THE PROCESSES OF HEATING AND COOLING IN A SECTION OF THE IRISH SEA

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Summary

The heat balance equation is applied to the Holyhead-Dublin Bay section of the Irish Sea. It is shown that the seasonal variation in the heat content of the section as a whole can be accounted for, largely, by the local effects of radiation and exchange of heat with the atmosphere. There is, however, a considerable transfer of heat from the coastal waters towards the centre of the channel in summer and in the reverse direction in winter. Attributing this transfer to mixing due to horizontal turbulence, an estimate is made of the horizontal eddy diffusivity.

I. Introduction

In any given region of sea, during a specified interval of time, the balance between the gain and loss of heat energy may be represented by the equation

\[ Q = Q_B + Q_E + Q_c + Q_r + Q_v. \]

where we denote by

- \( Q_B \) the heat absorbed from incoming solar radiation,
- \( Q_E \) the effective back radiation from the sea to the atmosphere and space,
- \( Q_c \) the heat energy lost by evaporation to the atmosphere,
- \( Q_r \) the heat lost by conduction to the atmosphere,
- \( Q_v \) the heat used in raising the temperature of the water within the region,
- and \( Q_r \) the heat energy transferred to adjacent regions of sea by advection and by horizontal mixing.

In dealing with coastal waters it may be necessary to add a term \( Q_L \) to the right-hand side of equation (1), representing heat transferred to the Earth, either through the sea bed or at the coast.

If the values of all but one of the terms are known, equation (1) may be used to evaluate the remaining term. The data available are rarely complete enough for this, however, and the cases in which the equation may be applied are usually selected so that one or more of the terms can be neglected. Schmidt (1) and Mosby (2) have considered mean annual conditions over the oceans, in which case \( Q_r \) and \( Q_v \) are both zero. Using data obtained by Helland-Hansen (3) for an area west of the Bay of Biscay, in which transport of heat by currents, \( Q_v \), could be neglected, Sverdrup (4) applied the energy equation to show the seasonal variation of \( (Q_B + Q_c) \), the total heat exchange with the atmosphere. The radiation terms could be estimated and \( Q_r \) was obtained from the seasonal variation of temperature at various depths. The method was extended to other areas of the North Atlantic and North Pacific by Jacobs (5).
Harvey (6) had previously applied a similar method to the seasonal variation of temperature at various depths at a single station in the English Channel. $Q_b$ could be calculated from the temperature data, $Q_s$ was estimated from the recorded solar radiation at Kew and in the absence of any considerable movement of water into or out of the area in the vicinity of the station, $Q_c$ was neglected. The equation then reduces to

$$Q_s - Q_b = Q_B + Q_E + Q_c.$$  

Harvey did not attempt to separate the effects of back radiation, evaporation and conduction, but assuming the predominance of evaporation during the six winter months, October to March, he obtained an upper limit for the evaporation during these months.

2. Seasonal Variation of Heat Balance

2.1. Description of Method.—An attempt is made in the present paper to apply equation (I) to the seasonal variation of temperature in that part of the Irish Sea forming a channel between Holyhead and Dublin Bay. Since 1934 May, observations of surface temperature have been made three times weekly at three stations on the steamer route between Holyhead and Kingstown. These observations are part of an investigation into the temperature and salinity of the Irish Sea being carried out by the Oceanography Department of the University of Liverpool, with the aid of a grant from the Development Commission. By reason of the frequency of the observations, the monthly mean values deduced from them probably give a more reliable representation of the seasonal variation of surface temperature than has hitherto been obtained for this area.

Matthews (7) found that in the Holyhead–Dublin Bay Section the temperature was very nearly uniform from surface to bottom at all seasons. His quarterly charts show little difference between surface and bottom temperatures, except for August, when a difference of 0.1 deg. C. occurred near Holyhead, 0.3 deg. C. in the centre of the channel and 0.6 deg. C. near Kish. The distribution of temperature with depth in this section does not appear to have been investigated systematically since Matthews' observations, but the few temperature soundings which have been made confirm the conditions which he found. As the vertical differences of temperature are small compared with the seasonal variation, it seems justifiable, as a first approximation, to take the surface observations as representing the temperature of a column of water extending from the surface to the bottom. Making this assumption, we are in a position to calculate the increase in heat content of the whole body of water in the section from one month to the next, i.e. $Q_b$ in equation (I).

The incoming solar radiation, $Q_s$, may be estimated from tables given by Kimball (8), corrected for the prevailing cloudiness. Similarly, the effective back radiation, $Q_B$, can be obtained from a diagram given by Sverdrup (9) using observed values of sea surface temperature and relative humidity, and again correcting for cloudiness. The details of these computations will be given later.

The tidal streams in this channel flow in a north–south direction, reaching a maximum speed of about 2 knots (100 cm./sec.). Superposed on the tidal currents there is believed to be a slow drift northwards, at a rate probably not exceeding 0.6 sea mile (1.1 km.) per day. The isotherms run nearly north and south at all seasons, as shown, for example, by the charts of Proudman, Lewis and
Dennis (10), so it seems unlikely that there will be any considerable net gain of heat from other areas, due to either the tidal or non-periodic currents, in the course of a month. Allowing for the mixing by tidal motion, our investigation may be considered to apply to a body of water extending completely across the channel and extending along the channel to a distance of, say, 20 km. both north and south of the Holyhead–Dublin section.

There remains the possibility of transfer of heat across the isotherms by horizontal diffusion, i.e. from the coastal areas towards the centre of the channel in summer and from the centre outwards in winter. We shall return to this point later, but its effect may be neglected at present as we are considering the section as a whole. We may therefore put $Q_v = 0$.

The possible magnitude of a term $Q_L$, representing heat transferred to, or received from, the Earth, needs consideration. Two different processes may be involved. Firstly, heat may be exchanged between the water and the sea bed by conduction, no other source of heat reaching the sea bed. Secondly, in shallow water, and more particularly on beaches or banks which dry at high tide, there may be direct absorption of radiation by the land, part of which is subsequently transferred to the water by conduction. In the depths of the oceans, where the bottom water is practically constant in temperature, it was pointed out by Helland-Hansen (3) that the transport of heat through the sea bed, from the interior of the Earth, was negligible.

In shallow seas the water at the bottom undergoes a seasonal variation in temperature and we must assume that the surface of the sea bed shares this variation. Heat will therefore be transferred from the water to the sea bed in summer and received back in winter. To get an idea of the quantity of heat involved in this transfer, let us take the following ideal case. Consider a vertical column of 1 cm$^2$ cross-section extending from the surface of the sea, through the bottom, into the sea bed. Let the water in the column be at a uniform temperature from the surface to bottom, and let the temperature undergo a simple harmonic variation with a period of one year. Then the temperature may be denoted by $\theta_0 = A \sin \omega t$, $\theta_0$ being measured from its annual mean, $t$ from the date when $\theta_0 = 0$ and $\omega = 2\pi$/year.

The alternating part of the heat content of the water will be

$$Q_w = \sigma p d A \sin \omega t,$$

where $\sigma$ is the specific heat and $p$ the density of the water and $d$ is its depth.

The temperature in the sea bed at a depth $z$ below the bottom will be given by

$$\theta_z = Ae^{-\alpha z} \sin (\omega t - \alpha z),$$

where $\alpha$ is given by $\alpha^2 = \omega^2 / 2K_b$, $K_b$ being the thermal diffusivity of the sea bed. The alternating part of the heat content will be

$$Q_b = \sigma_b \rho_b \int_0^\infty \theta_z dz,$$

where $\sigma_b$ is the specific heat and $\rho_b$ the density of the sea bed, which gives

$$Q_b = \sigma_b \rho_b \frac{A}{\sqrt{2\alpha}} \sin \left(\frac{\omega t - \pi}{4}\right).$$

The amplitude is therefore $|Q_b| = \sigma_b \rho_b A / \sqrt{2\alpha}$, and the effective thermal capacity for the annual variation is $|Q_b| / A = \sigma_b \rho_b / \sqrt{2\alpha}$.

The writer is not aware of any observations having been made at various depths in the sea bed, but Birge, Juday and March (11) investigated the seasonal variation
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of temperature at various depths in the bottom sediments of Lake Mendota. They found $\alpha = 5.7 \times 10^{-3} \text{ cm.}^{-1}$, corresponding to $K_b = 3.25 \times 10^{-3} \text{ cm.}^2/\text{sec.}$, the bottom being fine mud. Taking $\sigma_b \rho_b = 1 \text{ cal./cm.}^2 \text{ deg. C.}$, which is probably an upper limit, $|Q_b|/A = 124 \text{ cal./cm.}^2 \text{ deg. C.}$. This is for a soft bottom with, probably, a high water content. To go to the other extreme, we may take figures obtained from measurements of Earth temperatures at land stations. Keränen (12) gives a number of examples of these with $K_b$ ranging from $3.8 \times 10^{-3} \text{ cm.}^2/\text{sec.}$, while $\sigma_b \rho_b$ averaged 0.4. Taking $K_b = 10.7 \times 10^{-3}$ and $\sigma_b \rho_b = 0.4$ gives $|Q_b|/A = 94 \text{ cal./cm.}^2 \text{ deg. C.}$

We may say, therefore, that, as far as the seasonal variation is concerned, the effect of the sea bed is to increase the apparent thermal capacity of the water by an amount equivalent to an increase in its depth of the order of 1 m. In the present treatment of the Holyhead–Dublin Bay section we are justified in neglecting this effect.

As shown in a table given by Sverdrup (9, p. 56), in "average coastal water", 98.8 per cent of the incident radiant energy is absorbed in the first 10 m., so that in greater depths we may neglect any direct effect of radiation on the sea bed. In shallower water an appreciable proportion of the incident radiation may reach the bottom. Part of it will be reflected and scattered and much of this part will be absorbed in the water before it can re-emerge from the surface. Part will be absorbed by the sea bed (and increase its temperature), but as the temperature of the sea bed cannot rise appreciably above that of the water in contact with it, much of this part also will be transferred to the water by conduction and vertical mixing. When the thermal capacity of the sea bed, as discussed in the preceding paragraph, has been taken into account, the only part of the incoming radiation lost to the water is that which re-emerges from the surface after reflection and scattering at the bottom.

Conditions are different where large areas of foreshore are alternately exposed at low tide and flooded at high tide. In this case, the temperature of the exposed areas may rise considerably above that of the off-shore water in summer, so that with the incoming tide large quantities of heat are transferred to the water, to be carried away from the coast when the tide recedes. The reverse may occur in winter. On the Holyhead side of the section the area of shore exposed at low tide is very small and the sea bed slopes downwards steeply. The inter-tidal and shallow water zone is wider on the Dublin Bay side but it is still narrow compared with the width of the whole section. It therefore seems justifiable to neglect $Q_L$ to a first approximation.

We have now reduced equation (1) to the form

$$Q_A = Q_E + Q_c = Q_s - Q_B - Q_0,$$

where $Q_A$ represents the total energy transferred to the atmosphere and the three terms on the right-hand side may be computed for successive monthly intervals.

To separate the effects of evaporation and conduction, a knowledge of $Q_c/Q_E$, usually known as the Bowen ratio, is required. The data for a satisfactory determination of this quantity are not available, but tentative values have been obtained by combining the air temperature and relative humidity observations at Holyhead meteorological station with the sea surface temperature at the oceano-graphic station near Holyhead, the distance between the two stations being only 8 km. Using these values estimates can be made of the actual evaporation.
2.2. Details of Computations.—In carrying out the computations, the unit of volume is taken as a column of water, 1 cm\(^2\) in horizontal cross-sectional area, extending from the surface to the bottom. The unit of time is one calendar month, intervals being measured from the 15th of each month, i.e. January 15–February 15, etc. The oceanographic and meteorological data used are all mean monthly values for the twelve years 1935–1946.

The oceanographic data used are the mean monthly values of sea surface temperature \(\theta_{1}, \theta_{2}, \theta_{3}\), at three stations S1, S2, S3, which are 4 km., 40 km. and 80 km. respectively, west of North Stack, the total width of the section being 98 km. Assigning the monthly mean temperatures to the 15th day of each month, the monthly increments in temperature \(\Delta \theta_{1}, \Delta \theta_{2}, \text{ and } \Delta \theta_{3}\) from January 15–February 15, etc. for each station are found.

To simplify the computation of \(Q_{0}\), the monthly increment in heat content, the actual channel of irregular section has been replaced by three channels of rectangular section, as shown in Fig. 1; two coastal channels, I on the Holyhead side, 22 km. wide by 53 m. deep, and III on the Dublin Bay side, 28 km. wide by 30 m. deep, and a central channel II, 48 km. wide by 102 m. deep. The water throughout channel I is assumed to be at a constant temperature \(\theta_{1}\), that observed at station S1. Similarly, water in channel II is assumed to be at temperature \(\theta_{2}\) and that in channel III at temperature \(\theta_{3}\).

Then for the three channels separately, the monthly heat increment, assuming vertical uniformity of temperature, is given by

\[
Q_{0i} = \sigma \rho d_i \Delta \theta_i,
\]  
(3)
with similar expressions for $Q_{02}$ and $Q_{03}$, where $\sigma =$ specific heat and $\rho =$ density of the water, and $d_1$, $d_2$, $d_3$ are the depths of the three channels respectively. At a mean temperature 10 deg. C. and salinity 34.3 parts per thousand, $\sigma = 0.937$ and $\rho = 1.025$, so that $\sigma \rho = 0.96$. For the whole section,

$$Q_0 = \frac{b_1 Q_{01} + b_2 Q_{02} + b_3 Q_{03}}{b_1 + b_2 + b_3},$$

where $b_1$, $b_2$, $b_3$ are the respective widths of the three channels. The results of this computation are shown in Table I.

Values of the solar radiation (direct + diffuse), from a cloudless sky, incident upon a horizontal surface in various latitudes, have been given by Kimball (8, Table III) for the 21st day of each month. In cases where it has been possible to compare the computed values with measurements of radiation, Kimball states that the agreement is within a few per cent. By linear interpolation between his values for stations in $52^\circ$ N., $10^\circ$ W., and $56^\circ$ N., $7^\circ$ W., the values for $53^\circ$ N., the latitude of the Holyhead–Dublin section, have been found, ignoring the difference in longitude. From a smooth curve drawn through the points so obtained, the values of the total radiation for the intervals January 15–February 15, etc. were derived.

The reduction in intensity due to the presence of cloud is given by the equation

$$S = S_0 (1 - 0.071 C),$$

where $S_0$ is the radiation from a cloudless sky, $S$ is the radiation in the presence of cloud and $C$ is the cloud amount on the scale 0–10. $C$ has been taken as the average of the mean monthly values for Holyhead and Dublin for the period considered.

If $r$ is the mean reflectivity of the sea surface for solar radiation, the heat energy actually absorbed in the sea is

$$Q_a = (1 - r) S_0 (1 - 0.071 C).$$

Based on the observations of Powell and Clarke (13), a mean value of 8 per cent has been taken for $r$, so that $Q_2$ per cent of the incident radiation is absorbed.

The effective back radiation $B_0$ from the sea surface to a cloudless sky was obtained from the diagram given by Sverdrup (9, p. 59) based on data given by Ångström (14), as a function of sea surface temperature and relative humidity. The reduction due to cloud is computed, using the equation due to Asklöf (15),

$$Q_B = B_0 (1 - 0.083 C).$$

The sea surface temperature used in computing $B_0$ is the mean of $\theta_1$, $\theta_2$ and $\theta_3$ and the relative humidity is the mean of values at Holyhead and Dublin. $C$ is the mean cloudiness as before.

The monthly values of $Q_a$ and $Q_B$, so obtained, are given in Table I, which also gives the net gain of heat by radiation, $Q_R = Q_a - Q_B$, and the heat energy transferred to the atmosphere, $Q_A = Q_R - Q_0$.

The ratio of the heat used for conduction to the heat used for evaporation is given by the formula, first obtained by Bowen (16),

$$R = \frac{\rho \sigma_{pa} d\theta_a}{0.021 L \frac{de}{dz} \frac{d\theta_a}{dz}},$$

where $\rho =$ atmospheric pressure, $\sigma_{pa} =$ specific heat of the air at constant pressure, $L =$ latent heat of evaporation, $d\theta_a/dz$ and $de/dz$ are the vertical gradients of air
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temperature and vapour pressure respectively above the sea surface. Using the standard values, \( \sigma_{wa} = 0.237 \) cal./g. deg. C., \( L = 590 \) cal./g., \( p = 1013 \) mb., and measuring \( \theta_a \) in deg. C. and \( e \) in mb., the equation becomes

\[
R = 0.065 \frac{(\theta_w - \theta_a)}{(e_w - e_a)},
\]

(7)

where \( \theta_w \) and \( e_w \) are values at the sea surface and \( \theta_a \) and \( e_a \) are values at a standard height above the surface. \( e_w \) may be taken as the saturated vapour pressure at temperature \( \theta_w \).

To determine \( \theta_a \) and \( e_a \), the air temperature and relative humidity should be measured at the same position as the sea temperature and at the same time. The differences \( \theta_w - \theta_a \) and \( e_w - e_a \) are usually both small and, owing to horizontal gradients in the quantities, if there is any considerable distance between the positions at which \( \theta_w, e_w \) and \( \theta_a, e_a \) are measured, the differences will not be of much value. In the case of the Holyhead observations, the oceanographic station is approximately 8 km. from the meteorological station, which is 8 m. (26 ft.) above sea-level. It therefore seems worth while combining the observations at these two positions to obtain values of \( R \), which will give some indication of the ratio \( Q_c/Q_E \) to be expected over the section.

The data used in computing \( R \) are shown in Table II and the values of \( Q_E \) and \( E \), the quantity of water evaporated, in millimetres, at the foot of Table I.

2.3. Discussion.—The general course of the curves for \( Q_R, Q_b \) and \( Q_A \), shown in Fig. 2, is as expected. The fairly large negative values of \( Q_A \) from April–May to July–August, indicating a considerable transfer of heat from the atmosphere to the sea, appear improbable, however. The explanation may lie in an overestimation of \( Q_b \) due to the presence of slight vertical temperature gradients during these months. Another possible explanation is that some heat may flow into the area under investigation from other areas, e.g. from the land or from warmer waters to the north and north-east of the section, but it does not seem likely that any large quantities of heat would be gained in this way.

The sharp increase in \( Q_A \) in autumn, reaching a maximum in October–November, and falling off gradually during the rest of the winter, is in agreement with results from other areas. It corresponds to the variation of \( e_w - e_a \) for Holyhead (Table II) and is to be attributed mainly to evaporation. Using the values of \( R \) for Holyhead, \( Q_E \) and \( E \) (the evaporation in millimetres) are given at the foot of Table I for the winter months only. As the values of \( R \) may not be representative of conditions over the whole section, the figures for \( Q_E \) and \( E \) should be considered less reliable than those for \( Q_A \). The total evaporation for the six months October–March is estimated at 555 mm. Table II indicates negative values of \( R \) in the summer months, representing a loss of heat by evaporation but a gain by conduction from the air, which is quite probable. The values do not reach \(-1\) however and, unless \( R < -1 \), there cannot be a net gain of heat by the sea.

3. Seasonal Transfer of Heat by Horizontal Diffusion

Although the seasonal variation of \( Q_b \) follows the general course to be expected on the assumption that it is governed mainly by the local effects of radiation and exchange of heat with the atmosphere, the same is not true of the values of \( Q_{b1}, Q_{b2} \) and \( Q_{b3} \) taken separately. From Table I, it is seen that for a given month there may be large differences between them, the central channel showing a greater gain...
of heat ($Q_{o2}$) in summer than the two coastal channels ($Q_{o1}, Q_{o3}$) and a greater loss of heat in winter. This is consistent with the idea that there is a transfer of heat from the coastal areas towards the centre of the section in the summer and in the reverse direction in winter.

From equation (1), we may write for any of the three channels,

$$ Q_R = Q_A + Q_0 + Q_v $$

where $Q_R = Q_o - Q_e$ and $Q_A = Q_B + Q_C$, $Q_v$ being the heat removed from the channel by lateral diffusion. $Q_R$ and $Q_A$ may vary from one channel to another due to variations in the meteorological conditions, but it is unlikely that the variations will be of the same order of magnitude as those of $Q_{o1}, Q_{o2}$ and $Q_{o3}$ shown in Table I. If we assume that, to a first approximation, $Q_R$ and $Q_A$ are uniform across the section, we can estimate the lateral transfer of heat from channel I, to channel II and hence the effective horizontal diffusivity, and similarly for channels III and II.
### Table 1

**Seasonal Variation of Heat Balance**

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<tr>
<td>I Near Holyhead $\theta_1$</td>
<td>7.9</td>
<td>7.3</td>
<td>7.0</td>
<td>8.1</td>
<td>9.5</td>
<td>11.7</td>
<td>13.6</td>
<td>14.9</td>
<td>14.8</td>
<td>13.4</td>
<td>11.4</td>
<td>9.4</td>
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<tr>
<td>II Holyhead–Kish $\theta_2$</td>
<td>9.0</td>
<td>8.2</td>
<td>7.7</td>
<td>8.1</td>
<td>9.1</td>
<td>11.0</td>
<td>12.6</td>
<td>13.8</td>
<td>14.1</td>
<td>13.5</td>
<td>12.0</td>
<td>10.4</td>
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<tr>
<td>III Near Kish $\theta_3$</td>
<td>8.3</td>
<td>7.6</td>
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<td>Temp. Increments deg. C.</td>
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<tr>
<td>I $\Delta \theta_1$</td>
<td>-0.6</td>
<td>-0.3</td>
<td>1.1</td>
<td>1.4</td>
<td>2.2</td>
<td>1.9</td>
<td>1.3</td>
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<td>-1.4</td>
<td>-2.0</td>
<td>-2.0</td>
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<tr>
<td>II $\Delta \theta_2$</td>
<td>-0.8</td>
<td>-0.5</td>
<td>0.4</td>
<td>1.0</td>
<td>1.9</td>
<td>1.2</td>
<td>0.3</td>
<td>-0.6</td>
<td>-1.5</td>
<td>-1.6</td>
<td>-1.4</td>
</tr>
<tr>
<td>III $\Delta \theta_3$</td>
<td>-0.7</td>
<td>-0.3</td>
<td>0.8</td>
<td>1.5</td>
<td>2.0</td>
<td>1.6</td>
<td>1.1</td>
<td>-0.9</td>
<td>-1.6</td>
<td>-1.6</td>
<td>-1.5</td>
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| Heat Increments cals./cm.$^2$ |          |           |           |          |          |           |           |            |           |           |           |           |
| I $Q_{a1}$ | -3060     | -1530     | 5610      | 7130     | 11230    | 9690      | 6630     | -510      | -7140     | -10400    | -10400    | -7650    |
| II $Q_{a2}$ | -7840     | -4900     | 3920      | 9800     | 18620    | 15680     | 11760    | 2940      | -5880     | -14700    | -15680    | -13720   |
| III $Q_{a3}$ | -2030     | -870      | 2320      | 4350     | 5800     | 4640      | 3190     | -1160     | -2610     | -4640     | -4640     | -4350    |
| Whole section $Q_a$ | -5110     | -2990     | 3840      | -7640    | 13300    | 11180     | 8160     | 990       | -5230     | -10860    | -11340    | -9680    |

| Radiation cals./cm.$^1$ |          |           |           |          |          |           |           |            |           |           |           |           |
| Mean cloudiness | 7.1       | 6.9       | 6.5       | 6.1      | 6.3      | 6.9      | 7.0      | 6.6       | 6.7       | 6.9       | 6.9       | 6.9      |
| Mean sea temp. deg. C. | 8.0       | 7.5       | 7.7       | 8.8      | 10.4     | 12.3     | 13.7     | 14.3      | 13.8      | 12.4      | 10.7      | 9.2      |
| Mean rel. humidity per cent | 85         | 85        | 83        | 81       | 81       | 82       | 84       | 85        | 85        | 85        | 87        | 85       |
| Cloudless Solar rad. $S_0$ | 6110       | 9600      | 15750     | 20400    | 24120    | 23520    | 21230    | 16270     | 11100     | 7190      | 3900      | 3350     |
| Sky $\int$ Back rad. $B_u$ | 8000       | 7220      | 8030      | 8070     | 7770     | 7930     | 7590     | 7690      | 7470      | 7780      | 7560      | 7910     |
| Radiation absorbed $Q_r$ | 2810       | 4510      | 7820      | 10710    | 12220    | 11040    | 9760     | 7930      | 5310      | 3370      | 1830      | 1660     |
| Back radiation $Q_B$ | 3280       | 3110      | 3690      | 3810     | 3800     | 3260     | 3460     | 3290      | 3340      | 3280      | 3400      | 3400     |
| $Q_N = Q_a - Q_B$ | -470       | 1400      | 4130      | 6900     | 8420     | 7780     | 6500     | 4470      | 2020      | 30        | -1420     | -1740    |

| Transfer to Atmosphere |          |           |           |          |          |           |           |            |           |           |           |           |
| $Q_A = Q_B - Q_c$ cals./cm.$^3$ | 4640      | 4390      | 290       | -740     | -4880    | -3400     | -1660    | 3480      | 7250      | 10890     | 9920      | 7940     |
| $Q_B$ cals./cm.$^3$ | 3130      | 3600      | (350)     | ...      | ...      | ...      | ...      | (3510)    | 6020      | 8060      | 6840      | 5100     |
| $E$ mm. | 53        | 61        | (6)       | ...      | ...      | ...      | ...      | (59)      | 102       | 137       | 116       | 86       |
### Table II

**Values of $R = \frac{Q_g}{Q_{g'}}$ for Holyhead**

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</thead>
<tbody>
<tr>
<td><strong>Sea Temp. $\theta_w$ deg. C.</strong></td>
<td>7.6</td>
<td>7.1</td>
<td>7.5</td>
<td>8.8</td>
<td>10.6</td>
<td>12.7</td>
<td>14.3</td>
<td>14.9</td>
<td>14.1</td>
<td>12.4</td>
<td>10.4</td>
<td>8.7</td>
</tr>
<tr>
<td><strong>Air temp. $\theta_a$ deg. C.</strong></td>
<td>5.9</td>
<td>6.5</td>
<td>7.9</td>
<td>9.7</td>
<td>11.9</td>
<td>14.1</td>
<td>15.3</td>
<td>14.7</td>
<td>12.9</td>
<td>10.3</td>
<td>8.0</td>
<td>6.3</td>
</tr>
<tr>
<td><strong>$\theta_w - \theta_a$ deg. C.</strong></td>
<td>1.7</td>
<td>0.6</td>
<td>0.4</td>
<td>0.9</td>
<td>1.3</td>
<td>1.4</td>
<td>1.0</td>
<td>2.2</td>
<td>1.2</td>
<td>2.1</td>
<td>2.4</td>
<td>2.4</td>
</tr>
<tr>
<td><strong>S.V.P. at $\theta_w$, $e_w$ mb.</strong></td>
<td>1054</td>
<td>1011</td>
<td>1044</td>
<td>113</td>
<td>128</td>
<td>147</td>
<td>163</td>
<td>169</td>
<td>161</td>
<td>144</td>
<td>126</td>
<td>113</td>
</tr>
<tr>
<td><strong>V.P. in air $e_a$ mb.</strong></td>
<td>8.0</td>
<td>8.3</td>
<td>8.8</td>
<td>9.8</td>
<td>11.6</td>
<td>13.5</td>
<td>14.5</td>
<td>14.1</td>
<td>12.3</td>
<td>10.4</td>
<td>9.1</td>
<td>8.3</td>
</tr>
<tr>
<td><strong>$e_w - e_a$ mb.</strong></td>
<td>2.5</td>
<td>1.8</td>
<td>1.6</td>
<td>1.5</td>
<td>1.2</td>
<td>1.8</td>
<td>2.8</td>
<td>3.8</td>
<td>4.0</td>
<td>3.5</td>
<td>3.0</td>
<td></td>
</tr>
<tr>
<td><strong>$R = 0.66(\theta_w - \theta_a)/(e_w - e_a)$</strong></td>
<td>0.45</td>
<td>0.22</td>
<td>-0.17</td>
<td>-0.37</td>
<td>-0.71</td>
<td>-0.37</td>
<td>0.05</td>
<td>0.21</td>
<td>0.35</td>
<td>0.45</td>
<td>0.53</td>
<td></td>
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</tbody>
</table>

### Table III

**Lateral Transfer of Heat**

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</thead>
<tbody>
<tr>
<td>$Q_0 - Q_{01}$ cals./cm.²</td>
<td>-2050</td>
<td>-1460</td>
<td>-1770</td>
<td>510</td>
<td>2070</td>
<td>1490</td>
<td>1530</td>
<td>1500</td>
<td>1910</td>
<td>-460</td>
<td>-940</td>
<td>-2030</td>
</tr>
<tr>
<td>$\theta_1 - \theta_a$ deg. C.</td>
<td>-1.0</td>
<td>-0.8</td>
<td>-0.35</td>
<td>0.2</td>
<td>0.55</td>
<td>1.15</td>
<td>1.05</td>
<td>0.9</td>
<td>0.3</td>
<td>-0.35</td>
<td>-0.8</td>
<td>-1.05</td>
</tr>
<tr>
<td>$Q_0 - Q_{02}$ cals./cm.²</td>
<td>-3080</td>
<td>-2120</td>
<td>1520</td>
<td>3290</td>
<td>7500</td>
<td>6540</td>
<td>4970</td>
<td>2150</td>
<td>-2620</td>
<td>-6220</td>
<td>-6700</td>
<td>-5330</td>
</tr>
<tr>
<td>$\theta_2 - \theta_a$ deg. C.</td>
<td>-0.65</td>
<td>-0.5</td>
<td>-0.2</td>
<td>0.25</td>
<td>0.55</td>
<td>0.6</td>
<td>0.55</td>
<td>0.15</td>
<td>-0.35</td>
<td>-0.55</td>
<td>-0.6</td>
<td>-0.55</td>
</tr>
</tbody>
</table>
Assuming $Q_R$ and $Q_A$ are uniform across the section, we have, for the section as a whole,

$$Q_R = Q_A + Q_{v12}.$$  

For the channel I alone,

$$Q_R = Q_A + Q_{e1} + Q_{v12},$$  

where $Q_{v12}$ is the heat transferred from channel I to channel II. Hence

$$Q_{v12} = Q_0 - Q_{e1},$$  

remembering that all the values are in cals. per cm.$^2$ of surface per month. Thus for the total heat transferred across the vertical interface between channels I and II (see Fig. 1)

$$b_1(Q_0 - Q_{e1}) = -\alpha p K_{12} d_t (\frac{d\theta}{dx})_{12},$$

where $K_{12}$ is the effective coefficient of horizontal diffusion between channel I and channel II in cm.$^2$/sec., $t$ is the number of seconds in a month and $(d\theta/dx)_{12}$ is the horizontal temperature gradient at the interface. Replacing $-(d\theta/dx)_{12}$ by $(\theta_1 - \theta_2)/l_{12}$, where $l_{12}$ is the distance between the stations $S_1$ and $S_2$,

$$K_{12} = b_1 l_{12} (\frac{Q_0 - Q_{e1}}{\theta_1 - \theta_2}) = 594 \left(\frac{Q_0 - Q_{e1}}{(\theta_1 - \theta_2)}\right) \text{cm.}^2/\text{sec.},$$

when the numerical values are inserted.

Similarly, for channels III and II,

$$K_{23} = b_2 l_{23} (\frac{Q_0 - Q_{e3}}{\theta_3 - \theta_2}) = 1480 \left(\frac{Q_0 - Q_{e3}}{(\theta_3 - \theta_2)}\right) \text{cm.}^2/\text{sec.}.$$  

The values of $(Q_0 - Q_{e1})$ have been plotted against $(\theta_1 - \theta_2)$ in Fig. 3(a) and $(Q_0 - Q_{e3})$ against $(\theta_3 - \theta_2)$ in Fig. 3(b). The points show considerable scatter, which is not surprising, since, besides errors in the computation of $Q_0$, $Q_{e1}$ and $Q_{e3}$, $K_{12}$ and $K_{23}$ may themselves vary with time, and our estimates of the horizontal temperature gradients at the interfaces are based on observations in three positions only. However, the relation is sufficiently linear to justify the hypothesis that the

\[\times 10^3\]  

\[\times 10^2\]
process involved is primarily one of diffusion and to provide an estimate of the horizontal diffusivity. Taking the mean value of \( \frac{(Q_0 - Q_{01})}{(Q_1 - \theta_2)} \) to be 1900 cals./cm.\(^2\) month deg. C. and of \( \frac{(Q_0 - Q_{03})}{(\theta_3 - \theta_2)} \) to be 10,000 cals./cm.\(^2\) month deg. C., the final results are

\[
K_{12} = 1.1 \times 10^6 \text{ cm.}^2/\text{sec.},
\]
\[
K_{23} = 14.8 \times 10^6 \text{ cm.}^2/\text{sec.}
\]

Sverdrup (9, p. 25) has tabulated the values of the horizontal eddy diffusivity found by various investigators in other areas, which range from \( 2 \times 10^6 \) in the layer 200–400 m. deep in the California Current to \( 4 \times 10^7 \) in the surface layer of the north-western North Atlantic. In a shallow, partially enclosed area like the Irish Sea, smaller values would be expected than in the open oceans, as the formation of large horizontal eddies may be restricted. The thorough mixing indicated by values of \( K \) of \( 10^6 \) to \( 10^7 \) in this section of the Irish Sea is probably to be attributed to horizontal turbulence set up by the effects of bottom friction on the strong tidal currents. The much larger value of \( K \) found on the western side of the section is consistent with this view, as the bottom topography is more irregular there, with considerable areas of banks south of Kish.

The meteorological data for Holyhead from 1935 to 1946 and for Dublin from 1935 to 1940 have been taken from the *Monthly Weather Report*, published by the Meteorological Service of the Air Ministry. I am indebted to the Director of the Meteorological Service, Department of Industry and Commerce for Eire, for the data for Dublin from 1941 to 1946.

I wish to express my thanks to Professor J. Proudman for his helpful interest in this work.

Department of Oceanography,
Liverpool University:
1948 March 3.

References