expression for the mean pressure variation is therefore the same as if the free surface were the only moving boundary:

$$\frac{\bar{p} - p_0}{\rho} - \tfrac{1}{2} g\, \lambda_g = \frac{\delta^2}{\delta t^2} \frac{1}{2R} \int_{-R}^{R} \tfrac{1}{2}\zeta^2 \; dx. \quad (22)$$

Similarly in the three-dimensional case

$$\frac{\bar{p} - p_0}{\rho} - \tfrac{1}{2} g\, \lambda_g = \frac{\delta^2}{\delta t^2} \frac{1}{4R^2} \int_{-R}^{R}\int_{-R}^{R} \tfrac{1}{2}\zeta^2 \; dx\, dy, \quad (23)$$

that is

$$\frac{\bar{p} - p_0}{\rho} - \tfrac{1}{2} g\, \lambda_g = \frac{\delta^2}{\delta t^2} \frac{1}{4R^2} \int_{-\infty}^{\infty}\int_{-\infty}^{\infty} \tfrac{1}{2}\zeta'^2 \; dx\, dy, \quad (24)$$

since ζ' vanishes outside the square S. Now the expression on the right-hand side is closely related to the potential energy of the motion ζ', and can be simply expressed in terms of the Fourier spectrum—function A'. In fact (I Section 3.2)

$$\int_{-\infty}^{\infty}\int_{-\infty}^{\infty} \tfrac{1}{2}\zeta'^2 \; dx\, dy = R(\pi/k)^2 \int_{-\infty}^{\infty}\int_{-\infty}^{\infty} (A'A'^{\star} + A'A'_{-} e^{2i\sigma t}) \; du\, dv \quad (25)$$

where A'_- stands for $A'(-u, -v)$, and is the amplitude of the wave component opposite to $A(u, v)$. On substituting in (24) we have

$$\frac{\bar{p} - p_o}{\rho} - \tfrac{1}{2} g \lambda_g = - R \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} \sigma^2 \; A' A'_- \; e^{2i\sigma t} \, du \, dv \qquad (26)$$

This shows that fluctuations in the mean pressure p arise only from opposite pairs of wave components in the spectrum; that the contribution to p from any opposite pair of wave components is of twice their frequency and proportional to the product of their amplitudes; and that the total pressure fluctuation is the integrated sum of the contributions from all opposite pairs of wave components separately.

The necessary condition for the occurrence of second-order pressure fluctuations of this type is, therefore, that the sea disturbance should contain some wave-groups of appreciable amplitude which are "opposite," i.e. such that part at least of the corresponding region in the Fourier spectrum is opposite to some other part. For example, if Ω lies entirely on one side of a diameter of the (u, v)—plane, the mean pressure fluctuation, to the present order, vanishes.

An important case is when the disturbance consists of just two wave groups, corresponding to regions Ω_1 and Ω_2, and of equivalent amplitudes a_1 and a_2 (see figure 6). Ω_{1-} and Ω_{2-}, denote the regions opposite to Ω_1 and Ω_2 and Ω_{12} and Ω_{12} denote the regions common to Ω_1 and Ω_{2-} and to Ω_{1-} and Ω_2 respectively. Effectively, then, the integration in (26) is carried out over the two regions Ω_{12} and Ω_{12-}. When the spectrum is narrow an order of magnitude for the integral on the right-hand side of (26) can be obtained. It may be shown (see I Section 5.2) that

$$\frac{\bar{p} - p_o}{\rho} - \frac{1}{2} g \lambda_g \sim 2 a_1 \, a_2 \, \sigma_{12}^2 \; (\Omega_{12}/\Omega_1 \, \Omega_2)^{1/2} \, k e^{2i\sigma_{12} t} \qquad (27)$$

where σ_{12} is the mean value of σ in Ω_{12}. Thus the mean pressure on S increases proportionately to the square root of the region Ω_{12} of overlap of the wave groups, and inversely as the square root of Ω_1 and Ω_2 separately, for fixed values of a_1 and a_2.

10. Calculation of the ground movement—In order to estimate the movement of the ground, at great distances, due to waves in a storm area Λ, we suppose the storm area to be divided up into a number of squares S of side 2R such that S contains many wavelengths λ_g of the sea waves, but is only a fraction, say less than half, of the length of a seismic wave λ_s in the ocean and sea bed. This we may do, since the wavelengths of seismic waves are of the same order as the wavelengths of compression waves in water; therefore λ_g/λ_s is of the order of 10^{-2}. The mean pressure or total force on the base the gravity-layer can be calculated as in Section 9; the vertical movement of the ground δ' due to the waves in this square is of the same order as if the force were concentrated to a point at the center of the square, i.e.

$$\delta' \sim 4R^2 \,.\, 2 a_1 \, a_2 \, \sigma_{12}^2 \; (\Omega_{12}/\Omega_1 \, \Omega_2)^{1/2} \, k \, \bar{W} \, (2\sigma_{12}, r) \, e^{2i\sigma_{12} t} \qquad (28)$$

where r is the distance from the center of the storm and $\bar{W}(\sigma, r) e^{i\sigma t}$ is the movement of the ground at distance r due to a unit pressure oscillation $e^{i\sigma t}$ applied at a point in the mean free surface. The pressure can be considered to be applied in the mean free surface rather than at the base of the gravity-layer, since the latter is relatively thin compared with the length of the seismic waves. To find the total displacement δ from the storm we may add the energies from the different squares S, on the assumption that the contributions from the different squares are independent. Since there are $\Lambda/4R^2$ such squares in the whole storm area, this means that the disturbance δ' from each individual square is to be multiplied by $\Lambda^{1/2}/2R$. Hence we have

$$\delta \sim 4\pi \,.\, a_1 \, a_2 \, \sigma_{12}^2 \; (\Lambda \, \Omega_{12}/\Omega_1 \, \Omega_2)^{1/2} \, \bar{W} \, (2\sigma_{12}, r) \, e^{2i\sigma_{12} t} \qquad (29)$$

To calculate $\bar{W}(\sigma, r)$ we may consider the disturbance due to a force applied at the surface of a compressible fluid of depth h (representing the ocean) overlying a semi-infinite elastic medium (representing the sea bed). Although this model takes no account of variations in the depth of water, or of the propagation of the waves from the sea bed to the land or across geological discontinuities, it can nevertheless be expected to give a reasonable estimate of the order of magnitude of the ground movement.

The disturbance W (σ, r) $e^{i\sigma t}$ at great distances from the oscillating point source $e^{i\sigma t}$ consists of one or more waves of surface type, i.e. waves spreading out radially in two dimensions (see I Section 5.1). Thus

$$W(\sigma,r)\ e^{i\sigma t} = \frac{\sigma^{1/2}}{\rho_2\ \beta_2 (2\pi r)^{1/2}} \sum_m C_m e^{i[\sigma t - \xi_m r + (m+\frac{1}{4})\pi]} \tag{30}$$

where ρ_2 is the density of the elastic medium, β_2 the velocity of secondary waves in the medium, $2\pi/\xi_m$ is the wavelength of the mth wave and C_m is a constant amplitude depending on the depth of water and on the elastic properties of the fluid and the underlying medium. The first wave has no nodal plane between the free surface and the "sea bed," the second has one nodal plane, the third two, and so on. When the depth h of the water is small, only the first type of wave can exist; the others appear successively as the depth is increased. Graphs of C_1, C_2—have been computed for some typical values of the constants: ρ_1 (the density of the fluid) $= 1.0$ g./cm^3; c' (velocity of compression waves in water) $=1.4$ km./sec.; $\beta_2 =$ 2.8 km./sec., and with Poisson's hypothesis, that the ratio of the velocities of compressional and distortional waves in the medium is $\gamma\sqrt{3}$. The results are shown in figure 7, where C_1, C_2, C_3 and C_4 are plotted against $\sigma h/\beta_2$. C_1, for example, increases to a maximum when $\sigma h/\beta_2 = 0.85$, i.e. when $h = 0.27 \times 2\pi c'/\sigma$, or h is about one-quarter of the wavelength of a compression wave in water. This maximum may therefore be interpreted as a resonance peak. The amplitude, however, does not become infinite as in the case of the infinite wave-train discussed in Section 7, since now energy is being propagated outwards from the generating area. C_2, C_3, and C_4 have similar resonance peaks when $\sigma h/\beta_2 = 2.7$, 4.1 and 6.3, respectively, i.e. when the depth is 0.86, 1.31 and 2.0 times the length of a compression wave in water. A measure $\bar{W}$ of the total disturbance can be obtained by summing the energies from each wave. Thus

$$\bar{W} = \frac{\sigma^{1/2}}{\rho_2\ \beta_2^{5/2}(2\pi r)^{1/2}} \left(\sum_m C_m^2\right)^{1/2} \tag{31}$$

11. Practical examples—We have seen that a necessary condition for the occurrence of the type of pressure fluctuations studied in this paper is that the motion of the sea surface should contain at least some wave groups of the same wavelength traveling in opposite directions. We shall briefly consider some situations in which this may occur.

(a) **A circular depression.** The "eye" or center of a circular depression is a region of comparatively low winds; yet there are often observed to be high and chaotic seas in this region (which indicates the interference of more than one group of swell). Thus, the waves in the "eye" must have originated in other parts of the storm. Now the winds in a circular depression are mainly along the isobars, but in some parts of the storm they usually possess a radial component inwards. In addition, some wave energy may well be propagated inwards at an angle to the wind. This then may account for the high waves at the center of the storm.

If wave energy is being received equally from all directions, the energy in the spectrum will be in an annular region between two circles of radii $2\ \pi/\lambda_1$, and $2\ \pi/\lambda_2$, where λ_1 and λ_2 are the least and greatest wavelengths in the spectrum (see figure 8). This region may be divided into two regions Ω_1 and Ω_2 by any diameter through the origin. Let us take numerical values appropriate to a depression in the Atlantic Ocean. Suppose that the wave-periods lie between 10 and 16 seconds, so that $\lambda_1 = 1.54 \times 10^4$ cm., $\lambda_2 = 4.00 \times 10^4$ cm. and hence $\Omega_1 = \Omega_2 = \Omega_{12} = 2.15 \times 10^{-7}$ cm.$^{-2}$. Assuming $\Lambda = 1000$ km^2 (corresponding to a circular storm area of diameter 17 km.), $\sigma_{12} = 2\ \pi/13$ sec.$^{-1}$, $a_1 = a_2 = 3$m., $h = 3$ km. and $r = 2{,}000$ km. we find from (29) that $|\delta| = 3.2 \times 10^{-4}$ cm., or 3.2μ. The peak-to-trough amplitude of the displacement is 6.5μ. This is of the same order of magnitude as the observed ground movement.

(b) **A moving cyclone.** Consider a cyclone which is in motion with a speed comparable to that of the waves. Figure 9 represents the position of the cyclone at two different times. When the center of the storm is at A, say, winds on one side of the storm (marked with an arrow) will generate waves travelling in the direction of motion of the storm; these will be propagated with the appropriate group velocity. When the storm has reached B, winds on the opposite side will generate waves travelling in the opposite direction; and if the storm is moving faster than the group-velocity of the waves, there will be a region C where the two groups of waves will meet. Thus, in the trail of a fast-moving cyclone we may expect a considerable region of wave interference.

(c) **Reflection from a coast.** The extent of wave reflection from a coast is hard to judge, since the reflected waves are usually hidden by the incoming waves; but when the waves strike a coast or headland obliquely the reflected waves can sometimes be clearly seen. Effective wave interference will take place only on the parts of the coast where the shoreline is parallel to the crests of some wave components of the incoming waves, but refraction of the waves by the shoaling water will tend to bring the crests parallel to the shore.

If the incoming waves are represented by a region Ω_1 in the spectrum, then we may assume that the reflected waves are represented by a region Ω_2 which is the reflection of Ω_1 in the line through O parallel to the shoreline (see figure 10, in which the x-axis is taken parallel to the shoreline). $\Omega_{2'}$ is then the reflection of Ω_1 in the line through O perpendicular to the shoreline (the y-axis).

Suppose that the period of the incoming swell lies between 12 and 16 seconds, that its direction is spread over an angle of 30°, and that its mean direction makes an angle of 10° with the perpendicular to the shoreline. Then we find $\Omega_1 = \Omega_2 = 1.4 \times 10^{-8}$ cm.$^{-2}$, $\Omega_{12} = 1/3\ \Omega_1 = 0.47 \times 10^{-8}$ cm.$^{-2}$. If the effective shoreline is 600 km. in length and the region of interference extends, on the average, 10 km. from the shore, then $\Lambda = 6{,}000$ km^2. If also $a_1 = 2$m, $a_2 = 0.1$m (a reflection coefficient of 5%) and if r = 2,000 km., then we find from (26) (assuming h = 0) that $2|\delta| = 0.3\mu$. Since this amplitude is somewhat smaller than in case (a), we may conclude that coastal reflection does not give rise to the largest disturbances at inland stations, though it may be a more common cause of microseisms near to the coast.

Besides the examples given above there is another possible class of cases, namely when a swell meets an opposing wind. For example, coastal swell may be subject to an offshore wind, or there may be a sudden reversal of the direction of the wind at the passage of a cold front.* The wind will doubtless tend to diminish the amplitude of the original swell, but it may also tend to generate waves travelling in the opposite direction, the amplitude of which may increase rapidly on account of the roughness of the sea surface. However, in none of the first three cases discussed above is it necessary to assume that such action takes place.

12. Observational tests—The present theory suggests several possible kinds of experimental investigation. The first is a comparison of the periods of microseisms and of the sea waves possibly associated with them, (which should be about twice the microseism periods). There is a general agreement between the periods, in that the range of microseism periods is from about 3 to 10 seconds while the periods of high sea waves vary from about 6 to 20 seconds. Further, the periods of both microseisms and sea waves both increase, in general, during a time of increased disturbance. The close two-to-one ratio between the periods of sea waves and of the corresponding microseisms which was found by Bernard (1937 and 1941) and related by Deacon (1947) and Darbyshire (1948) is highly suggestive, though not conclusive. A similar, though less detailed study by Kishinouye (1951) during the passage of a tropical cyclone, has not confirmed the relationship. Comparisons of this kind are, however, inconclusive, unless it can be shown that the microseisms can be associated uniquely with the recorded sea waves. The meteorological conditions are rarely so simple, and the recording stations so well placed, that it is possible to be certain of the connection; the examples selected by Darbyshire (1950) were, however, chosen with this requirement in mind.

Figure 7 shows that the displacement of the "sea bed" may vary by a factor of the order of 5, depending on the depth of the "ocean." Although the model chosen is extremely simplified, we can nevertheless infer that the amplitude of microseisms should, on the present theory, depend considerably on the depth of water in the path of the microseisms; the depth in the generating area itself, where the energy-density is greatest, should be of the most critical importance. Comparisons between the microseisms due to storms in different localities would therefore be of considerable interest. It should be noticed that the unequal response of the ocean to different frequencies may result in a displacement of the spectrum towards those frequencies for which the response is a maximum.

The nature of the frequency spectrum of sea waves under various conditions is of fundamental importance, and further studies should be undertaken. The wavelengths and directions of the components of the spectrum, both for swell and for waves in the generating area, could be studied by means of aerial photographs or altimeter records taken from an airplane. An estimate of the amount of wave reflection from a coast might be obtained by techniques similar to those which were used in the model experiments described in Section 7, that is, by comparing the frequency spectra of pressure records taken at different depths in the water, or off different parts of the same coast where the bottom gradient varied. The effect of an opposing wind on a swell might be investigated on a model scale, by generating progressive waves in the usual manner and then exposing them to an artificial wind; the growth of the opposing waves would be measured by means of the second-order pressure fluctuations deep in the water.

It would be of great interest to record the pressure fluctuations on the ocean floor directly, if the practical difficulties of making measurements at such depths can be overcome. A pressure recorder has been designed for this purpose by F. E. Pierce, of the National Institute of Oceanography.

* See also the author's comment on the paper by Frank Press.

13. Historical notes—It was known to FARADAY (1831), who refers to earlier work by Oersted, Wheatstone and Weber, that fluid resting on a vibrating elastic plate will form itself into short-crested standing waves. Faraday was the first to show, by an ingenious optical method, that the period of the standing waves is twice that of the vibrations of the plate. The waves that he used were mostly "ripples," controlled predominantly by surface tension, since their wavelength lay between ¼ and ⅜ inch. In the same paper (1831) Faraday describes many other interesting experimental studies of waves in water, mercury and air.

About fifty years later Rayleigh (1883 b) repeated Faraday's experiments and verified, by a slightly different method, the doubling of the period. In a theoretical paper (1883 a) Rayleigh gives general consideration to the problem of how a system can be maintained in vibration with a period which is a multiple of the period of the driving force. He refers in particular to Melde's experiment, in which a stretched string is made to vibrate by the longitudinal oscillation of a tuning fork attached to one end; such a phenomenon is sometimes called "subharmonic resonance."

Neither Faraday (1831) nor Rayleigh (1883) evaluated the second-order pressure fluctuations associated with standing waves. This, however, was done by MICHE (1944) in a different connection, using a Lagrangian system of coordinates. Miche noticed the unattenuated terms, and, though he does not mention microseisms, he remarks, "on peut aussi se demander si ces pulsations de pression, malgré leur faible intensité relative, n'exercent pas une action non négligéable sur la tenue des fonds soumis au clapotis." (1944, p. 74.)

The wave interference theory seems to have arisen as follows. In 1946 Deacon, following similar studies by Bernard (1937, 1941 a) compared the period and amplitude of swell off the coast of Cornwall, England, with the corresponding microseisms at Kew, and found a two-to-one ratio between the periods (Deacon 1947). F. Biesel, then visiting England, pointed out to Deacon Miche's theoretical work on standing waves. Miche's results, however, cannot be applied directly to sea waves, since exact standing waves do not occur in the ocean. Moreover, his method is not easily generalized, since it involves a complete evaluation of the second approximation to the wave motion. A very simple proof of Miche's result, however, which depended essentially on the idea of the vertical motion of the center of gravity of the whole wave train, was found by Longuet-Higgins and Ursell (1948); the advantage of this method was that the second-order pressure fluctuations on the bottom could then be obtained immediately from the first approximation to the surface elevation. It then became possible to extend the results to much more general and realistic types of wave motion. A complete theory, giving the necessary conditions for the occurrence of this type of pressure fluctuation, taking into account the compressibility of the ocean, and determining the order of magnitude of the ground movement, was given by Longuet-Higgins (1950).

It is interesting that Bernard (1941 a, b) had suggested, with intuitive reasoning, that microseisms might be caused by the standing-type waves observed to occur at the center of cyclonic depressions:

"J'ai cru qu'on pourrait trouver la raison de cette particularité dans le charactère que présentent les mouvements de la mer au centre des dépressions cycloniques: la houle s'y dresse aux vagues pyramidales constituant un clapotis gigantesque dont les points de plus ample oscillation peuvent être autant des sources de pression périodique sur le fond de la mer, pression qui donnera naissance à un mouvement oscillatoire de même période du sol . . . "

"Un clapotis analogue, avec oscillations sur place du niveau de l'eau, se produit lorsque la houle, se reflèchissant sur un obstacle, vient interférer avec les ondes incidentes . . .

"Au contraire, dans le cas d'un train d'ondes de front continu et de déplacement constant, les points ou les mouvements sont de phase opposée donneront sur le fond de la mer des pressions de sens contraire, et la longeur d'onde des oscillations microséismiques étant beaucoup plus grande que celle de la houle, les mouvements transmis par le sol à une certaine distance seront pratiquement simultanés, mais opposés, et ils interféreront, de sorte que l'effet total du train de vagues a l'exterieur sera nul." (BERNARD, 1941 a, p. 7.)

However, Bernard did not apparently see that the corresponding pressure fluctuations must have a frequency twice that of the waves; for he suggests other causes for the observed doubling of the frequencies in the case of coastal waves." (Bernard, 1941a, p. 10.)

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Discussion

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The wave-interference theory explains, for the first time, how energy sufficient to generate long, regular, microseisms is communicated to the ground. It has been clear for a long time that the occurrence of microseisms is associated with the presence of sea waves, but it could not be proved that the waves played an essential part in the energy transfer.

Although each breaker, as it crashes on the coast, must cause a local disturbance, and has been shown to do so, the variations in the moment of impact along a stretch of coast, and the shortness of the wavelength compared with that of 3 to 10 second microseisms, make it most unlikely that the actual beating of surf on a coast could produce the long microseismic waves that can be detected far from the coast.

The exponential decrease in wave movement with depth was sufficient reason why a train of progressive waves should not disturb the sea bottom at great depths, and at lesser depths the contributions from different parts of the sea bed would tend to cancel each other out. Taking account of the compressibility of the water made no significant difference to this conclusion.

If the conviction held by many who had studied microseisms, that sea waves are directly concerned in the generation of microseisms were to be confirmed, we had to find a theory which showed that sea waves were modified in such a way that they were able to cause regular changes in pressure, acting simultaneously over large areas of the sea bed. During the past few years it has, in addition, become necessary to explain why the periods of the microseismic waves are half those of the sea waves, and how the effect of wind and wave-height could vary with the depth of water, being sometimes greater in deep water than in shallow.

The new wave-interference theory seems to fill these requirements, and to be capable of withstanding the test of more precise and well-directed observations.

It is not easy for the non-mathematician to understand the precise demonstration that two trains of waves of the same wavelengths, meeting each other in opposite directions, will cause variations in pressure on the sea bed with twice the frequency of the surface waves, but Dr. Longuet-Higgins has done his best to explain it in non-technical terms. The deduction is simplified by considering the vertical movements of the centre of gravity of a water mass bounded by two vertical nodal planes, and by a comparison with the changing tension in the string of a pendulum. It is perhaps not very difficult to accept the result intuitively, as Bernard (1941) did, particularly if we remember the convincing agreement between theory and observation obtained by measurements in a tank.

There is also confirmation of the mean pressure changes and their ability to produce microseisms that can be detected far from the coast, in the work of Darbyshire (1950). As Dr. Longuet-Higgins says in his paper, confirmation of the two to one relationship between wave and microseism periods does not completely verify the theory, but when, as Darbyshire showed, the trend of a band of swell from long to short periods was exactly paralleled by proportionate changes in the microseism periods there is little room to doubt that the waves caused the microseisms.

If the previous literature is re-examined, bearing the wave-interference theory, and what we already know about waves, in mind, some of the apparent contradictions to which emphasis has been given appear explainable. The example given by Whipple and Lee (1935) of almost identical isobaric charts of two depressions south-east of Greenland, one associated with intense microseismic activity and the other with practically none, is not such an obstacle when the previous histories of the two depressions are studied. One had moved rapidly northwards over the ocean, with plenty of opportunity for wave interference, whereas the other had developed over the land. Similar attempts to estimate wave interference might explain why less microseismic activity was found with a depression over the mouth of the St. Lawrence river and an anticyclone over the Great Lakes than when the positions of the depression and anticyclone were reversed; or why, with a shallow depression off the east coast of Japan, the microseisms were larger on the coast of China while the wind was stronger off the coast of Japan.

There is, however, not much to be gained by studying cases which are not fully documented. We must, as Dr. Longuet-Higgins emphasizes, learn more about the conditions which give rise to wave interference; we must select examples in which the metorological conditions are sufficiently simple for us to be certain of the connection between the storm and the microseisms, and we must measure the waves and the microseisms as precisely as modern techniques will allow. It is possible that some of the present misunderstanding is due to faulty interpretation of records from seismometers that are highly tuned to the short-period end of the microseism range, and faulty estimation of the sea surface or wave and microseism recordings, in which the size of a long period oscillation can be underestimated owing to the interruption of its swing by minor, shorter, waves.

The wave-interference theory is, to say the least, an excellent working hypothesis, and if it is subjected to further question and experiment, of the standard set by Dr. Longuet-Higgins and his co-workers, we must move rapidly towards a full solution.

It seems to me that the subject has now been put on a systematic basis, and that its progress must be more rapid. In spite of some setbacks we shall soon be in a better position to take full advantage of the practical possibilities.

I think that Dr. Longuet-Higgins's historical note gives a proper account of the development of the new theory.

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Discussion

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As a discussion of the theoretical paper "Can Sea Waves Cause Microseisms," I should like to present some of the data and interpretations obtained by the Naval Research Laboratory on various field trips during the hurricane seasons of the past several years.

The data considered here is concerned with hurricanes which have followed paths in the Western Atlantic and Caribbean. It has been a primary objective of this work to obtain evidence which might help to determine where the area of microseism generation is with respect to the hurricane center and to determine under what condition a hurricane can generate microseisms. In furthering this objective it has become of interest to study the data in the light of various theories to see if the data lends support to any of these theories.

During the hurricane seasons of 1948-1951 records of microseisms have been obtained at points in the Bahamas, Florida, North Carolina and Washington D. C. as various hurricanes have followed varying paths in the Western Atlantic. The following observations have in general been true for all these hurricanes:

(1) Storms which generate in the Middle Atlantic and approach the seismograph locations do not produce appreciable microseismic activity until the storm moves over the continental shelf or, over the shallower waters surrounding the Islands of the Caribbean Sea. This same observation is pointed out by Donn (1952).

(2) As the storm recedes, the microseisms continue at a much higher level of amplitude as compared to the same distance from the seismograph location during the approach of the storm.

(3) The point of nearest approach is not necessarily the time of maximum amplitude.

The above observations can be interpreted as giving evidence that the storm must move over the shallower waters of the continental shelf before microseisms are recorded and that the wake of the storm continues to be important in the generation of microseisms. This and similar observations in the light of the Longuet - Higgins (1950) theory, together with the word of Deacon (1947) and Darbyshire (1950), prompted the Naval Research Laboratory group to conduct field experiments during the 1951 hurricane season designed to obtain data which could assist in determining whether any correlation appears to exist between microseisms and hurricane-generated ocean waves.

The installations of the field experiments included the following:

(1) A tripartite station on the West End of Grand Bahama.

(2) The installation of two wave gages at Cocoa Beach, Florida, through the cooperation of the Beach Erosion Board and the University of California. These gages were of the pressure-sensitive type; the one was similar to the type developed by Woods Hole, and used quite extensively by the Beach Erosion Board, and the other was developed by the University of California. These gages were in water depths of about 29 and 46 feet respectively.

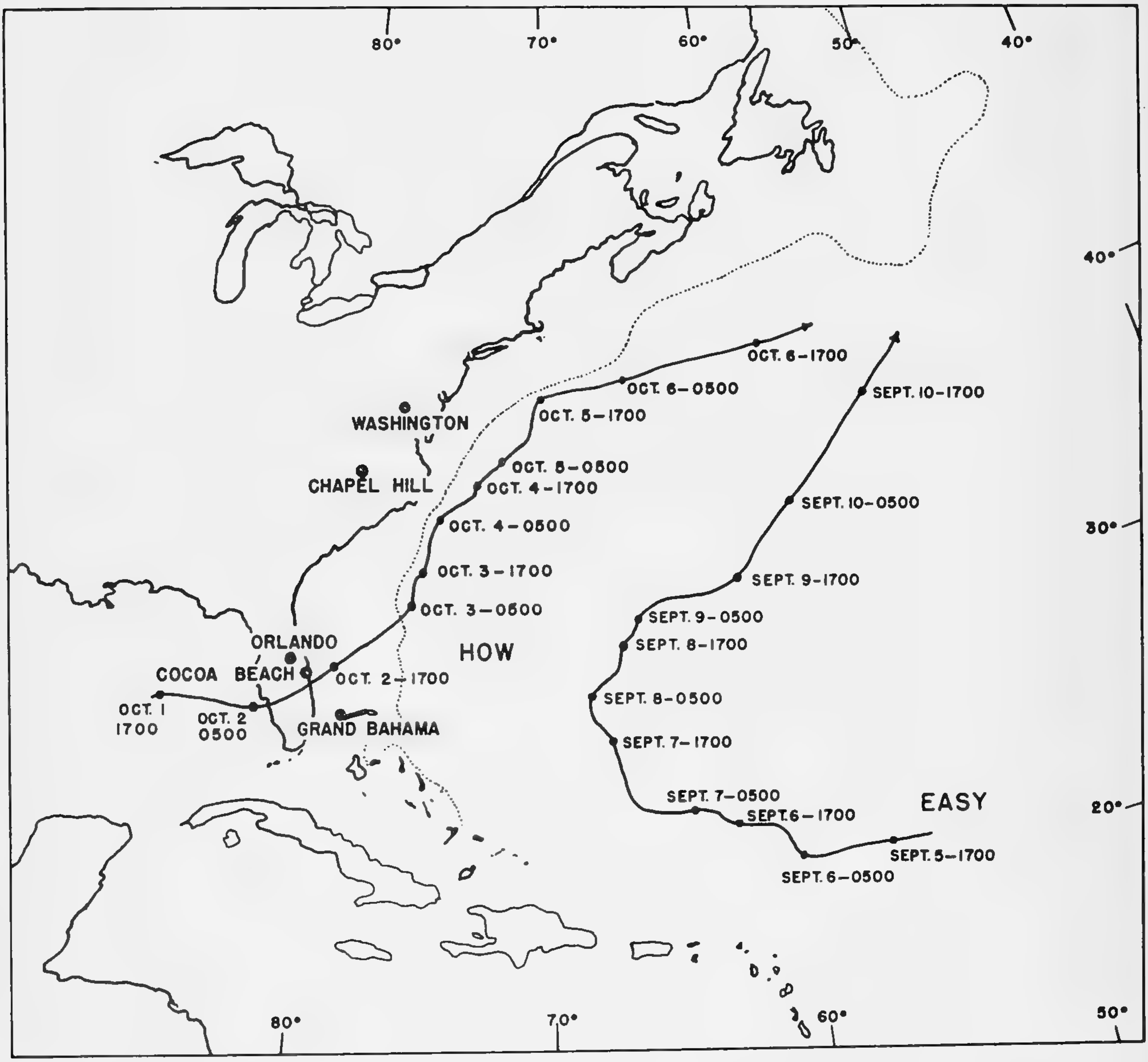

Figure 1. Map Showing Paths of Hurricanes "Easy" and "How"

(3) A single horizontal-component seismograph was placed on the grounds of the U. S. Navy Underwater Sound Reference Laboratory at Orlando, Florida. This location is approximately 50 miles inland from Cocoa Beach, and therefore can be considered isolated from local surf vibrations, which can cause high seismic noise near the shore.

The simultaneous data of microseisms and water waves obtained by these installations during the two hurricanes of the 1951 season is of special interest in that the paths of the storms were radically different. Figure 1 shows the paths of the two storms "Easy" and "How." "Easy" followed a path which was well out over deep water during its entire course (except near its end when it moved over the Banks of Newfoundland). Its nearest approach to Florida was about 650 miles. Hurricane "How" generated in the Gulf of Mexico, rapidly moved across Florida, and entered the Atlantic with the center passing slightly to the south of the wave-recorded location. Both of these storms produced high waves on Florida but the character of the waves was considerably different and the microseismic activity was greatly different. The two storms therefore provide an interesting comparison.

Figure 2 gives results of the simultaneous recordings of microseisms and water waves throughout the period hurricane "Easy" was in existence. The wave-gage data was analyzed by the Beach Erosion Board to give the significant wave height and period plotted as curves C and D respectively. A measure of the amplitude of the microseisms was obtained by measuring the area enclosed by the envelope of the microseisms during a 15 minute interval, an interval being used every two hours and in

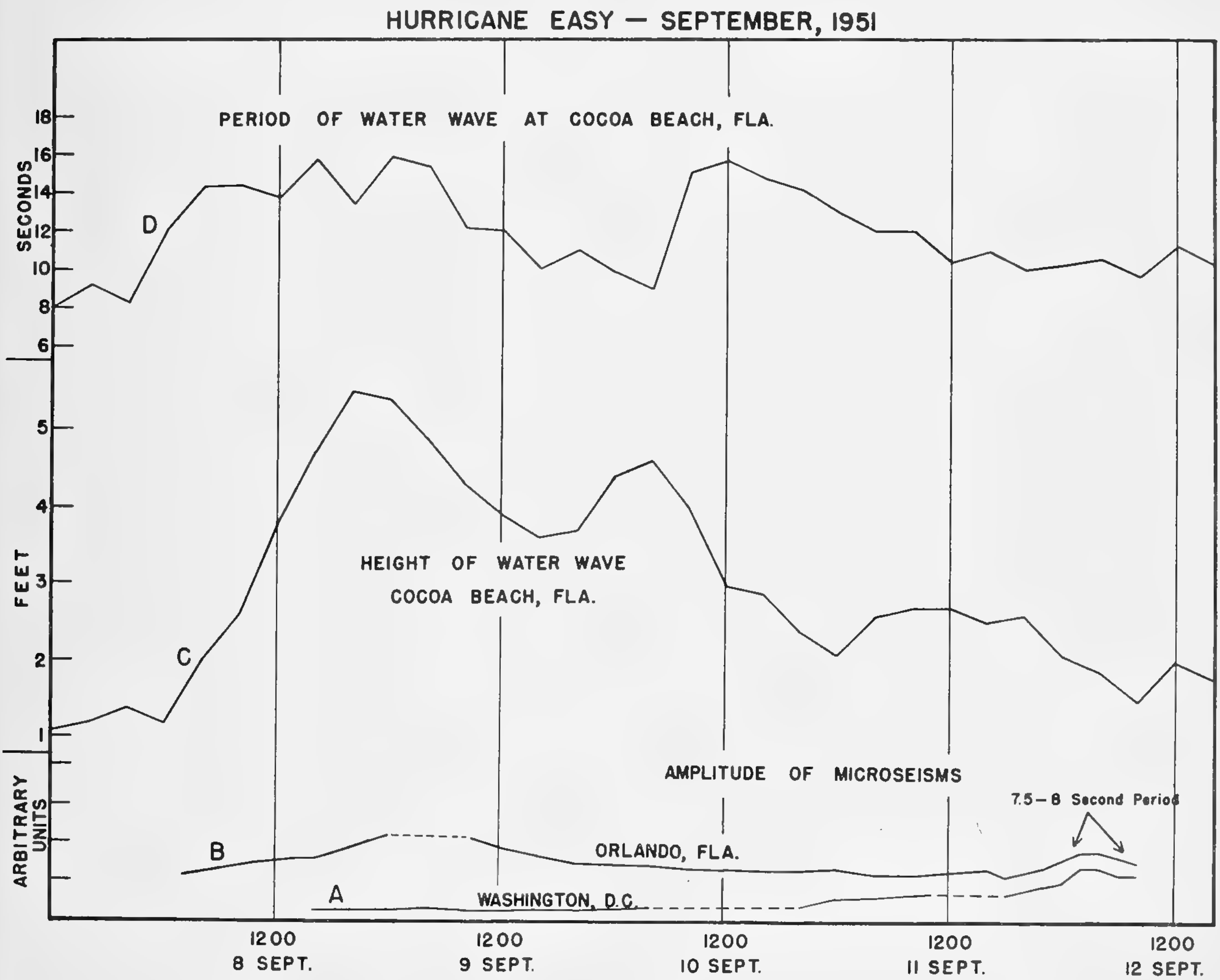

Figure 2. Microseismic and Water Wave Activity During Hurricane "Easy".

some parts of the record every hour. The relative position of curves A and B has no significance since the two curves have been shifted with respect to each other. However, the value of the arbitrary units for A and B is the same.

The sharp increase in both wave height and period as shown in curves C and D on the morning of September 8 accompanied the arrival of the swell from "Easy." Data from a Beach Erosion Board gage at Cape Henry and a report from Weather Ship H, several hundred miles east of Charleston, N. C., also gives added evidence that the wave activity shown by curves C and D on Sept. 8 is associated with the arrival of swell from "Easy." The microseisms as recorded at Orlando on 8 Sept. show some increase in amplitude at approximately the same time as the maximum wave activity at Cocoa Beach. This increase in amplitude was not at all pronounced; in fact this particular period of microseisms normally would not have received any attention as being an indication of anything unusual. The record was too erratic to permit an analysis of the most pro-

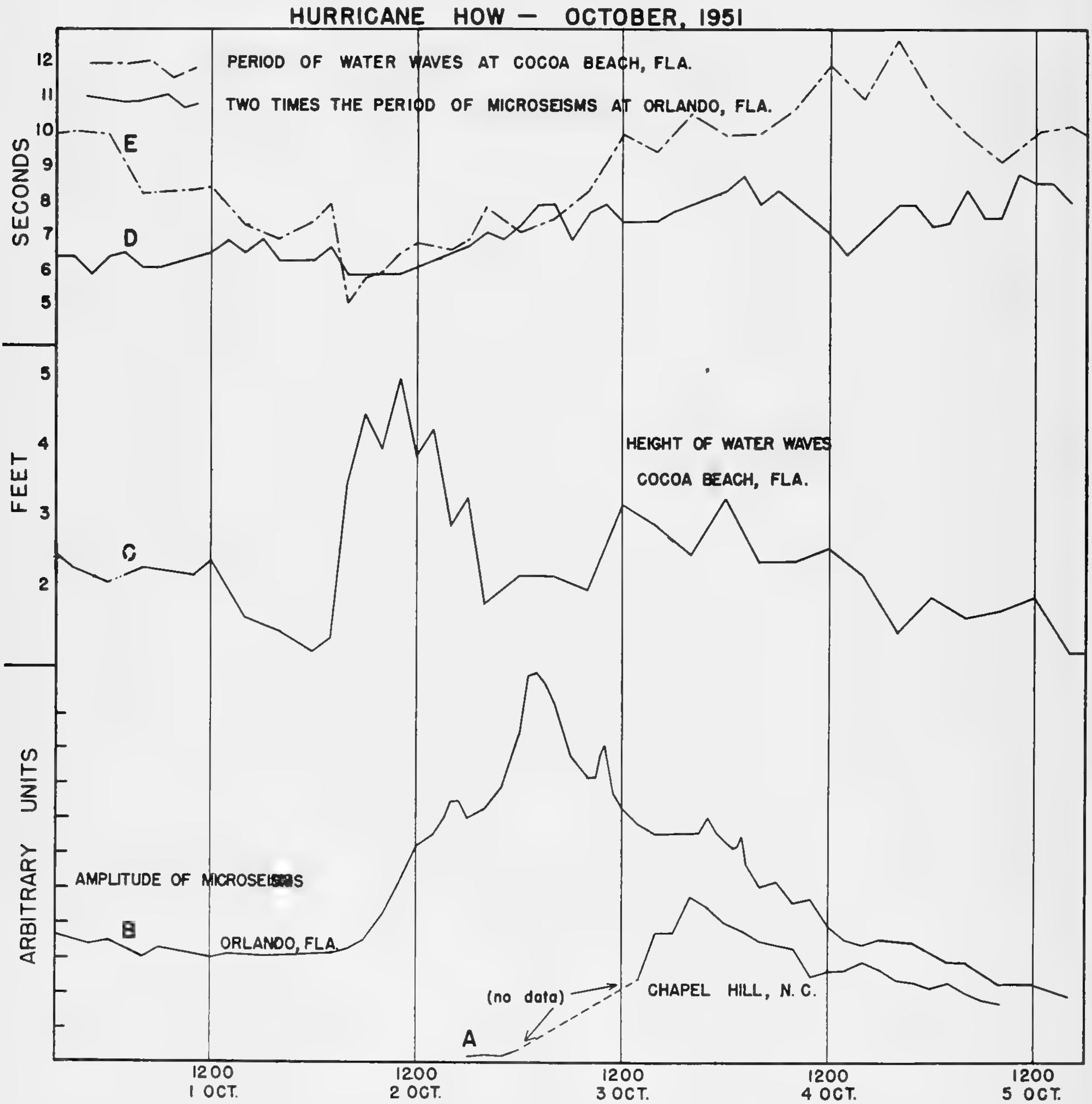

Figure 3. Microseismic and Water Wave Activity During Hurricane "How".

nounced period. The slight increase in microseisms during the wave activity can be interpreted as being associated with the swell rather than being generated directly under the storm for these reasons:

(1) No simultaneous increase in microseisms occurred in Washington.

(2) Microseisms generated under the storm should also have shown increased activity before the arrival of swell.

According to the Longuet-Higgins theory, a standing-wave pattern is required to transfer the water wave energy to microseisms. A standing wave pattern can conceivably be established upon reflection of the incoming swell by a sufficiently steep coast. The low level of microseismic activity during the swell from "Easy" would indicate, if the Longuet-Higgins theory is of importance, that the reflected wave energy along the Florida coast is very small. Because of the very gradual slope of the shore along Florida one would indeed expect low reflections.

The fact that no microseisms of any consequence were recorded during the period this intense storm remained over deep water indicates either one of two things: (1) microseisms were not generated by any method or (2) the generated microseisms were almost completely attenuated before reaching the continent. The data obtained by NRL is unable to resolve which of these two factors is the important one. Carder (1951) has presented evidence to indicate that the attenuation of microseisms propagated through the floor of the Western Atlantic is much greater than the attenuation over continental land masses. If attenuation is the important factor, then the attenuation may vary with the nature of the ocean floor and thus the results could be different in various parts of the world. Darbyshire (1950), Banerji (1935) and others have presented evidence that microseisms are generated in deep water and have been recorded at distant points in the case of storms over the Eastern Atlantic, the mid-Bay of Bengal, and the Pacific. In view of these observations which contrasts with the observations in the Western Atlantic it may be inferred that attenuation is a much greater factor in the Western Atlantic than in certain other parts of the world.

It is of interest to point out the fact that longer period (7.5 to 8.0 second) mocroseisms are evident on curves A and B, Figure 4, as occurring at Washington and Orlando on the morning of 12 September. The records of these microseisms were nicely formed and of a regular nature. The simultaneity in time and period of these microseisms at Washington and Orlando would indicate a common area of generation. The fact that the storm at this particular time was dissipating itself over the shallow areas off the coast of Newfoundland is further evidence that a storm moving from deep water to shallow water begins to generate microseisms. Intense winter microseisms are frequently observed when low-pressure areas move over this portion of the North Atlantic.

Let us consider Figure 3 which shows simultaneous data on wave and microseismic activity obtained during hurricane "How." The wave gage was fortuitously placed in a strategic location slightly to the north of the area where the storm entered the Atlantic. We may therefore assume that, if waves are responsible for the generation of microseisms, the waves as measured at this time should yield the best possible correlation inasmuch as the waves were confined to the water areas near the gages. Let us therefore compare the water wave amplitude and the position of the storm. We note an abrupt increase in wave amplitude during the early morning of 2 October, reaching a maximum about 1200 and dropping off abruptly about 2000. Referring again to Figure 1 we see that the forward part of the storm entered the Atlantic in the morning of 2 October with strong winds blowing from south-southeast and bringing waves toward Cocoa Beach. At about 1200 the center of the storm moved into the Atlantic and by 2000 the winds in the trailing part of the hurricane were from the north, thus effecting a reversal of wind as it existed 20 hours previously over this area. This reversal of wind is evident on the wave records by a rather abrupt decrease in wave amplitude. On Figure 3 we see from curve C that the maximum microseisms occurred just after the wind reversal. From curves D and E we observe that during the period when the water wave activity was confined to an area near the wave recorder the period of the water waves was closely two times the period of the microseisms. It should also be pointed out that the magnitude of the arbitrary units used as a measure of microseismic amplitude on the Orlando records during "Easy" and "How" are the same. It is apparent that, although the height of water waves recorded during the two storms is about the same, the amplitude of the microseisms during "How" was five or six times as large as the amplitude during "Easy" and in the case of "How" the amplitude was very outstanding above the normal background.

From the above facts one may make the following interpretations:

(1) The correlation between one half the period of the waterwave and the period of the microseisms during "How" lends support to the Longuet-Higgins theory.

(2) The reversal of wind and the setting up of waves in a direction more or

less in opposition to the waves generated a few hours previously may be a very effective method of producing the necessary standing wave system.

One may also refer here to the association of microseisms with cold fronts to support the thought that a relatively sudden reversal of wind over shallow water areas provides a condition for microseism generation. Typical weather conditions off the eastern North American coast, prior to the arrival of a cold front, include moderately strong southernly winds. These winds would develop waves travelling in a northerly direction of relatively small amplitude and short period. Following the passage of the cold front the wind direction normally changes abruptly to the northwest. It is reasonable that at some time, shortly after the passage of the front, waves developed by the northwest winds will have periods and wavelengths nearly equal to that of the dying swell from the south. Thus, a standing wave component could exist which would have the potential for excitation of microseisms in accordance with the Longuet-Higgins theory.

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Discussion

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The existence of an unattenuated pressure variation in the ocean was already suspected by Whipple and Lee (1935) and some years later Bernard (1941) also suggested that a standing wave-system produced in some way microscisms, but the well-known exponential decrease of gravity waves precluded any understanding of the process. However, in 1942 Miche proved that in the case of standing gravity waves in an incompressible ocean a second order pressure variation exists which is not essentially influenced by the depth. Moreover, as the frequency of this variation is twice that of the ocean waves and as Bernard had observed that the period of microscisms is roughly half that of sea waves, Longuet-Higgins and Ursell (1948) supposed that this second order effect is the primary cause of some microscisms.

The formula obtained by Miche can be derived by a small extension of the theory of gravity waves. Consider the irrotational motion in an incompressible ocean of infinite depth; for simplicity's sake we suppose the movement to be two-dimensional.

The horizontal (u) and vertical (w) components of the velocity are determined by a velocity-potential:

$$u = -\partial\phi/\partial x \text{ and } w = -\partial i/\partial z.$$

From the equations of motion

$$Du/Dt = -\frac{\partial p}{\partial x} \text{ and } \frac{Dw}{Dt} = -\frac{\partial p}{\partial z} + g\rho$$

where D/Dt = a differentiation following the motion of the fluid, and p = the pressure, we obtain

$$\frac{\partial\phi}{\partial t} - \frac{1}{2}q^2 + gz = \frac{p - p_o}{\rho}$$

with $q^2 = u^2 + w^2$ and p_o = the constant pressure at the free surface.

Placing the origin in the undisturbed surface the equation of this surface is

$$z = \zeta, \quad g\zeta = \left(-\frac{\partial\phi}{\partial t} + \tfrac{1}{2}q^2\right)_{z=\zeta}$$

The potential ϕ has to satisfy the equation of continuity $\Delta\phi = 0$ and the boundary condition

$$D/Dt\left(\frac{\partial\phi}{\partial t} - \frac{1}{2}q^2 + gz\right) = 0 \text{ for } z = \zeta, \text{ or}$$

$$\frac{\partial^2\phi}{\partial t^2} = g\frac{\partial\phi}{\partial z} + \frac{\partial q^2}{\partial t} + \tfrac{1}{2}q\,\Delta q^2 \text{ for } z = \zeta \quad \text{...............(1)}$$

A wave system consisting of two plane progressive waves travelling in opposite directions:

$$\phi_1 = \frac{g}{\nu}\left\{a_1 \sin(kx - \nu t) - a_2 \sin(kx + \nu t)\right\} e^{-kz} \text{ with } a_1 \approx a_2$$

fulfils $\Delta\ \phi = 0$ and satisfies the boundary condition equation 1, to a first approximation if $ka \quad 1$ and $\nu^2 = gk$. For a second approximation we put $\phi = \phi_1 + \nu a^2 f$ where $d \approx d_1$; neglecting terms of third and higher order in ka we obtain $a^2\ f = a_1\ a_2\ \sin 2\nu t$.

The corresponding surface elevation $\zeta = \zeta_1 + \zeta_2$, with

$$\zeta = a_1 \cos(kx - \nu t) + a_2 \cos(kx + \nu t)$$

$$\zeta_2 = -\frac{1}{2}k \left\{ a_1^2 \cos 2(kx - \nu t) + a_2^2 \cos 2(kx + \nu t) + 2a_1 a_2 \cos 2kx \right\}$$

and the pressure $= p_o + g\rho z - g\rho\zeta_1 e^{-kz} - \frac{1}{2}\rho\nu^2 (a_1^2 + a_2^2 - 2a_1 a_2 \cos 2t)e^{-2kz}$

$$- 2\rho a_1 a_2 \nu^2 \cos 2\nu t.$$

Obviously at large depths ($kz \gg 1$) the varying part of the pressure is

$$p = 2\rho a_1 a_2 \nu^2 \cos 2\nu t \ldots\ldots\ldots\ldots (2)$$

which is the result obtained by Miche for a standing wave system ($a_1 = a_2$).

Considering a rather general irrotational movement LONGUET-HIGGINS (1950) was able to generalize equation (2) and to calculate the amplitude of microscisms caused by an arbitrary wave-like motion of the ocean. His final formula (his equation 198) may be interpreted in the following (inexact) way.

The Miche force of the square λ^2, where λ = the mean wavelength of the interfering progressive waves, is according to (2) equal to

$$2\rho a_1 a_2 \nu^2 \lambda^2$$

If the microscismic amplitude caused by a concentrated unit force with frequency 2ν at a distance r is denoted by $w(2\nu, z)$ the total amplitude will be

$$2\rho a_1 a_2 \nu^2 \lambda^2 W(2\nu, z)$$

Supposing the phases of ocean waves at points separated by a distance of a wavelength to be uncorrelated the amplitude generated by a storm with an area A will be of the order

$$\left(\frac{A}{\lambda^2}\right)^{\frac{1}{2}} 2\rho a_1 a_2 \nu^2 \lambda^2 W(2\nu, z)$$

With $A = 10^3 \text{km}^2$ and $\lambda = 0.25\text{km}$ ($\nu = \frac{1}{2}$) the vertical amplitude at a distance of 3000 km. appears to be 9.4μ, which is of the order of the observed amplitudes. The detailed investigation of Longuet-Higgins shows that this has to be multiplied by a factor which depends on the frequency spectrum of the wave system. For instance, if the energy of the movement is uniformly divided in every direction within a range of wave lengths between λ_1, and λ_2 this factor is

$$\left\{\frac{\pi}{2}\left(\frac{\lambda_1}{\lambda_2} - \frac{\lambda_2}{\lambda_1}\right)\right\}^{\frac{1}{2}}$$

the numerical value of this quantity is about 0.54 if $\lambda_1 = 400$ meters and $\lambda_2 = 154$ meters. The vertical amplitude is then 5μ, and the horizontal 3μ.

This theory undoubtedly explains the phenomenon of microscisms in a straightforward way. The only difficulty which it encounters is the fact that microscisms occur very often, while it is a matter of considerable doubt whether standing waves of rather large amplitudes are as common.

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Discussion from the Floor

Haskell. (Questioning Longuet-Higgins.) Ocean waves are coherent over more than just one wave length, so shouldn't the area of generation be subdivided into areas that are larger than one wave length on a side—perhaps the wave lengths? (*Longuet-Higgins* answered, perhaps so.)

Longuet-Higgins. (In answer to Press's question, "what if the wave periods on the surface occur off the peak of your resonance curve?") The sea waves must be considered as possessing not a single period, say 12 seconds, but a frequency spectrum of a certain width, say 8-16 seconds (the pressure fluctuations would then be from 4 to 8 seconds period.) The spectrum of the microseisms should be a combination of the spectrum of the pressure variations and that of a response curve. If the most prominent period of the pressure variations occurs off the peak of the resonance curve, the most prominent period of the microseisms would be expected to be displaced towards the peak.

STORM AND SURF MICROSEISMS

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The practical meteorologist is interested in the so-called microseismic phenomenon because it may provide a potential for giving early warning of the existence of large destructive storms and because it may permit a type of direction-finding which will track such storms once they have formed.

If microseismic techniques can realize one or both of these potentialities, the meteorologist will have a means of coping with the apparent unpredictable nature of typhoons and hurricanes and will thereby provide the storm information necessary to reduce hazard and damage to a minimum. Storm information at the present time is obtained only at a great cost and considerable risk. Meteorological art is advancing to the point of permitting the forecasting of typhoon-prone conditions. But not all such conditions develop into typhoons. Conversely, an occasional typhoon develops in an unclassical situation which gives no clue as to an incipient storm development. The track which a storm will take, once developed, is also still shrouded in mystery and the most effective method of tracking still remains the purely visual one employing aircraft or radar to penetrate the eye of the storm, or both.

As scientists, meteorologists are interested in the theory of microseismic generation and propagation. That interest, however, is a "pure" interest to be contrasted with their interest in what microseisms can do for meteorologists. Theirs is essentially the pragmatic approach.

This paper deals entirely with the pragmatic approach and while, necessarily, the theory of microseismic generation must be touched on its consideration is limited to how the various theories affect the potential usefulness of microseisms as a meteorological tool.

The history of microseismic research in this country and abroad has been so thoroughly covered that any repetition is sheer redundancy. Let it suffice to say that during the middle years of World War II, the Navy established a microseismic network to exploit the possibilities of storm detection and tracking in the Caribbean. In so doing, the Navy was accepting the theory that microseisms originated within the storm and that they proceeded instrument-ward not through water but through the crust of the earth at the ocean bottom. The Microseismic Research Program proceeded through the years with just enough success to warrant its continuation but without sufficiently clear-cut results to establish it on a firm operational basis. Under Mr. Gilmore's direction, the network in both the Atlantic and the Pacific has been expanded and results for numerous storms have been tabulated, studied and published.

As a separate endeavor, the Naval Research Laboratory instituted a program to improve the instrumentation used in microseismic research. Under the direction of Drs. Kammer and Dinger a method was devised which much simplified the methods of calibration and interpretation of microseismic records. In testing out their equipment, the Naval Research Laboratory scientists also recorded the progress of storms in the Western Atlantic and unlike Mr. Gilmore, reached the conclusion that in these cases coastal action was responsible for the tremors which affected their microseismic installations.

As far as the operating forces of the Navy are concerned, the question of the value of microseismic research was thrown wide-open. If microseisms originate within or near a storm, the possibility of early warning and tracking remains real. If microseisms are the result of a local coastal effect, an observer on the coast watching the incoming surf might prove a reasonable substitute for a microseismic network. At best, if the latter theory is the correct one, an oceanographic tool was being developed.

Many published papers were studied in an effort to decide between storm and surf microseisms. Before lining up the evidence on each side, it might be well to define in what sense the words "storm" and "surf" are being used here. By storm microseisms, I mean a crustal disturbance which is produced in the vicinity of the storm, transmitted downward through the water to the ocean bottom and then transmitted through solid matter to the land block on which the seismometer rests. The definition of surf microseisms is somewhat less clear-cut. In the

* In view of the limited scope of this paper, no attempt has been made to cite fundamental or supporting investigations.

vicinity of the storm, an oceanographic disturbance is established which radiates through the ocean in the form of ocean swell until it reaches a continental boundary—be it continental shelf or actual coast. There, the disturbance is transmitted to the solid continental block on which is erected the microseismic recorder. It should be noted that this definition of surf microseisms in no way adopts the usual definition of surf—the breaking of the sea against the coast. The mechanism by which the microseisms are generated is undefined in both cases. The distinction is merely whether the initial generation is near the storm or at a considerable distance from the storm.

Let us examine some of the evidence favoring the one theory or the other in much the same way that we in Naval Operations examined it.

For a number of years, Gilmore used the tripartite method for storm tracking. The fact that he was unable to track certain storms he attributed to the presence of geological barriers in the ocean bottom which refracted and reflected the microseismic disturbance to the extent that the direction from which the microseisms apparently came bore no simple relationship to the direction of original propagation. On scientific and theoretical grounds, I am not in a position to contest this theory—as indeed I am not in a position to contradict any of the prevalent theories. Two things seem apparent, however, in reviewing the earlier Gilmore work. First, if the ocean bottom is really as discontinuous as is indicated, successful tracking of typhoons and hurricanes would appear to be improbable. A half-century of storm tracks indicate that those are so nonreproducible that it would be rare for two to follow the exact same path. Newly apparent barriers would be appearing all the time and no large confidence factor could be given to microseismic storm tracking.

The second thought concerning the results must probably be labelled a more-or-less philosophical one. Time and again it has been demonstrated in the history of science that when an accepted theory results in practice in more exceptions to the rule than cases which follow the rule, the theory has been inadequate or incorrect. Certain geological barriers are well established but as more and more unsuspected ocean barriers appear, the possibility that some additional factor not taken into account by Mr. Gilmore becomes more plausible.

It seems possible that the unknown factor might be surf microseisms. On the other hand, of course, the ocean barrier theory may be entirely sound.

The work at the Lamont Geological Observatory would indicate also that microseismic disturbances are produced only in the vicinity of the storm. Three dominant arguments are presented. The first is a number of instances when microseismic level was high and observed swell was low. The second is the reverse of this picture—observed swell was high but microseismic response was low. The third involves cold frontal passages off the east coast of the United States when the seismometers did not record the relatively strong winds preceding the passage of the front off the coast but began to respond only when the front itself progressed over water. Lower wind velocities behind the front did not apparently affect the microseisms adversely.

Kammer and Dinger, on the other hand, indicate that microseisms occur when the storm affects shallow water and that the magnitude of response to such "surf action" is such as to mask any direct storm response.

An individual not directly involved in microseismic research must accept all of the published data as valid. He must make a choice between two courses of action. He must either abandon further consideration of the problem until the experts reach agreement or he must attempt to derive some logical explanation which will provide consistency in apparently divergent findings. Abandoning the field to the experts would undoubtedly be more discreet. Unfortunately, as administrators of the Navy Microseismic Program, that would immediately involve withdrawal of Naval Aerological support. A less wise but more practical solution is to attempt to resolve the differences.

As a starting point, let us adopt the assumption that microseisms can only be produced when two trains of swell intersect and produce stationary or quasi-stationary waves. This phenomenon may occur in the vicinity of the storm or along a coast. Simple surf or shore pounding would, therefore, not produce microseisms. The absence of microseisms in some of the cases when surf measurements showed high waves would thus be explained.

The conventional cold front is preceded by southwesterly winds which over a water area would be persistent enough to produce appreciable swell. Following the cold front, the winds are northwesterly. The interaction of the swells produced by these winds and those produced ahead of the front may well produce microseisms. The cold front case might be explained in this way.

Hurricanes paralleling the eastern coast of the United States could be expected to produce large areas of swell which would first reach the continental shelf and then secondarily produce coastal surf. The time lag observed by Donn might conceivably be the result of the time lag between the primary effect and the secondary effect. Thus microseismic activity would build up before the observed waves at the beach. Donn's observations are mainly those taken along the New England coast when the storms are traveling rapidly north and east. The Dinger-Kammer observations are made further south where most of the storms are traveling slowly first west and then north. Under these conditions, sea and storm are able to approach the coast at an almost simultaneous rate, particularly when the slowing down resulting from the recurvature of the storm is taken into account.

While nothing definitive is established by this analysis, it would appear that the question of what produces the microseismic phenomenon is still completely unresolved and that some hope still remains that the meteorologist may find a tool for studying hurricanes.

It would seem that the first test for determining whether microseisms originate in or near a storm is to seek a case when it is well-established that no surf action is occurring but when microseismic activity is marked. In finding such a case, it is obvious that East Coast hurricanes are unsatisfactory. The numerous land masses of the Caribbean and the long coast of the United States make it difficult to prove the absence of interfering coastal action anywhere along the storm's path. The microseismic installation on Guam seemed more appropriate in establishing the test case. Except for the chain of islands of the Marianas themselves, Guam is at least 800 miles removed from the nearest appreciably land mass—the Philippines to the west and New Guinea to the south. If no surf were observed at Guam, it would be reasonable to assume that microseismic activity was not produced by surf action only.

It is not a simple matter to determine whether the island of Guam was under the influence of surf action at a given time or not. There have been no instrumental records available. Observers, at one time or another, have recorded visual observations. Due to the difficulty of making an accurate visual estimate and also due to location difficulty—observations were made at fixed points—individual records must be viewed with some question and simultaneous measurements do not agree.

Table 1 shows two sets of data recorded by different observers at different places on Guam at the same time. The lack of agreement is obvious. The somewhat more detailed record made by Observer B lends more credence to his observations. Using approved sea swell forecasting techniques, a hindcast of the swell reaching Guam was made. The values obtained are in last column of the table and compare quite favorably with those of Observer B.

Table 2 shows a similar comparison of observed swell and calculated swell reaching Guam in connection with two other storms. While I am not willing to defend the relative merits of either the observed values or the calculated values, it is worth noting that the calculated values tend to be higher than the observed. Thus any deduction made concerning the effects of swell would presumably be biased in the direction of exaggeration.

Table 1

Typhoon Marge -- August 1951

Date Time (GCT)	Observed Swell (A) (ft)	Observed Swell (B) (ft)	Calculated Swell (ft)
111200	7	Moderate sea 3-5	--
121200	6	Rough sea	12
131200	3	Confused 6-8, 3-5	--
141200	1	Moderate sea 12	13
151200	2	Heavy 5-6	9
161200	2	Moderate 8-12	9

Table 2

Date Time (GCT)	Observed Swell (ft)	Calculated Swell (ft)	Micro Amplitude (mm)
Typhoon Ruth			
071200	1	4.5	11
081200	3-4	6	36
091200	3	7	64
101200	3	7	169
111200	2	8.5	144
121200	2-3	4	147
131200	1-2	3	180
141200	2	3	141
Typhoon Nora			
281200	3	3	17
291200	2	3	20
301200	1	3	29

Only those storms were studied which presented a relatively simple meteorological picture. This choice was made by necessity since the procedure for forecasting swell became too involved when more than one storm appeared on the map of the western Pacific. In all, ten storms were found which met the requirement of simplicity. For each of these, the swell was calculated corresponding to the time of microseismic observation, and the development of the storm in a meteorological sense was reconstructed from post-analyzed maps, pilot reports, etc.

Four of the storms developed far enough away from Guam as to produce insignificant swell at Guam. The swell, microseismic response and center wind force of these storms are illustrated in figures 1, 2, 3, and 4 and Tables 3, 4, 5, and 6.

Table 3

Typhoon Ruby -- October 1950

Date Time (GCT)	Distance from Guam (mi)	Center Intensity (kts)	Swell at Guam (ft)	Micro Amplitude (mm)
271200	900	40	8	20
280000	900	55	5	32
281200	890	70	5	42
290000	740	90	4	25
291200	720	100	3	40
300000	720	110	3	55
301200	840	100	4	48
310000	1140	65	5	43

Table 4

Typhoon Iris -- April-May 1951

Date Time (GCT)	Distance from Guam (mi)	Center Intensity (kts)	Swell at Guam (ft)	Micro Amplitude (mm)
291200	390	40	3	39
300000	420	55	3	48
301200	450	60	3	45
010000	470	70	5	42
011200	540	95	5	45
020000	600	70	5	36
021200	700	65	5	38
030000	810	100	5	53
031200	850	130	5	65
040000	930	130	4	80
041200	1020	110	4	85

Table 5

Typhoon Nora -- August 1951

Date Time (GCT)	Distance from Guam (mi)	Center Intensity (kts)	Swell at Guam (ft)	Micro Amplitude (mm)
280000	270	30	3	19
281200			3	17
290000	545	40	3	20
291200			3	20
300000	800	45	3	27
301200			3	29
310000	1050	75	3	28

Table 6

Typhoon Ruth -- October 1951

Date Time (GCT)	Distance from Guam (mi)	Center Intensity (kts)	Swell at Guam (ft)	Micro Amplitude (mm)
070000		15	3	11
071200		20	4.5	11
080000		30	4.5	15
081200		40	6.0	36
090000		50	6.5	57
091200	260	75	7	64
100000		85	7	80
101200	540	100	7	169
110000		110	9	158
111200	750	110	8.5	144
120000		120	5.0	134
121200	870	120	4	147
130000		110	3	150
131200	1080	120	3	180
140000		130	3	169
141200	1170	110	1	141

Each of these storms illustrate the point that microseismic activity can be significantly above noise level even when no oceanographic activity reaches the coast. The first three of these storms were almost completely unattended by surf effects and the microseismic activity reached a maximum of 85. It would appear that the first crucial test was passed.

Of the remaining storms which did pass close enough to Guam to produce significant oceanographic effects, two are presented here. The first illustrates the case when center intensification and swell arrival did not coincide, Figure 5. The second shows Typhoon Allyn which passed directly over Guam with both the storm and swell reaching maximum amplitude simultaneously, Figure 6. See also Tables 7 and 8.

Table 7

Typhoon Allyn -- November 1949

Date Time (GCT)	Distance from Guam (mi)	Center Intensity (kts)	Swell at Guam (ft)	Micro Amplitude (mm)
131200	960	35	3	30
140000	850	60	3	30
141200	750	65	3	38
150000	645	70	4	44
151200	540	105	8	70
160000	480	100	9	77
161200	420	110	13	99
162000		115	14	164
162200		120	22	186

Table 8
Typhoon Doris -- November 1949

Date Time (GCT)	Distance from Guam (mi)	Center Intensity (kts)	Swell at Guam (ft)	Micro Amplitude (mm)
061200	450	60	3	30
070000	360	70	3	40
071200	300	80	10	50
080000	290	85	22	70
081200	150	90	27	90
090000	100	110	27	140
091200	180	115	27	135
100000	270	120	17	100
101200	300	115	5	90

From these cases, it would appear that the microseismic phenomenon usually recorded, is essentially a dual one, with part of the effect being the result of generation within the storm and part as the result of secondary surf effects. From the order of magnitude of response observed on Guam, it appears further that the surf effect is the dominant one having a magnitude at least twice that of the storm-induced microseism.

Despite the paucity of data available, it seemed worthwhile to attempt to find some sort of relationship between the various figures available. One such attempt involved plotting central wind intensity against microseismic response for the three storms which showed

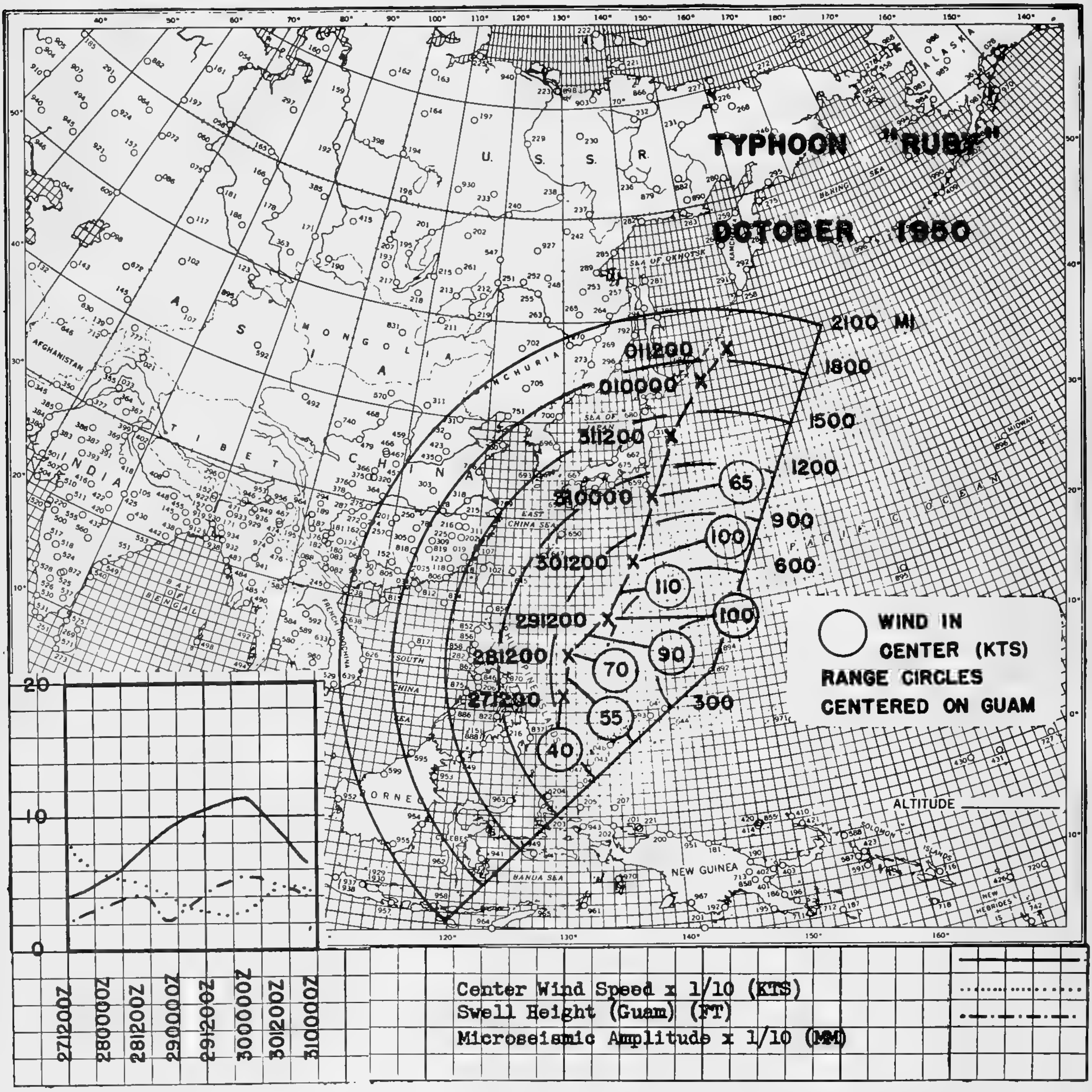

Figure 1

no swell at Guam. The composite graph, ignoring distance, as well as the individual graphs plotting the same values in various distance categories are presented as Figure 7.

This rather pleasing result is somewhat soured by the fact that an attempt to subtract the amplitude of the microseism produced by the storm as shown by Figure 7 from the total effect on a storm-surf microseismic record did not produce any relationship between swell amplitude and microseismic activity. This may be the result of incorrect hypothesis, poor swell forecasting or the result of the fact that the microseismic amplitude as measured is not a simple additive function of the two disturbances.

Microseismic records seem to indicate that at least two phenomena are being measured simultaneously. Figure 8 shows a construction which seems to bear this out. A simple disturbance having a period of 4 units and a crest-to-trough amplitude of 18 units was

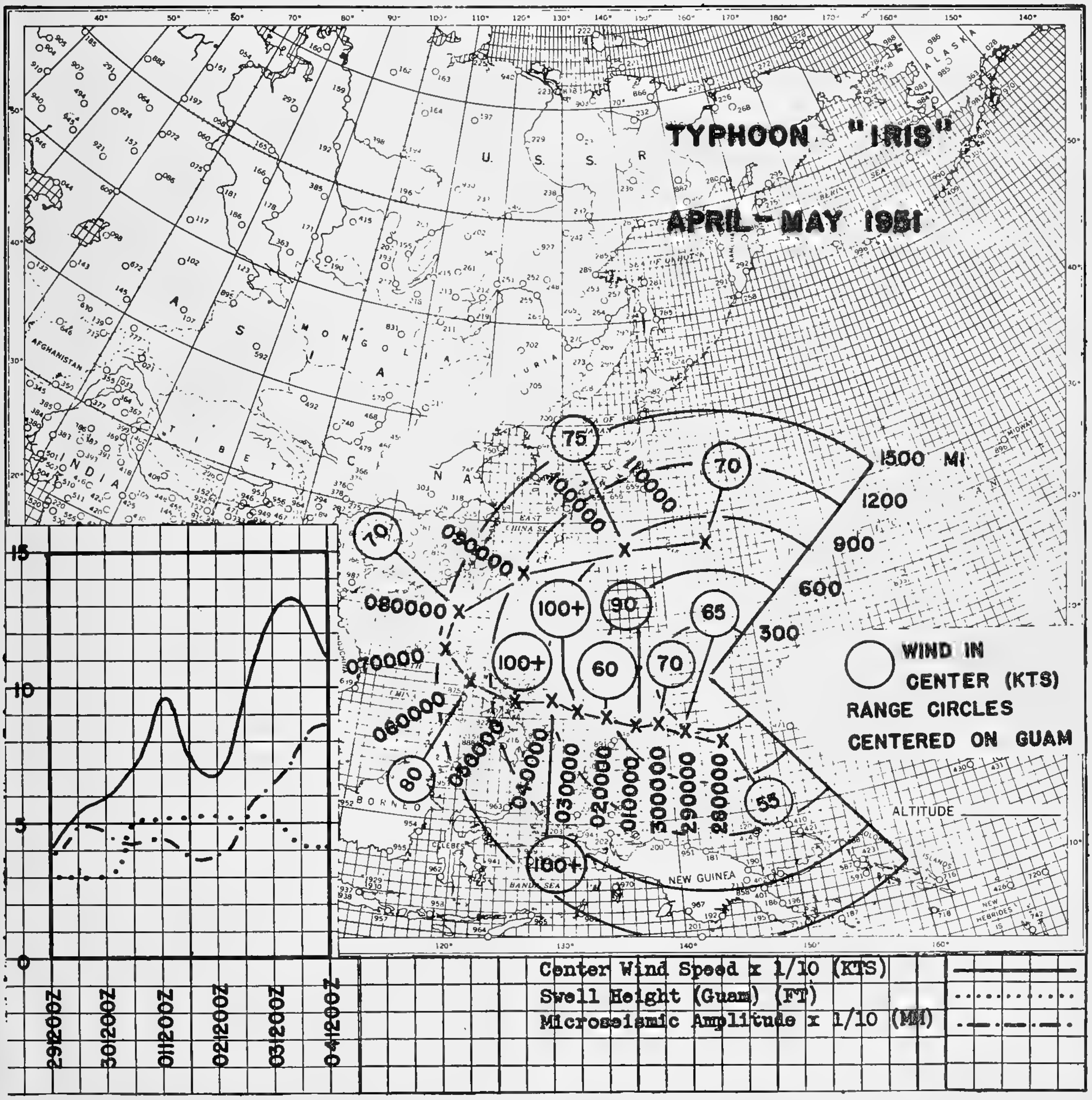

Figure 2

added to another disturbance with a period of 5.5 and an amplitude of 30. The individual disturbances are represented by (A) and (B) in the figure while the sum of the disturbances is represented by (c). A sample microseismic record is included as (d). Although the values taken for the simple cases were selected at random a considerable parallelism seems to exist between the composite record and the microseismic record.

Another interesting deduction can be made from this construction. Although the original periods were 4 and 5.5, the most marked period on the composite is 5—representing neither the shorter nor the longer wave. Moreover, the amplitude, depending upon where and how it is measured could be recorded as a value between 39 and 43.

It would seem from this, that if the reasoning has been valid to this point, much of the microseismic data collected thus far are inconsistent due to the fact that they do not represent analysis into the component parts.

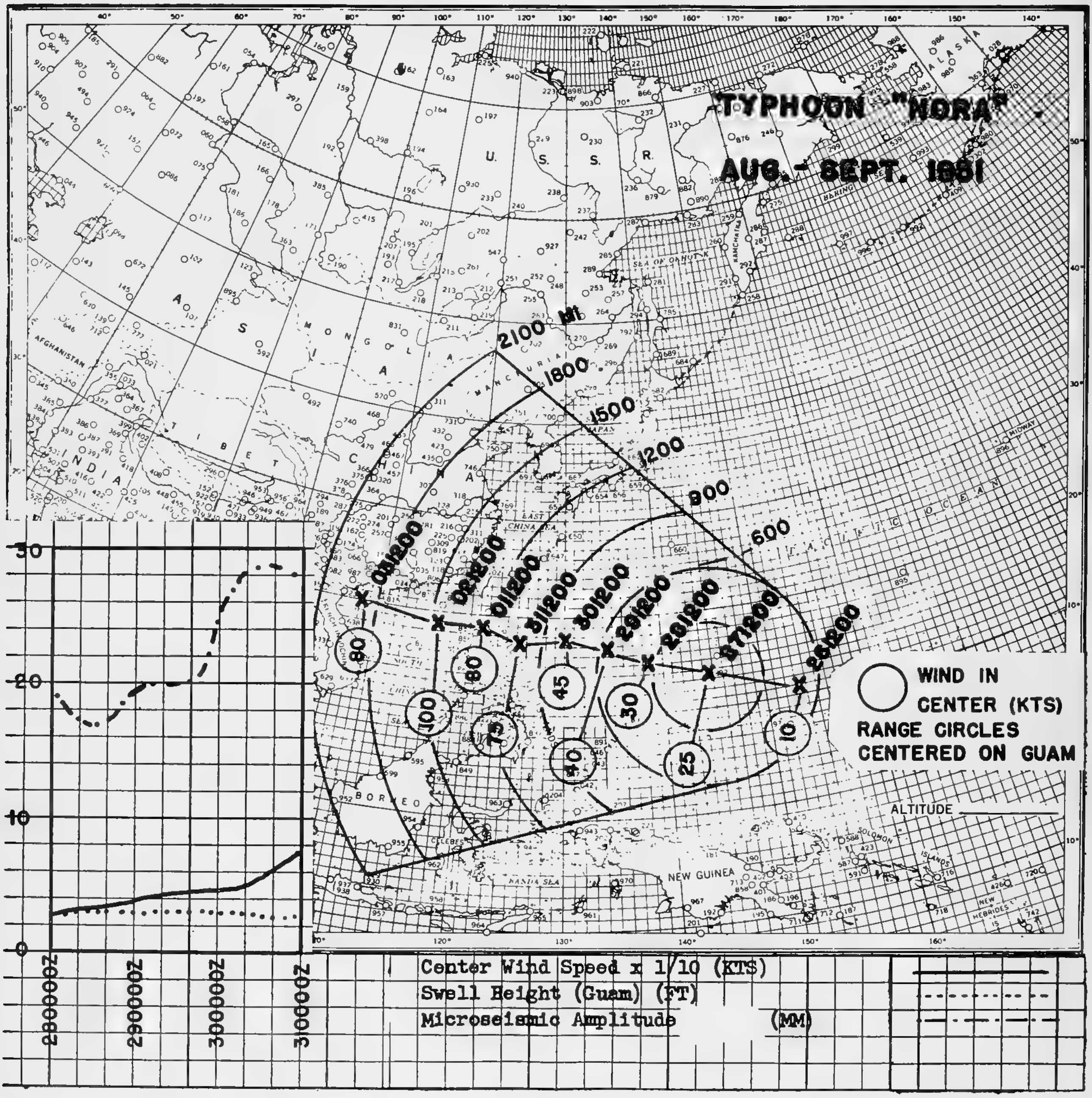

Figure 3

These speculations have led to the development of the 1952 program for microseismic research for the Navy. Wave recorders have been installed around Guam. The amplitude and period data are being so recorded that machine subtraction from the microseismic record will be possible. Perhaps by a frequency analysis the question of where microseisms originate will be solved. To amplify the wave recorder data, further visual observations are being made. Aerologists aboard weather reconnaissance planes are using their drift meters to measure the wave length of the swell approaching the coast as well as that leaving the storm.

Discussion

B. Gutenberg
California Institute of Technology

Dr. van Straten in her careful investigation came to the conclusion that hurricane microseisms are partly generated within the storm area, partly by surf effects, and that the latter are dominant. As a result of a recent investigation of microseisms connected with non-tropical storms approaching the Pacific coast of the United States, the present author has reached the similar conclusion that these microseisms derived their energy mainly from

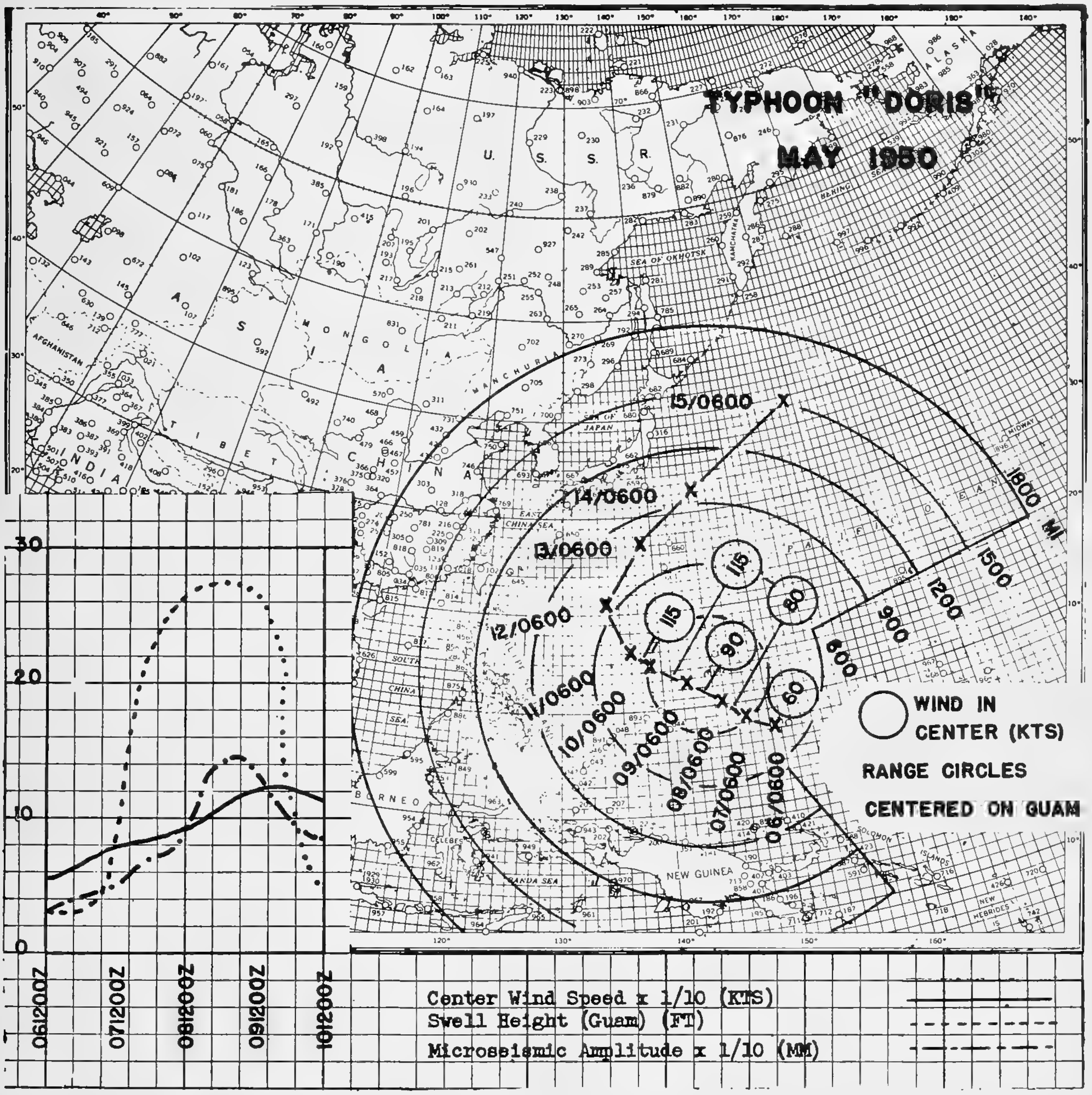

Figure 4

high ocean waves near the recording stations. Sources of energy near the storm center appeared to play at best a minor role and even that only as long as the storm was over the ocean and not too far from the recording station. More detailed results of this investigation which was sponsored by the Geophysical Research Division of the Air Force Cambridge Research Center, are to be published in the Transactions of the American Geophysical Union.

In several microseismic storms recorded at stations in California, the State of Washington and in British Columbia during November-December, 1951, the increase and decrease of the microseismic amplitudes were more and more delayed with increasing distance of the recording station from the storm center. In Southern California the time of the largest microseismic amplitudes lagged the time at which the storm center passed the coast (usually in British Columbia or in the State of

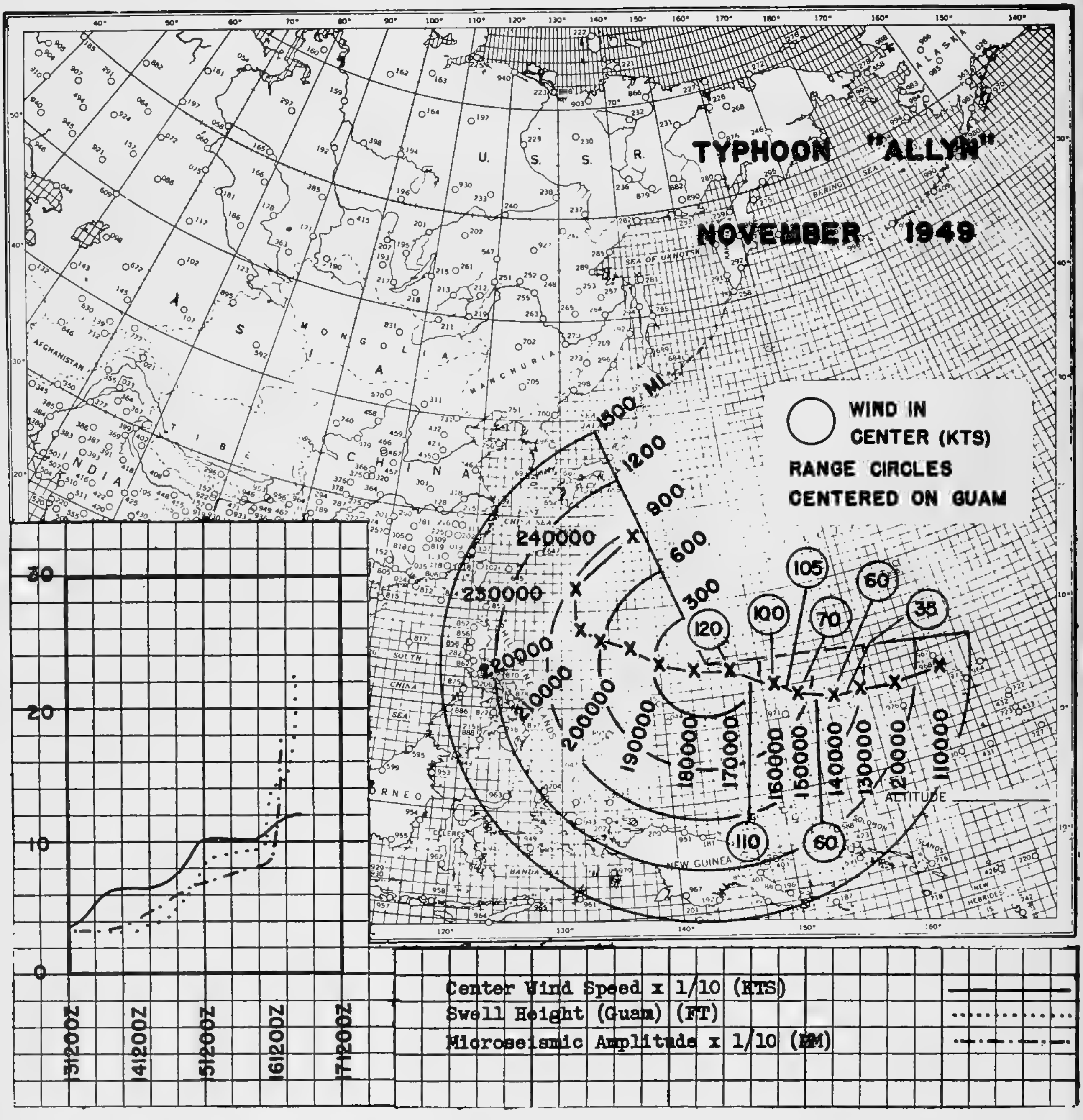

Figure 5

Washington) by about two days, but they usually coincided approximately with the highest breakers and ocean waves recorded at or observed at three points in Southern California. For an example, see Fig. 1A.

Microseisms traveling along the mountain chains and fault systems which follow the Pacific coast, decreased rather rapidly along their paths. In some instances these "barriers" reduce large microseismic maxima recorded at some stations to practically normal size while the seismic waves travel for a few hundred miles in the geologically disturbed area.

The periods of the ocean waves and those of the microseisms usually showed no parallelism (Fig. 1B); to the contrary, the microseisms usually reached their largest periods at about the time of the maximum amplitudes, while the ocean waves frequently exhibited relatively short periods when their amplitudes were large. This is a consequence of the well-known fact that the periods of ocean waves usually increase with the distance from the source.

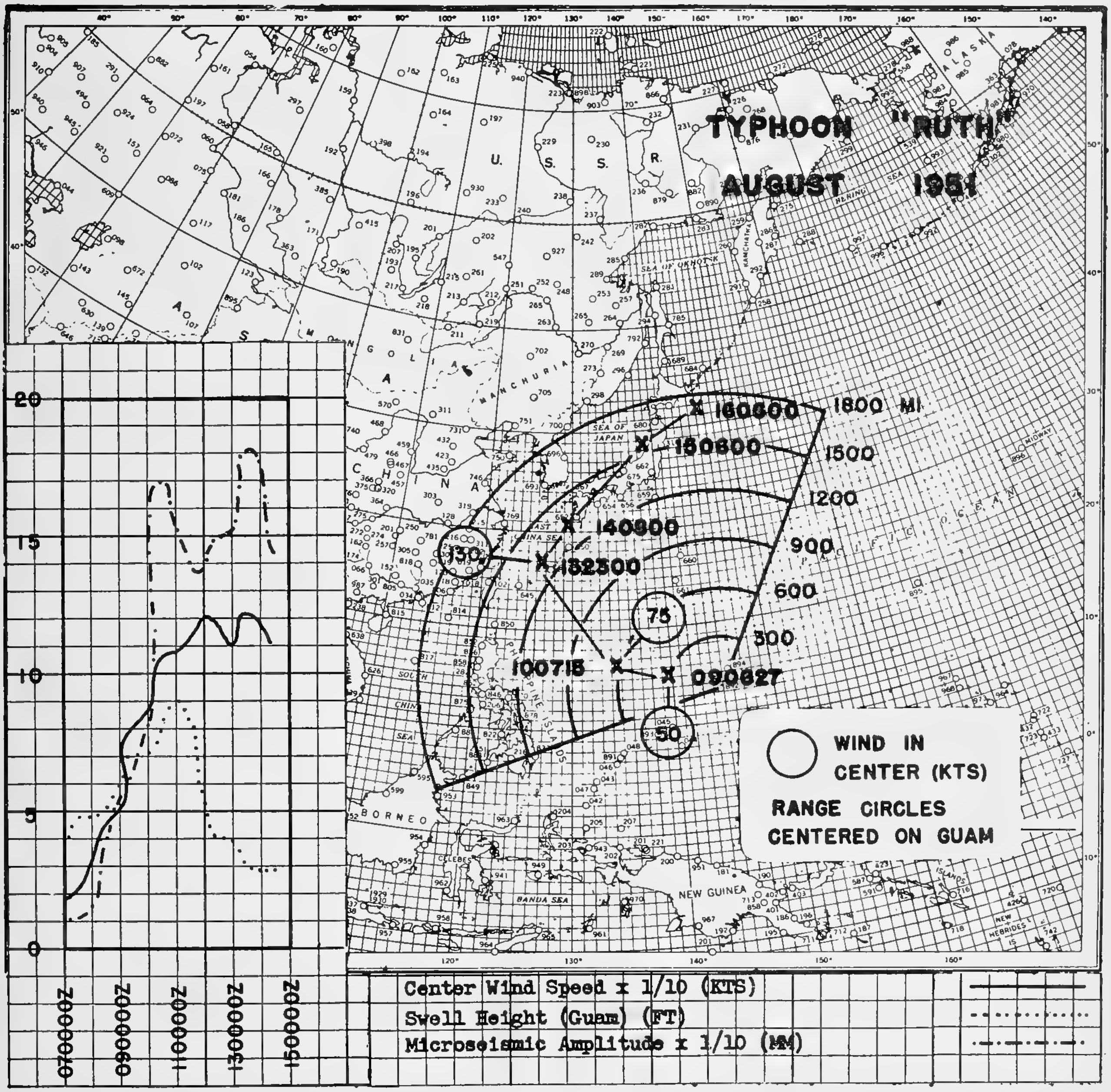

Figure 6

The fact that in Southern California the microseisms usually reach their maximum at a time when the storm center has moved rather far inland and when its intensity is decreasing makes it easier to investigate their correlation with the meteorological effects there than at the east coast of North America where the storm is usually moving out to sea and frequently intensifying at the time of the microseismic maximum. It also seems to be easier at the west coast to distinguish between the rather regular microseisms with periods from 4 to 10 or more seconds (Fig. 2B) and the irregular microseisms with periods near 4 seconds (Fig. 2A). While microseisms of the regular type usually increase and decrease rather slowly and, during their maxima, have periods of at least 6 seconds, the amplitudes of the irregular microseisms frequently increase from small values to large maxima within a few hours with the periods remaining close to 4 seconds. This type of irregular microseisms seems to be correlated with strong local winds, especially after passage of cold fronts. Microseisms with periods of about 2 seconds are frequently superposed (Fig. 2A), especially during local rain. In the author's opinion, the fact that the irregular microseisms with periods of about 4 seconds are frequently confused with the regular microseisms (they do not always differ as much as the examples in Fig. 2) is the main reason for the failure to reach a conclusion as to the source of energy for these microseisms in spite of extensive investigations for about 50 years.

Discussion

J. JOSEPH LYNCH, S.J.
Fordham University

Dr. van Straten has presented a thought provoking paper. She has courageously attempted to reconcile two schools of thought on the origin of group microseisms—the school that holds that microseisms originate at the center of a storm and the school that holds that microseisms cannot originate at the center of a storm over deep water but rather at some distance from the center in shallow water, as a

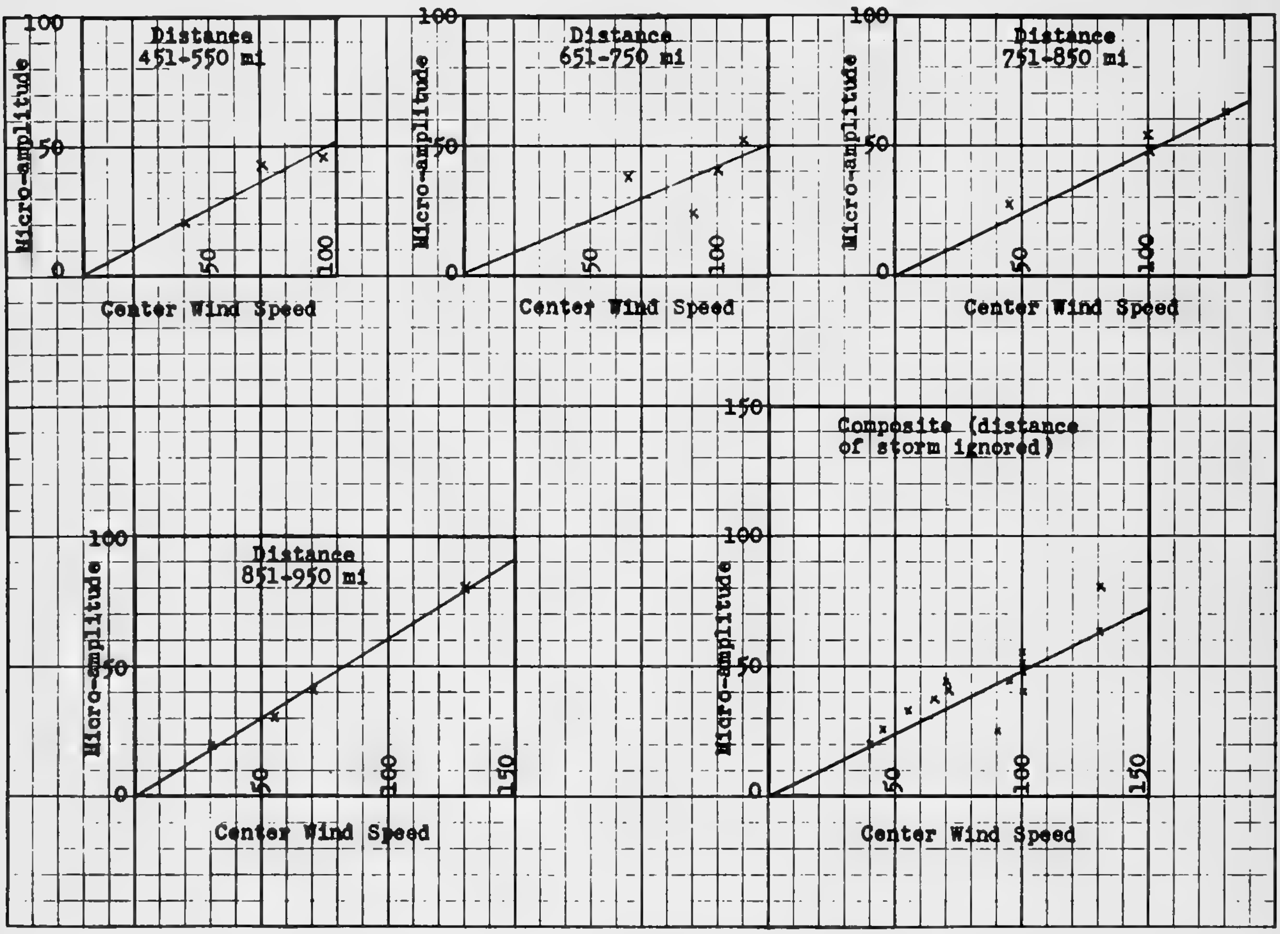

Figure 7

result of intermediary surf action very broadly understood. She moreover clearly points out that it is only the practical aspects of the problem that she is considering.

Much as I agree with her paper as a whole, I feel obliged as devil's advocate to disagree with or at least to challenge her on some minor points.

As the first proponent of the storm center school Dr. van Straten cites the work of Marion Gilmore. She does not mention the work of Fr. Ramirez, presumably because she felt that all in the field are aware that Gilmore's work was based on that of Ramirez and because Gilmore has provided the most extensive application of the theory of Ramirez.

In criticizing Gilmore's work Dr. van Straten refers to the indifferent results that he obtained. It is not quite clear whether she is classifying all of Gilmore's results as indifferent or whether she is criticizing those which he admitted were not too successful. Either way, Gilmore did successfully track some hurricanes following their storm center. This successful part of his work must be kept in view in any evaluation of his results. In trying to track others he was unsuccessful and in an attempt to account for his lack of success he ventured a possible explanation suggested by Dr. Gutenberg—namely geological ocean barriers. Dr. van Straten objects to this explanation, (1) that since the paths of hurricanes differ widely from year to year, the ocean barriers encountered would differ widely from year to year and hence "No large confidence factor could be given to such storm tracking," and (2) that as a theory which in the words of Shakespeare is "more honoured in the breach than in the observance" it should be discarded as inadequate or incorrect.

I disagree both with the logic and the wisdom of Dr. van Straten's criticism. When a marksman scores more misses than hits he should not be dissuaded from trying to hit the

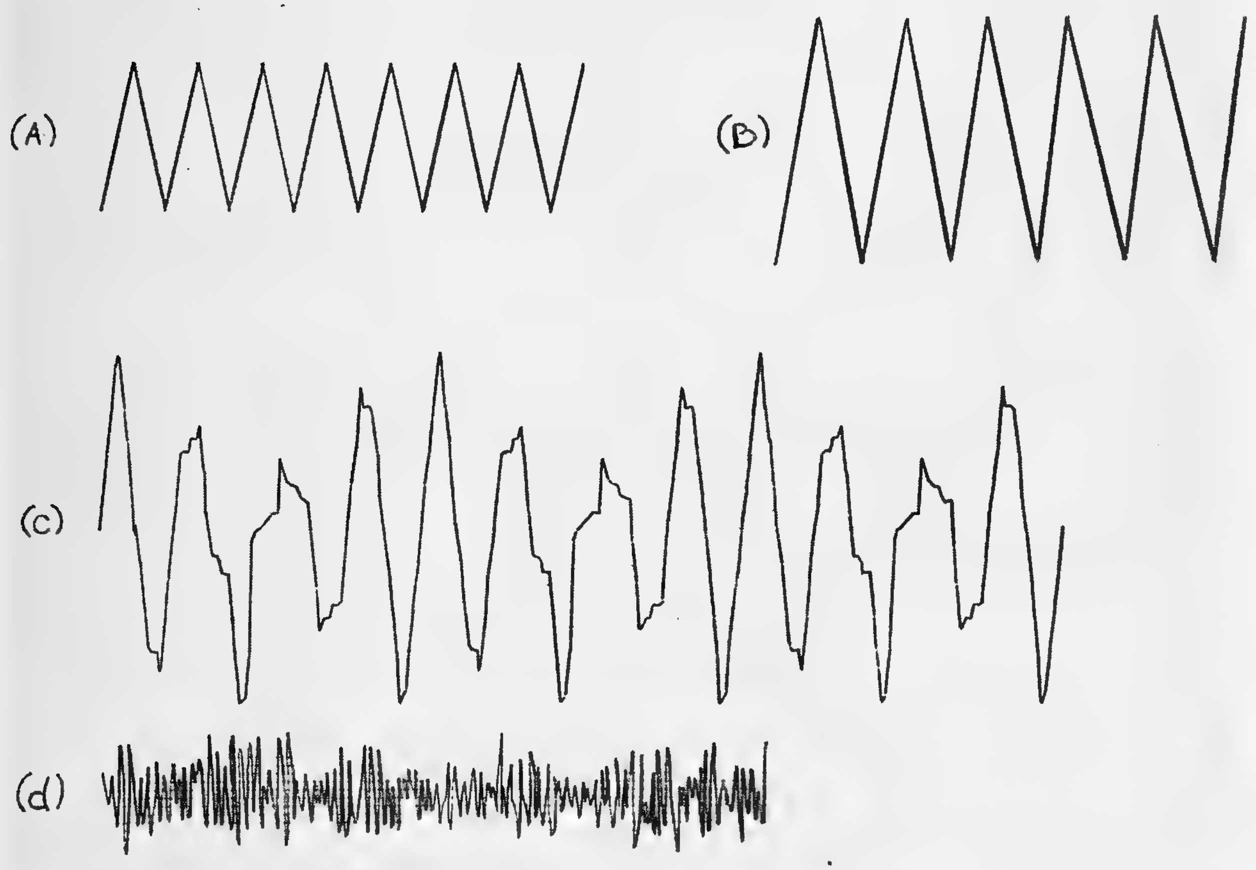

Figure 8

target but should be encouraged to find the reasons for his misses.

When Sir Isaac Newton first measured the velocity of sound in air by measuring the pressure and density, he obtained a value ridiculously far from the accepted value. He erroneously attributed his erratic result to the fact that air is not a perfect gas. Since no actual gases are perfect gases, he should, following Dr. van Straten's logic have given up the method since it would fail in more cases than it would succeed. Fortunately for Physics, further investigations were conducted. These finally led LaPlace to point out that not the method but Newton's explanation of his failure was incorrect. This may well be true in Gilmore's case. Newton's erratic result arose from his treating sound as an isothermal process whereas actually it is an adiabatic process. With this correction the method has been used successfully to measure the velocity of sound in all gases. The analogy is far from perfect but applying the same reasoning here, I would disagree with Dr. van Straten's criticism of Gilmore's work "that it does not permit a decision one way or another" and say rather that since it has proved successful in some cases, it represents progress and should be continued until the reason for its failure in other cases is definitely established beyond question.

As the second proponent of the storm center theory, the work of the Lamont Observatory is cited—presumably the work of Wm. L. Donn. Since Donn's work differs from Gilmore's chiefly in that Donn assigns a definite theory of the origin of microseisms, I shall pass over this section of Dr. van Straten's paper. I should like to mention however that it is unfortunate that Ramirez, Gilmore and Donn all used average values of time intervals in determining the direction of the microseismic source. It would have been more satisfying if direction had been obtained from time intervals of individual waves. The resulting directions could then be grouped into the most prominent ones and the presence of more than one seismic source would at once become apparent.

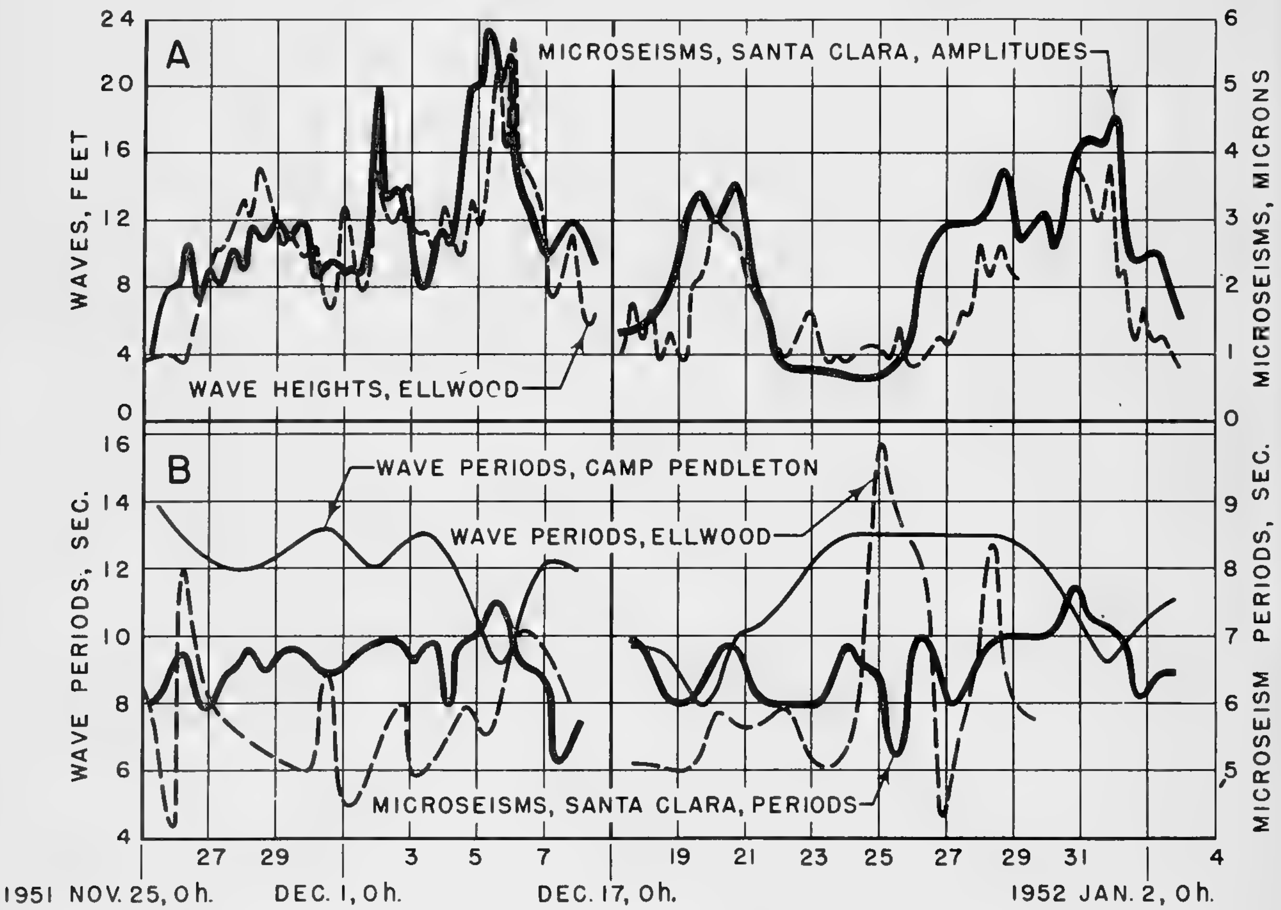

Figure 1. (A) Amplitudes of regular microseisms with periods of 5-8 seconds recorded at Santa Clara University, California, during November-December, 1951, and highest recorded ocean waves at Ellwood, California. (B) Periods of microseisms at Santa Clara and periods of ocean waves recorded at Ellwood and at Camp Pendleton, California.

Passing now to the proponents of the surf center theory loosely so called, the work of Kammer and Dinger is cited. Quite properly Dr. van Straten states that she must accept all of the published data as valid. It is one thing to accept the data as valid—it is quite another thing to accept author's conclusions from data as valid.

Dr. van Straten states that "Kammer and Dinger demonstrate that microseisms occur only if surf action is appreciable and that the magnitude of response to surf action is such as to mask any direct storm response." I think that is a fair statement of the conclusion of Kammer and Dinger, but I emphasize that it is their conclusion from their data and not merely a statement of their data.

Since it is Kammer and Dinger who are being criticized, it seems fair to quote them verbatim. Their conclusion in NRL Memorandum Report No. 3 reads: "No microseisms can be identified as being propagated through the earth from the storm center when the storm is over deep water. Therefore the early warning value of microseisms in the Atlantic and Caribbean seems to be non existent!"

I cannot agree that this necessarily follows from their data. The old logicians had a saying "QUI NIMIS PROBAT, NIHIL PROBAT!" He who proves too much proves nothing.

That Kammer and Dinger did not get any bearings on the center of the storm is an extremely interesting and important fact. That their comparatively meagre experiments demonstrate that no bearings can be obtained from the center of the storm is too sweeping an assertion. Gilmore did obtain bearings on the center of the storm in deep water. The interesting problem to be solved now is—"Why did Kammer and Dinger fail to get bearings while Gilmore succeeded?" Gilmore too failed on occasion. What are the conditions that cause failure?

The problem of the two conflicting views on microseisms is very much like the problem of the nature of light that confronted physicists during the past few decades. Is light a corpus-

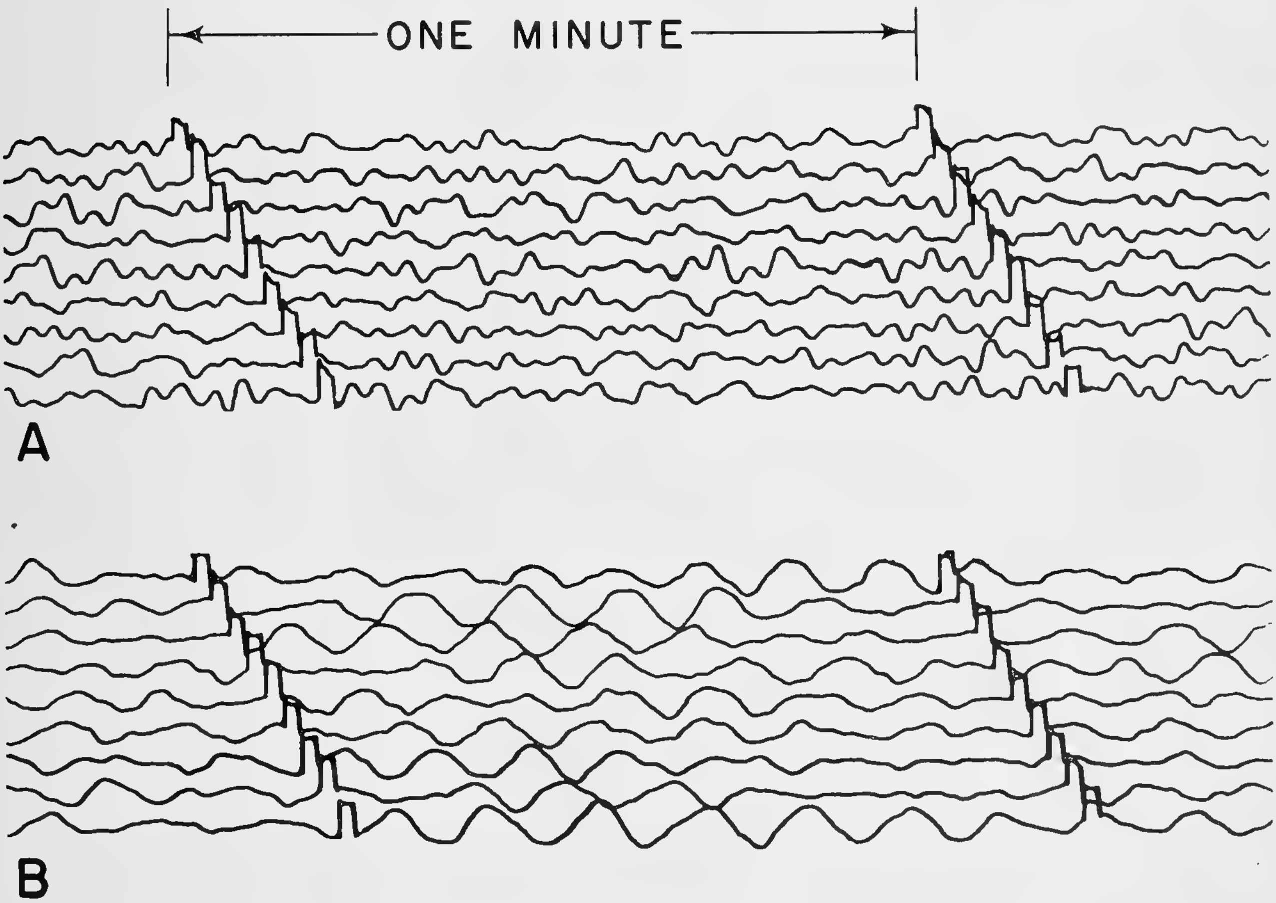

Figure 2. Microseisms recorded by Benioff vertical seismograph at Pasadena.

cular or a wave phenomenon? The physicist now finds it is both! Not enough data is available at present for settling this microseism problem. I hope that both Gilmore and Dinger will continue their hurricane tracking independently so that with increased data an effort can be made to solve the baffling problem.

Dr. van Straten's suggestion that we take as a starting point that microseisms are generated by interfering swells is an excellent one and should be the object of intense experimental work. Our work with cold front microseisms has convinced us that these frontal microseisms originate in the Great Lakes, possibly, as Dr. van Straten suggests as the result of interfering swells. Further and more pronounced activity is then noticed when the cold front enters the Atlantic. This may well be the masking action that Kammer and Dinger observed in hurricane micros. Our reason for the conviction that the Great Lakes are a source of micros is the following: When we moved our tripartite station to Poughkeepsie to get away from New York traffic, we noticed a persistent source of two second micros to the West. We recorded from many directions, but there was a persistent source to the West. Since the Hudson River is a sizable body of water to the West we felt it necessary to test this as a possible source of local micros. We set up a station at West Park on the West side of the Hudson, opposite Poughkeepsie. The source of the micros was still to the West, eliminating the Hudson River as a possible origin. Moreover, there were two persistent time intervals indicating two persistent directions—one due West and one North West. Lake Erie was due West of our Station and Lake Ontario, North West. This threw suspicion on the Lakes as the source of our micros. As a first test we set up a third station in the Western part of North Carolina, at Hot Springs. This is due South of Lake Erie and as far South of the Lake as Poughkeepsie is to the East of the Lake. At Hot Springs the source of our micros was due North. Since Lake Erie is due North of Hot Springs and due West of Poughkeepsie we are now convinced that Lake Erie is the source of our persistent two second cold front microseisms. We propose in the near future to record both water and ground activity on the shores of one of the Great Lakes. In this connection it might be worth mentioning in passing that while working on frontal microseisms in Fort Schuyler, one instrument set up on a concrete pier in the Sound recorded the propeller pattern of each passing tug.

We agree with Dr. van Straten that most micros are complex waves. Manly has given a simple method of analyzing such wave trains by inspection. He uses the fact that when two waves of different periods combine, the resulting wave assumes the period of the wave of greater amplitude, the amplitude oscillating, as in ordinary beats between the sum and difference of the amplitudes of the combining waves. In the resultant wave, if the separation of peaks at the maxima is greater than the separation of peaks at the minima, the frequency of the component of greater amplitude is greater than the frequency of the component of less amplitude, otherwise the reverse is true. Suppose we have a microseismic wave train in which the time interval between two group maxima is one minute. Suppose there are 12 peaks between the maxima. Then the frequency of the component of greater amplitude is 12 per minute and the period of this component is therefore 5 seconds. If the separation of peaks at the maxima is less than the separation of peaks at the minima, the frequency of the lesser component is greater than that of the major—namely 13 per minute. Hence the period of the lesser component is 4.6 seconds. Suppose the amplitude of the maxima is 11.4 mm and that of the minima is 2.2 mm. Since these are respectively the sum and difference of the constituent amplitudes, the constituent amplitudes are 6.8 and 4.6 mm respectively. Hence our microseismic wave can be analyzed into two waves of periods 5 and 4.6 seconds and of amplitudes, 6.8 and 3.6 mm.

When three waves combine the treatment becomes more complicated, but it is given by Manley (1945).

In conclusion I wish to thank Dr. van Straten for the privilege of having been able to pre-digest her very masterful resumé of storm and surf microseisms.

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THE OCEAN AS AN ACOUSTIC SYSTEM

By Frank Press and Maurice Ewing
Columbia University

Introduction—A cursory examination of the voluminous writing on the subject indicates that there is almost as much disagreement on the observational data of microseisms as there is on the question of their origin. Progress toward a solution of the problem can only be made by first deciding what are the data to be explained. We begin our discussion with a list of what we believe are basic facts derived primarily from data of east coast stations. The list pretends to be neither complete, nor generally applicable to other localities. It is our belief, however, that a successful theory of the origin of microseisms must satisfactorily explain these data (Donn, 1951a, 1951b, 1952a, 1952b) in addition to the observations from other localities. We realize that some of the data disagree with other observations reported at this meeting. However we are convinced that observations on this coast forces one to these conclusions.

1. Frontal microseisms are generated very soon (often abruptly) after a cold front passes seaward from land, with no obvious correlation to prior wind and sea conditions.

2. A relatively narrow spectrum of periods appears to be generated by a front, cyclone, or hurricane at a given time when the disturbance is over an area of uniform water depth. Characteristic periods of microseisms can be related to generating areas in the ocean.

3. As a front recedes from shore, the spectrum gradually shifts to longer periods, and becomes fairly constant after deep water is reached.

4. Cold fronts and air masses following them can generate microseisms whereas warm air masses preceding the cold fronts fail to generate microseisms even when strong onshore winds are present.

5. In many cases there are no obvious correlations between swell and surf conditions and microseisms.

6. Microseism energy is dissipated by a profound crustal discontinuity at the edge of the continental shelf. Hurricanes crossing the edge suddenly generate larger microseisms.

It is our opinion that no published theory of microseisms satisfactorily explains all of these observed data. The authors' theory (Press and Ewing 1948) advanced some years ago utilizing the Airy phase associated with stationary values of group velocity requires long, homogeneous, propagation paths. The work of Donn and others shows that this is not the case for many microseism storms. The work of Longuet-Higgins, and others on stationary gravity waves appears to explain satisfactorily how pressure fluctuations of sufficient magnitude to account for microseisms may be communicated to the sea floor. That stationary waves capable of generating microseisms occur in the open ocean has not been demonstrated to the satisfaction of many investigators and cannot be reconciled with many of the observations of Donn and others. Many difficulties are found with the theory of surf pounding. The authors have at present no theory which can account for amplitudes and periods of microseisms but feel that the data requires one in which the properties of the ocean-rock acoustic system under the generating area are significant in determining microseism periods. A theory should also account for the observation that only certain air masses appear capable of generating microseisms.

The Ocean as an Acoustic System—Seismic refraction measurements and earthquake surface wave studies (Ewing et al, 1950, Ewing and Press 1950, Ewing et al, 1952, Ewing and Press in press, Officer et al, 1952) indicate that the ocean basins are underlain by about 1 km of mud with acoustic properties much closer to those of sea water than the underlying crystalline rock. The mud velocity is about 5500 ft/sec with density about 1.5 gm/cm whereas the crystalline rock velocity is about 22,000 ft/sec with density 3.0 gm/cm (Donn 1952 b). This is to be compared with a velocity of 5000 ft/sec and unit density for sea water. It is seen that a great impedance contrast exists between the water-mud layer and the crystalline floor. A single set of acoustic parameters can be used to specify the unique properties of the ocean-crystalline basement system over a large area of the ocean basin. In many cases, however, microseisms are generated on the continental shelf or near the continental edge and the paths do not cross this excellent acoustic system.

Seismic refraction measurements on the submerged continental shelf off the eastern U. S. reveal that a wedge of sediments, thickening seaward, overlies the crystalline basement rock. The details of the variation of sedimentary thickness with distance from the shore vary along the coast. An upper sedimentary layer with acoustic properties similar to those of water and a lower layer acoustically intermediate between sea water and the crystalline rock below are the major constituents of the sedimentary wedge.

That these acoustically unique features of oceans are significant has been demonstrated from dispersion studies of earthquake Rayleigh waves in the period range 16-40 seconds. The coherent, sinusoidal oscillations showing a simple, orderly dispersion point up the excellence of the acoustic system for these periods. A simple theory can account for the entire sequence arrivals of first mode Rayleigh waves having oceanic paths in terms of normal mode propagation over long distances in the water-crystalline rock system. Predictions of the water and sediment thickness as well as the nature of the crystalline rock underlying ocean basins have been verified by seismic refraction measurements (Ewing and Press 1950, Ewing and Press in press, Officer et al. 1952).

J. E. Oliver has been studying shorter period surface waves which propagate across the oceans with periods 7-12 seconds. These oscillations appear on transverse as well as radial and vertical components, and are best recorded on islands. Preliminary results suggest that they consist of both Love waves and second mode Rayleigh waves which are strongly refracted and almost entirely absorbed at the continental margins. This is in accord with the great discontinuity known to exist at the continental margin. The dependence of the degree of absorption and refraction on period can be demonstrated from surface wave studies. Investigation of Mantle Rayleigh waves with periods greater than 60 seconds (Ewing and Press 1953) indicates that negligible absorption and refraction occurs. Comparison of absorption of first and second mode Rayleigh waves from the same tremor indicates that the shorter waves are much more strongly attenuated by the discontinuity. Evernden (1952) has shown that Rayleigh waves with periods less than about 35 seconds are significantly refracted by the continental margin. It seems probable from these results that these effects are even more pronounced for the shorter period microseisms. It is not surprising that refraction effects and barriers are among the most significant features noted by those studying the data of tripartite stations in view of the length and irregularity of the continental margin.

It seems significant that the shortest period surface waves from earthquakes in the Atlantic Ocean are above the periods generally observed for microseisms. That this is not primarily due to the spectrum of the source is suggested by the fact that body waves occur with microseism periods, and the T-phase often present, appears with even shorter periods.

Atmosphere-Ocean Coupling — ROSCHKE (1952) in a recent paper reports that micro-oscillations in the atmosphere of periods less than one minute reach their maximum amplitudes in the post-cold front interval and that streams of cold air are more efficient producers of microoscillations than warm air. Data from Columbia microbarographs are in agreement. In view of the previous observation that the type of air mass over the ocean is a significant factor in microseism generation these results strongly suggest that pressure fluctuations in the atmosphere may provide energy for microseisms in a manner as yet unknown to us. More data is needed on the areal extent of these oscillations as well as their oceanic amplitudes.

It has been suggested that vertical oscillations of the water column analogous to "organ pipe" vibrations may well be a significant feature of the ocean-rock acoustic system. Use of this concept to explain microseisms is not new (Banerji 1935). On a seismic prospect in shallow water (Burg et al. 1951) where the bottom was composed of smooth hard rock, the predominant signal obscuring all other waves on short spread seismograms consisted of a repetitive pattern of the "organ pipe" modes of vibration of the water layer. In some cases all the modes but one could be filtered revealing a long train of sinusoidal oscillations with the proper frequency for that mode and water depth. Another aspect of vertical compressional oscillations of the water column is revealed by a simple calculation of the vertical displacements on the ocean floor originating from steady vertical oscillations applied to the surface. The results show, as might be expected from the general theory of transmission through plates, that the ocean is an extremely sharp filter for transmission of compressional waves from the surface to the bottom—the sharpness originating in the high impedance contrast between the water and mud and the crystalline basement. The peak periods, T, for waves transmitted to the crystalline basement are given by

$$T = \frac{2H}{2nv} \qquad n = 1,\ 2,\ 3$$

Where H is the water-unconsolidated sediment thickness, v is about 5000 ft/sec. Calculations by Dr. Jardetzky have shown (as one might expect from the general theory of filters) that a transient impulse applied to the sea surface appears at the bottom as trains of damped sinusoidal waves having periods corresponding to the "organ pipe" modes. Although these waves can explain microseism periods they cannot be propagated horizontally to any significant distances due to downward leakage of energy out of the system as body waves.

Recently it has been shown that under certain circumstances a resonant transfer of energy from the atmosphere to the earth's surface can occur despite the tremendous impedance mismatch between the two media (Haskell 1951). This phenomenon has now been observed for coupling between compressional waves in the atmosphere and Rayleigh waves on the earth's surface (Press and Ewing 1951a), flexural waves on floating ice and tsunami (Press et al. 1951b, Press and Ewing 1951c). When viewed from the elementary standpoint of the theory of travelling disturbances, resonant coupling occurs when a disturbance travels along the surface of a medium at a velocity close to that of a free wave in the medium. If the free wave is dispersive the energy from the disturbance goes into those waves whose periods are such that the phase velocity is close to the velocity of the disturbance. The resonance is especially sharp for large density contrasts between the two media as is the case with the atmosphere and the earth. The possible connection between this mode of coupling of atmosphere to ocean and microseisms is being investigated. One obvious feature is that pressure oscillations in the atmosphere striking the sea surface at an almost vertical angle and maintaining coherence over a large area do not fully satisfy the conditions for resonant coupling since "organ pipe" oscillations in the sea column are not free due to the small leakage at each boundary.

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Discussion

N. A. HASKELL

Air Force Cambridge Research Center

I have tried to make a crude order of magnitude estimate of the amplitudes to be expected from the generation of microseisms via excitation of the "organ pipe" modes in the ocean by atmospheric pressure oscillations. The results seem to me to indicate that this mechanism is of questionable quantitative significance. In Roschke's (1952) study of atmospheric pressure oscillations he classifies oscillations of periods less than 1 minute as "large" when the double amplitude is greater than about 2 dynes/cm². In the illustrations he gives of typical high microbarometric activity immediately following the passage of a cold front the double amplitude appear to run around 6 dynes/cm². The same order of magnitude has been quoted for pressure fluctuations having periods in the neighborhood of 5 sec. observed at the Signal Corps Engineering Laboratories (Daniels. 1952).

Now if the ocean is excited in one of the vertical "organ pipe" compressional modes by a pressure oscillation of amplitude P_0 at the surface, the pressure amplitude at the bottom is greater than P_o in the ratio $\rho_b c_b / \rho_w c_w$ where ρ_b and ρ_w are the densities and c_b and c_w are the compressional wave velocities in the bottom and water respectively. If we take $\rho_w = 1$, $\rho_b = 3$, $c_w = 5000$ ft/sec, $c_b = 22{,}000$ ft/sec, this ratio is 13.2, giving about 80 dynes/cm² at the bottom for a 6 dyne/cm² amplitude at the surface. The displacement amplitude at the bottom for an oscillation of period T will be of the order of $TP_b/2\pi\rho_b c_b = TP_o/2\pi\rho_w c_w$ which is about 0.3 micron for T = 5 sec and the other quantities having the values assumed. This should be a value characteristic of the immediate area of generation and the amplitude at a seismic station at some distance should be considerably less, yet the amplitudes of microseisms attributed to the passage of cold fronts over deep water may run to more than 2 microns. There seems to be a discrepancy by a factor of 10 or more.

I rather doubt that the idea of resonant coupling between elastic waves in two different media due to coincidence between the phase velocities of different wave types will turn out to have a great deal to do with the coupling of atmospheric pressure oscillations to the ocean bed. Where it has been possible to correlate wave forms of microbarometric waves across a tripartite array, the apparent phase velocity has usually come out to be very much less than sound velocity and comparable with the wind velocity at some moderate altitude. Gravity surface waves on the ocean also have phase velocities of the same order as wind velocities, so that if microbarometric oscillations are coupled with anything in the ocean it is presumably with gravity rather than compressional waves.

Gravity waves only 1 meter high would give pressure oscillations of the order of 10^5 dynes/cm² near the surface and bottom pressures exceeding the 80 dynes/cm² estimated for the direct excitation of the "organ pipe" modes for all water depths less than 1.25 wave lengths, or about 160 feet for waves of 5 sec. period. So far as pressures go, surface waves in shallow water seem to be adequate to generate observable microseisms.

However, there seems to be a good deal of statistical evidence that at least some microseismic activity, and perhaps most of it in some areas, has a deep water origin. Whipple and Lee (1935), investigating Banerji's (1930) suggestion that gravity waves should generate compressional waves that were not attenuated exponentially with depth, showed that the compressional wave travelling with the velocity of gravity surface waves would necessarily have an exponential attenuation rather than a sinusoidal variation with depth.

If therefore appears to me that neither the direct action of atmospheric pressure oscillations nor indirect coupling via gravity waves in the first order linear approximation are adequate to explain the generation of microseisms in deep water, and the second order term in the expression for the bottom pressure as discussed by Longuet-Higgins (1950) is the only mechanism that has been proposed so far that looks quantitatively adequate. The failure of some observers to verify the two-to-one ratio between the periods of ocean waves and of microseisms as deduced from this theory may indicate nothing more than that the wave periods observed on a swell recorder in shallow water near the coast are not necessarily the same as the periods of the interfering wave systems that produce microseisms in the storm area. A deep-water bottom pressure recorder should throw a great deal of light on this question.

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Discussion from the Floor

Lynch. Since the purpose of this conference is to reconcile contradictory views on microseisms as far as possible, I should like to call attention to a contradiction or at least an apparent contradiction presented by the opening part of Dr. Press' paper. He states, on the basis of Dr. Wm. Donn's work that microseismic activity on the East Coast begins only when the cold front enters the Atlantic. The speaker in a paper yesterday states that he and his colleagues feel positive that microseismic activity on the East Coast from cold fronts originates in the Great Lakes. Here then we have an apparent contradiction—one author claims the activity originates in the Great Lakes, another author claims the activity originates in the Atlantic—to the casual listener surely a contradiction!

This, however, is one contradiction that we can easily reconcile. I should like to point out that the time taken for a microseismic wave to reach New York from the Great Lakes is a matter of minutes. I should like to point out also that the position of Cold Fronts on weather maps is only a rough geographical location —and good only within a period of a few hours. It is therefore impossible to state within minutes just when a cold front first enters the Atlantic. That pronounced microseismic activity is recorded when a cold front enters the Atlantic we most heartily agree. One statement of yesterday therefore in no way contradicts those of Dr. Press. We do state, however, that a matter of hours before this (the precise number of hours depending on how fast the cold front is advancing) we record microseismic activity caused by the front as it passes over the Great Lakes. We record this in a matter of minutes after the front has reached the Lakes—giving the seismograph a definite warning value in the case of fronts.

In a sentence, the contradiction is explained by pointing out that we are recording waves from the frontal activity over the Lakes, whereas Press and Donn are referring to waves from the frontal activity over the Atlantic. I merely wish to emphasize that we are all in agreement on the microseismic activity as the front passes over the Atlantic.

Longuet-Higgins. (1) The response curve for the movement of the ground shown in Figure 1 is considerably sharper than that shown in my paper (Figure 7). The reason is probably as follows: the first curve is the response to a horizontal plane oscillation of infinite extent, which causes energy to be propagated vertically downwards into the ground; the other is the response to a pressure distribution of finite extent from which the waves are propagated outwards horizontally. The first waves are relatively difficult to generate, being subject to less constraint.

(2) Dr. Press has pointed out the rather sudden onset of microseisms at the time that a cold front crosses the coast. I think that there is no difficulty at present in supposing that this is due to wave interference. As the weather maps show, there is then a very sharp change in the direction of the wind. It is not necessary that the wind should be exactly reversed in direction, because a given wind will probably generate waves which, when analyzed, will be found to have some wave components travelling at a considerable angle to the mean direction. N. F. Barber has shown by an optical diffraction method that even a regular swell has components spread over an angle of 30°; for an irregular sea the angle would be greater. The rapidity with which the microseism amplitude is built up may be explained by the fact that, if once the original progressive system of waves is established (from which no microseisms would be expected), only a small amount of wave energy travelling in the reverse direction would be sufficient to produce the necessary pressure fluctuations. Data at present available for the rate of growth of waves under a wind refers to waves growing gradually under a following wind; it is quite conceivable that the rate of growth of waves travelling downwind, but in the presence of an opposing swell, is greater, on account of the roughness of the sea surface. Observations of the rate of growth should be obtained. Controlled experiments could also be made on a smaller scale, using a laboratory wave tank and an opposing artificial wind.

(3) The amount of wave reflection from the New England coast is probably very small, since the shore in most places is not steep. The exact value of the reflection coefficient cannot be assumed to be the same as for laboratory experiments with a beach of the same slope, since the scaling, for waves of different period, is uncertain; also in the laboratory experiments the motion was laminar, while in the sea turbulence may play a part in the energy dissipation. However, it may be possible actually to determine the extent of reflection from different parts of a coast by a comparative study of the spectra of pressure fluctuations on the bottom, just offshore.

(*Bath* pointed out that on the Norway coast the effect is when the front crosses the coast and not the edge of the shelf. After *Haskell's* formal discussion, *Melton* asked if the sudden increase in microseisms and the *Longuet-Higgins* theory may not be consistent due to the reversal of winds. *Longuet-Higgins* pointed out the waves will then be short. *Donn* pointed out cases where the sea has been calm but there are microseisms. *Byerly* commented on the fact that at least one seismologist believes microseisms result from winds against mountains. *Gutenberg* replied to this that because of the location the wind may actually be on the shore. *Press* described a swell observed on the New England coast in instances of a large swell and no microseisms. *Longuet-Higgins* blamed this on a low reflection coefficient. *Deacon* inquired if during some of the swell described by Press, which was of eighteen seconds period, there were any nine second microseisms, and was told no.)

ON THEORIES OF THE ORIGIN OF MICROSEISMS

by J. G. Scholte

In the last decennium the concept that many microseisms are generated by a storm at sea has been more and more generally accepted; detailed investigations as for instance by Bernard (1941) as well as the successful detection of hurricanes by means of tripartite stations prove the validity of this view beyond any doubt (Gutenberg 1952).

It is however still uncertain by which process these seismic movements come into existence; the observations often point in different directions and it is therefore not possible to formulate a theory covering all observed data.

Perhaps the most useful way to treat this matter theoretically is to ignore various meteorologic and oceanographic circumstances and to start from the undisputed fact that a disturbance at the surface of the ocean causes at a distance of the order of 10^8cm. microseisms with an amplitude of say 5 μ and a mean period of about 6 seconds.

The movement of the ocean in the vicinity of the storm area has an amplitude which is of course several times greater than 5 μ and as the vertical motion at the bottom has to be continuous the same is true for the movement of the water. In view of the well known fact that the amplitude of gravity waves decreases exponentially with the depth, it is evident that the motion of the water which generates at the bottom the seismic waves is not a gravitational but a compressional wave and that we may neglect the effect of gravity on this process.

Consequently we have to consider waves of compression in a purely elastic system consisting of a fluid layer of finite depth h covering a solid body. Consider a cartesian coordinate system with the x axis in the free surface parallel to the direction of propagation and the z axis vertically downward. In order to avoid complications which are irrelevant to this problem we suppose this body to be semi-infinite. Denoting the horizontal component of the movement by u and the vertical one by w,

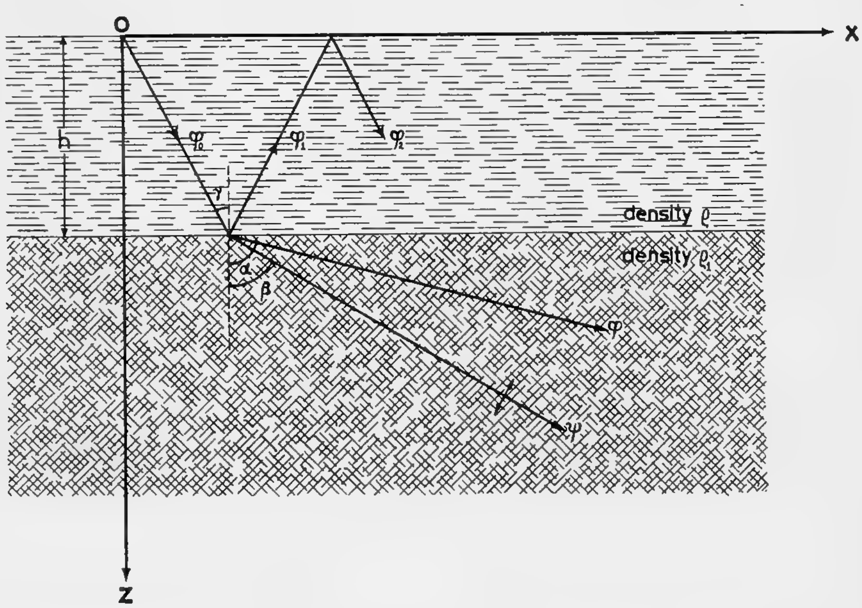

Figure 1

a compressional wave travelling in the liquid in the direction γ is described by

$$u = \frac{\partial \phi_o}{\partial x}, \quad w = \frac{\partial \phi_o}{\partial z}, \quad \phi_o = A \exp\left\{ i\nu\left(\frac{x \sin\gamma + Z \cos\gamma}{c} - t \right) \right\}$$

where ν = the frequency, c = the velocity of sound in water, and γ is an angle of incidence measured from the vertical.

At the bottom $z = h$ three other waves are excited:

the reflected wave $\qquad \phi_1 = RA \exp\left\{ i\nu\left(\frac{x \sin\gamma - z\cos\gamma}{c} - t \right) \right\}$

and two refracted waves:

the longitudinal one: $\qquad u = \frac{d\phi}{dx}, \quad w = \frac{d\phi}{dy}$, with the potential

$$\phi = D_l A \exp\left\{ i\nu\left(\frac{x \sin\alpha + 2\cos\alpha}{a} + \frac{h\cos\gamma}{c} - \frac{h\cos\alpha}{a} - t \right) \right\};$$

and the transverse wave: $\qquad u = \frac{\partial\psi}{\partial z}, \quad w = \frac{\partial\psi}{\partial x},$

with the vector-potential ψ in the y-direction:

$$\psi = D_t A \exp\left\{ i\nu\left(\frac{x \sin\beta + Z\cos\beta}{b} + \frac{h\cos\gamma}{c} - \frac{h\cos\beta}{b} - t \right) \right\};$$

a and b are the velocities of longitudinal and transverse waves in the solid. The quantities α and β are angles of refraction for compressional and sheer waves in the solid respectively.

The coefficients of reflection and refraction (R and D) are determined by the condition that the tensions T_{ZZ}, T_{ZX} and the vertical motion w have to be continuous across the plane $z = h$;

$$w: \quad \frac{\cos\gamma}{c} - R\frac{\cos\gamma}{c} = D_l \frac{\cos\alpha}{a} + D_t \frac{\cos\beta}{b}$$

$$T_{zz}: \quad \rho + R\rho = D_l\ \rho_1 \cos 2\beta + D_t\ \rho_1 \sin 2\beta$$

$$T_{zx}: \quad 0 = D_l \frac{\sin 2\alpha}{a^2} - D_t \frac{\cos 2\beta}{b^2}$$

where ρ and ρ_1 are the densities of the liquid and solid respectively.

The solution is:

$$R = \left(\cos^2 2\beta + \frac{b^2}{a^2} \sin 2\alpha \sin 2\beta - \frac{\rho c \cos \alpha}{\rho_1 a \cos \gamma}\right) \frac{1}{\Delta},$$

$$D = \frac{2\rho \cos 2\beta}{\rho_1\ \Delta}, \quad D_t = \frac{2\rho b^2 \sin 2\beta}{\rho_1\ \Delta a^2}$$

$$\text{with:} \quad \Delta = \cos^2 2\beta + \frac{b^2}{a^2} \sin 2\alpha \sin 2\beta + \frac{\rho c \cos \alpha}{\rho_1 a \cos \gamma}$$

Arriving at the surface $z = 0$ the reflected wave ϕ_1 gives rise to a new wave ϕ_2 in the $+\gamma$ direction; as the normal pressure caused by ϕ_1 and ϕ_2 has to disappear at $z = 0$ we have

$$\phi_2 = -RA \exp\left\{iv\left(\frac{x \sin \gamma + z \cos \gamma + 2h \cos \gamma}{c} - t\right)\right\}$$

Thus the twice reflected wave ϕ_2 is equal to ϕ_0 multiplied by $-R \exp (2 iq)$, with $q = (vh \cos \gamma)/c$; it follows that the wave Φ_1 travelling in the γ direction is given by

$$\sum_{n=o}^{\infty} \phi_{2n}, \text{ or } \phi_o \sum_{n=o}^{\infty} \left\{-R \exp (2iq)\right\}^n. \quad \text{Hence}$$

$$\Phi_+ = A \frac{\exp\left\{iv\left(\frac{x \sin \gamma + Z \cos \gamma}{c} - t\right)\right\}}{1 + R \exp (2iq)} \quad \ldots\ldots\ldots\ldots\ldots (1)$$

and in the same way the reflected wave system $\boldsymbol{\Phi}_-$ is

$$\boldsymbol{\Phi}_- = RA \frac{\exp\left\{i\nu\left(\dfrac{x \sin\gamma - Z\cos\gamma + 2h\cos\gamma}{c} - t\right)\right\}}{1 + R\exp(2iq)}$$

similar expressions for the refracted waves are easily obtained.
From these potential-functions we derive the movement of the ocean's bottom:

$$u = w \tan\alpha \left(\cos 2\beta - 2\frac{b}{a}\cos\alpha\cos\beta\right) \text{ and}$$

$$w = 2i\nu A \frac{\rho\cos\alpha}{\rho_1 a} \frac{\exp\left\{i\nu\left(\dfrac{x}{c}\sin\gamma - t\right)\right\}}{\Delta\left\{\exp(-iq) + R\exp(iq)\right\}}$$

which after some reduction can be written as

$$u = w \sin\beta\left\{\cos 2\beta\left(\frac{b^2}{a^2} - \sin^2\beta\right)^{-\frac{1}{2}} - 2\cos\beta\right\} \text{ and}$$

$$w = \frac{i\nu\rho A}{b\rho_1 N}\exp\left\{i\nu\left(\frac{x}{b}\sin\beta - t\right)\right\}, \qquad \text{(2)}$$

where

$$N = \left\{\cos^2 2\beta\left(\frac{b^2}{a^2} - \sin^2\beta\right)^{-\frac{1}{2}} + 4\sin^2\beta\cos\beta\right\}\cos q - 2i\frac{\rho}{\rho^1}\left(\frac{b^2}{c^2} - \sin^2\beta\right)^{-\frac{1}{2}}\sin q$$

$$\text{and } q = \left(\frac{\nu h}{b}\right)\left(\frac{b^2}{c^2} - \sin^2\beta\right)^{\frac{1}{2}} \qquad \text{(3)}$$

The primary wave φ_0 which excites this whole wave system, causes a pressure T_{zz} at the free surface $z = o$ equal to $-\rho\nu^2\phi_0$; the pressure T_{zz} of the secondary waves ϕ_i cancel out at $z = o$. Supposing ϕ_0 to be generated by a pressure

$$P = p \exp\left\{i\nu\left(\frac{x}{c}\sin\gamma - t\right)\right\}$$

uniformly applied on the plane $z = o$, the amplitude A of ϕ_0 is equal to $P/\rho\nu^2$.

In actual circumstances this periodic pressure—which is in any case necessary to obtain waves of compression—is confined to a finite area; in order to obtain a function which describes the actual conditions better than the function

$$p \exp\left\{i\nu(\frac{x}{c}\sin\gamma - t)\right\}$$

we change this into

$$pJ_0\left(\nu \frac{r}{c} \sin\gamma\right) \exp(-i\nu t)$$

where J_0 is a Bessel function and

$$r = (x^2 + y^2)^{1/2}.$$

The motion of the bottom is then given by the same expressions (2), if we change the factor $\exp(i\nu x/c \sin\gamma)$ into $iJ_1(\nu r/c \sin\gamma)$ for the horizontal (radially directed) component and into $J_0(\nu r/c \sin\gamma)$ for the vertical one.

Remembering the discontinuous factor of Weber

$$\int_0^\infty J_0(\xi r)\, J_1(\xi r_0)\, d(\xi r_0) = \begin{cases} 0 & \text{if } r>r_0 \\ 1 & \text{if } r<r_0 \end{cases}$$

It will be seen that the pressure function $p J_0(\nu r/c \sin\gamma) \exp(-i\nu t)$ changes into a function which is equal to $p \exp(-i\nu t)$ for $r < r_0$ and vanishes for $r > r_0$ if we apply the operator

$$\int_0^\infty J_1\left(\nu \frac{r_0}{c} \sin\gamma\right) d\left(\nu \frac{r_0}{c} \sin\gamma\right)$$

The parameter r_0 is arbitrary; following Lamb's procedure (1904) we diminish r_0, at the same time increasing p in such a way that the total force $\pi p r_0^2$ exerted on the plane $z = 0$ remains constant ($= Q$). In the limiting case $r_0 = 0$ the normal force Q is evidently concentrated in the point O and is expressed by

$$\frac{Q}{2\pi}\int_0^\infty J_0(\xi r) \cdot \xi\, d\xi.$$

Consequently by applying the operator

$$\frac{Q}{2\pi}\int_0^\infty \frac{\nu \sin\gamma}{c}\, d\left(\frac{\nu \sin\gamma}{c}\right)$$

to the expressions for u and w we obtain the motion at the bottom of the ocean, generated by a force Q concentrated in one point of the surface. We readily find

$$u = -\frac{\nu Q}{2\pi b^3 \rho_1} e^{-i\nu t} \int_0^\infty \frac{\sin^2\beta}{N} \left\{ \cos 2\beta \left(\frac{b^2}{a^2} - \sin^2\beta \right)^{-\frac{1}{2}} - 2 \cos \beta \right\}$$

$$J_1 \left(\nu \frac{r}{b} \sin \beta \right) d \sin \beta$$

$$w = i \frac{\nu Q}{2\pi b^3 \rho_1} e^{-i\nu t} \int_0^\infty \frac{\sin\beta}{N} J_0 \left(\nu \frac{r}{b} \sin\beta \right) d \sin \beta$$

It can be shown that the main value of these types of integrals is for large values of r contributed by the residues of the integrand at the zeroes of N. The microseismic movements at large distance from the generating force $Qe^{-i\nu t}$ are therefore:

$$u = -\Sigma_m W_m \sin\beta_m \left\{ \cos 2\beta_m \left(\frac{b^2}{a^2} - \sin^2\beta_m \right)^{-\frac{1}{2}} - 2 \cos\beta_m \right\}, \quad w = \Sigma_m W_m$$

$$W_m = Q \frac{\nu^{\frac{1}{2}}}{\rho_1 b^{\frac{5}{2}} (2\pi r)^{\frac{1}{2}}} \frac{(\sin\beta_m)^{\frac{1}{2}}}{(\partial N/\partial \sin\beta)_m} e^{i\nu \left(\frac{r}{b} \sin\beta_m - t \right) + \frac{1}{4}\pi i} \quad \ldots\ldots\ldots\ldots (4)$$

From (3) we see that the equation $N = 0$ determines for each value of $\nu h/c$ one or more values β_m; writing this equation in the form of the denominator in (1):

$$-R \exp (2 i q) = 1 \quad \ldots\ldots\ldots\ldots (5)$$

it is obvious that the values γ_m corresponding to the roots β_m determine the directions γ for which the reflected elementary wave ϕ_i is identical to ϕ_{i-2} (the phase shift caused by the two reflections at the boundaries cancels the difference in phase due to twice transversing the layer). Therefore the main part of the motion is caused by constructive interference of the plane waves in which the spherical wave originating at the origin can be decomposed (Press and Ewing 1948).

It follows from (5) that, as $|R| = 1$, γ_m has to be greater than the angle of total reflection for the tranvserse wave; hence $b > c$ and $\sin \beta_m > 1$.

Again if $\sin \beta < b/c$ the quantity q is real; for each value of sin between 1 and b/c we obtain therefore an infinite series of values $\nu h/b$ satisfying $N = 0$. In the diagram (fig. 2) several of these modes are shown (we have used the numerical values $a/b = \sqrt{3}$, $b/c = 2$ and $\rho/\rho_1 = 0.4$).

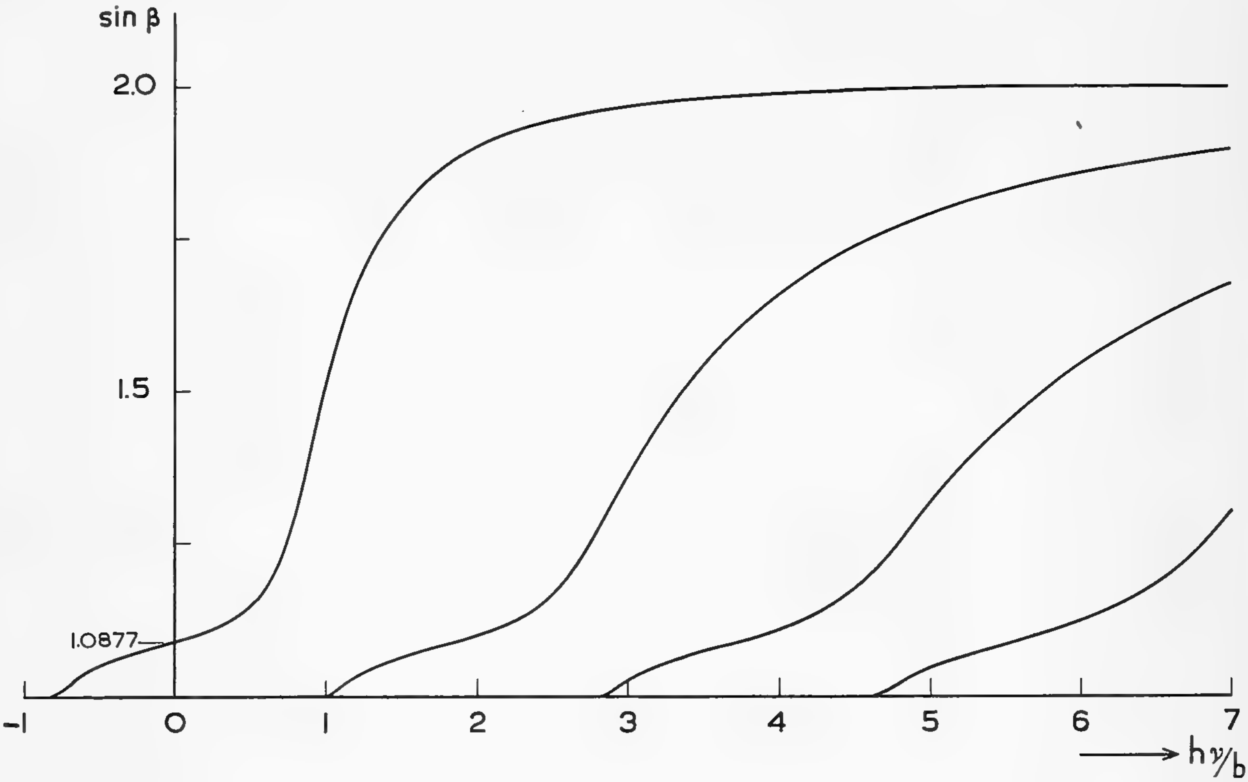

Figure 2

With the exception of the first mode each curve starts at $\sin\beta = 1$ and approaches $\sin\beta = b/c$ asymptotically. The curve of the first mode shows two peculiarities:

if $h = \phi$ the equation $N = O$ changes into the simple Rayleigh equation for the suboceanic medium

$$\cos^2 2\beta + 4\sin^2\beta\cos\beta\left(\frac{b^2}{a^2} - \sin^2\beta\right)^{1/2} = 0$$

with the root $\sin\beta = b/S_1$, where S_1 = the velocity of Rayleigh waves. At smaller values of $\sin\beta$ (in fig. 2 for $\sin B < 1.0788$) the equation $N = O$ yields a negative value of $\nu h/b$.

In the second place: if $h = \infty$ the equation changes into the equation for Stoneley waves in the two semi-infinite media (water and suboceanic rock):

$$\cos^2 2\beta + 4\sin^2\beta\cos\beta\left(\frac{b^2}{a^2} - \sin^2\beta\right)^{1/2} = 2i\frac{\rho}{\rho_1}\left(\frac{b^2}{a^2} - \sin^2\beta\right)^{1/2}\left(\frac{b^2}{c^2} - \sin^2\beta\right)^{-1/2}$$

with the root $\sin\beta = b/S_2$, where S_2 = the velocity of these waves. Hence the first curve starts at $\sin\beta = b/S_1 > 1$ and tends asymptotically to $\sin\beta = b/S_2 > b/c$.

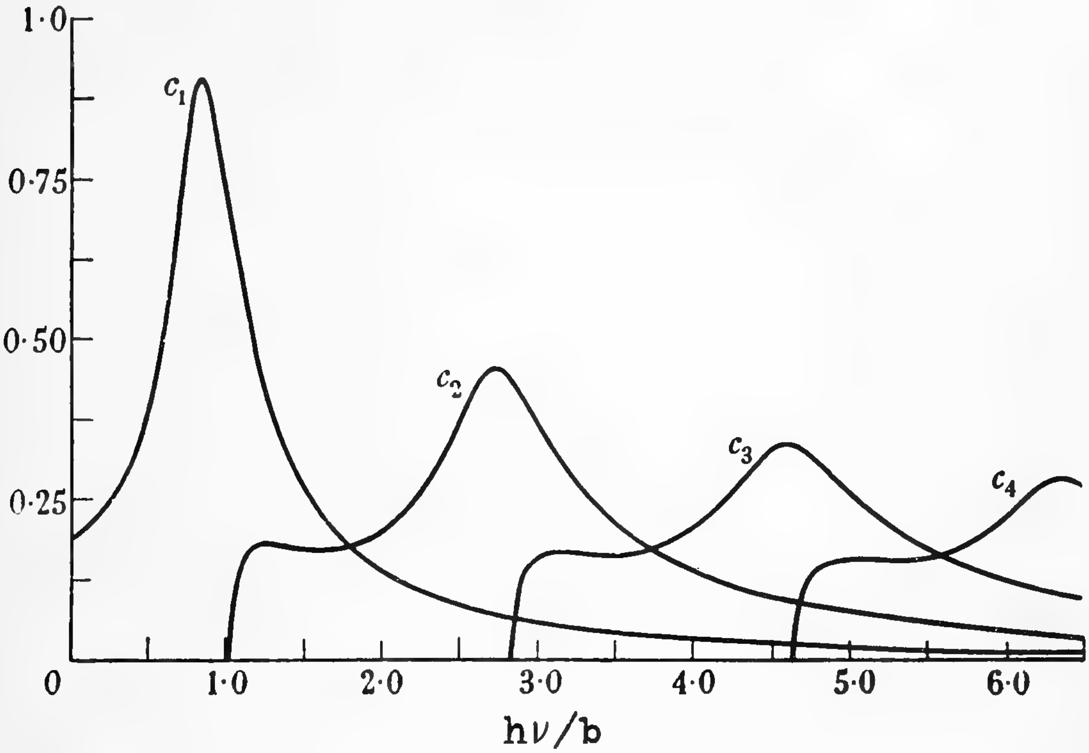

Figure 3

The values of $C_m = \dfrac{(\sin\beta_m)^{1/2}}{\left(\dfrac{\partial N}{\partial \sin\beta}\right)_m}$

have been calculated by Longuet-Higgins (1950) and are plotted in fig. 3 against $h\nu/b$. In fig. 4 the ratio

$$\frac{u}{w} = \sin\beta\left\{\cos 2\beta\left(\frac{b^2}{a^2} - \sin^2\beta\right)^{-1/2} - 2\cos\beta\right\}$$

is shown as a function of $\sin\beta$.

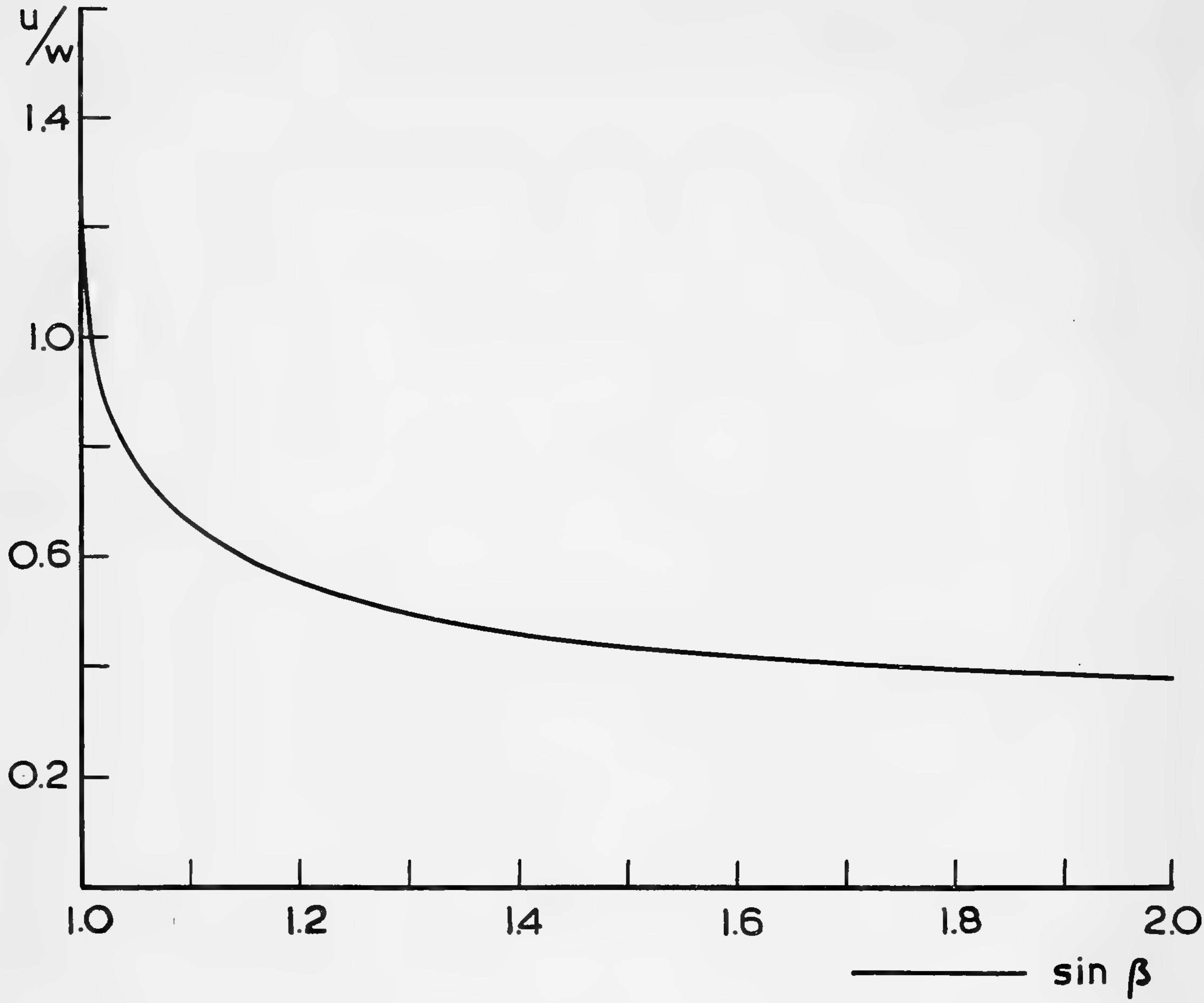

Figure 4

The maximum of the first mode appears at $\nu h/b \approx 0.85$; supposing $b = 2.8$ km/sec and $h = 3$ km we have $\nu = 0.79$ (period of about 8 sec.). At a distance of 3000 km the amplitude of the vertical component $= Q \times 1.5 \times 10^{-19}$ cm. In order to obtain a microseismic amplitude of 5 μ the total force Q has to be about 3.10^{-15} dynes; assuming the radius of the storm area $\approx$ 18 km it appears that a mean pressure variation of 1/3 mb is necessary to produce the observed microseisms.

In this calculation it has been assumed that the pressure variations at widely separated points are correlated; as this will not be the case in actual circumstances, the obtained value of 1/3 mb has to be interpreted as the effective pressure variation.

Supposing the phases of these pressure oscillations at points separated by a distance greater than a to be incorrelated the effective pressure of the storm area b^2 is about a/b times the mean pressure.

An explanation of such a pressure variation has to be found either in the atmospheric or in the hydrodynamical circumstances during a storm. With regard to the first and most obvious explanation we refer to the observations at Wei-ka-wei by Gherzi (1921); if the atmospheric pressure in the "eye" of the typhoon changes periodically with an amplitude of 0.5 mb this would be sufficient to generate microseisms at large distances of the track of the storm (Scholte 1943). It is perhaps possible to obtain more data about this phenomenon by placing a network of microbarographs in the regions where typhoons often occur.

Recent observations by Donn (1951) may also elucidate the connection between atmospheric disturbances and microseisms in the western hemisphere.

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Discussion

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Dr. Scholte is to be complimented for his concise discussion of Theories of the Origin of Microseisms. We can only add a few remarks on certain aspects of the problem. It is convenient to discuss separately (a) the nature of the source (b) the mode of transmission over oceanic paths (c) the mode of transmission over continental paths.

(a) *The Nature of the Source*—The wave interference theory of Longuet-Higgins (see his paper in this volume) is the only published treatment which can quantitatively account for energy transfer from the atmosphere to the ocean. Little work has been done on the possibility that pressure fluctuation and gustiness present in turbulent air masses can transfer energy directly to compressional waves in the ocean. In this connection the impulsive modification of sea waves by wind gusts (WHIPPLE AND LEE, 1935) may be important.

(b) *The Mode of Transmission Over Oceanic Paths*—There seems to be general agreement among investigators on the manner of transmission of elastic energy across the oceans (Stoneley 1926, Press and Ewing 1948, 1950a, b, Scholte 1943, Longuet-Higgins 1950). That transmission peaks occur for certain periods has been pointed out on several occasions. In order for these transmission peaks to develop fully, propagation paths of the order of 100 wavelengths or more are needed, a condition not fulfilled by many microseismic situations. It seems therefore that these transmission peaks can play only a secondary role in any theory on the origin if microseisms. It is particularly disturbing that surface waves from oceanic earthquakes contain comparatively little energy in the microseism period range.

(c) *The Mode of Transmission Over Continental Paths*—It has been observed that short period surface waves (with periods in the microseism range) propagate with surprisingly little attenuation over large continental paths. (Press and Ewing 1952). Although no theory has been presented to account for the details, it is apparent that the continental crust behaves as an homogeneous sialic plate for these waves. It seems possible that microseisms, once past the barrier at the continental margin, may well be transmitted in a manner similar to these earthquake phases.

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Discussion from the Floor

(*Dinger* asked if the absence of periods of one to seven seconds across the western Atlantic was also true of the Pacific, and *Press*, replied yes, on the Aleutians to Hawaiian path.)
Båth. Dr. Scholte mentioned that the ratios of horizontal and vertical amplitude of microseisms at De Bilt were larger than could be explained by his theory. The reason is obviously the very loose ground at De Bilt with around 9 km. of sediments (in accord with the theory of A. W. Lee).

(*Haskell* pointed out that where there is a big contrast in velocity between the surface layer and the underlying medium, there may be large ratios of horizontal to vertical amplitudes. *Romney* pointed out that there are some microseisms with periods as great as 20 seconds, but with much lower amplitudes than those of shorter period.)

Longuet-Higgins. The occurrence of microseisms in "groups" appears to be an example of a very general phenomenon which is observed whenever a disturbance can be considered as the sum of a large number of disturbances of about the same frequency. Such a sum was first considered by Rayleigh in 1880 in connection with sound from many different sources—this is sometimes called "the bee-hive problem." He showed that the probability P(a)da of the sound wave amplitude being between a and a-da was given by

$$P(a) = \frac{2a}{(\bar{a})^2} e - \frac{a^2}{\bar{a}^2}$$

where ā was the r.m.s. amplitude. The height of sea waves (defined as the difference in elevation between a crest and the preceding trough) has been shown to obey the same statistical law. In this case the generating area of the swell can be considered as divided up into regions, each large compared with the length of a sea wave, and each giving a sinusoidal contribution of independent phase. A similar concept probably applies to microseisms. An analysis of the statistical distribution of microseismic amplitude over a fairly short interval of time would be of interest.

(*Donn* commented that frontal microseisms are weaker in the summer than in the winter. *Carder* replied that recently there have been some intense summer cold front microseisms in Washington.)

Bath. Cold fronts cannot be located sufficiently accurately by interpolation from weather maps, especially not when they pass over oceans. Hydrographic records of pressure and temperature have to be used.

There is no microseismic effect observed when cold fronts pass the limit of the continental shelf outside the Norwegian coast, whereas there is generally rapid increase of the microseisms when the cold fronts pass the coast itself.

Jardetzky. It is yet difficult to understand in all details the mechanism of transmission of a disturbance in the air to be a recorder of microseisms. There are three media involved. The wave propagation in the ocean bottom (or coast) does not present any difficulty, but there is no agreement about the kind of disturbance at the sea surface and the behaviour of the sea. The observations are interpreted in different ways and the surf, the strong wind in the cold front of a cyclone, the atmospheric pulsations, the ocean swell or the interference of gravity waves are made responsible for microseisms in different theories. Neither of them seems to be convincing when all observations are taken into account, but each one might be true for a corresponding group of microseisms. It is difficult to see whether a sufficient amount of energy is communicated to the sea by the air masses directly in the form of compressional waves or it is transmitted or increased by the action of gravitational waves. There is no doubt more that the movement of air masses in a cyclone has to be taken as a primary disturbance. The atmospheric pulsations can be one of characteristics of this movement. On assuming the existence of these pulsations (or some other cause producing compressional waves) at the interface air-water, one can determine the part played by the latter. The signed has computed from the theory of propagation of a plane wave in the vertical direction a curve mentioned by Dr. Press. This curve represents the amplitude of the vertical displacement at the sea bottom in terms of $\frac{\omega}{\alpha}$ H, where H is the depth, α the velocity of sound in water, ω the circular frequency). The shape of this curve suggests that the ocean acts as a filter. For example, if α = 1.430 m/sec., H = 1000 or 3000 m. periods of waves, which will reach the bottom with amplitudes not changing essentially, will vary from 1.5 to 4 sec. Making clear such an interpretation of the behaviour of the water layer, this result does not explain the conditions at the sea surface. It seems that far more systematized data should be correlated with each of the factors involved in order to clarify those conditions.

(*Longuet-Higgins* brought up the subject of gusts again. He made a plea for measurement of their intensity at sea. *Van Straten* pointed out that there is a strong land-sea breeze on the East coast, and suggested it might make a difference between the day and night frontal microseisms.) *Deacon* agreed with the remarks about the possibility of microseisms of all periods up to a certain maximum being caused by one source, although they were ground oscillations of short and long periods produced by other causes such as traffic, wind on mountains, buildings, etc. Sixteen second ocean waves produced 8 second microseisms at Kew, and it seemed very likely that the 2 second microseisms observed by Father Lynch might be caused by 4 second waves on the Great Lakes. Dr. *Longuet-Higgins* had used ½ second waves to reproduce ¼ second microseisms in the bottom of his tank.

With regard to the common explanation of microseisms associated with weather disturbances it was likely that when the explanation was found it would be universally satisfactory. The position at present was very difficult to understand. On the eastern side of the ocean the microseism records looked like the frequency spectrum of the waves if the scale was divided by two. There was also a theory which appeared to be very satisfactory. On the western side of the ocean great emphasis was placed on the possible effect of microbarometric oscillation which looked nothing like the microseism records; if they had any period it was the wrong one, and the theory used to explain the energy transfer required confirmation at many points. The theory used on the eastern side of the ocean was held to be unsatisfactory.

It might be useful to concentrate more effort on the study of the phenomenon when it appeared to be simplest. It would be very useful if a seismologist from the United States of America could come to England to work on wave and microseism recordings made in the United Kingdom; he would be sure of a warm welcome and plenty of material to work on.