THEORY OF MICROSEISMS.†

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Received February 19, 1935.

1. Introduction.

In the paper* on microseisms associated with disturbed weather in the Indian Seas, I gave definite experimental evidence which showed that microseisms were due to the disturbance of pressure at the bed of the sea produced by sea-waves. The most important evidence was that microseisms were recorded as soon as a storm was formed either in the mid-Arabian Sea or in the mid-Bay of Bengal, five or six hundred miles away from the coast. In the storm area, tremendous waves were produced, which must have caused in some way or other disturbance of pressure at the bed of the sea, giving rise to the microseisms recorded. The microseisms could not be due to the waves travelling over shallow water near the coast, because the storm area was so far away that the waves would take 2 or 3 days to reach shallow water; the microseisms were recorded soon after the storm developed and not 2 or 3 days after it formed.

It was shown in the paper referred to above that if the assumption was made that the sea-waves caused a disturbance of pressure at the bed, then expressions could be obtained for the microseisms, from theoretical considerations, which explain correctly their observed periods and amplitudes at a distant station. The expressions also give correctly the ratio of the horizontal to the vertical component of microseisms (see Appendix). Some difficulty was, however, experienced in explaining the mechanism by which the disturbance of pressure was communicated to the bed of the deep sea. The ordinary hydrodynamical theory, which assumes irrotational motion and no viscosity or compressibility, shows that there is no disturbance of pressure at the bed of a deep sea. When eddy viscosity is taken into consideration, it is found that the disturbance due to the waves will take several hours to reach the bottom of the sea. In the treatment adopted in the paper referred to above the sea was divided into two parts:

† The publication of this paper has been delayed for nearly two years owing to my other pre-occupation.

(1) a superficial layer (marked 'A' in Fig. 1), and
(2) the sea underneath right down to the bed (marked 'B' in Fig. 1).

It was considered that in the superficial layer the surface waves behave in the same way as in a shallow sea so that on the horizontal plane $X''$ underneath this layer, there is a disturbance of pressure equal to $g\rho (h + \eta)$ where $\eta$ is the 'elevation' of the surface waves and $h$ the depth of the plane below the equilibrium surface. It was assumed that the sea underneath this superficial layer transmitted this disturbance of pressure to its bed in some way or other, almost undiminished in intensity. The pressure at the bed $X$ of the sea was therefore taken to be simply equal to $g\rho (d + h + \eta)$, $d$ being the depth of the sea below the superficial layer.

The above consideration clearly assumed that the disturbance was transmitted like that of sound by the property of compressibility, but when that paper was written, it was not quite clear to me how this compressibility of water was brought into play.

Since then I have carried out a series of experiments† in order to determine the effect of viscosity and compressibility in the transmission of disturbance of pressure to the bed of the sea. The best way of performing these experiments appeared to be to imitate the conditions in the sea in a small masonry tank and photograph the disturbance of pressure at various depths. Accordingly experiments were made in several tanks and mainly

† A preliminary note on the subject was published in Current Science, July 1932.
in a circular masonry tank of diameter 210 cm. and depth 108 cm. This tank was of very strong construction, the walls being 18 inches thick so that vibrations due to the movements of observers were not communicated into the water. The experiments showed that even when the waves were of such small wave-length, as 3 or 4 cm. only, there was a definite disturbance of pressure at the bed of the tank and that the disturbance takes only a fraction of a second to reach a depth of one metre. The manner in which the disturbance of pressure decreased up to a certain depth and then increased clearly suggested that the compressibility of water, which is neglected in hydrodynamical equations, plays a very important part in the communication of disturbance to various depths.

That the compressibility of water plays a very important part in the propagation of waves in water becomes abundantly clear from many experiments. If in a liquid enclosed by fixed boundaries, an arbitrary variation of pressure is caused at some point of the boundary, then the pressure at all points rises or falls by equal amounts, the disturbance being propagated with the velocity of sound. It is this compressibility of water which has made the sound echo method successful in the sounding of depths of sea.

Accordingly the theory of microseisms has been worked out by introducing into the usual hydrodynamical equations the terms involving compressibility. It will be seen that the amplified theory gives a satisfactory explanation of all the observed facts.

Since the previous paper was written, two Milne-Shaw seismographs have been installed at Calcutta and at an inland station, Agra, for the purpose of recording microseisms, and particularly to find out the effect of travel on microseisms over extensive land areas. While the magnification of the N-S and E-W components at Bombay were 250 and 350 respectively, and that of N-S component at Calcutta 250, the Agra instrument, which is about 700 miles away from the nearest sea-coast, had to be maintained at a magnification of about 650 times in order to have the recorded amplitudes of microseisms comparable with those at Calcutta and Bombay. With this arrangement, it was found that the microseisms due to storm either in the Bay of Bengal or in the Arabian Sea were recorded by instruments at the three stations. When a storm is in the Bay of Bengal, the Calcutta instrument records microseisms of a complex type, with irregular variations of amplitude, but at Bombay and Agra, the minor complexities are wiped out as a consequence of distance of travel, but the irregular variations of amplitude remain. Similarly when there is a storm in the Arabian Sea, the
Bombay records are of complex type with irregular variations in amplitude, while in the Calcutta and Agra records the minor complexities are wiped out but the irregular variations of amplitude remain. The local geological conditions differ widely at Bombay, Calcutta and Agra and the observations show that their effect on microseisms is of a minor nature. The minor complexities are wiped out at distant stations as a pure effect of dissipation owing to distance of travel. Typical records of microseisms at Calcutta, Bombay and Agra due to disturbed weather in the Bay of Bengal have been given in Plate XXVI.

2. "True" and "False" Microseisms.

It does not appear to be realised by every worker that unless extreme care is taken in the installation and the working of a seismograph, the records of microseisms may very easily be faulty. Quick-run records† of gusts of wind indicate that these generally have the same periods as the microseisms (Fig. 2). Consequently these constitute very important disturbing factors in the case of most seismographs.

† "The Structure of Wind over Level Country," London M. O. Geophysical Memoirs, 1932, 54.
How a seismograph can be profoundly affected by rapid pressure variations in the air was made vividly clear to the present writer when he observed some years ago that each one of the 31 gun-fires made in salutations of the outgoing and incoming Viceroy respectively from a battery couple of hundred yards away produced definite oscillations (Fig. 3) in two most carefully installed seismographs in an underground room in Bombay, calculated to be amply protected from sudden pressure fluctuations in the air. That the impulses recorded by the seismographs were due to actual pressure pulses reaching the instruments and not to reaction with the ground of the salutation guns, which are mounted on wheels, is confirmed by observations of
artificial vibrations* produced in the ground. Such vibrations do not ordinarily penetrate more than 4 or 5 feet below the surface and their amplitude diminishes rapidly with distance.

Later on experiments were made with seismographs which were insufficiently protected from the effects of gusts of wind and these showed that

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under such conditions the microseisms recorded were largely "false" micro-
seisms produced by gusts of wind. Figs. 4a and 4b give the records ob-
tained with two seismographs (magnifications of ground movements: 275 and
1000 respectively) installed in an over-ground room, with thick wooden
walls, on pillars isolated from the floor of the room. The variation in the
intensity of microseisms in both cases followed the variation in the intensity
of the wind and the periods of microseisms agreed with the periods of the
gusts of wind. The effect of pressure fluctuations which have direct access
to a seismograph should not be confused with the effect of free vibrations
which might be excited in buildings or trees. The latter has periods less
than 0.1 sec., but the former has larger periods and behaves in the same way
as the slow application of a pressure on the seismograph pillar, such as a
weight slowly added to and withdrawn from the pillar.

It is clear that if we are to protect a seismograph from the effect of
pressure fluctuations due to gusts of wind, we must provide sufficient "dead" air all round it.

The best site for the installation of a seismograph is clearly a double-
walled underground room. The "dead" air provided in this way in the
channel round the instrument room acts as a damper to the communication
of pressure fluctuations. It is, however, doubtful whether the provision
of a single layer of "dead" air in this way completely eliminates the effects
of pressure fluctuations. In the Colaba Observatory, the two Milne-Shaw
seismographs are installed with elaborate precautions as shown in the above
diagram and in spite of this they recorded the salutation gun-fires. That
the provision of the underground "dead" air space had a considerable
damping effect on the pressure fluctuations was, however, made perfectly
clear by installing temporarily a Milne-Shaw seismograph in a single-walled overground room.

It does not appear that as a rule adequate precautions are taken in observatories for the protection of their seismographs against rapid pressure fluctuations in the air. The microseisms recorded appear in many cases to show unmistakable signs of gust effect. The effect of gusts which is mixed up with true microseisms is often underestimated.

3. Effect of geological structure on microseisms.

A. W. Lee* has recently studied in some detail the effect of geological structure on microseisms. He employs Love's analysis for a superficial layer with a view to show that the horizontal amplitudes of waves having the period of microseisms are affected more than the vertical amplitudes by such a layer and that their ratio depends upon the wave-length (or periods) and upon the composition and thickness of the layer. Even with the most favourable assumption relating to the composition of the layers he could not get the ratio of the horizontal to the vertical amplitudes to be greater than 4 : 3, whereas in many observatories, the observed ratio is as great as 3 : 1. Moreover at Kew and Abisko, the ratio was found to show practically no variation with periods and to show a very small variation at De Bilt only, agreeing very crudely with theory, which might be purely an accident. Lee's analysis is applicable only to layers which are of infinite horizontal extent. In practice it is made applicable to layers which vary from locality to locality and are of extent small in comparison with or nearly equal to the wave-length of Rayleigh waves. This appears to be unsound.

In analysis of this type we have to remember that there is an enormous increase of pressure and temperature with depth. The properties of the discontinuities under such conditions would be quite different from those in an isotropic non-gravitating elastic solid. Even if we ignore the variation of temperature, we have a fundamental difficulty. In fact, Love¹ says,

"A body in equilibrium under the mutual gravitation of its parts is in a state of stress, and when the body is large the stress is enormous. The Earth is an example of a body which must be regarded as being in a state of initial stress, for the stress that must exist in the interior is much too great to permit of the calculation by the ordinary methods, of strains reckoned from the unstressed state as unstrained state." Chree² emphasised the

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² Chree, Phil. Mag., 1891, Ser. 5, 32.
difficulty and pointed out that in the case of the earth, the internal stress is much too great to permit of the direct application of the mathematical theory of superposable small strains. One way of evading this difficulty is to treat the material of which the earth is composed as homogeneous and incompressible. Later on, Lord Rayleigh devised a method for dealing with the complication, which was as follows:

"The earth should be considered to be a body in a state of initial stress; this initial stress may be regarded as a hydrostatic pressure balancing the self-gravitation of the body in the initial state. The stress in the body when disturbed, may be taken to consist of the initial stress compounded with an additional stress. The additional stress may be taken to be connected with the strain, measured from the initial state as unstrained state by the same formulæ as hold in an isotropic elastic solid body slightly strained from a state of zero stress." There is a certain ambiguity in this theory, but according to Love, the initial stress at a point of the body which is at \((x, y, z)\) in the strained state should be the pressure in the initial state at that point which is displaced to \((x, y, z)\) when the body is strained. For the investigation of transmission of waves through a superficial layer, Love accordingly takes the equation of motion to be of the form

\[
\rho \frac{\partial^2 u}{\partial t^2} = -\frac{\partial P}{\partial x} + \mu \nabla^2 u; \quad \rho \frac{\partial^2 w}{\partial t^2} = -\frac{\partial P}{\partial z} + \mu \nabla^2 w
\]

where \(P\) denotes a hydrostatic pressure. Love, however, assumes the material to be incompressible. Lee, on the other hand, assumes the material to be compressible, but neglects the hydrostatic pressure, and takes the equation of motion to be of the form

\[
\rho \frac{\partial^2 u}{\partial t^2} = (\lambda + \mu) \frac{\partial \Delta}{\partial x} + \mu \nabla^2 u; \quad \rho \frac{\partial^2 w}{\partial t^2} = (\lambda + \mu) \frac{\partial \Delta}{\partial z} + \mu \nabla^2 w
\]

\(\Delta\) being the dilatation.

If gravity and initial stress were taken into consideration, a correction will be required in Lee's results but it appears from Love's investigation of the form of this correction that it will be small. It would thus appear that geological formation would not explain the observed ratio of the horizontal and vertical amplitudes, which varies from 0.6 to 3.

On the other hand, working on the theory that microseisms are due to the disturbance of pressure at the bed of the sea, it was shown by me\(^5\)

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Love, Some Problems of Geodynamics, 1911, p. 89.)

4 Some Problems of Geodynamics, 1911, p. 165.

5 Phil. Trans. Roy. Soc. (A), 1930, 229, 316.
that the ratio of the horizontal to vertical displacements of microseisms will be 3:1 in the bed of the sea and its neighbourhood (i.e., stations near the sea-coast), and that this ratio should decrease as we proceed inland to 0.7:1, in very close agreement with the observed ratios (vide Appendix). The geological formation will slightly augment this ratio at inland stations.


Any disturbing factor which will give rise to persistent vibrations in a seismograph will be recorded as microseisms. It has already been indicated that if pressure variations due to gusts of wind have direct access to a seismograph, they will produce vibrations which will have all the characteristics of microseisms. On the other hand, even if a seismograph has been adequately protected from the effects of gusts of wind, these may shake the buildings in which it is housed or the neighbouring trees. The vibrations which will be communicated to the ground in this way will, however, be of short periods, less than 0.1 sec., and cannot therefore be confused with microseisms.

How are then these world-wide vibrations produced? There are six important possibilities:—

(1) Pressure variations due to sea waves over shallow water.
(2) Surf breaking along the coast line and transferring the energy of the waves into the ground.
(3) Effect of strong winds on the uneven surface of the earth.
(4) Setting up of natural free vibrations of tracts of country by breakers.
(5) Some kind of emanation of energy from the interior of the earth setting the surface into continuous vibrations.
(6) Communication of the disturbance of pressure due to sea waves to the bed of a sea by some mechanism and the production of forced elastic oscillations which travel as microseisms.

Besides these there are several less important possibilities but we need not discuss them here.

In regard to (1), it has been pointed out by me\(^6\) that sea waves over shallow water have usually large periods, varying from 10 to 30 secs. The disturbance of pressure produced by such waves cannot therefore explain the production of microseisms of periods 4 to 10 secs. Such waves will, however, produce microseisms of large periods.

Although Gutenberg\(^7\) has discussed the order of magnitude involved in the process of impact of "breakers" on steep rocky coasts, and concluded that the energy transferred to the coast by them is large enough to cause microseisms, the conditions postulated do not exist along all coast lines. Undoubtedly such breakers will produce ground movements, but the non-simultaneous action of innumerable breakers along a coast line will produce movements of very complex type. Moreover, if we compare the movements which will be produced by such impacts with the artificial vibrations\(^8\) produced in the ground by dropping a weight, we can conclude that the vibrations produced by an individual breaker will scarcely have periods exceeding a second or two unless such breakers are able to set a large tract of country into free vibrations. Such a process contemplates a line source of energy which though large in total is, nevertheless, very small per unit length, and cannot therefore cause appreciable microseismic movements in the transverse direction. While therefore at stations near such coast line such impacts will explain the superposed small period movements, they cannot be the sole cause of the movements of periods 4 to 10 secs. over extensive areas.

With regard to the question whether microseisms could be produced by the effect of strong winds on the uneven surface of the earth, I have already pointed out that the vibrations produced in a building have periods less than 0.1 sec. The period, however, depends on the dimensions of the solid on which the wind is acting. If therefore wind is able to set a ridge (say 10 km. long, 2 km. broad and 200 metres high) into oscillations, its periods may be comparable with those of microseisms. But if the rigidity of such a ridge be very great, it is doubtful whether the small fluctuating force due to wind will be able to cause appreciable vibration. Darwin\(^9\) has shown that if the difference of barometric pressure between consecutive regions of high and low pressures be 5 cm. of mercury and if the centre of the "high" and "low" be 1500 miles apart, then, as a consequence of the yielding of the ground, it will be 9 cm. higher under the barometric depression than under the elevation. If therefore we consider the small period oscillations of barometric pressure produced by gusts of wind to be of amplitude \(10^{-2}\) cm. of mercury, and wave-length 1/10th of a mile, the amplitude of the movements produced in the ground will be \(1.2 \times 10^{-6}\) cm. (using Darwin's formula and assuming the rigidity of ground to be equal to \(3 \times 10^{11}\) C.G.S.

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\(^8\) Banerji, "On the Artificial Vibrations of Ground," *Ind. Journ. of Phys.*, 1933, 8, Part II.

\(^9\) *Scientific Paper*, 1, 448.
units). This is considerably smaller than the observed amplitude of microseisms. The amplitude will be comparable with the amplitude of microseisms only if the rigidity is as low as $10^8$ C.G.S. units. In certain localities, the rigidity may be as low as this, but cannot be so everywhere. It would thus appear that in localities with soft ground or clay, microseisms may be produced by pressure fluctuations caused by the gusts of wind. The Agra instrument which was maintained at a large magnification recorded microseisms of large periods during the passage of "western disturbances" over Northern India in winter and these were apparently due to the action of wind on the uneven surface in its neighbourhood (including Rajputana).

Wiechert suggested long ago that the existence of definite periods in earthquake records may be due to the setting up by the shock of the natural free vibrations of tracts of country. Knott pointed out, after Omori, that the records of Japanese earthquakes always showed a preponderance of vibrations of periods, 4-6 seconds, in the preliminary tremors, and that these were also the common periods of microseisms observed at Tokyo, and accordingly suggested that this was a period of vibration natural to the plain in which Tokyo lies. Wadate supposed that the free vibrations were excited by the impact of sea waves on the coast. The free vibrations of tracts of country would be different in different parts of the world, and if microseisms were due to this cause, they would show widely different characteristics in different parts of the world. This is, however, not found to be the case. In localities where free vibrations of certain periods can be excited by breakers, microseisms having those periods would undoubtedly be recorded, but this would hardly seem to be the cause of world-wide microseisms.

Our knowledge about the physical conditions of the earth's interior and also of the material which composes it is still so meagre that it is not possible to make any definite pronouncement about the possibility of the earth's surface being set into vibrations by the emanation of some kind of energy (thermal, radio-active, etc.) from its interior. But it seems unlikely that if there were vibrations due to such a cause, their nature would have escaped detection.

None of the hypotheses discussed above will explain the production of the special type microseisms which are associated with storms in the mid-Arabian Sea or mid-Bay of Bengal, far away from the coast and with storms over sea areas in other parts of the world. In such cases the conclusion is inevitable that there must be some mechanism by which the disturbance of pressure over the surface of the sea produced by the storms is
communicated to its bed. To understand this mechanism an analysis was made, both theoretically and experimentally, of the influence of (1) eddy viscosity and (2) compressibility, in the transmission of disturbance of pressure to the bed.

We will discuss these two factors separately.

5. Disturbance of pressure at the bed of the sea.
   Effect of eddy viscosity.

The best method of obtaining a solution to this question is to find out how a velocity communicated to the surface of water would be propagated downwards. The problem was first discussed by Zöppritz. Let the x-axis be drawn vertically downwards and y-axis in a horizontal direction at the surface. Now let V be the velocity of a horizontal current suddenly communicated to the surface water. The velocity \( v \) at any point will satisfy the differential equation

\[
\frac{\partial v}{\partial t} = \frac{1}{\rho} \frac{\partial}{\partial x} \left( K \frac{\partial v}{\partial x} \right)
\]

subject to the condition

\[
\begin{align*}
  v &= V, \text{ at } x=0, \ t > 0, & \quad (2) \\
  v &= 0, \text{ at } t=0, \ x > 0, & \quad (3)
\end{align*}
\]

If we consider the eddy viscosity \( K \) and the density to be constant, the solution can be expressed in the form

\[
v = V \left(1 - \frac{2}{\sqrt{\pi}} \int_{0}^{x} e^{-\xi^2} d\xi\right)
\]

From this we can see that the time required for attaining half the surface velocity at a depth \( x \) cm. is given by

\[
\frac{x}{2} \sqrt{\frac{\rho}{Kt}} = 0.48
\]

and \( \frac{1}{10} \)th of the surface velocity by

\[
\frac{x}{2} \sqrt{\frac{\rho}{Kt}} = 1.16.
\]

Taking \( \rho = 1 \), and \( K \) to have the value of molecular viscosity \( =0.144 \text{ C.G.S.} \), we get \( t = 239 \) years for \( x = 100 \) metres. This is the value given by Zöppritz. This result means that after 239 years the disturbance at the depth of 100 metres will reach half the surface value. Assuming the relationship (5) to

\footnote{K. Zöppritz, "Hydrographische Probleme in Beziehung zur Theorie der Meeresstromung," Wiedemanns Annalen der Physik und Chemie, Ne Folge, Bd. 3, 582-608.}
remain approximately true even in a case when the surface of the sea is highly disturbed and turbulent in character, and therefore K to have large values of the order of $10^4$ C.G.S. units, we see that the disturbance at a depth of 100 metres will reach half the value at the surface after an interval of about 2 hours and 50 mins. A smaller value than this will of course be attained much earlier. For instance 1/10 of the surface disturbance will be attained at the depth of 100 metres after an interval of 30 mins. It will thus be seen that the disturbance will take at least 2 or 3 hours to reach the bottom of the deep sea (1000 or 1500 metres deep) with appreciable intensity. The eddy viscosity therefore plays a comparatively minor part in the transmission of disturbance of pressure to the bed of the sea.

6. Disturbance of pressure at the bed of a sea.—Effect of compressibility.

(a) Experiments in tanks in imitation of the conditions in the sea.—Experiments were made in a galvanized iron tank, 162 cm. × 132 cm. × 84 cm., in a rectangular masonry tank 190 cm. × 190 cm. × 95 cm., and finally in a circular masonry tank of diameter 210 cm. and depth 108 cm. with walls about 50 cm. thick. Waves were generated on the surface of water by means of a small electrical vibrator, in which a small metal blade, 10 cm. × 5 cm., was fixed. The vibrations were excited by a pair of electromagnets, the periods being altered by altering the length of the vibrator. The blade was made to dip only 2 or 3 cm. into the water. Experiments were usually made with waves of length 2 to 6 cm.

The disturbance of pressure at various depths was recorded by sinking vertically downwards a rigidly supported tube of diameter 4 cm. A small quantity of coloured oil (iodine dissolved in kerosene) was poured into the tube so as to bring the upper surface above the general surface of water and make it visible. The communication of disturbance of pressure into the tube is clearly through its lower end, and before an observation was taken a test was always made to see that the oil surface remained undisturbed when the lower end of the tube was closed. With a tube of diameter 4 cm. and over, the periods of the oil column were always found to be the same as the periods of the waves outside.

The oscillations in the oil column, when the tube was sunk to various depths, were photographed. The records for one particular positions of the tube and vibrator are shown in Plates XXVII and XXVIII. The results obtained for other positions are similar in character. It will be seen that the disturbance, starting from the maximum value at the surface, diminishes up to a certain depth and then increases to another but lower maximum at the bed.

The curves in Fig. 6 show the manner in which the disturbance of pressure decreases with depth. The minimum disturbance of pressure at the depth
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Fig. 6. Masonry Tank (180 cm. x 190 cm. x 95 cm.)
Wave-Length : 6 cm. Vibrator : 18 cm. from centre of one edge.

Table 1.
Tube in centre of circular tank.

<table>
<thead>
<tr>
<th>Distance of Vibrator from tube</th>
<th>Depth of tube</th>
<th>Time</th>
<th>Distance of vibrator from tube</th>
<th>Depth of tube</th>
<th>Time</th>
</tr>
</thead>
<tbody>
<tr>
<td>cm.</td>
<td>cm.</td>
<td>secs.</td>
<td>cm.</td>
<td>cm.</td>
<td>secs.</td>
</tr>
<tr>
<td>97.5</td>
<td>25.0 - 48.0</td>
<td>5.5</td>
<td>15.0</td>
<td>17.5 - 41.0</td>
<td>1.0</td>
</tr>
<tr>
<td></td>
<td>96.5</td>
<td>7.5</td>
<td></td>
<td>63.5 - 96.5</td>
<td>1.2</td>
</tr>
<tr>
<td>76.0</td>
<td>25 - 70</td>
<td>5.0</td>
<td></td>
<td>17.5 - 34.5</td>
<td>1.0</td>
</tr>
<tr>
<td></td>
<td>100</td>
<td>5.4</td>
<td>7.5</td>
<td>56.0 - 82.5</td>
<td>1.2</td>
</tr>
<tr>
<td>40.0</td>
<td>19 - 69</td>
<td>2.0</td>
<td></td>
<td>105.3</td>
<td>1.4</td>
</tr>
<tr>
<td></td>
<td>100</td>
<td>2.5</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
of 50 cm. is clearly due to the interaction of the waves travelling downwards and those reflected upwards from the bed of the tank. The experiment therefore clearly shows that the compressibility of water plays very important part.

Starting the vibrator at definite instants the time taken by the disturbance to reach various depths was carefully noted by means of a stopwatch. These are given in Table I.

As it was not possible to secure absolute rest for the water in the tank, before each observation commenced, no great accuracy is claimed for the figures noted in the above table. The error in noting time was roughly ± 0.2 sec. The liquid in the tube was found to show no appreciable oscillations as long as the surface waves emanating from the vibrator did not reach the centre of the tank. But the moment they reached the centre, the disturbance travelled in a fraction of a second to the depth to which the tube was sunk. It would thus appear that there is no downward transmission of disturbance below a point on the surface until the fluctuation in the elevation has commenced at the point. Calculation shows that in the case of a disturbance produced by an electric vibrator of the type used in the above experiment, the eddy viscosity is approximately 1 C.G.S. unit. Consequently, the theoretical formula would indicate that the disturbance will reach half the surface value at the depths given below after the time stated against them.

<table>
<thead>
<tr>
<th>Depth (cm)</th>
<th>Time (sec)</th>
</tr>
</thead>
<tbody>
<tr>
<td>10</td>
<td>100</td>
</tr>
<tr>
<td>25</td>
<td>625</td>
</tr>
<tr>
<td>100</td>
<td>10,000</td>
</tr>
</tbody>
</table>

On the other hand, 1/10 of the surface values will be reached at the following depths at the times stated against them.

<table>
<thead>
<tr>
<th>Depth (cm)</th>
<th>Time (sec)</th>
</tr>
</thead>
<tbody>
<tr>
<td>10</td>
<td>19</td>
</tr>
<tr>
<td>25</td>
<td>116</td>
</tr>
<tr>
<td>100</td>
<td>1,852</td>
</tr>
</tbody>
</table>

When we compare these figures with the results of observations given in Table I, we can conclude that the disturbance is transmitted to various depths not by eddy viscosity but by some other factors.

(b) **Forced oscillations.**—In the above experiment, there is one possibility, which requires examination, namely, the possibility of the whole
water in the tank being set into forced oscillations. The energy communicated by a small blade of breadth 4 cm., dipping 2 or 3 cm. into water, is so small that it appeared unlikely that it could set the water in the tank as a whole into forced oscillation. In any case, certain test experiments were performed and these do not indicate that the system as a whole was set into forced oscillations.

If a cylinder of galvanized iron, CC, be fixed inside the tank as shown in Fig. 7, and if observations be taken inside and outside with two tubes T₁ and T₂, rigidly supported from the same frame, and sunk to the same depth, then the oil column in tube T₂ shows conspicuous oscillations, but no perceptible oscillations can be observed in the oil column in tube T₁, whatever be the depth to which the tubes are sunk.

If on the other hand, we fix a plate, PP, horizontally, 8 or 10 cm. below the water surface and sink a tube T₁ through a hole in its centre to a small depth below its surface, minute oscillations can be observed in the oil column in the tube, but the amplitude of these is a very small fraction of those of the oscillations in an outside tube sunk to the same depth (Fig. 8). The minute oscillations in tube T₁ appear to be due to diffraction of pressure disturbance round the plate, PP.

In my experiments with self-recording evaporimeter the float to which recording mechanism was attached worked inside a guide placed inside the evaporation tank 4' x 3' x 2½'. Although the recording mechanism was amply protected from the direct action of the wind, the ripples produced on the surface of water by wind caused oscillations in the float even though

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the communication with the water in the tank was by means of a small opening at the bottom of the guide. When the tank was covered by an iron plate so as to cut off the action of the wind on the water surface, no oscillations in the float were recorded. This can be seen from the record reproduced in Fig. 9. The records clearly suggest that the disturbance of pressure due to the ripples communicated to the bed of the evaporation tank and finds access into the guide through the small opening at its bottom.

(c) Analysis based on analogy.—In the first place, let us adopt an approximate treatment. If we refer to Fig. 1, and confine our attention to the superficial layer, and assume that the layer below is non-existent, then in consideration of the small depth of this layer, the waves may be regarded to be virtually the same as "long" gravity waves. The equations of these waves have therefore the form
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\[ \frac{\partial u}{\partial t} = -g \frac{\partial \eta}{\partial x}, \quad \eta = -h \frac{\partial \xi}{\partial x} \quad \ldots \quad \ldots \quad (6) \]

where \( \xi = \int u\,dt \) and \( \eta = \) elevation.

We may compare the equations (6) with equation of plane waves, namely,

\[ \frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} = -\frac{1}{\rho} \frac{\partial p}{\partial x} = -\frac{1}{\rho} \frac{\partial \rho}{\partial x} \frac{\partial \rho}{\partial x} \]

and the equation of continuity

\[ \frac{\partial \rho}{\partial t} + \frac{\partial (\rho u)}{\partial x} = 0, \]

which become

\[ \frac{\partial u}{\partial t} = -\frac{K}{\rho} \frac{\partial s}{\partial x} \quad \text{and} \quad \frac{\partial s}{\partial t} = -\frac{\partial u}{\partial x} \quad \ldots \quad \ldots \quad (7) \]

on writing \( \rho = \rho_0 (1+s) \), \( s \) being the condensation, and

\[ K = \left[ \rho \frac{\partial \rho}{\partial \rho} \right] \rho = \rho_0 \]

Equation (6) becomes the same as equation (7) if we write

\[ \frac{\eta}{h} = s, \quad gh = \frac{K}{\rho_0} = c^2 \quad \ldots \quad \ldots \quad \ldots \quad (8) \]

\( \eta/h \) in the superficial layer thus behaves in the same as "condensation". The layer "B" transmits to the bed of the sea this "condensation wave" in the same way as sound wave is transmitted through water. Since for long waves the disturbance of pressure is \( g\rho \eta \) on plane \( X' \), this is transmitted by the compression effect to the bed of the sea almost undiminished in intensity.

We will now give a more exact treatment.

(d) More exact analysis.—In making an analysis of the effect of compressibility we have to distinguish between two important features, namely, the production of gravity waves by wind over a thin stratum and their transmission downwards by the effect of compressibility. If the \( z \)-axis be drawn vertically downwards, and the axes of \( x \) and \( y \) on the surface of water, then the gravity waves are determined by the equations

\[ \frac{\partial^2 \phi}{\partial x^2} + \frac{\partial^2 \phi}{\partial y^2} + \frac{\partial^2 \phi}{\partial z^2} = 0, \quad \text{when } z = 0, \]

\[ \frac{\partial \eta}{\partial t} = -\frac{\partial \phi}{\partial x}, \quad \text{when } z = 0, \]

and \( \frac{\partial \phi}{\partial \eta} = 0, \quad \text{when } z = d. \]

\[ \ldots \quad \ldots \quad (9) \]
φ being the velocity potential and η the elevation and d the depth of the sea. The periods T, wave-length λ and the velocity V of these waves are, according to the usual solution of the above equations, connected by the relationship

\[ V = \frac{\lambda}{T} \left( \frac{g \lambda}{2\pi} \operatorname{tanh} \frac{2\pi d}{\lambda} \right) \]  

and this is equal to simply \((g\lambda/2\pi)^{\frac{1}{2}}\) in the case of deep sea.

This velocity is considerably smaller than the velocity of compressional waves in water. The disturbance is confined to a superficial stratum and we have to find out how compressibility transmits it downwards. When compressibility is taken into account, the equations of motion are:

\[ \rho \frac{\partial^2 u}{\partial t^2} = \sum \left\{ \rho \epsilon^2 \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} \right) + \rho (Xu + Yv + Zw) \right\} - X \left\{ \frac{\partial (\rho u)}{\partial x} + \frac{\partial (\rho v)}{\partial y} + \frac{\partial (\rho w)}{\partial z} \right\} \]  

and two similar equations, where

\[ \frac{D\rho}{Dt} = c^2 \frac{Dp}{Dt} \]  

and X, Y, Z are the components of extraneous force, c is the velocity of sound in water and is approximately equal to 1430 metres/sec. It is quite different from V, the velocity of gravity waves. If X, Y, Z have a potential, the equilibrium pressure will be a function of \(\rho_0\), that is to say,

\[ \hat{p}_0 = f(\rho_0) \]

Equation (11) therefore reduces to the form

\[ \frac{\partial^2 u}{\partial x^2} = \epsilon^2 \left\{ \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} \right\} + Xu + Yv + Zw \]

The disturbed motion is therefore not in general irrotational, but if (12) reduces to the form*

\[ c^2 = \frac{d\rho}{d\rho} = f'(\rho), \]

then the equation becomes

\[ \frac{\partial^2 u}{\partial x^2} = \frac{\partial}{\partial x} \left\{ \epsilon^2 \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} \right) + Xu + Yv + Zw \right\} \]  

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* This will be so if the undisturbed state be one of convective equilibrium or of uniform temperature.
The three equations of type (14) are satisfied by
\[ u = -\frac{\partial \Phi}{\partial x}, \quad v = -\frac{\partial \Phi}{\partial y}, \quad w = -\frac{\partial \Phi}{\partial z} \]
provided
\[ \frac{\partial^2 \Phi}{\partial t^2} = c^2 \nabla^2 \Phi + \left( X \frac{\partial \Phi}{\partial x} + Y \frac{\partial \Phi}{\partial y} + Z \frac{\partial \Phi}{\partial z} \right). \tag{15} \]
In the present case \( X = 0, \) \( Y = 0, \) and \( Z = g, \) and therefore \( \Phi \) satisfies the equation
\[ \frac{\partial^2 \Phi}{\partial t^2} = c^2 \left( \frac{\partial^2 \Phi}{\partial x^2} + \frac{\partial^2 \Phi}{\partial y^2} + \frac{\partial^2 \Phi}{\partial z^2} \right) + g \frac{\partial \Phi}{\partial z} \tag{16} \]
For solving this equation, we assume
\[ \Phi = \cos (ct - lx - my) e^{\lambda z} \tag{17} \]
and we get on substitution
\[ \lambda^2 + \frac{g}{c^2} \lambda + (1 - l^2 - m^2) = 0 \tag{18} \]
Therefore
\[ \lambda = -\frac{g}{2c^2} \pm \frac{1}{2} \left[ \frac{g^2}{c^4} - (4 - 4l^2 - 4m^2) \right]^\frac{1}{2} \tag{19} \]
Since \( g^2/c^4 \) is of the order \( 10^{-14}, \) it is very small in comparison with \((4 - 4l^2 - 4m^2).\) Therefore
\[ \lambda = -\frac{g}{2c^2} \pm i (1 - l^2 - m^2)^\frac{1}{4}. \tag{20} \]
As a consequence of this approximation, the solutions given below are also approximate solutions and not the true solutions. We can write
\[ \Phi = e^{-\frac{g}{2c^2} z} \cos (ct - lx - my) \times \left[ A \cos (1 - l^2 - m^2)^\frac{1}{4} z + B \sin (1 - l^2 - m^2)^\frac{1}{4} z \right]. \tag{21} \]
\( \Phi \) thus assumes any of the forms
\[ e^{-\frac{g}{2c^2} z} \cos [ct - lx - my + (1 - l^2 - m^2)^\frac{1}{4} z + \epsilon] \]
\[ e^{-\frac{g}{2c^2} z} \cos [ct - lx - my - (1 - l^2 - m^2)^\frac{1}{4} z + \epsilon'] \]
\[ e^{-\frac{g}{2c^2} z} \cos [ct + lx + my + (1 - l^2 - m^2)^\frac{1}{4} z + \epsilon''] \]
\[ e^{-\frac{g}{2c^2} z} \cos [(ct + lx + my) - (1 - l^2 - m^2)^\frac{1}{4} z + \epsilon'''] \tag{22} \]
In the case of standing vibrations on the surface of the sea, \( \Phi \) will be of the form
All these expressions represent waves travelling with the velocity $c$ towards the bed of the sea or waves reflected from the bed of the sea and travelling upwards. Identifying these expressions with the expressions for gravity waves at the surface, we see that \(2\pi/l\) and \(2\pi/m\) represent the wave-lengths of the gravity waves in the directions of $x$-axis and $y$-axis respectively. $l$ and $m$ are therefore less than unity.

To satisfy the boundary condition at the bed of the sea, we take $\Phi$ to have the form

\[
\Phi = e^{-\frac{g}{2c^2} z} \left[ A \cos lx \cos my \cos \left(1 - l^2 - m^2\right)\frac{1}{2} z + \sin \left(1 - l^2 - m^2\right)\frac{1}{2} z \right] \cos (ct + a) \ldots (24)
\]

Since $g/2c^2$ is of the order $10^{-7}$, $e^{-\frac{g}{2c^2} z}$ will become $e^{-1}$ at a depth of 100,000 metres. We thus see that $e^{-\frac{g}{2c^2} z}$ undergoes very slow decrease with depth. For depth up to 5,000 metres, we can take it to be approximately equal to unity. At the bed of the sea ($z = d$), the normal velocity should be zero. We therefore get

\[
A \sin \left(1 - l^2 - m^2\right)\frac{1}{2} d - B \cos \left(1 - l^2 - m^2\right)\frac{1}{2} d = C \ldots (25)
\]

Therefore

\[
\Phi = C \cos (ct + a) \cos lx \cos my \left[ \frac{\cos \left(1 - l^2 - m^2\right)\frac{1}{2} z}{\sin \left(1 - l^2 - m^2\right)\frac{1}{2} d} \right] e^{-\frac{g}{2c^2} z} \ldots (26)
\]

The expression for the pressure is

\[
\frac{p}{\rho} = \frac{\partial \Phi}{\partial t} + g z - \frac{1}{2} \left[ \left(\frac{\partial \Phi}{\partial x}\right)^2 + \left(\frac{\partial \Phi}{\partial y}\right)^2 + \left(\frac{\partial \Phi}{\partial z}\right)^2 \right] + \text{const.} \ldots (27)
\]

Neglecting as usual the square of the velocity and noting that the elevation $\eta$ is given by

\[
\eta = \frac{1}{g} \left(\frac{\partial \Phi}{\partial t}\right)_{z = 0},
\]

we see that the pressure at the bed of the sea can be expressed in the form

\[
\frac{p}{\rho} = g \rho d + \frac{g\rho \eta}{\cos \left(1 - l^2 - m^2\right)\frac{1}{2} d} \ldots (28)
\]

We thus see that there is a finite disturbance of pressure at the bed of a deep sea as a result of compressibility of water and that this disturbance
of pressure is proportional to $g \rho h$, exactly as postulated by a rough analysis in the paper in *Phil. Trans. Roy. Soc.*, 1930, 229A, 287–328.

7. **Effect of Resonance.**

If we consider a vertical column of water from the surface right down to the bed of the sea, we can compare the motion in this column with that in an organ pipe. In a column like this, the upper end is an open end in which forced oscillations of certain definite periods are maintained by wind. At the bed of the sea we have a closed end. If the period of oscillations at the open end be such that it will correspond to an anti-node at the bed instead of a node, when the column is regarded as an organ pipe, we will clearly get the maximum disturbance of pressure at the bed. With this analogy, the period is given by

$$\frac{\text{Depth of sea}}{\text{Velocity of sound in water}}$$

The average depth of the Arabian Sea and the Bay of Bengal is about 1,500 metres; so that the bed of the sea will be an anti-node if the period of the waves is about a second. On the other hand, the bed will be a node, if the period of the waves is about 2 seconds. For gravity waves of periods 4 to 10 seconds, there is therefore no resonance in a sea of this depth, and there is no node or anti-node either at the bed or anywhere in the column. As shown in the previous article there is only a slow decrease of motion with depth according to the law $e^{-\frac{g}{2c^2}z}$.

8. **Explanation of Discrepancies.**

Some observers have noted that two storms of similar nature and of more or less the same barometric depression and occupying very nearly the same position in the sea do not always produce microseisms of the same intensity at an observing station. This would constitute a real discrepancy, if sea-waves were the only factors which would produce movements in the pillar on which seismographs are mounted at a distant station. I have already pointed out that there are many factors other than sea-waves which might produce movements in such a pillar. The pressure pulses produced by gusts of wind are the most effective and these have the same periods as microseisms and are therefore easily confused with microseisms. Unless therefore an instrument has been set up so that it will record only the movements which are produced by sea-waves and no others, no comparison of the movements produced by two similar storms is possible. The criteria for this are:—
(1) that the instrument should be so installed that no microseisms are recorded when the sea over a distance of say 1,000 miles all round (depending on the critical sensitiveness adopted) is in a normally undisturbed state.

(2) when the sea becomes disturbed anywhere over this area, microseisms should be immediately recorded.

The two Milne-Shaw seismographs installed in the underground constant temperature room in the Colaba observatory very nearly satisfied the above conditions as indicated in the paper already referred to and no discrepancy of the above kind was noted. With storms of more or less the same intensity and also with the same storms the amplitude of microseisms decreased with the distance of the centre of the storm according to a law which should be expected theoretically.

The discrepancies noted by other observers are probably to be attributed to the fact that their seismographs did not satisfy the above two conditions.

Summary.

In a previous paper on the subject, it was pointed out that storms in the mid-Arabian Sea and mid-Bay of Bengal produce microseisms of a definite type and these must therefore be due to communication of disturbance of pressure to the bed of the sea. In the analysis adopted in that paper, the sea was divided into two parts, a thin superficial stratum, in which the gravity waves were confined, and the water below this stratum, which transmitted the disturbance in the stratum by some mechanism. Since then experiments were made in tanks in imitation of the conditions in the sea in order to understand this mechanism and these show that the disturbance is transmitted by the effect of compressibility of water. A theoretical analysis of the effect of compressibility confirms the experimental results.

The various causes which might produce microseisms are discussed. The gusts of wind have the same periods as the microseisms; the pressure fluctuations due to these behave as a slow acting force on the seismograph pillar if these have direct access to the instrument. Observations show that it is not very easy to protect a seismograph from the effect of pressure fluctuations due to gusts of wind and most instruments therefore record "true" microseisms combined with "false" microseisms. The best situation for a seismograph seems to be an underground room in which a channel of "dead" air has been provided all round, which acts as a damper to pressure

fluctuations. When an instrument has received a critical adjustment so that it records no microseisms when the neighbouring seas are disturbed, but records them as soon as they are disturbed, the intensity of microseisms is found to be proportional to the intensity of the disturbances on the surface of the sea.

My thanks are due to Mr. S. S. Joshi for assisting me in taking observations of the disturbance of pressure in tanks.

APPENDIX.

Ratio of the amplitude of the horizontal to the vertical component of microseisms.

The ratio of the observed amplitude of the horizontal to the vertical component of microseisms varies from 0.6 to about 3, and as a rule the larger values hold for stations near the coast and the smaller values for inland stations.

All the theories of microseisms postulate the action of some pressure (wind, sea-waves and breakers) for the excitation of vibrations in the ground. If a load is placed on a horizontal plot of ground, so that it is equivalent to a pressure $P$ in the direction of $z$-axis drawn vertically downwards, the horizontal displacement is

$$\frac{P \sin \theta}{4\pi \mu r} \left( \cos \theta - \frac{\mu}{\lambda + \mu} \frac{1}{1 + \cos \theta} \right)$$

and the vertical displacement is

$$\frac{P}{4\pi \mu r} \left( \frac{\lambda + 2\mu}{\lambda + \mu} + \cos^2 \theta \right)$$

where $\theta$ is the angle the radius vector $r$ makes with the vertical.

When $\lambda = \mu$, the ratio of the horizontal to the vertical component of the displacement at a point on the surface, not too near the origin, is $1 : 3$. The observed ratio in the case of microseisms is never as low as this, so that the ratio given by the statical theory is too low.

If the movement has been recorded at a place which is far away from the region where a periodic force is acting, the waves will assume the characteristic of Rayleigh waves and the ratio of the horizontal to the vertical component will therefore be about 0.7. If the crust is stratified, then as shown by Lee, the largest value which can be obtained is about 4 : 3 for waves of periods 2$\pi$ secs. propagated through granite covered by a layer of clay 1 km. in thickness. This result is only applicable in the case where the layer of clay is of uniform thickness and of large horizontal extent, large in comparison with the wave-length of Rayleigh waves. Actually, however,
a layer of clay is seldom of uniform thickness and of large horizontal extent except in special localities. Consequently theory would suggest that the ratio should be less than $4:3$ and nearer the usual ratio for Rayleigh waves.

In the region where the pressure is acting or in its neighbourhood, the conditions are different. There the waves are forced waves. If the elastic solid is bounded by the plane, $z = 0$, and lies on the positive side of this plane, the displacements, as shown on pp. 306–307 of *Phil. Trans. Roy. Soc.*, 1930, 229A, can be expressed in the form.

$$
u = (A\xi e^{-\alpha z} - B\beta e^{-\beta z}) e^{\xi x} e^{\nu t},$$

$$w = (-\alpha A e^{-\alpha z} - i\xi B e^{-\beta z}) e^{\xi x} e^{\nu t},$$

where $A$ and $B$ are two arbitrary constants. When the pressure disturbance has the form

$$[Z_x]_0 = Pe^{\xi x} e^{\nu t}, \quad [X_x]_0 = 0,$$

$A$ and $B$ have the values

$$A = \frac{2\xi^2 - k^2 P}{F(\xi)} \mu, \quad B = \frac{2i\xi a P}{F(\xi)} \mu.$$

The pressure disturbance, whether it is caused by sea-waves or gusts of wind, will have wave-lengths considerably smaller than those of free elastic waves on the surface of the earth. Consequently as shown on pp. 315–316 of the paper referred to above, $k$ and $h$ are very small in comparison with $\xi$, and, as $A$ and $-iB$ are equal to

$$\frac{2\xi^2 - k^2 P}{F(\xi)} \mu, \quad \frac{2\xi^2 - k^2 P}{F(\xi)} \mu,$$

they are both approximately equal to

$$\frac{2\xi^2 P}{F(\xi)} \mu.$$

The expressions for $u$ and $w$ can therefore be expressed in the form

$$u = \frac{2\xi^2 P}{F(\xi)} \mu (\xi e^{-\alpha z} - \beta e^{-\beta z}) \sin \xi x \cos \nu t,$$

$$w = \frac{2\xi^2 P}{F(\xi)} \mu (\alpha e^{-\alpha z} - i\xi \beta e^{-\beta z}) \cos \xi x \cos \nu t,$$

taking into account only the real parts.

At the surface $z=0$, the ratio of $u$ to $w$ becomes $3:1$ (as shown on p. 315, *loc. cit.*).

The above approximation assumes (1) the pressure disturbance to be simple harmonic waves in the direction of the $x$-axis, and (2) the ground or the sea bed over which it is acting to be a horizontal plane. Actually neither of these conditions holds. But it is easy to see that the effect of complexity
of waves and the unevenness of ground or sea bed would be to make the arbitrary constants $A$ and $-iB$, if the solution be expressed in the above form, still more equal to each other.

If the pressure acted tangentially to the surface, that is to say, if it had the form

$$[Z]_0 = 0, \quad [X]_0 = P et^x e^{ipt},$$

the constants $A$ and $B$ in the expressions for displacements will have the values

$$A = -\frac{2\xi^2}{F(\xi)} \frac{P}{\mu}, \quad B = \frac{2\xi^2 - k^2}{F(\xi)} \frac{P}{\mu},$$

or

$$-\frac{\xi A}{F(\xi)} = \frac{2\xi^2 - k^2}{\mu} \frac{P}{\mu}, \quad B = \frac{2\xi^2 - k^2}{F(\xi)} \frac{P}{\mu},$$

when $k/\xi$ is small, and these are equal to each other.

When the ground is uneven, the problem is approximately equivalent to the pressure having components in the direction of $z$-axis as well as $x$-axis. It is clear from the above result that the presence of a horizontal component in the pressure disturbance will tend to equalise the constants still more.

It will be seen from the above expressions that the tangential stress contributes more to horizontal displacement than the vertical displacement. Indeed, it is the unevenness of the ground and sea bed, which makes the tangential stress more important than the vertical stress. The ratio of the horizontal to the vertical component in the region where the pressure is acting or in its neighbourhood will therefore be $3:1$, and will decrease to $0.7$ as we move away from the region. Stratification in the surface crust will tend to augment this ratio slightly. These results appear to be in agreement with observations.
FIG. 1. June 8, 1932.
Depression over head Bay of Bengal.

FIG. 2. October 21, 1932.
Moderate Storm in centre Bay of Bengal.
Circular Tank of diameter 210 cm. and depth 108 cm.
Vibrator at centre; measuring tube 62 cm. from edge.
Circular Tank of diameter 210 cm. and depth 108 cm.
Vibrator at centre; measuring tube 62 cm. from edge.