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FACTORS AFFECTING THE TEMPERATURE OF THE SURFACE LAYER OF THE SEA

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FACTORS AFFECTING THE TEMPERATURE OF THE SURFACE LAYER OF THE SEA

A study of the heat exchange between the sea and the atmosphere, the factors affecting temperature structure in the sea and its forecasting

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ABSTRACT

Earlier attempts to predict the heat budget and temperature in the sea are reviewed, and the problems to be solved for successful prediction of the temperature in surface layers are discussed.

Surface water masses are defined for practical forecasting purposes on various geographical and physical bases, corresponding roughly to the definitions of air masses in the atmosphere. Furthermore, the classification of optical water masses has been revised.

A general heat budget is established, and the terms in this budget are evaluated for computing local changes of surface temperature. Some examples of measured daily incident radiation during clear and cloudy days are presented, and the following empirical formula for estimating insolation from the noon altitude of the sun, the length of the day and the cloudiness is derived (symbols see Chapter 4):

$$Q_s = 0.014\, A_n t_d\, (1 - 0.0006\, C^3) \quad [\text{g cal cm}^{-2}\,(24\text{ h})^{-1}]$$

This formula is valid to $A_n = 75°$; above this value Q_s remains constant.

For short periods, the following formula can be used:

$$Q_{os} = 1.9 \sin \bar{a} \quad [\text{g cal cm}^{-2}\text{ min}^{-1}]$$

It is concluded that average cloudiness (and also low visibility) during the day can be estimated from pyroheliometer measurements by the use of the following formula:

$$C = \sqrt[3]{\frac{Q_{os} - Q_s}{0.0006\, Q_{os}}} \qquad \text{(tenth of sky covered)}$$

For estimating radiation reflected from the sea surface, the following empirical formula, which is valid only for daily computations, is derived:

$$Q_r = 0.15\, Q_s - (0.01\, Q_s)^2 \quad [\text{g cal cm}^{-2}\,(24\text{ h})^{-1}]$$

For the computation of short term (hourly or three-hourly) albedo, a simplified formula can be used:

$$Q_r = Q_s \frac{300}{\bar{a}} \qquad (\text{g cal cm}^{-2}\text{ min}^{-1})$$

For computing effective back radiation, the linear formula of LÖNNQUIST is adopted and a graph is constructed for the estimation of effective

back radiation (Figure 12). The effective back radiation is corrected for the effect of cloudiness with Möller's formula:

$$Q_b = Q_{ob}\,(1 - 0.0765\ C) \quad [\text{g cal cm}^{-2}\ \text{min}^{-1}]$$

Various theoretical and empirical formulas for estimating evaporation from the sea are compared and the modified formula of Rohwer is found to be the most accurate one.

$$E = (0.26 + 0.077\ V) \quad (0.98\ e_w - e_a) \quad [\text{mm (24 h)}^{-1}]$$

The change of wind speed with height over the sea in various stability conditions is discussed and the average change is given in Figure 13.

Convective transfer of sensible heat is computed with the formula:

$$Q_h = 39\ (0.26 + 0.077\ V)\ (T_w - T_a) \quad [\text{g cal cm}^{-2}\ \text{(24 h)}^{-1}]$$

The possibilities for estimation of the changes in the temperature of the air moving over the ocean are pointed out.

Diurnal variations of sea surface and air temperatures are discussed and the selection of proper values for use in the above formulas is recommended.

When the differences $(T_w - T_a)$ or $(0.98\ e_w - e_a)$ are negative, sensible heat is transferred to the sea or condensation of vapour takes place on the sea surface. In these conditions high stability of the air close to the sea surface is expected and therefore the following modified formulas are proposed:

$$Q_c = 0.077\ V\ (0.98\ e_w - e_a)\ L_t \quad [\text{g cal cm}^{-2}\ \text{(24 h)}^{-1}]$$
$$Q_h = 3\ V\ (T_w - T_a) \quad [\text{g cal cm}^{-2}\ \text{(24 h)}^{-1}]$$

The heat budget of seas covered with ice is discussed, and a formula for the computation of heat conduction through the ice is given.

The factors affecting the thermal structure in a given locality in the sea and procedures for the computation and prediction of temperature structure are critically reviewed, and the methods, formulas and theories applicable for prediction of temperature changes in given localities are set out in Part II.

Tables are computed for the determination of the amount of radiation absorbed in different layers of the sea and by different optical water masses. Synthetic procedures are described and formulas given for the computation of temperature changes at various depths in the continuous density model and in the two-layer system, using the above-mentioned tables.

The changes of temperature caused by mixing are discussed, and the difficulties of the estimation of the austausch coefficient are reviewed. Convective stirring and mixing by wave action are considered as the most important mixing processes in the surface layers, and formulas for the estimation of the thickness of the mixed layer caused by these factors are given. Wave forecasting formulas are reviewed and a new simplified formula is proposed, in which the length of the fetch, the duration of the wind, its speed and the difference between the sea and air temperatures enter as parameters (Formula 46).

The periodic short- and long-term fluctuations of the thermocline depth and the difficulties of their prediction are discussed.

The horizontal transport of heat by currents must be accounted for by predicting the temperature changes in a given locality. The separation of wind currents from permanent flow is recommended, and procedures for the estimation of the direction and speed of surface currents from wind and from changes of atmospherie pressure data are described.

The influences of tidal currents on the temperature in shallow water are outlined. The principles for the estimation of the movement of current divergences and convergences are established.

The influence of the changes of sea level, caused by changes of barometric pressure and the piling up action of wind, on the water temperature and the depth of the mixed layer are described, and existing formulas for computing sea level variations are reviewed.

In Part III the application of the formulas and procedures for the forecasting of sea temperatures are illustrated with hindcasts in three different sea areas. General rules are given for the selection of data (both meteorological and oceanographic) to be used for the computation of forecasts. The hindcasts are presented in tabular form, and the procedures of estimation in case of lacking or deficient data are briefly described. The accuracy of the forecasts and the local factors affecting the accuracy are pointed out. At the end, an autoevaluation of the whole work is given and some problems to be solved in the future, in order to improve the accuracy of the forecasts, are listed.

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PART I

HEAT EXCHANGE BETWEEN THE SEA AND THE ATMOSPHERE

1. INTRODUCTION

Successful conduct of human activities at sea (navigation, fishing, etc.) requires the ability to predict the behaviour of natural phenomena which influence these activities.

One of the most important practical aims of science is to provide predictions, made possible by scientific descriptive and experimental procedures and by interpreting the vast amount of data gathered on the behaviour of elements in the natural system.

Forecasts of various oceanographic conditions have long been made, and some have been extremely successful (e.g. prediction of tides, waves etc.). In recent years, attempts have been made to extend forecasting to other conditions which especially affect fisheries and navigation (TERADA and HANZAWA, 1957, and others). LAEVASTU (1960) has summarized the general principles and procedure of complete hydroptic forecast.

Hydropsis (= hydro synopsis) (LYMAN, 1958) is roughly comparable to synoptic meteorology. Hydropsis is defined as analysis of oceanic conditions in a given area, based upon more or less »synoptic» data obtained within a few days, and predictions based upon this analysis. (The term *hydroclime* was selected by LYMAN (1958) to indicate average hydrographical conditions, based on statistical treatment of, for example, monthly data, collected over many years).

This paper analyzes and summarizes the principles and methods for forecasting temperature and its changes in the surface layers of the sea. New approaches and significant improvements to the earlier procedures are proposed. Emphasis is put on the construction of empirical formulas, based on the consideration of physical cause-effect principles, which can be used with ease for practical predictions. The parameters and factors in these formulas are selected so that they are either contained in routine meteorological and oceanographic observations, or can easily be derived from them.

2. EARLIER WORK ON THE PREDICTION OF THE TEMPERATURE AND HEAT BUDGET OF THE SEA

The temperature hydroclime of the sea surface is relatively wellknown and is shown on monthly maps in many atlases. However, considerable differences occur between the average temperatures at different seasons and in different years. The big temperature variations from year to year in the surface water of the Norwegian Sea, and their causes, were analysed by HELLAND-HANSEN and NANSEN (1920). They concluded that the major causes of these variations are to be found in the variations in solar radiation, that the change in air temperature goes ahead of the change in water surface temperature, and that the winds are the principal cause of the temperature variations of the surface layers. The wind action also causes the surface waters to pile up, and there is, therefore, a positive mathematical correlation between the water level along the European coast and the temperature.

On the other hand it has long been recognized in Japan and in maritime countries elsewhere that the temperature conditions in the surrounding seas govern the air temperature over the land and influence the annual field crops. Many attempts have been made in Japan to forecast sea surface temperature (HAYASHI, 1935, SUDA, 1938, TAKENOUTI, 1957).

SHULEIKIN (1953) considered in detail the processes of energy exchange and summarized the Russian papers on the prediction of sea water temperature. The most complete treatment of forecasting the thermal structure of the sea is that by SCHULE (1952).

Attempts to predict water temperatures may be divided into two groups covering the long term trends and the short term changes.

2.1. PREDICTION BY CORRELATION AND LONG TERM TRENDS

HATANAKA (1948) found that the secular variations of coastal water temperature in the north-eastern sea region of Japan had a 9-year period. WATANABE and HIRANO (1955) tried the »long term trends» method by computing the correlation between the resemblance and the temperature anomalies for four different areas west of central and southern Japan and predicted the temperatures by correlation. MCLELLAN and LAUZIER (1956) considered the cyclic nature of the variations of temperature in past years around eastern Canada and concluded that long term trends could be forecast with some confidence.

MILLAR (1952) showed the possibilities and limitations of statistical temperature forecasts for 8 days, in the Great Lakes of North America, based on one past temperature observation and the normal annual curve. He concluded that predictions could be improved by using more past observations and an elaborate regression formula.

KOLESNIKOV (1947, 1953) derived a complicated theoretical formula for determining annual variations of temperature; it had 6 arbitrary constants and 6 auxiliary equations from which the constants could be determined. These formulas required the use of computers and have not been applied for practical prediction.

It can be concluded from the study of the above and other similar attempts that the predictions utilizing seasonal temperature curves for restricted localities are unsatisfactory and do not allow any greater accuracy. Considerable variations occur from year to year, as pointed out by HELLAND-HANSEN and NANSEN (1920), and it is important to know these as well as shorter term variations.

2.2. SHORT TERM PREDICTIONS BY CONSIDERING THE HEAT BUDGET

The heat budget method has been used by several workers for the computation of seasonal averages of heat-exchange components (JACOBS, 1951; MASUZAWA, 1952; NEUMAN and ROSENAN, 1954). Almost identical formulas have been used by the workers mentioned. These formulas were worked out by KIMBALL (1928), MOSSBY (1936) and SVERDRUP (1945). HELA (1951) developed a slightly different approach for the computation of energy exchange between the sea and the atmosphere in the Baltic Sea. He analysed the exchange processes and adapted DEVIK's (1932) formulas for the computation of the exchange. In his work, HELA pointed out the relation between the direction of the wind and the sea surface temperature in the Baltic Sea and discussed other factors pertinent to hydropsis in coastal waters and semi-closed seas. JUNG and GILCREST (1955) and WATANABE (1955) investigated the heat budget of a water column in the North Atlantic and the North Pacific respectively, taking into consideration the advection and change of thermocline depth.

The aim of the present paper is to perfect existing methods and formulas for the prediction of the temperature structure in the sea and its changes caused by the changes in the heat exchange between the sea and the atmosphere in a given locality and by the advection of heat by currents.

3. PROBLEMS OF FORECASTING TEMPERATURE IN THE SEA AND AN OUTLINE FOR A FORECASTING PROCEDURE

There are many processes acting simultaneously in the sea and tending to change the temperature of the water. It is often difficult to ascertain the varying influences of different factors separately because of the difficulties in carrying out controlled experiments and the expense of collecting data at sea. In order to facilitate investigation, the problems can be grouped according to several principles. In the present paper the following sets of problems are considered:

I. The influence of factors from outside the sea which affect local changes of water temperature: that is, the exchange of heat between the sea and the atmosphere (insolation, back radiation, evaporation, precipitation, etc.).

II. Absorption of radiation in the sea.

III. Temperature structure in a given locality as determined by turbulent mixing and vertical convection. Here two different models must be considered (SCHULE, 1952):

(1) Density as a continuous function of depth;

(2) The two-layer system.

In the two-layer system it is necessary to determine:

(a) the average thickness and temperature of the upper mixed layer;

(b) the sharpness of the thermocline and related pycnocline; and

(c) periodic and non-periodic variations in the depth of the thermocline.

IV. Horizontal transport of heat by upper currents and run-off and horizontal mixing.

V. Other factors affecting the temperature structure (variations of sea level, piling-up, convergences and divergences of currents, upwelling, ice cover, etc.).

The exchange processes between the sea and atmosphere take place at the sea surface, where also the deep water masses are formed, which are later modified by mixing. Furthermore the activities of man are mainly concerned with the surface waters. Therefore, in hydropsis, the main emphasis is on the prediction of changes in the surface waters, caused by the day-to-day changes in meteorological conditions.

For application in hydropsis, it is necessary to have formulas in terms of readily measurable physical parameters, preferably those used already in routine observations. Therefore, the solution of the problems in this paper must, in the main, be sought on an empirical basis with a consideration of cause-effect principles.

Relatively little advanced statistical treatment of data has been performed in the present work, mainly because of the availability of only a limited number of samples and observations on heat exchange problems. Furthermore, emphasis has been put on the explanation of the dependence of happenings, especially in relation to routine meteorological and oceanographical measurements, and not on the determination of the probability of occurrence. KINSMAN (1957) pointed out that the data in geophysical problems, in contrast to laboratory problems, are characterized by small samples which cannot be readily extended. He suggested that the small sample theory makes feasible a division of the data into two groups, one to be used in formulating hypotheses and the other to be used as test material. In this way, the validity of any hypothesis may be established without the long delays inherent in gathering more data.

Where the physical or dynamical processes are known, the formulas have been constructed and the corresponding argument derived on the basis of known relations. If a statistical treatment of a problem has been necessary and wherever the nature and amount of the data have allowed, the regression curves have been constructed and the numerical arguments derived at first hand according to conventional statistical methods (see EZEKIEL, 1941).

In most cases, the nature of the problem and the data, and the already existing background knowledge, have made the application of routine statistical procedures undesirable. In some cases, curve fitting and the determination of empirical arguments have been carried out through trial fitting of curves with the simplest possible formulas which would yield the same or higher accuracy than the standard statistical curves. Formulas (10) and (17) can be given as examples of this kind of fitting.

4. NOTATIONS AND UNITS USED

A	— Austausch coefficient
A_m	— Amount of air (g)
A_n	— Noon altitude of the sun (degrees)
a_t	— Amplitude of the tide (m)
$a_{w(1)}$	— Percentage of energy absorbed in a layer 1
a_λ	— Atmospheric transmission
C	— Cloudiness (in tenth of the sky)
D	— Optical air mass ($D = 1$ with the sun at zenith)
$\Delta D_A - \Delta D_B$	— Difference between the dynamical depth of two stations (or of anomaly of dynamic height)
D_m	— Average depth of the thermocline (m)
ΔD_m	— Change of the depth of the thermocline (m)
D_{om}	— Thickness of the mixed homogeneous surface layer (depth to the upper limit of the thermocline) (m)
D_{tm}	— Depth of the deeper end of the thermocline [$(D_{tm} - D_{om})$ = thickness of the thermocline) (m)]
d_a	— Density of the air (≈ 1.24 g l^{-1})
d_i	— Thickness of the ice (cm)
E	— Evaporation [mm $(24\ h)^{-1}$]
e_a	— Water vapour pressure of air (mb)
e_w	— Saturated water vapour pressure at the temperature of the water surface (mb)
F	— Length of the fetch (km)
g	— Acceleration of gravity
H_u	— Humidity factor (2.0 by ICAO Standard Atmosphere)
ΔH	— Elevation of sea surface (in cm per 100 km)
H_g	— $(\varphi - \alpha_a p_h)$ — geostrophic potential
H_w	— Depth of the water (m)
$H_{1/3}$	— Average height of the 1/3 of the higher waves (significant wave height) (m)
h	— Specific heat of the air (≈ 0.238 g cal g^{-1})

Δh_s — Change of sea level (mm)

I_λ — Intensity of monochromatic energy transmitted through cloudless atmosphere

$I_{0\lambda}$ — Intensity of solar radiation on a horizontal surface outside the earth's atmosphere

K — Constant, proportionality factor in general

K_1 — Proportionality factor (explained in text where numerical values are given)

K_2 — Stefan-Boltzman constant (8.13×10^{-9} cal cm^{-2} °K^{-4}min^{-1})

K_3 — Evaporation constant (explained in text)

K_4 — Constant for Bowen's ratio (given in text)

K_5 — Conversion factor for conversion of water vapor pressure to grams of water in the air

K_h — Specific heat conductivity of the ice (given in Smithsonian Meteorological Tables)

L — Latitude

L_t — Latent heat of evaporation (g cal)

l — Thickness of the layer (cm)

l_n — Distance in nautical miles

$\left. \begin{matrix} l_1 \\ l_2 \end{matrix} \right\}$ — Thickness of layer 1 and 2 (m)

l_m — Mixing length by tidal currents

m — Percentage of reflected direct radiation

n — Percentage of reflected sky (diffuse) radiation

ΔP — The change of atmospheric pressure (mm Hg)

p — Fraction of total radiation from the sun (direct radiation)

p_a — Air pressure (mb)

p_h — Pressure

P_s — Amount of precipitation (snow) in mm. of water

Q — Amount of heat in general (g cal.)

Q_b — Effective back radiation from the sea surface (long wave radiation)

Q_{bl} — Black body radiation

Q_B — Heat from the bottom of the sea

Q_c — Heat transfer by condensation of water vapour

Q_e — Heat used for evaporation

Q_f — Heat transferred by fresh water run-off

Q_h — Net convection of sensible heat to and from the atmosphere

Q_k	—	Heat released by chemical processes
Q_l	—	Heat used in an area for the local change of water temperature (residual heat)
Q_{ob}	—	Back radiation by clear sky
Q_p	—	Heat transferred by precipitation
Q_r	—	Reflection back from the sea surface (albedo of the sea surface)
Q_s	—	Total incoming radiation (solar and sky)
Q_{os}	—	Total incoming radiation by clear sky
Q_v	—	Heat transported in or out of the area by currents
Q_{wt}	—	Heat from the dissipation of wind and tidal energy (transformation of kinetic energy)
q	—	Fraction of total radiation from the sky (diffuse)
q_w	—	Specific heat of sea water
R	—	Bowen's ratio
r	—	Percentage of total incoming radiation reflected
T	—	Temperature in general
T_a	—	Temperature of the air (°C)
T_{aw}	—	Absolute temperature of the sea surface ($= T_w + 273$)
T_{db}	—	Dry bulb temperature (°C)
T_{dw}	—	Temperature of deep water below thermocline (°C)
T_i	—	(1 and 2) Temperature of the opposite surfaces of the ice (°C)
T_{ow}	—	Initial temperature of the water (°C)
$T_{ow(m)}$	—	Average initial temperature of the surface mixed layer above the thermocline (°C)
T_p	—	Temperature of precipitation (°C)
T_t	—	Normal, average surface temperature for the date under consideration
T_w	—	Temperature of sea surface (°C)
T_{wb}	—	Wet bulb temperature (°C)
$T_{w(1)}$	—	Temperature of the water in layer 1 (°C)
$T_{w(m)}$	—	Average calculated (new) temperature of the surface mixed layer above the thermocline (°C)
$T_{w(2.5)}$	—	Temperature of the layer 0 to 2.5 m. (°C)
$T_{wa(2.5)}$	—	Temperature of the surface layer as computed with Formula 41 (no heat losses accounted for) (°C)
t	—	Time, in general
t_{at}	—	Period of tidal current

t_d — Length of the day from sunrise to sunset (minutes)
t_e — Time in days
t_h — Time in hours
t_p — Period
t_{sec} — Time in seconds
U_m — Volume of heavier water passing through unit interface in unit time
U_o — Relative humidity (%)
V — Wind speed (m sec^{-1})
V_2 — Wind speed in miles per day at the height of 2 m above the water surface
W — Velocity of the current (cm sec^{-1})
W_c — Critical velocity of mixing
W_o — Velocity of the surface layer
W_t — Mean velocity of a tidal current
W_{max} — Maximum velocity of a tidal current
x — Distance from the coast (km)
Z — Zenith distance of the sun
z_w — Depth
α_a — Specific volume anomaly (constant along δ_t — surface)
$\bar{\alpha}$ — Average solar altitude (degrees)
γ_m — Temperature lapse rate (°C km^{-1}) (6.5 by ICAO Standard Atmosphere)
ϱ — Density of the water
$\left.\begin{matrix}\varrho_1\\ \varrho_2\end{matrix}\right\}$ — Densities of layers 1 and 2
φ — Anomaly of the dynamic height

5. CHARACTERISTICS OF WATER MASSES

The surface water masses need to be defined, if possible, in terms similar to those defining air masses, for the use of general forecast in the sea. The term »surface water» refers usually to the water above the thermocline and/or to a depth of 200 m. In this paper water types have been defined in a generalizing way only, sufficient for hydropsis. The definitions are made on the basis of salinity, temperature, range of seasonal changes and »coastal influence». A separate classification of water types on the basis of optical properties has also found to be necessary, because of the variations in the absorption of insolation.

Table 1. *Characteristics of oceanic and coastal surface waters*

Property of the water mass 1	Oceanic surface waters 2	Coastal surface waters 3
Temperature	Determined by heat exchange between air — water, vertical mixing and transport by currents.	As well as col. 2, the changes are caused by intensive mixing, transport by fresh water and bottom influence.
Salinity	Uniform; determined by $E-P$ balance, vertical mixing and locally by melting of ice.	Usually lower, owing to runoff from land.
Turbidity	Low; mainly caused by plankton.	High; influenced by runoff and upwhirling of sediment from bottom by wave action.
Seasonal and Diurnal changes	Small; depends on latitude	Large, influenced by land masses.
Currents	Wind currents and permanent flow.	Tidal currents predominate; currents influenced by the morphology of the coast.
Fertility	Uniform and low, (higher in upwelling zones).	High, influenced by mixing, upwelling and runoff.

Detailed characteristics of the oceanic and coastal waters are given in Table 1. A further division of oceanic water masses can be made on a geographical basis, considering salinity and temperature and their seasonal changes. The values of these properties, given below, are approximate average criteria.

Polar waters of low temperature and low salinity ($< 8°C$ and $< 34^0/_{00}$); relatively small seasonal changes in salinity and temperature ($< 5°C$).

Boreal (or temperate) waters of medium temperature and salinity. Relatively large seasonal changes.

Tropical waters of high temperature and usually high salinity ($> 20°C$ and $> 35^0/_{00}$). In the doldrums, however, the surface salinity may be relatively low, because of precipitation. Small seasonal changes ($< 5°C$).

A further subdivision of coastal waters is unnecessary, as these waters can easily be classified by optical properties and salinities.

Mixtures of different water masses have been described by means of various terms. For hydropsis the following terms have been selected:

Subpolar — a mixture of polar and boreal water masses.

Subtropical — a mixture of boreal and tropical water masses.

(Boreal mixed — a mixture of polar and tropical water masses).

Slope water — a mixture of any offshore and coastal water masses.

It is necessary to know how much radiation is absorbed by various water masses before temperature changes can be computed, and so the optical properties of the water masses must be classified. These properties also indicate the productivity of the waters and, sometimes, their origin. The optical classification, presented in Table 2, is largely based on the works of JERLOV and KULLENBERG (1946) and JERLOV (1951). It was found necessary to combine several optical water masses defined by the

Table 2. *Optical water masses*

No.	Water mass	Characteristics
1	Oceanic, clear.	»Old» clear oceanic waters in low-productive areas (especially in low latitudes). Water colour 0 to 2 (Forel Scale). (J. Oceanic I).
2	Oceanic, normal.	Medium-productive oceanic waters in medium and low latitudes. Water colour 2 to 5. (J. Oceanic II and III).
3	Oceanic, turbid and Coastal, clear.	High-productive oceanic areas, especially during plankton bloom. Tropical coastal waters, especially over deep shelves. Water colour 5 to 8. (J. K. Coastal 1—3).
4	Coastal, normal.	Normal, medium-productive coastal waters and waters over shallow shelves. Water colour 8 to 10. (J. K. Coastal 5).
5	Coastal, turbid.	Estuarian and coastal waters during intensive plankton bloom and waters close to the coast where much sediment has been whirled up by wave action. Water colour 10. (J. K. Coastal 9).

Abbreviations: J. K. — Jerlov and Kullenberg, 1946
J. — Jerlov, 1951.

Table 3. *Values of some small terms in the heat budget*

Heat term	Value	Locality and other remarks	Author
Q_B — Heat flow through the bottom of the sea.	$50-80$ g cal cm^{-2} year^{-1}		Helland-Hansen, 1930 (from Sverdrup, Johnson and Fleming (1949). Bullard, 1954.
Q_{wt} — Heat from dissipation of wind and tidal energy	1050 g cal cm^{-2} year^{-1} (=0.002 g cal cm^{-2} min^{-1})	Irish Channel	Taylor, 1919. (from Sverdrup *et al.*, 1949).
	0.002 g cal cm^{-2} min^{-1}	Bay of Fundy (tidal energy only).	Recalculated from McLellan, 1958.
	1 g cal cm^{-2}*	Dissipation of wave energy, generated by 16.5 m sec^{-1} wind	Olson, 1959.
Q_k — Heat bound and/or released by chemical processes (mainly photosynthesis).	235 g cal cm^{-2} year^{-1} (= 0.00045 g cal cm^{-2} min^{-1}).	Assuming organic production 250 g Carbon m^{-2} year^{-1}.	Present author.

* Dissipated in unspecified time in area equal to area of generation.

above authors, because (a) a very detailed classification is only possible when actual measurements are made, and (b) the optical properties of water masses can change relatively rapidly to a certain degree (e.g. through plankton bloom, upwhirling or sedimentation of minerogen suspension), which makes the detailed classification superfluous.

6. HEAT BUDGET OF THE SEA

The amount of heat used in the change of temperature in a given locality and time can be represented by the formula below (for notations and units used, see Chapter 4):

$$(1) \quad Q_s + Q_B + Q_{wi} + Q_k + Q_f + Q_c - Q_b - Q_r - Q_h - Q_e + Q_p + \\ + Q_v = Q_l$$

The amounts of Q_B, Q_{wi} and Q_k are found in the major part of the oceans to be < 1 per cent of Q_s or even very much smaller and can therefore be ignored for practical purposes (Table 3). The formula can be reduced to:

$$(2) \quad Q_s + Q_f + Q_c + Q_p - Q_b - Q_r - Q_h - Q_e + Q_v = Q_l$$

Q_f can be ignored in offshore areas and Q_p needs be taken into account if there is a considerable amount of precipitation, if the temperature of the precipitation differs considerably from the temperature of the sea surface, or especially if the precipitation comes down as snow or hail. Q_c and Q_e are computed with the same formula, where the negative values give Q_c. In the following, a critical examination is made of the terms in equation (2) and formulas for computing these terms are derived and/or existing formulas revised.

7. INSOLATION

7.1. EARLIER WORK ON THE DETERMINATION OF INSOLATION

In most of the earlier work, the intensity of parallel monochromatic energy transmitted through the cloudless atmosphere was computed from the values of solar radiation on a horizontal surface outside the earth's atmosphere, according to the formula:

$$(3) \quad I_\lambda = I_{0\lambda}\, a_\lambda^D$$

The solar radiation on a horizontal surface outside the earth's atmosphere ($I_{o\lambda}$) [g cal cm^{-2} (24 h)$^{-1}$] is given in LIST 1951, Smithsonian Meteorological Tables. According to the latest investigations, the average solar constant is between 1.90 and 1.94 g cal cm^{-2} min^{-1} and can vary from 2 to 3%, either in very short or very long waves (DRUMMOND, 1958). Furthermore, according to DRUMMOND (1958), the solar radiation varies with the earth's distance from the sun, being *circ.* 7 per cent greater during the northern winter. The atmospheric transmission (a_λ) depends mainly on the turbidity of the air caused by dust, water vapour and water particles. The unit optical air mass (D) can be presented with the values of sec Z except for large zenith angles.

KIMBALL (1928) computed the daily totals of solar radiation received on a horizontal surface on the earth, in the absence of clouds, for various regions between 90°N and 60°S and for the 21st day of every month. He corrected the values for average atmospheric turbidity. These data have been used by the majority of workers on heat budget studies as a basis for computing insolation.

MOSBY (1936) has given an empirical formula for the determination of insolation by a clear sky:

$$(4) \qquad Q_{os} = 0.0275\bar{\alpha} \quad (\text{g cal cm}^{-2}\ \text{min}^{-1})$$

The factor 0.0275 varies slightly with the air mass and turbidity. MOSBY (1936), SVERDRUP (1945) and JACOBS (1951) have used KIMBALL's (1928) Formula (5) for the calculation of insolation by varying cloudiness:

$$(5) \qquad Q_s = Q_{os} \quad [0.29 + 0.71(1 - C)]$$

ANDERSON (1954) found that MOSBY's insolation formula (Formula 5) gives *circ.* 15 per cent too low values.

Relatively few radiation data over the sea areas were available when KIMBALL and MOSBY derived their radiation formulas. No suitable formula exists at present for the computation of insolation on a daily basis. In the following such a formula is derived, using the 102 full-day radiation measurements made on board USS »Rehobot» at different latitudes and during different seasons and kindly given to the author by the U.S. Navy Hydrographic Office. An examination of the accuracy of the insolation computed is also made, together with an examination of the factors affecting the variations in insolation.

7.2. EXAMPLES OF MEASURED DAILY INCIDENT RADIATION

The incident radiation was read from continuous pyroheliometer recordings on board USS »Rehobot» (U.S. Navy Hydrographic Office, 1955) and presented as g cal cm^{-2} for the past hour.

Examples of measured incident radiation are given in Figures 1 and 2, where cloudiness and the types of clouds are also indicated. Figure 1

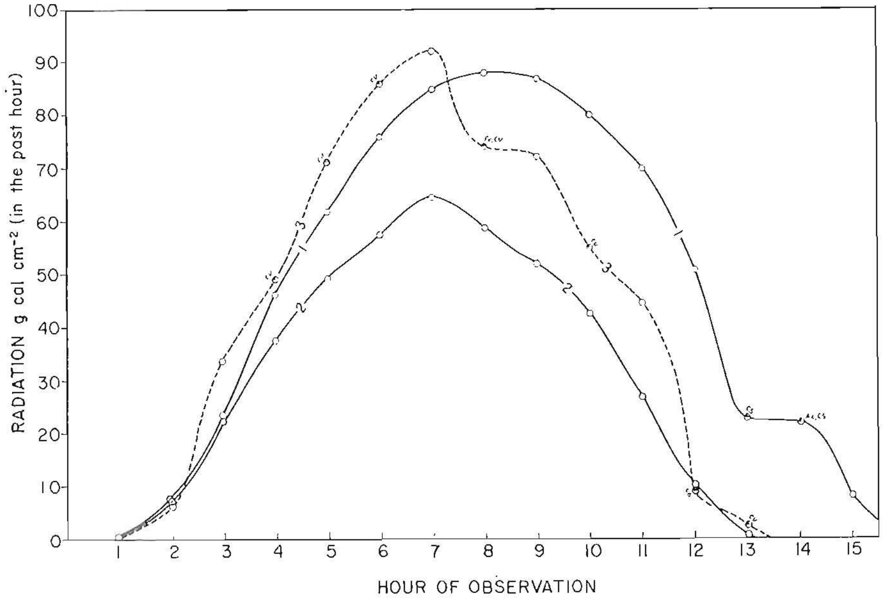

Figure 1. Examples of measured incident radiation (clear days).

1. Data for the curves

Number	Lat.	Long.	Date	Total incident radiation g cal cm^{-2} day^{-1}
1	58°N	9°E	12. III. 53	728.7
2	40°N	11°E	12. III. 51	429.2
3	23°N	57°W	5. III. 53	596.4

2. Cloudiness symbols

Cloudiness	Symbol	Cloudiness	Symbol
0 —3	○	8.1— 9.0	●
4 —5	◔	9.1—<10	●
6 —7.0	◑	10	●
7.1—8.0	◕		

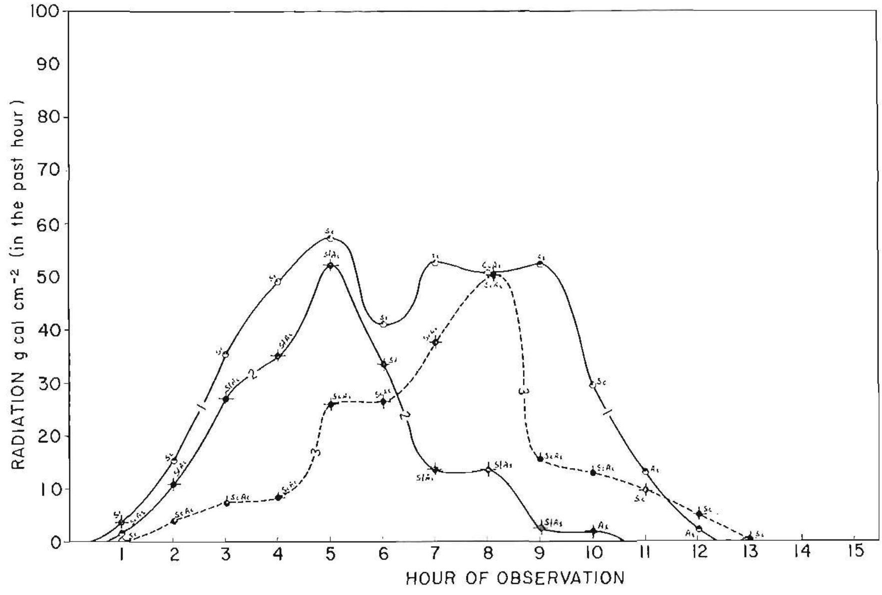

Figure 2. Examples of measured incident radiation (cloudy days).

1. Data for the curves

Number	Lat.	Long.	Date	Total incident radiation g cal cm^{-2} day^{-1}
1	41°N	11°W	28. III. 51	401.3
2	43°N	51°W	9. IX. 51	190.9
3	42°N	29°W	28. III. 52	204.6

2. Cloudiness symbols see Figure 1.

presents the insolation under a predominantly clear sky. The irregularities in curve 1 during the 13th and 14th hour of the readings are caused by changes in the cloudiness. The same cause is operative in curve 3 during the 8th hour of reading, when the temperature of the dry bulb dropped 3°F, the wet bulb temperature remaining the same as during the previous hour; consequently, the relative humidity of the air increased. As a further consequence, the visibility decreased from 37 to 13 kilometres. During the 12th hour of reading on the same day (curve 3), the dry bulb temperature dropped 2°F and the wet bulb remained the same. There was also an increase of cloudiness and a decrease of visibility (from 37 to 18 kilometres). In both cases (during the 8th and 12th hour of readings) the insolation decreased because of the increase of the cloudiness and the increase of the relative humidity, with the accompanying decrease of visibility and increase of the turbidity of the air.

Figure 2 shows examples of measured incident radiation during cloudy days and days with variable cloudiness. On curve 2 a cold front passed after the 5th hour of reading and caused considerably lower insolation for the remainder of the day, probably because of the higher relative humidity of the cold air immediately behind the front. Curve 3 on Figure 2 represents insolation during a stormy day, during which visibility varied between 3.5 and 15 kilometres.

7.3. EMPIRICAL DETERMINATION OF INSOLATION WITH A CLEAR SKY

Obviously the magnitude of the daily insolation on the ocean surface by a clear sky depends mainly on the length of the day and the noon altitude of the sun. The latter factor accounts partly for the factor a_λ^D used in formula (3). Excluding so far the corrections for cloudiness and turbidity of the air, which are considered in a later paragraph, the formula for determining incident radiation per 24 hours with a clear sky may be given as

$$Q_{os} = K_\lambda A_n t_d \tag{6}$$

The values for A_n and t_d can be computed using nautical almanacs.

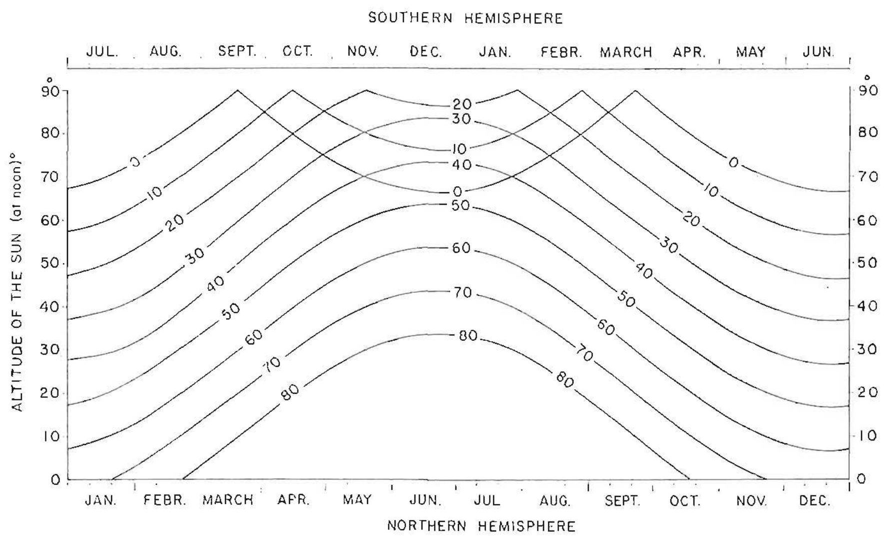

Figure 3. Noon altitude of the sun.

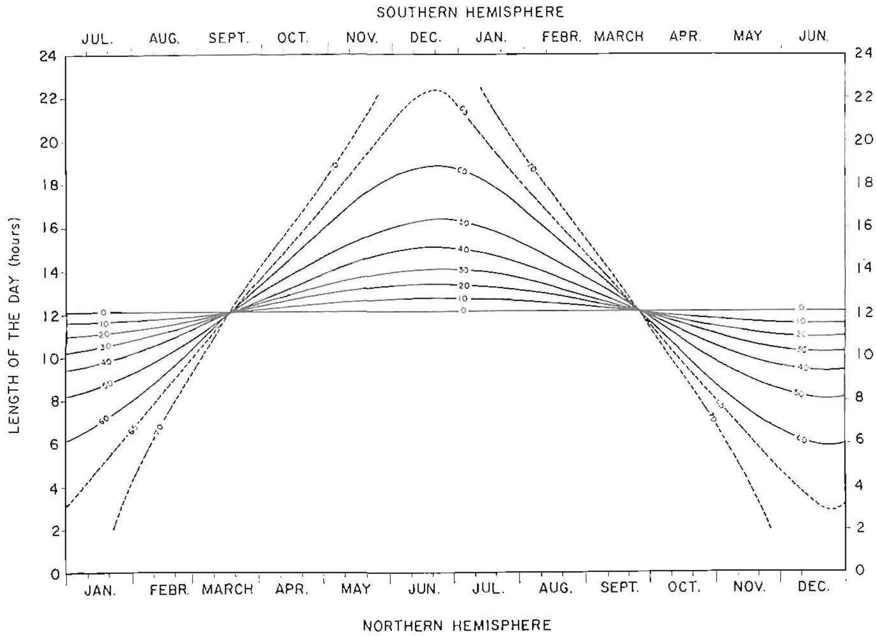

Figure 4. Length of the day.

For convenience and to save time, these values are given in graphical form in Figures 3 and 4, from which the noon altitude of the sun and the length of the day can be taken with sufficient accuracy for given dates and latitudes. The U.S. Navy Hydrographic Office (1959) prepared a more advanced solar altitude nomogram, which is especially useful for short-term (e.g. hourly or three-hourly) determination of insolation.

The proportionality factor K_1 also contains indirectly the turbidity factor of the air. As will be shown later, the variations of this turbidity factor with place and time in general affect the insolation much less than cloudiness, and therefore variations in the turbidity factor can be neglected for the present purpose. In order to determine the average value of proportionality factor K_1 empirically, the measured total daily incident radiation (g cal cm^{-2}) was divided by the length of the day (in minutes) and the factor so derived was plotted on the abscissa with the noon altitude of the sun as ordinate. The available insolation data with the cloudiness less than five tenths are plotted in Figure 5. Figure 6 contains data for cloudiness 6 to 9, and Figure 7 the insolation data with cloudiness greater than nine tenths.

There is considerable scattering of the values on the figures due to:

(1) Difficulties in determining visually the exact amount of the cloudiness.

(2) The variation of the amount, thickness, height and types of clouds from hour to hour.

(3) Variation of the water vapour content of the air and other meteorological factors.

(4) Human and instrument errors in reading the pyroheliometer and recording and computing the data.

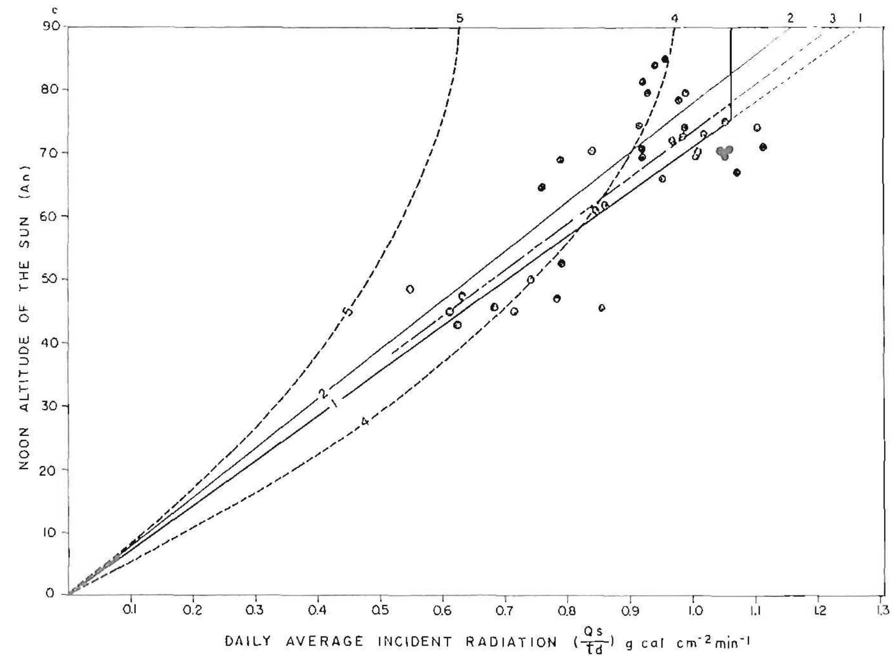

Figure 5. Average daily incident radiation at different noon altitudes of the sun (cloudiness 0 to 5).

1. Data for the regression lines

Line 1 — Cloudiness 0 [Formula (11)]
» 2 — Cloudiness 5 [Formula (11)]
» 3 — Cloudiness 3 [Formula (11)]
» 4 — Cloudiness 0 (from Kimball, 1928, Table 3)
» 5 — Cloudiness 5 (computated from Kimball's Table 3 and calculated with the formula: $Q_s = Q_{os}[0.29 + 0.71\,(1 - C)]$

In considering the plotted data (Figures 5 to 7), it is evident that the relation between the factor derived by dividing the total measured insolation by the length of the day and the noon altitude of the sun can be conveniently presented by a straight regression line with the formula:

$$Q_{os} = 0.014\, A_n t_d \quad [\text{g cal cm}^{-2}\ (24\ \text{h})^{-1}] \tag{7}$$

If more data become available, the factor 0.014 could be expressed more exactly. With a solar altitude greater than *circ.* 75° and with a clear or slightly cloudy sky, the relation seems to be non-linear. This deviation might be due to three factors: (a) the higher turbidity of the air in the tropics; (b) decrease in reflected radiation from the sky by small zenith distances of the sun; and because (c) the value of the solar constant is approached at the high noon altitude of the sun. Therefore linear Formula

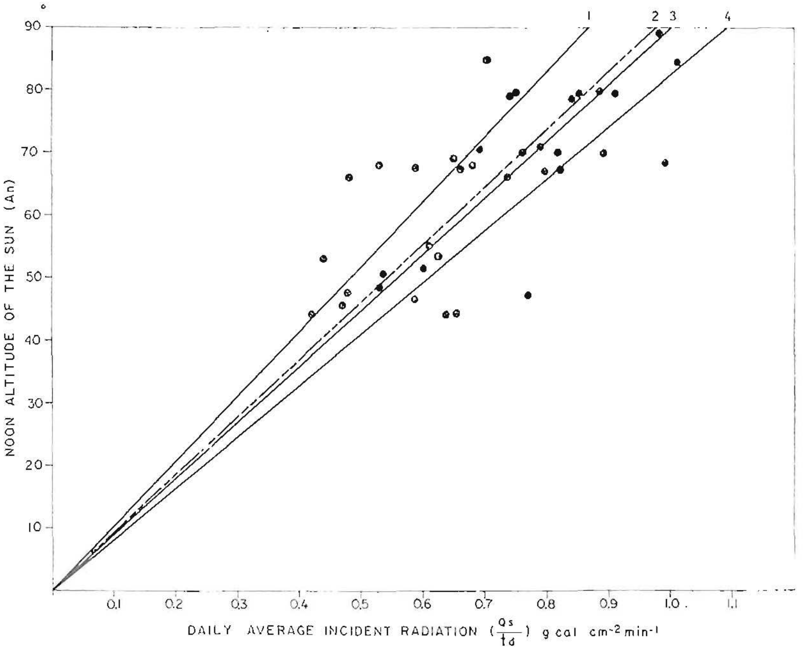

Figure 6. Average daily incident radiation at different noon altitudes of the sun (cloudiness 6 to 9).

1. Data for the regression lines

Line 1 — Cloudiness 8 [Formula (11)]
» 2 — Cloudiness 7.2 [Formula (11)]
» 3 — Cloudiness 7 [Formula (11)]
» 4 — Cloudiness 6 [Formula (11)]

(7) is valid up to about 75° of the noon altitude of the sun. Above this altitude, the product $K_1 A_n$ can be taken as constant, in which case the formula reads:

(8) $$Q_{os} = 1.06\, t_d \qquad [\text{g cal cm}^{-2}\ (24\ \text{h})^{-1}]$$

The relation of the incoming radiation to the noon altitude of the sun as computed from KIMBALL's (1928) Table 3 is given in Figure 5. His theoretical values are considerably smaller when the noon altitude of the sun lies between 50° and 80°. This might be due to the fact that the reflection of the radiation from the sky considerably increases the insolation on clear days at medium solar altitudes. The theoretical formula

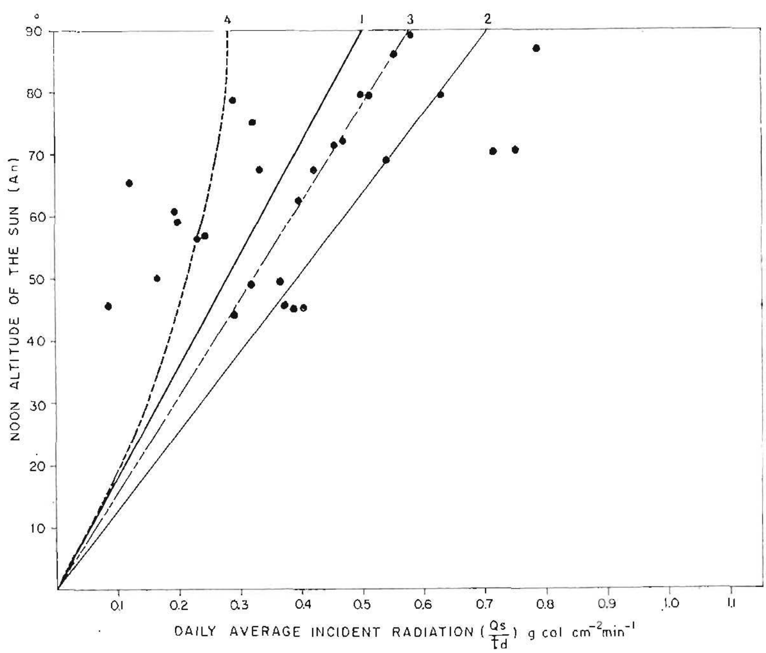

Figure 7. Average daily incident radiation at different noon altitudes of the sun (cloudiness > 9 to 10).

1. Data for the regression lines

Line 1 — Cloudiness 10 [Formula (11)]
» 2 — Cloudiness 9 [Formula (11)]
» 3 — Cloudiness 9.6 [Formula (11)]
» 4 — Cloudiness 10 (from Kimball's Table 3 and calculated with the formula: $Q_s = Q_{os}\,[0.29 + 0.71\,(1 - C)]$

used by KIMBALL does not account for this reflection and radiation from the sky.

Incident radiation measurements per minute by a clear sky, taken from the USS »Rehobot», have been plotted against the average solar altitude in Figure 8. According to this figure, the insolation can be expressed by:

$$(9) \qquad Q_{os} = 1.9 \sin \bar{\alpha} \qquad (\text{g cal cm}^{-2}\ \text{min}^{-1})$$

or by linear Formula (4) using factor 0.028, which is close to MOSBY's value. The linear formula is valid to about 67.5° average solar altitude, Q_{os} remaining constant above this altitude (= 1.9 g cal cm^{-2} min^{-1}).

The factors in Formulas (7) and (4) are not directly comparable because Formula (7) uses the noon altitude of the sun and gives the insolation per day, but Formula (4) uses average solar altitude, giving insolation per

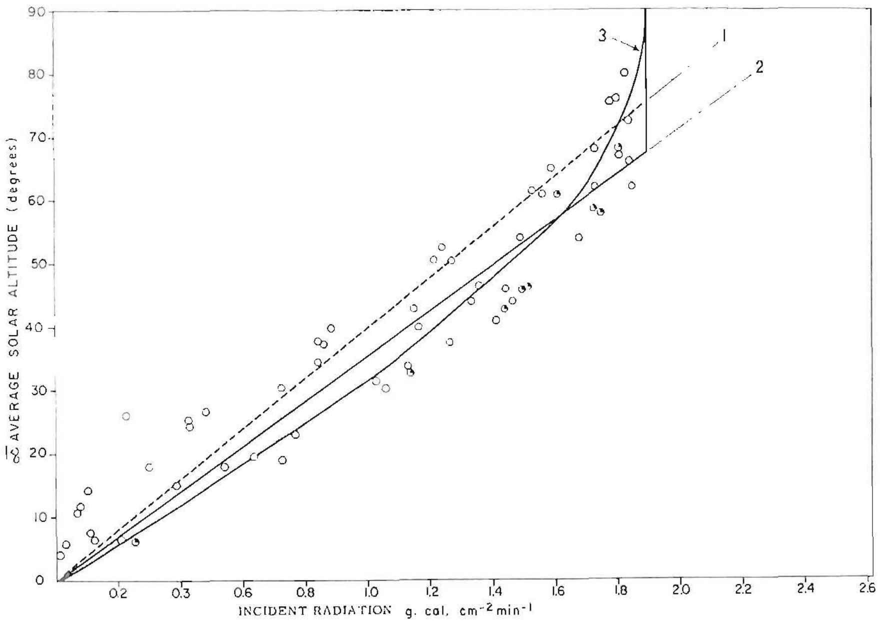

Figure 8. Incident radiation (solar and sky) with clear sky at various solar altitudes (g cal cm^{-2} min^{-1}).

1. Data for the regression lines

Line 1 — $Q_{os} = 0.025\ \bar{\alpha}$ (Mosby's formula)

» 2 — $Q_{os} = 0.028\ \bar{\alpha}$ (Mosby's formula, with modified constant)

» 3 — $Q_{os} = 1.9 \sin \bar{\alpha}$ [Formula (9)] minute. Formula (7) can be used for hydropsis for daily values and Formula (4) for three-hourly insolation reports (in connection with weather reports), taking $\bar{\alpha}$ as average solar altitude during these hours. Furthermore, the following observations could be made during the plotting of Figure 8:

(1) The factor K varies slightly both with latitude at the same solar altitude and with solar altitude at the same geographical latitude, decreasing slightly with increasing solar altitude and latitude. This variation is probably caused by: a) variation of predominant types of clouds, b) variations of atmospheric turbidity, and c) seasonal variations of the solar constant. The available data are insufficient for numerical determinations of the relations of these variations.

(2) The insolation with the same solar altitude seems to be higher in the morning than in the afternoon. The reason for this change can probably be sought in the changes of relative humidity during the day. (Variations in the amounts of precipitable water in the atmosphere can cause a $\pm$ 10 per cent variation of insolation [Olson, 1959]).

7.4. INFLUENCE OF THE CLOUDS ON INSOLATION

As seen from Figures 5 and 7, Formula (5) gives considerably lower values of insolation than the observed ones, especially with slight clouds. However, the influence of cloudiness on insolation for the available plotted data can best and simplest be represented by the following formula (see Figures 5 to 7):

$$(10) \qquad Q_s = Q_{os}\ (1 - 0.0006\ C^3) \qquad [\text{g cal cm}^{-2}\ (24\ \text{h})^{-1}]$$

Combining Formulas (7) and (10) we can compute the total insolation:

$$(11) \qquad Q_s = 0.014\ A_n t_d\ (1 - 0.0006\ C^3) \qquad [\text{g cal cm}^{-2}\ (24\ \text{h})^{-1}]$$

As seen from this formula and Figures 5 to 7, the influence of slight cloudiness on the insolation is relatively small. When cloudiness is slight and the clouds are scattered, the insolation may be even at times greater than when the sky is clear, because of the reflection from the clouds. The extent of the cloudiness is difficult to determine exactly by the currently used visual estimations. Slight cloudiness also varies considerably within a short time. Using Formula (7) for estimation of the insolation on clear and partly cloudy days (cloudiness up to 6), and ignoring the cloudiness terms, relatively small errors are introduced in the results.

The USS »Rehobot» data reveal that the amount, type, thickness and height of the clouds at sea vary considerably from hour to hour. This variation, especially around midday, probably influences the scattering of the insolation values as seen in Figures 5 to 7. Cloudiness, higher than say 5, associated with fronts and cyclones is usually easier to forecast and must be taken into account when computing the insolation. Cloudiness higher than 5 can also be estimated with greater accuracy from the insolation measurement than can cloudiness below this value.

DRUMMOND (1958) has stated that the variability in the amount and type of clouds is the principal factor in determining the value of the diffuse total radiation from hour to hour and day to day at a particular spot. HOURWITZ (from LIST, 1951) gave the ratio of total short wave radiation with overcast skies of different types of clouds to the total short wave radiation for cloudless skies. This ratio is smallest with fog, nimbus and stratus clouds (15—25 per cent), and greatest with cirrus and cirro-stratus clouds (65—85 per cent).

From the available data, it is not possible to ascertain the influence of the types of clouds on the insolation when cloudiness is less than 8. This effect of slight cloudiness is masked with the hour-to-hour variability in the amounts, types and thickness of the clouds, as mentioned earlier. With cloudiness greater than 8, and especially with cloudiness 10, it can be recognized that nimbus and low, thick stratus clouds suppress insolation more than other types of clouds. Small insolation can be especially found when the visibility is below 9 kilometres and on foggy days. The little available data do not permit exact determination of the magnitude of the influence of fog, nimbus and low stratus clouds, but it can be estimated that the correction to be subtracted is between 10 — 20 per cent.

The height of the clouds also seems to influence the insolation. With the same type and amount of clouds, insolation is smaller when the clouds are low and higher when the clouds are high. Although no definite quantitative conclusions can be drawn, on the basis of the available data, about the influence of the water vapour content of the air on insolation, it seems that high relative humidity slightly decreases insolation, especially during cloudy days.

Formula (11) was used to compute the insolation at the places and during the days where full day radiation measurements were made by

USS »Rehobot». Leaving out 26 rainy, foggy or hazy days with cloudiness higher than eight tenths, the computed values of Q_s differed $\pm$ 5 per cent or less from the measured values during 41.6 per cent of the days and $\pm$ 10 per cent or less during 51 per cent of the days. The average differences were + 8.2 per cent and — 12.3 per cent, indicating that the constant 0.014 in Formula (11) might be *circ.* 2 per cent too low. However, it is not possible to establish a more accurate constant with the available data. During the 26 rainy and foggy days the computed values were on the average 5.5 per cent higher than measured. However, among these values lower ones also occurred. During rainy days the instrumental errors are expected to be higher than during dry days. The seasonal variation of the solar constant (*circ.* $\pm$ 3.5%) has not been accounted for and therefore the overall accuracy can be considered as good.

7.5. POSSIBILITIES OF USING PYROHELIOMETER MEASUREMENTS FOR DETERMINATION OF CLOUDINESS

It is obvious that if higher accuracy in the measurement of the amount of insolation is desired, it must be measured directly. These direct insolation measurements will also permit estimation of cloudiness instrumentally. Since insolation is closely related to cloudiness, the latter can therefore be computed if the insolation measurement for a location and time is given:

$$C = \sqrt[3]{\frac{Q_{os} - Q_s}{0.0006\, Q_{os}}} \quad \text{(tenth of sky)} \tag{12}$$

Formula (12) is derived from Formula (10). The greater the amount of cloud, the more accurate will be the determination of the cloudiness. This accuracy is also desirable from the point of view of aviation, meteorological forecasts, etc. If the value of the cloudiness computed with Formula (12) is higher than 10, fog, haze, rain and the associated low visibility can be expected to prevail at the location of measurement.

It is anticipated that the above indirect measurement of the cloudiness by means of pyroheliometers will be more accurate and consistent than visual observation.

7.6. SUMMARY OF CHAPTER 7

Earlier existing formulas for computing insolation have been found inaccurate. In particular, the factors expressing the influence of the cloudiness on the insolation have been badly evaluated in the existing formulas.

The insolation measurements made at sea by the U.S. Navy Hydrographic Office have been used for deriving a new, improved formula for computing daily insolation.

Examples of daily insolation during clear and cloudy days are given. For the computation of daily insolation Formula (11) has been established, which gives insolation values which may vary on an average *circ.* $\pm$ 10 per cent from the measured values. The variations are mainly caused by the variations in the cloudiness and humidity. Seasonal variation of solar constant has not been taken into account.

For the estimation of short term insolation Formula (9) has been established, which gives approximately the same results as Formula (4), using factor 0.028. With higher altitudes of the sun, Formulas (11) and (4) reach a constant value.

A pyroheliometer can be used for measurements of the cloudiness (Formula 12) and during cloudy days can also indicate rain and fog. Properly constructed pyroheliometers are therefore also useful instruments in self-recording and reporting (telemetring) buoys.

8. RADIATION REFLECTED FROM THE SEA SURFACE

8.1. EARLIER WORK ON THE DETERMINATION OF REFLECTED RADIATION

Part of the incoming radiation is reflected from the sea surface. Referring to SCHMIDT (1915), SVERDRUP (1945) stated that the reflected amount of the direct solar radiation varies with the altitude of the sun, being 34.8 per cent at a solar altitude of 10° and 2.0 per cent at a solar altitude of 90°. SVERDRUP also presented a table for the percentage of the total incoming radiation from the sun and sky which, on a clear day, is reflected from a horizontal smooth water surface at different altitudes of the sun. Furthermore, SVERDRUP (1945) quoted POWELL and CLARKE, whose measurements on clear days were in agreement with the data from SCHMIDT though on overcast days the reflection of diffuse radiation amounted to 8 per cent of the incoming radiation. SVERDRUP gave the following formula for the computation of reflected radiation; the direct radiation from the sun and the scattered radiation from the sky were considered separately:

(13) $$r = mp + nq$$

JACOBS (1951) used the values for reflected radiation derived by SCHMIDT (1915), which are given in tabulated form also by MOSBY (1936) and show a variation with latitude from a minimum of 3.3 per cent at the equator to 8.0 per cent at the poles. These values are invalid, as shown by BURT (1953). NEIBURGER (1954) found that the reflection from the sea surface is probably 9.5 per cent of the incident radiation. GODSKE, BERGERON, BJERKNES and BUNDGAARD (1957) quoted the same data as SVERDRUP (1945) but seem to have overlooked the fact that BURT (1953) discovered that SCHMIDT made an error in his computations.

SIMOJOKI (1949) assumed the reflection from the surface in the Baltic Sea to be 7 per cent in the period May to September and 10 per cent at other times. These assumptions are close to the real values as shown later. ANDERSON (1954) found that the average reflection in Lake Hefner was 6 per cent. He presented the reflection as a function of the solar altitude:

(14) $$r = 1.18\, \bar{\alpha}^{-0.77}$$

He furthermore found that the constants 1.18 and — 0.77 vary only slightly with the height and amount of the clouds.

Considering the variety of views and formulas for computing the amounts of radiation reflected from the sea surface, a re-examination of this problem is necessary.

8.2. EXAMPLES OF MEASURED DAILY REFLECTED RADIATION

Some examples of reflected radiation measured on board USS »Rehobot» (U.S. Navy Hydrographic Office, 1955) are given in Figure 9. In the morning and evening hours, when the altitude of the sun is low, the amount of reflected radiation increases rapidly with the increasing insolation, reaching a maximum value *circ.* 3 hours after sunrise (or 3 hours before sunset). On clear days (curve 3) the total amount of reflected radiation decreases at midday. On partly cloudy and cloudy days it remains more or less constant during the day between *circ.* 3 hours after sunrise and 3 hours before sunset, or changes with the varying insolation caused by cloudiness.

8.3. EMPIRICAL DETERMINATION OF THE PERCENTAGE OF REFLECTED RADIATION

Two separate ways must be selected in determining empirically reflected radiation: if the reflected radiation is determined per 24 hours, it

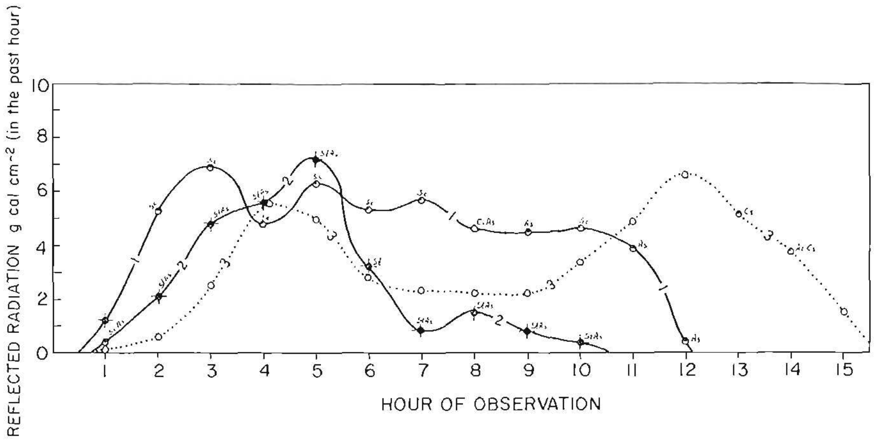

Figure 9. Examples of measured reflected radiation.

1. Data for the curves

Number	Lat.	Long.	Date	Reflected radiation g cal cm^{-2} day^{-1}
1	41°N	1°W	28. III 51	53.6
2	43°N	51°W	9. IX 51	26.8
3	58°N	9°E	12. VIII 53	48.7

can be related to the total incoming radiation; if the reflected radiation is determined per minute (e.g. for 3 hourly values), it must be related to the average solar altitude.

8.3.1 *Reflected radiation in 24 hours*

A trial was made to relate the daily reflected radiation to the noon altitude of the sun, the length of the day and the cloudiness. Although there seems to be an indication of quantitative relation, as expected theoretically, between the daily percentage of reflected radiation and the noon altitude of the sun, the relatively scattered values, and the lack of values, when the noon altitude is less than 42°, do not allow statistical treatment.

Another attempt to relate the reflected radiation to incident radiation during the same day is shown in Figure 10.

From this figure two alternative tentative solutions can be offered for the computation of the amount of daily reflected radiation in relation to daily insolation:

(1) Assuming a linear relation, the data presented can be expressed with the formula:

(15) $$Q_r = 0.087\, Q_s \quad [\text{g cal cm}^{-2}\ (24\ \text{h})^{-1}]$$

This formula gives a higher value than the percentage reported by SCHMIDT (1915) and used by JACOBS (1951) and than that reported by ANDERSON (1954) but is slightly less than the average found by NEIBURGER (1954).

(2) Assuming a nonlinear relation, the average for given data can be expressed by:

(16) $$Q_r = 0.15\, Q_s - (0.01\, Q_s)^2 \quad [\text{g cal cm}^{-2}\ (24\ \text{h})^{-1}]$$

Both formulas are valid only for daily reflected radiation, and the regression lines of these formulas are shown in Figure 10.

A comparison of the reflected radiation, computed with Formulas (15) and (16) using computed Q_s, with the measured radiation shows

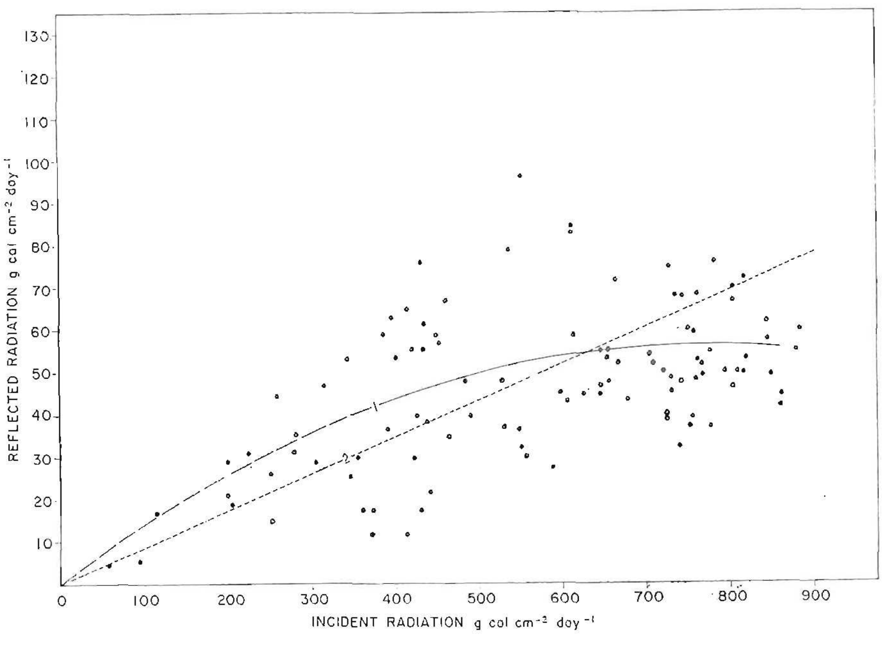

Figure 10. Reflected daily radiation versus incident daily radiation,

1. Data for the regression lines

Line 1 — $Q_r = 0.15\ Q_s - (0.01\ Q_s)^2$ [Formula (16)]

» 2 — $Q_r = 0.087\ Q_s$ [Formula (15)]

that the reflected radiation, computed with Formula (15), varies in 48.5 per cent of cases less than $\pm$ 25 per cent from the measured values, and the average variations are + 28.1 per cent and − 45.3 per cent. (A $\pm$ 25 per cent variation in the reflected radiation causes a $\pm$ 2.2 per cent variation in the total insolation.) The values computed with Formula (16) vary in 59.2 per cent of the cases less than $\pm$ 25 per cent from the measured values, and the average variations are + 27.4 per cent and − 29.9 per cent. Formula (16) is therefore more accurate and gives a higher percentage reflected when the incident radiation is smaller, compensating therewith the effect of neglected solar altitude factor in the reflected radiation. The incident radiation is smaller when the noon altitude of the sun is low and the degree of cloudiness usually high in higher latitudes. In both cases, data support the theory that the percentage of *daily* reflected radiation *is higher*. Formula (16) gives approximately the same shape of curve, with very slightly higher values, as that of the reflected radiation measured on Lake Hefner by ANDERSON (1954).

The data do not permit a quantitative determination of the relation between the cloudiness, or the types of clouds, and the daily percentage of reflected radiation, although there is a slight qualitative indication, and also a theoretical reason (see Figure 9) why the percentage of reflected radiation is slightly higher during cloudy days than during clear days.

The influence of cloudiness on the reflected radiation is relatively small as can be seen from ANDERSON's (1954) work. The cloudiness and the height and types of clouds vary considerably during a day and are difficult to forecast with accuracy. Therefore the cloudiness factor can be neglected in computing the average daily reflected radiation.

There seems to be some relation between the length of the day and the radiation reflected during clear days with the same noon altitude of the sun, because the percentage of reflected radiation is lower between the 3rd hour after sunrise and the 3rd hour before sunset during clear days.

8.3.2 *Reflected radiation during short periods*

The percentage of reflected radiation during one hour versus the average solar altitude during the same period are plotted in Figure 11. Generally, data for a clear sky are used. The scattering of values is as great as in Figure 10, due to many factors (e.g. instrument errors, influence of the state of the sea surface, etc.). During the plotting it was found that the percentage reflected was slightly lower in the morning than in the afternoon, as is also indicated in Figure 9 (curve 3).

The following tentative empirical relation could present the percentage of reflected radiation at various solar altitudes (minimum *circ.* 7.5°):

$$r = \frac{300}{\bar{\alpha}} \tag{17}$$

This formula gives higher values than that of SCHMIDT (1915) but agrees reasonably well with the values reported by ANDERSON (1954). However, Formula (17) fits present data better and is more convenient to use than that presented by ANDERSON (Formula 14). A detailed evaluation of the reflected radiation during short periods is in progress in the U.S. Hydrographic Office (OLSON, M.S.).

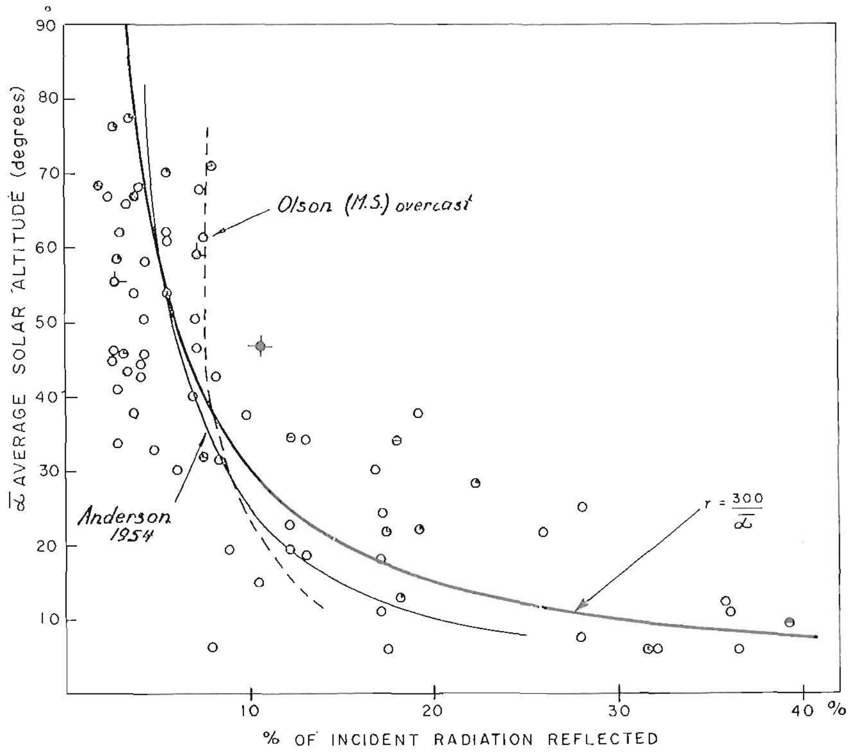

Figure 11. Relation between percentage of reflected radiation and average solar altitude.

8.4. SUMMARY OF CHAPTER 8

Several widely used formulas for the computation of the radiation reflected from the sea surface have been shown to be in error. No generally valid formula for the computation of the daily reflected radiation existed previously. On the basis of the available measurement of the daily reflected radiation on board USS »Rehobot» Formula (16) for the estimation of the daily reflected radiation has been given, as well as a tentative Formula (17) for the estimation of albedo during short periods. The factors influencing the albedo of the sea surface are briefly discussed and the relative accuracy of the formulas is indicated.

9. EFFECTIVE BACK RADIATION FROM THE SEA SURFACE

The measurement of long-wave radiation over the sea presents some technical difficulties. However, this radiation must be taken into account in the heat budget studies of given water masses, because the loss of heat by long wave radiation from the sea surface is considerably greater than the long wave radiation received by the sea surface. This difference might account, under certain conditions, for more than 10 per cent of the incoming radiation from the sun and sky.

The sea surface emits long-wave radiation as an almost black body. There are no new measurements of this radiation available for discussion in this paper, and therefore a short summary of the present knowledge and controversies thereon will be presented here.

Q_{bl} for sea water is given by WATANABE (1955) with the formula also used by SVERDRUP (1945):

$$Q_{bl(sea)} = 0.94\, K_2 T_{aw}{}^4 \quad (\text{g cal cm}^{-2}\ \text{min}^{-1}) \tag{18}$$

ANDERSON (1954) has used the same formula, except that his factor is 0.970. However, he also considered reflected long-wave radiation, which amounts to *circ.* 3 per cent and makes approximately the difference between these two factors. The sea surface also receives long-wave radiation from the sky. No applicable formula exists for the computation of this atmospheric long-wave radiation. The best approach is therefore to account for the »effective back radiation» leaving the sea surface. ÅNGSTRÖM (1920) (quoted by SVERDRUP, 1945) has computed the effective back radiation with a clear sky, taking into consideration the humidity of the air and making allowance for the fact that back radiation from the sea surface is about 6 per cent less than radiation of a black body. SVERDRUP (1945) also mentioned that ÅNGSTRÖM's data are somewhat greater than the values obtained by BRUNT's empirical formula:

$$Q_{ob} = Q_{bl}\,(1 - 0.44 - 0.8\sqrt{e_a}) \tag{19}$$

ANDERSON (1954) found that ÅNGSTRÖM's modified straight-line relationship expressed the relation between atmospheric radiation and local vapour pressure adequately.

For the correction of back radiation for cloudiness effect MOSBY (1936) and SVERDRUP (1945) used the factor $(1 - 0.083\,C)$. According to DIETRICH and KALLE (1957) the latest cloudiness factor of MÖLLER (1953) is:

$$Q_b = Q_{ob}\,(1 - 0.0765\,C) \quad (\text{g cal cm}^{-2}\ \text{min}^{-1}) \tag{20}$$

The effective back radiation depends also upon the altitude and density of the clouds. With a sky completely covered with cirrus, the effective back radiation is 0.75 Q_{ob}, with altostratus 0.4 Q_{ob} and with strato cumulus 0.1 Q_{ob}. The back radiation Q_{ob} has relatively little variation between day and night (except the differences caused by diurnal variations in T_w and e_a).

LÖNNQUIST (1954) has reviewed the existing measurements and formulas for effective back radiation (long-wave radiation). The true magnitude of this radiation by the ICAO Standard Atmosphere is still unknown. Eight plausible empirical values (LÖNNQUIST, 1954) give an average of 0.144 g cal cm^{-2} min^{-1}. The linear formula of LÖNNQUIST for effective back radiation (for clear sky) is in mw cm^{-2}:

$$Q_{ob} = 16.2 - 0.09\,T_w - 0.046\,U_o + 0.12\,\gamma_m - 1.3\,H_u \tag{21}$$

If upper air observations are not available, we can replace γ_m and H_u with the average values for the ICAO Standard Atmosphere (6.5 and 2.0 respectively). Furthermore, we can convert mw cm^{-2} to g cal cm^{-2} min^{-1} and we get the following formula:

$$Q_{ob} = \frac{14.28 - 0.09\,T_w - 0.046\,U_o}{69.72} \quad (\text{g cal cm}^{-2}\ \text{min}^{-1}) \tag{22}$$

This formula gives a value of 0.145 g cal cm^{-2} min^{-1}, for the ICAO Standard Atmosphere, which is in agreement with the plausible empirical value and is therefore the best formula available. The graph in Figure 12 has been computed with the use of this formula, taking various temperatures and humidities. Besides humidities, which have already been considered in Formula (22), the effective back radiation is also dependent on cloudiness and height and type of the clouds. Although there is a considerable scattering of the measured values, reported by ANDERSON (1954), they suggest, however, that the formula used by MOSBY (1936) is approximately correct. Taking into consideration the latest improvements by MÖLLER, the present author suggests the use of Formula (20) for the correction of the effective back radiation for the cloudiness effect.

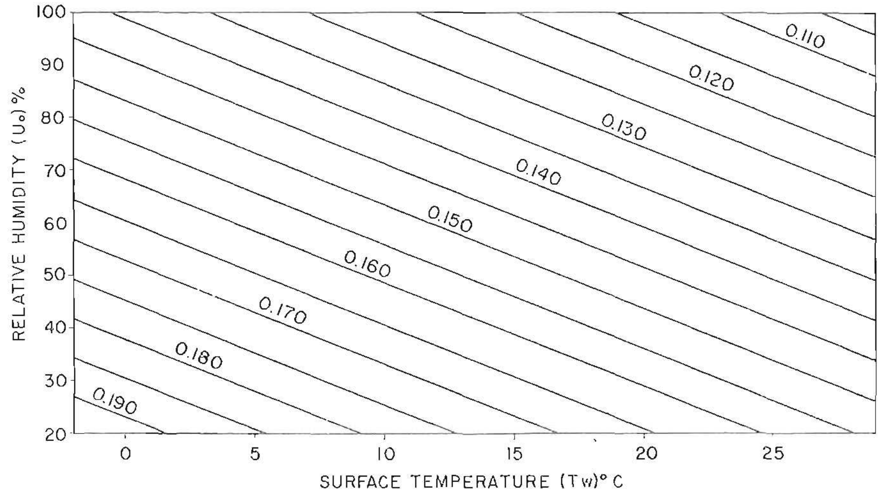

Figure 12. Effective back radiation from sea surface to a clear sky (Lönnquist's formula, modified).
g cal cm^{-2} min^{-1}

9.1. SUMMARY OF CHAPTER 9

The earlier practices for the estimation of long-wave radiation to and from the sea have been reviewed. LÖNNQUIST'S Formula (21) for the estimation of effective back radiation has been found most accurate. When substituting the temperature lapse rate and humidity factor in his formula with those defined in the ICAO Standard Atmosphere and converting the formula to g cal cm^{-2} min^{-1} [Formula (22)], it gives a value for the ICAO Standard Atmosphere which is nearly identical with the mean plausible empirical value. The cloudiness factor of MÖLLER [Formula (20)] is recommended for use.

10. LOSS OF HEAT FROM THE SEA BY EVAPORATION

10.1. EARLIER WORK ON THE DETERMINATION OF EVAPORATION FROM THE SEA

The rate of evaporation depends on (a) the water vapour deficit of the air above the water $(e_w - e_a)$, and (b) the factors affecting the removal of the moisture saturated air above the surface (wind speed and thermal convection).

Several formulas for the determination of the annual and seasonal evaporation from the heat balance and from the surface salinity values have been used by many workers, but as they cannot be used for the purposes of this paper, they will not be discussed here. DEACON, PRIESTLEY and SWINBANK (1958), and SVERDRUP (1951), have given various theoretical and hydrodynamical approaches (including mass transport formulas) for evaporation calculations and have described the practical difficulties met in the use of these approaches. SVERDRUP (1951) concluded that different theoretical approaches lead to different evaporation factors and are useless until valid arguments can be advanced for accepting one of the many models. Also MARCIANO and HARBECK (1954) found that all theoretical evaporation formulas are unsatisfactory.

There are several empirical formulas for the estimation of evaporation which can be expressed in the general form (DALTON's formula):

$$E = K_3(e_w - e_a)V \tag{23}$$

The comparison of the values for the constant K_3 as given by SVERDRUP (1945) [3.6 cm year^{-1} or 0.099 if expressed in mm (24 h)$^{-1}$] and the one given by JACOBS (1951) as 0.143 [mm (24 h)$^{-1}$] reveals that the former is *circ.* 31 per cent smaller. The above-mentioned constants are applicable only to climatic data, although they have also been used by several authors for short periods. According to SHULEIKIN (1953), MIKHALEVSKII has used a factor of 0.157 in Formula (23) [mm (24 h)$^{-1}$]. In an earlier work, JACOBS and CLARKE (1943) used the value 0.149 when e_a and V were measured 3.6 metres above the sea level.

JUNG and GILCREST (1955) used a modified DALTON's formula, incorporating the theoretical approach used by MONTGOMERY, SVERDRUP and others in the study of diffusion of water vapour. The basic idea of the modifications was that the sea surface is changed from smooth to rough by a wind speed of approximately 6.5 m sec^{-1} and that the evaporation from a smooth surface is different than from a rough one. PRIVETT (1958) doubted whether evaporation is considerably increased when the sea surface changes from smooth to rough. The same doubt has been expressed by DEACON, PRIESTLEY and SWINBANK (1958): »Sverdrup's aerodynamical consideration that the rate of evaporation from a rough water surface may be two to four times that from smooth water must be open to serious doubt». MARCIANO and HARBECK (1954) also concluded from observations on Lake Hefner that a natural water surface is always to be considered as rough. However, SVERDRUP (1951) also found that the empirical evidence is too meagre to allow a firm conclusion that the sea surface is hydrodynamically smooth at low wind speeds.

It is often pointed out that a considerable amount of spray is brought into the air by higher wind speeds and that in consequence the evaporation must be considerably increased by such winds. However, quantitative consideration of the amounts of the sea salts in the precipitation, on an average 6.2×10^{-4} g of NaCl m^{-2} per day (JUNGE and GUSTAFSON 1957), shows that thiscon tribution by spray is relatively unimportant. In the Lake Hefner studies ANDERSON (1954) also found that the effect of spray and the stability of the air are negligibly small. According to JACOBS (1951) the same conclusion has been reached by SVERDRUP.

DEACON *et al.* (1958) advocate the latest revised DALTON formula by PENMAN (1948, 1956), who has compared the available evaporation measurements in different conditions and derived the following formula (water vapour pressure being here in mm Hg):

$$(24) \qquad E = (e_w - e_a)\, 0.35\, (1 + 9.8 \times 10^{-3}\, V_2) \qquad [\mathrm{mm}\ (24\ \mathrm{h})^{-1}]$$

JACOBS and CLARKE (1943) found through evaporation measurements on board the »Carnegie» that the data showed an increasing rate of evaporation with increasing air temperature. WÜST (1957) presented a modified DALTON's formula (a slightly modified version of the one in his earlier (1920) fundamental work on the subject) where the temperature of the air was taken as a factor. HELA (1951) used the formula deduced by DEVIK (1932), where the temperature was also considered:

$$(25) \qquad Q_e = 3.02\, \frac{T_w}{P_a} \sqrt{V + 0.3\, (e_w - e_a)} \qquad (\mathrm{g\ cal\ cm^{-2}\ h^{-1}})$$

(P is in mb and e in mm Hg). Later works have shown no direct relation between evaporation and temperature, except that e_a is dependent on T_a.

One of the most exhaustive and careful works on the evaporation from free water surfaces during 24 hour periods is by ROHWER (1931), but it seems to have escaped the attention of those working on this subject, probably because it was published in an agricultural series. ROHWER reviewed all the existing empirical formulas and arrived, after several years of careful experimental work and analyses of voluminous amounts of evaporation measurements in various conditions, at the following empirical formula (recomputed for the $c\,g\,s$ system, with the altitude term eliminated for the measurement of evaporation at sea level):

$$E = (0.26 + 0.154\,V)\,(0.98\,e_w - e_a) \quad [\text{mm} \quad (24\ \text{h})^{-1}] \tag{26}$$

KOHLER (1954) used a similar formula for the Lake Hefner evaporation studies. By averaging the results of the measurements in his four different evaporation pans and recomputing for the $c\,g\,s$ system, the following formula can be derived:

$$E = (0.47 + 0.225\,V)\,(e_w - e_a) \quad [\text{mm}\ (24\ \text{h})^{-1}] \tag{27}$$

The direct lake evaporation, as computed from heat and water budget studies, gave:

$$E = (0.13 + 0.138\,V)\,(e_w - e_a)\ [\text{mm}\ (24\ \text{h})^{-1}] \tag{28}$$

These formulas are not directly comparable with ROHWER's Formula (26), because ROHWER used wind speed at the water surface and KOHLER used average wind speed at higher levels. Therefore a mean for the reduction of the wind speeds to the speeds at the water surface is sought in the following section.

10.2. WIND SPEED PROFILES OVER THE SEA

The constants in the DALTON's type formulas depend on the height above the sea surface at which e_a and V are measured. ROHWER (1931) presented a graph for the reduction of wind speeds measured at different heights above the ground to the velocity at the water surface.

A revaluation of ROHWER's graph was found necessary, because the lapse rate of the wind speed over the sea might differ from that of the wind speed over the land. Furthermore, it was found necessary to evaluate the wind speeds measured at different levels to the »normal» level of 8 metres, to which the Beaufort scale usually refers. Several available estimates of wind speeds over the water are summarised in Figure 13 and an average has been selected.

The water surface speed has been defined for practical reasons as the wind speed 5 cm above the water surface. This selected lower level is approximately the upper boundary of a great temperature instability layer close to the water surface, as the measurements of ROLL (1952) show.

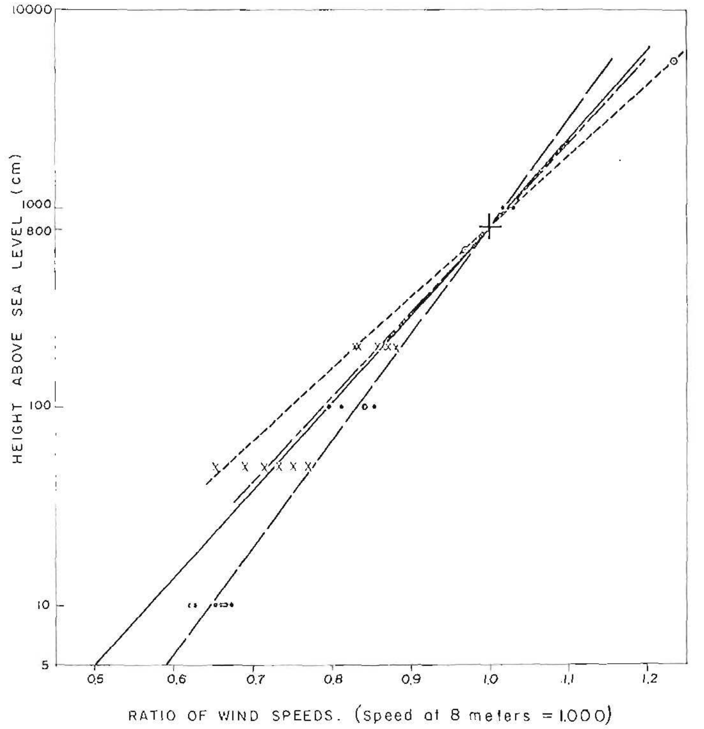

Figure 13. Variation of wind speed with height above sea level.

—·— ×	Hay (1955)	— — — —⊙	Texas Tower No. 2
— — ∘	Roll (1949)	————	Selected average

The wind speed is assumed to increase in neutral conditions linearly with the logarithm of the height (ROLL 1949, CHARNOCK 1956). MARCIANO and HARBECK (1954) found on Lake Hefner that the wind profiles were logarithmic regardless of the stability of the air. The relations of wind speeds at 8 and 2 metre levels varied from 1.239 to 1.269.

Several investigators have pointed out that the lapse rate varies slightly with the stability of the air. BROOKS and BROOKS (1958) have investigated the accuracy of wind speed estimates at sea and conclude that »colder water produces a more or less stable layer of chilled which air inhibits wind thrusts to the surface, especially in winds not exceeding 4 Beaufort (13—16 knots). It cools the entire turbulent layer of air and, thereby, reduces the general turbulence and increases also the vertical gradient of wind speed so that the wind at the surface will actually be considerably less than that on the mast». However, TAKAHASHI's (1958) measurements of wind speeds at various levels over a bay show the opposite relations. His data reveal that the ratio of the wind speed

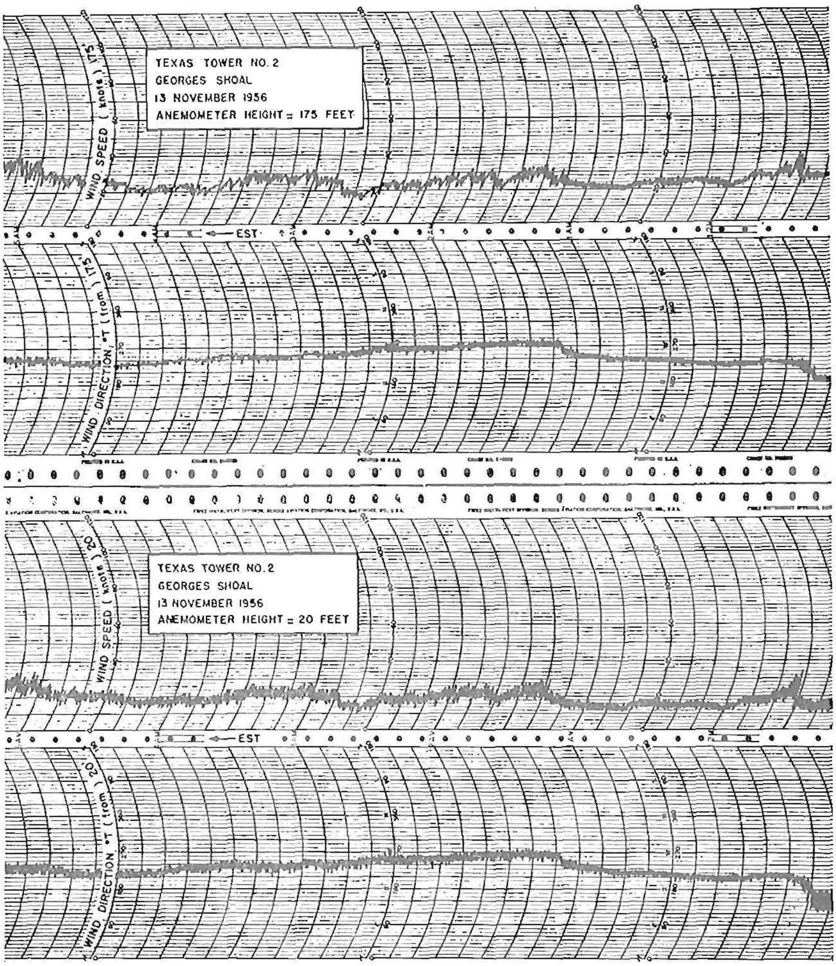

Figure 14. Examples of continuous wind registrations at 20 and 175 feet levels in Texas Tower No. 2.

at 0.25 and 4 metres levels in stable conditions is 1.34, in neutral conditions 1.39 and in unstable conditions 1.47.

The U.S. Navy Hydrographic Office has kindly put at the author's disposal four wind registrations at heights of 20 and 175 feet taken at Texas Tower No. 2 on Georges shoal (see examples Figure 14).

From these registrations the following additional observations can be made:

(1) At higher wind speeds, the lapse rate of the speed is greater than at low wind speeds. (see also Roll, 1949).

(2) The gustiness of the wind is greater at a lower (20 feet) level than at a higher (175 feet) level, but the gusts have longer duration at a higher level; if the speed is low, the gustiness is nearly the same at both levels.

(3) There is a slight difference of wind direction between the two levels. The short-period variations of wind direction are greater at the lower level.

Considering the great variability of wind speed and the results of other investigations described above, it can be concluded that no great error in the estimation of wind speeds at various levels is introduced when the stability conditions are left out of consideration. It is estimated from Figure 13 that the wind speed at the water surface is about half of the speed at the 8 metre level. The profiles of e_a and T_a vary similarly with height above the sea. e_a and T_a are usually measured on the ships bridge the height of which is usually not reported and no conversion to »standard» height has yet been established.

10.3. COMPARISON OF VARIOUS EMPIRICAL FORMULAS FOR ESTIMATING EVAPORATION

A change in evaporation with a change in the roughness of the sea surface has not been established by the measurement of evaporation. Although the amount of spray might slightly influence the evaporation locally, it has been shown to be relatively unimportant, as discussed above. In consequence, the formulas of Jung and Gilcrest (1955) and others are unnecessarily complicated. Although the influence of air temperature on evaporation has been observed, the bulk of the evaporation data does not show any direct correlation with the actual temperature of the air, and these formulas are not considered in the comparison below. Formulas (27) and (28), although very similar to

Formulas (24) and (26), were derived specially for one lake and are therefore not considered in the comparison.

Formulas (24) and (26) are very close to each other, if expressed in the same units. Both are based on numerous evaporation measurements. Formula (26), however, is based on a very large amount of data from various conditions and is considered to be better suited for evaporation estimates from natural water surfaces. It remains to compare Formulas (26) and (23).

From Figure 13 and from discussions in Chapter 10.2 it can be concluded that if the wind speed is estimated by force (on the Beaufort scale) and converted to m sec^{-1} ROHWER's formula can be revised to:

$$(29) \qquad E = (0.26 + 0.077\ V)\,(0.98\, e_w - e_a)\ [\text{mm cm}^{-2}\ (24\ \text{h})^{-1}]$$

The e_w and e_a have diurnal variations, and the selection of the proper 24-hour average value is described in a later chapter. If the depression of the wet bulb temperature is known, the e_a is taken from proper tables or derived by multiplying the relative humidity with the vapour pressure at saturation at the same temperature. The factor 0.98 can be eliminated if the saturated vapour pressure for sea water is taken from MIYAKE's (1952) table.

A comparison between Formulas (23) and (29) is made in Table 4, using the actual evaporation measurements made at sea on board the »Carnegie» (JACOBS and CLARKE, 1943). It should be noted that the measurements do not present the actual evaporation values from the sea surface and that the correction factor 0.53 applied by WÜST is not well established. This correction factor was adopted by WÜST in 1936 when relatively few data on the subject were available. SVERDRUP (1951) found it also uncertain and WÜST agrees (in a personal communication) that according to recent estimates this correction factor should be somewhat higher. KOHLER (1954) found the pan coefficients of the four pans used in the Lake Hefner studies to vary between 0.69 and 0.91. These pan factors vary from type to type. Considering the above, and, in addition, the fact that the evaporation measurements on board ship are likely to give higher values than the measurements made by KOHLER on the lake, the measured evaporation values in Table 4 have been corrected with the factor 0.65. It is also noted that JACOB's and CLARKE's computed values are increased by a factor of 1.33, because they used mm. instead of mb. and did not adjust the factor accordingly. As the wind speed factor in

Table 4. *Twenty-four-hour values of sea water evaporation*
(Measurements made on board »Carnegie» 1928—29)

Run No.	From Jacobs and Clarke 1943: Mean relative humidity %	Mean wind speed m sec^{-1}	Total 24 h measured evaporation mm	Present author: Corrected values, factor 0.65 mm	Calculated after Sverdrup $E=0.149(e_w-e_a)V$ mm	$E=(0.26+0.116V)(e_w-_a)$ mm
1	2	3	4	5	6	7
1	90	3.8	1.99 (a)	1.30	0.90	1.12
2	76	4.8	6.68	4.35	6.74	7.70
3	75	5.4	9.60	6.24	8.10	8.95
4	79	6.1	10.12	6.60	6.50	6.95
5	85	4.5	6.13	3.99	3.38	3.96
6	82	1.3	5.91	3.85	1.16	2.47
7	83	1.8	5.98	3.89	2.14	3.84
8	83	5.7	6.34	4.12	6.66	7.25
9	83	5.0	2.49	1.62 (b)	5.43	5.58
10	81	4.0	5.06	3.30 (c)	2.84	3.48
11	73	3.2	7.31	4.75 (d)	3.10	4.13
12	80	2.4	4.75	3.09 (e)	1.33	2.01
13	83	3.1	2.84	1.85 (f)	1.53	2.07
14	75	4.3	8.50	5.53	4.16	4.96
15	80	3.8	7.25	4.72	3.52	4.39
16	72	4.2	9.94	6.45	5.14	6.17
17	78	3.2	7.21	4.70	2.66	3.48
18	73	4.7	4.11	2.67	3.90	4.43
19	74	2.3	5.72	3.72	2.03	3.17
20	91	5.2	3.38	2.20	0.72	0.80
21	88	3.6	3.96	2.58 (g)	0.93	1.18
22	67	0.4	6.48	4.22	0.53	2.75
23	81	3.0	6.36	4.14	1.85	2.52
Means:	79.7	3.73	6.22	3.91	3.27	4.06

(a) — Salt spray in air; (b) — Evaporimeter not moved on account of spray; (c) — Heavy dew during night; (d) — Volume of vessel 1791 cc.; (e) — Little direct sunshine during run; (f) — Vessel upset, sample collected at once; (g) — Heavy dew during first night.

Formula (29) 0.116 V has been used, because the wind speed presented is an average of those measured at the evaporation pan and those at the rail. It can be noted that Formula (29) gives slightly higher values than the measured amounts corrected with the factor 0.65. The values given by Formula (29) deviate less from the actual measurements, especially at low wind velocities, than those given by Formula (23).

MARCIANO and HARBECK (1954) advocated that the extra term in DALTON's type formulas is unnecessary, because evaporation in the absence of wind is a slow molecular diffusion process. However, in the opinion of the present author, the air over the ocean is scarcely absolutely quiet if the sea surface is warmer than the air; convective movements are caused by this temperature difference. Furthermore formulas without an extra term are for use with climatological data. Therefore the extra term present in ROHWER's modified formula should remain for 24 h formulas. However, in certain cases the extra term can and should be neglected when the difference $e_w - e_a$ is negative and the wind speed low, as shown in Chapter 12. In addition, Formula (29) is based on numerous measurements and it gives more consistent values as compared to actual measurements at sea. Therefore it is concluded that ROHWER's modified Formula (29) is the best available for the computation of evaporation from the sea surface.

10.4. SUMMARY OF CHAPTER 10

Earlier existing formulas for the computation of evaporation from natural water surfaces have been reviewed. The theoretical evaporation formulas have been found unsatisfactory. It is concluded that the effect of spray on the evaporation can be neglected.

The formula of ROHWER has been modified for the computation of evaporation from the sea. For the modification it was necessary to estimate the wind speeds at various levels over the sea. It was concluded that the effect of stability of air can in general be neglected and that the speed of the wind at the surface is about half of the speed at the 8 metre level.

For conversion of the evaporation measurements at sea to actual evaporation a conversion factor of 0.65 is adopted and the modified ROHWER Formula (29) has been found the most accurate and suitable for the computation of evaporation from the sea.

11. EXCHANGE OF SENSIBLE HEAT BETWEEN THE SEA AND THE ATMOSPHERE

11.1. EARLIER WORK ON CONVECTIVE TRANSFER OF SENSIBLE HEAT

Measurement of the convective transfer of sensible heat between the sea and the atmosphere presents difficulties. However, this transfer may considerably modify the thermal characteristics of the water and air masses above it and must be taken into consideration in heat budget studies. Also an estimate of the modification of air masses is most important for forecasting weather in coastal areas (TERADA, 1956), and MOSBY (1957) has shown the possibilities of estimation of temperature changes of the air which moves over the oceans.

BOWEN (1926; from JACOBS, 1951) has related the convective transfer of sensible heat to evaporation by the following formula:

$$\frac{Q_h}{Q_e} = R = K_4 \frac{(T_w - T_a)}{(e_w - e_a)} \frac{P_a}{1000} \tag{30}$$

The constant K_4 depends on the units used and is in our case 0.66. The term $\frac{P_a}{1000}$ is very close to unity and is usually neglected. This formula has been used by most workers concerned with the transfer of sensible heat. ANDERSON (1954) found in the Lake Hefner studies that BOWEN's ratio is valid and gives in general consistent results, except when the difference between the vapour pressure of the atmosphere and that of the water surface is small.

SVERDRUP (1945) developed a theoretical formula, but because the temperature and humidity gradients in this formula were difficult to measure, he came to the same formula as BOWEN.

SHULEIKIN (1953) used no wind factor in his formula for computing the exchange of sensible heat, if *the water was warmer* than the air:

$$Q_h = 30.24\,(T_w - T_a) \quad [\text{g cal cm}^{-2}\ (24\ \text{h})^{-1}] \tag{31}$$

If *the water was cooler* than the air, SHULEIKIN (1953) used the following formula where wind speed was considered:

$$Q_h = 0.432\,(T_a - T_w)\,V \quad [\text{g cal cm}^{-2}\ (24\ \text{h})^{-1}] \tag{32}$$

11.2. DEVELOPMENT OF A FORMULA FOR CONVECTIVE TRANSFER OF HEAT

The convective transfer of heat can be approximately related to the evaporation with the following assumptions:

All the air which has been in contact with sea surface obtains the same water vapour pressure and temperature as the sea surface; therefore the amount of air (in grams) per unit area which has been in contact with sea surface is:

$$(33) \qquad A_m = \frac{100\, E\, d_a}{(0.98\, e_w - e_a)\, K_5} \qquad [\text{g cm}^{-2}\ (24\ \text{h})^{-1}]$$

The factor K_5 for the conversion of water vapour pressure to the amounts of water in the air is slightly variable with temperature; but the variations are relatively small and for all practical purposes it can be taken as 1 if $(e_w - e_a)$ is expressed in mm Hg, and as 0.76 if it is expressed in mb (variations: 0° — 0.805; 10° — 0.772; 20° — 0.745).

The convective heat transfer can be expressed according to the assumption:

$$(34) \qquad Q_h = A_m\, (T_w - T_a)\, h \qquad [\text{g cal cm}^{-2}\ (24\ \text{h})^{-1}]$$

Substituting A_m with (33) and E with (29) we get:

$$(35) \quad Q_h = 39.0\ (0.26 + 0.077\ V)\ (T_w - T_a) \qquad [\text{g cal cm}^{-2}\ (24\ \text{h})^{-1}]$$

From BOWEN's ratio we get:

$$(36) \quad Q_h = 38.9\ (0.26 + 0.077\ V)\ (T_w - T_a) \qquad [\text{g cal cm}^{-3}\ (24\ \text{h})^{-1}]$$

The results from Formulas (35) and (36) are practically identical if for the latent heat of evaporation in (36) the surface temperature is assumed to be *circ.* 10°C. Additional confirmation for the assumptions on which Formula (35) is based can also be found in the fact that there is usually a good correspondence between humidity and temperature profiles over the sea.

When the difference $T_w - T_a$ is negative, the sensitive heat is transferred from the air to the water. This transfer is described in Chapter 12. There are diurnal changes in sea and air temperature which affect the transfer of sensitive heat. These changes are described in Chapter 11.4 and suggestions are made for the selection of proper values for use in Formula (35).

11.3. CHANGES IN THE TEMPERATURE OF AIR WHICH MOVES OVER WATER WITH A DIFFERENT TEMPERATURE

If the heating of the air by direct radiation and the release of the latent heat of evaporation are neglected, one could compute the change in the temperature of the mixed surface layer of the air from the convective transfer of heat when the air moves over the ocean.

Assuming in the following example that the thickness of the mixed layer is 500 m, the wind speed 5 m sec^{-1} and the sea surface 3°C warmer than the air, computation of the exchange of sensible heat gives that by this exchange the mixed layer of air will get warmer by *circ.* 1°C during 6 h when the air has moved *circ.* 108 km. Increased wind speeds will of course reduce the time of temperature adaptation. A visual examination of daily weather maps shows that such rapid warming up (and/or cooling) actually takes place. MOSBY (1957) also found by theoretical and empirical consideration that the temperature difference between the sea and air is reduced to 10 per cent of its initial value during 5 — 6 hours and that the temperature adaptation is faster in a warm wind than in a cold wind. The latter case is probably due to the change in the thickness of the mixed layer. There is of course, a continuous exchange of air between the lower mixed layer and the upper layer, and for a more precise calculation this eddy conductivity must be taken into consideration.

The estimation of the temperature changes of surface air moving over the oceans seems to be important for daily and especially for long-range weather forecasts, and a further elaboration of this subject seems desirable, in connection with the large, slow-moving and long-lived patterns of temperature anomalies. This subject, however, is outside the scope of this investigation.

11.4. DIURNAL CHANGES IN SEA SURFACE AND AIR TEMPERATURE AND THE EFFECTS OF THESE CHANGES ON THE ESTIMATION OF EVAPORATION AND CONVECTIVE TRANSFER OF HEAT

In Formulas (29) and (35), the selection of $T_w - T_a$ and $e_w - e_a$ presents some difficulties because these quantities vary diurnally.

In the case of scarcity of actual data, any available good diurnal mean has to be used in approximate calculations. For more precise calculations, and when more daily measurements are available, care should be taken in the selection of values.

From numerous studies (e.g. LIU 1934, HAY 1956 and others) we know that the sea surface temperature varies less than 1°C diurnally and that there are local minute to minute fluctuations of the same range as the diurnal fluctuations. The diurnal temperatures are usually lowest at 06^{00} hours and highest at 15^{00} local hours. The diurnal variation in both sea and air temperature is greatest on quiet, fair days in summer. The variation is generally small in autumn and winter.

Knowledge of the convective transfer of heat and evaporation has great meteorological implications, and so, when selecting the values for sea surface and air temperature, one has to adhere to the international standard hours of observation. It is therefore suggested that the temperature and water vapour pressure values obtained from measurements close to 00^{00} and 12^{00} local hours should be used, as they will represent more closely the night and day averages.

11.5. SUMMARY OF CHAPTER 11

Assuming that all the air which has been in contact with the sea surface attains the same water vapour pressure and temperature as the sea surface, a new Formula (35) has been derived for the computation of convective transfer of »sensible» heat. This formula gives nearly identical results to those derived from BOWEN's ratio. It is suggested that in Formulas (29) and (35) the sea surface and air temperature and water vapour pressure values at 00^{00} and 12^{00} local hours be used, as these are representative values of average day and night conditions.

The rapid changes in the temperature of air moving over the sea with different temperature are demonstrated with an example.

12. TRANSFER OF HEAT BY PRECIPITATION AND CONDENSATION OF VAPOUR ON THE SEA SURFACE

The transfer of heat to the sea by precipitation and condensation has not been considered quantitatively in earlier works in heat budget studies; however, the possible downward convection of heat has been qualitatively pointed out by several workers [e.g. BUNKER, HAURWITZ, MALKUS and STOMMEL (1949) and WATANABE (1955)].

The transfer of heat to the sea by rain can usually be ignored if the amount of the rain and the temperature difference between the rain and sea surface are small.

If snow or hail fall, a considerable amount of heat might be needed to melt the particles in the sea. For the computation of the heat needed the following selfexplanatory formula is given:

$$(37) \qquad Q_p = 7.97\,P_s + 0.1\,P_s\,(T_w - T_p) \qquad (\text{g cal cm}^{-2})$$

Loss of heat by melting of snow must be allowed for during late autumn and winter in the higher latitudes because it is an important factor in the formation of ice and the prediction of icing conditions (for example, in the Baltic Sea).

When the values $T_w - T_a$ and/or $e_w - e_a$ are negative, convective transfer of sensible heat from the air to the sea surface and/or condensation of vapour on the sea surface are taking place. In this case it is doubtful whether evaporation follows Formula (29) or convective transfer Formula (35), because the cooled air causes a high stability close to the sea surface. This doubt is especially serious when wind velocities are low. Some data from Mosby (M.S.) also indicate the validity of this doubt.

It is therefore proposed to omit factor 0.26 when the product $T_w - T_a$ in Formula (35) is negative, and the new formula for convective transfer of heat to the sea is:

$$(38) \qquad Q_h = 3\,V\,(T_w - T_a) \qquad [\text{g cal cm}^{-2}\ (24\ \text{h})^{-1}]$$

A slow vernal warming and a rapid autumnal cooling of surface water (Lauzier, 1957) might also indicate that the above-proposed modification of the formula is justified. The formula gives about seven times higher values than the corresponding Formula (32) by Shuleikin (1953). Empirical test of these formulas is difficult, except in future practical evaluation and forecasting tests.

When the difference $(e_w - e_a)$ is negative, condensation of vapour on the sea surface is expected. The additional »convection» factor should also be eliminated from the evaporation formula in the case of negative values, on the basis of the discussions above; and the new formula for transfer of heat by condensation is:

$$(39) \qquad Q_c = 0.077\,V\,(e_w - e_a)\,L_t \qquad [\text{g cal cm}^{-2}\ (24\ \text{h})^{-1}]$$

12.1. SUMMARY OF CHAPTER 12

For the computation of the heat taken from the sea water for melting hail and snow, Formula (37) is given. Formula (38) has been suggested for the convective transfer of heat to the sea, if the sea surface is cooler than the air, and for the computation of heat released by the condensation of vapour on the sea surface another Formula, (39), is suggested. The two last formulas have been derived through physical reasoning, because an empirical test of these formulas is extremely difficult.

13. THE HEAT BUDGET IN ICE-COVERED SEAS

In addition to the formulas established in previous chapters, the following points should be observed in computing the heat budget in ice-covered seas:

(1) The albedo (Q_r) of ice and snow is *circ.* 70 per cent (List 1951, Smithsonian Meteorological Tables).

(2) For computing evaporation, the saturated water vapour pressure over ice can be taken from List (1951). As this pressure is low, condensation often occurs on the ice surface.

(3) If the surrounding temperature of the water is higher than the melting point of the ice, melting occurs. The latent heat of melting and the specific heat of sea ice are given in the tables mentioned above. Formula (37) can be used for computing the heat loss from the water by melting ice.

(4) The conduction of heat through ice can be computed by the following formula:

$$Q_h = \frac{K_h\,(T_{i2} - T_{i1})\,t_{sec}}{d_i} \qquad (\text{g cal cm}^{-2}\ \text{sec}^{-1}) \tag{40}$$

The thickness of the ice d_i should also include the thickness of the snow on it (Uusitalo 1957). The specific heat conductivity K_h can be taken from List (1951).

PART II

FACTORS DETERMINING TEMPERATURE CHANGES AND STRUCTURE IN THE SEA

14. FACTORS AFFECTING THE THERMAL STRUCTURE IN A GIVEN LOCALITY IN THE SEA

The procedures and formulas for the computation and/or forecasting of the heat exchange between the sea and the atmosphere have been established in Part I. Part II will deal with the possibilities of forecasting the temperature structure and its changes in the sea, using the heat balance and the factors affecting the temperature structure.

The main influencing and determining factors are: 1) The absorption of heat in different layers, 2) mixing and 3) advective transport of heat. These factors affect the sea temperature in different ways, some directly, some indirectly. These factors are in no way simple or uniform. Therefore some of them have been divided into additional steps; for example, the mixing is divided into three categories: convective stirring, mixing by wave action, and turbulent mixing by currents. In the following chapters they are explained in detail with emphasis upon their significance to the main problem of this study.

Beside the use of factors affecting the temperature and its structure in the sea, some indirect ways and means must be used: for example, correlation between sea level and temperature; or certain general procedures of estimation must be established: for example, determination of the depth of the thermocline and prediction of the movement of current convergences and divergences.

For prediction of the temperature in the sea use must be made of several other important but complex processes in the sea, which themselves are subject for separate forecasts: waves and currents, for example. Simplified methods and procedures for forecasting these phenomena have also been worked out in the following chapters.

In Part III the procedure for forecasting temperature in the sea will be tested in some localities where the method of application of some of the established formulas is also illustrated.

15. ABSORPTION OF ENERGY IN THE SEA

The first step in the estimation of changes in temperature structure is obviously the computation of the radiant energy absorbed in the different layers. The extinction coefficient of the radiant energy in the sea varies with variations in the optical properties of the water and with the wave length of the energy. Exact mathematical procedures would therefore require lengthy computations. In order to facilitate these computations, the method used by SCHULE (1952) and LÖNNQUIST (1954) (the synthetical method) will be employed here.

Table 5. *Transmission of energy, % per metre, in various optical water masses*

Spectral region	% of total radiant energy	Optical water mass				
		1	2	3	4	5
$<0.490\,\mu$	11	92	75	50	25	5
$0.490-0.525\,\mu$	7	98	91	85	76	42
$0.525-0.635\,\mu$	13	89	82	76	68	55
$0.635-0.690\,\mu$	9	67	61	52	44	38
$0.690-3.0\,\mu$	60	(10)*	(6)*	(3)*	(0.5)*	(0)*

* Rough estimates only (no accurate data available).

The spectral composition of solar energy on the earth's surface for an average air mass of 1.5, given in first two columns of Table 5, has been taken from the data given by DRUMMOND (1958). Variations in the spectral composition with varying length of the path of the incoming radiation in the atmosphere are relatively small, and possible errors introduced into Table 6 by using the values of spectral composition given in Table 5 as overall averages are less than 1 per cent of the values presented in Table 6, even in cases when the air mass is 3.

The optical water masses are defined in Part I (see Table 2) and the corresponding transmission values are given in Table 5. They are interpolated from data given by JERLOV (1951). The absorption of total energy is computed for every one metre layer. A grouping of these layers by depth intervals, convenient for application to standard hydrographical data in the oceanic areas, is given in Table 6.

A high percentage of energy is absorbed in the upper 2.5 metres and it could be argued that this layer should be divided into two or more units. However, this layer can, for all practical purposes, be considered, in most cases, as a thoroughly mixed layer, and no further subdivision is necessary.

Table 6. *Absorption of total energy (%) in various layers of the sea*

Layer in meters from surface	Optical water mass				
	1	2	3	4	5
0 — 2.5	71.4	78.2	84.4	89.6	95.1
2.5— 5	6.8	9.1	8.1	6.5	4.0
5 — 10	7.2	7.3	4.7	3.2	0.9
10 — 20	6.6	3.7	2.0	0.6	
20 — 30	3.0	0.9	0.3	0.1	
30 — 40	1.3	0.5	0.1		
40 — 50	0.9	0.2			
50 — 75	1.1	0.1			
75 — 100	0.6				
100 — 150	1.1				

The percentage of absorption of total energy, as shown in Table 6, is considerably higher in the surface layers than that given by JERLOV (1951) for the corresponding optical water masses. JERLOV has omitted part of the long-wave radiation. This long-wave radiation is absorbed quickly in the first metre below the surface and part of it is radiated back into the atmosphere as »effective back radiation». MOSBY (M.S.) also concluded that the diurnal temperature variations in calm water, due mainly to radiation, are confined to the first metre below the surface.

Diffuse, upward scattered radiation in the water has not been considered by computation in Table 6, because it is relatively small and depends on several factors, difficult to define. This radiation is in the wave length of 350 — 550 mμ (JOSEPH, 1950) and amounts approximately 3 to 5 per cent of the total radiation in this wave length range. The solar radiation in the above mentioned range is approximately 15 per cent of the total and therefore the error introduced would be around 1 per cent of the values in Table 5 if the upward radiation were not considered. However, this neglect of the upward-reflected radiation does not influence the total heat budget of the water column, because this upward radiation leaving the surface layer is indirectly included in the computation of Q_r (reflected radiation).

15.1. SUMMARY OF CHAPTER 15

Exact mathematical procedures for computing the absorption of energy in the sea reguire lengthy computations. Table 6 has been computed to facilitate the estimation of the absorption of energy in the different layers. The absorption values given in this table are slightly different from those given by JERLOV (1951) because the long-wave radiation has been included in the present work. The possible small errors introduced in the simplification of the estimation of the absorption in the sea have been briefly discussed and found to be negligible in the practice.

16. DETERMINATION OF LOCAL TEMPERATURE CHANGES CAUSED BY HEAT EXCHANGE AND MIXING

Several approaches have been tried in the past for the prediction of temperature variations in a given locality in the sea. FJELDSTAD (1933) treated the temperature changes by a continuous density model and derived an equation with a number of possible solutions. However, no satisfactory theoretical thesis on the solution of the problem was found. SVERDRUP *et al.* (1949) gave a formula for the computation of temperature changes where the average temperature was taken to be a linear function of depth, and variations of surface temperature were represented by means of a series of harmonic terms. The term A/ϱ was assumed to be constant. No higher accuracy can be expected from SVERDRUP's formula, because the surface temperature variations do not follow a harmonic curve exactly, and the eddy conductivity varies with locality, depth and time, as it depends on many factors (waves, currents, stability, etc.).

In the following discussions it is assumed that an initial temperature structure is available from actual measurement and will be repeated after certain time intervals (e.g. weekly, or monthly). If actual measurements of temperature structure are not available, an approximate temperature structure can be constructed from seasonal hydroclime data. The accuracy of the hydropsis is adversely influenced in this case, and it will be used chiefly to predict changes of temperature differences from the hydroclime.

The computation and prediction of temperature structure in the sea, using a heat exchange budget, is slightly different when using the continuous density model as opposed to the two-layer system, and so these two models are treated separately below. With the continuous density model the absorption of energy in the different layers is found by means of Table 6, and the heat losses are subtracted from the heat in the upper 2.5 metre layer. With the two-layer system the depth of the thermocline is estimated first, and the surface mixed layer is treated as one layer. The methods are also given for the estimation of the changes of thermocline depth caused by convective stirring, which is usually a result of negative heat exchange in the surface layer, and by wave action.

16.1. CONTINUOUS DENSITY MODEL

The theoretical daily increase of temperature in various layers due to insolation can be computed with the following formula, taking appropriate data from Table 6:

$$T_{w(1)} = \frac{a_{w(1)}(Q_s - Q_r)}{100\, q_w l\, \varrho} + T_{ow} \qquad (°C) \tag{41}$$

The specific heat of sea water at different salinities can be found in the Smithsonian Meteorological Tables (see LIST 1951). From the temperature of the upper 2.5 metre layer the following components are subtracted or added:

$$T_{w(2.5)} = T_{w(a2.5)} - \frac{Q_b + 0.1EL_t + Q_h}{q_w l \varrho} \qquad (°C) \tag{42}$$

If necessary, corrections should be also made for Q_f and Q_p.

It can be assumed that the upper 2.5 metres are practically always thoroughly mixed. The surface layer might become unstable through decrease of temperature and through excess of evaporation, which causes an increase in salinity. Therefore, after computing $T_{w(2.5)}$ and the change of salinity caused by evaporation, one should check that the density of the upper 2.5 metre layer has not exceeded that of the 2.5 to 5 metre layer. Notes in Chapter 16.3 on convective stirring explain further steps to be taken in the computation of the new temperature structure caused by instability.

16.2. TWO-LAYER SYSTEM

When the depth of the thermocline is known, the surface layer above it may generally be considered as thoroughly mixed, especially when there has been excessive cooling (heat balance negative). This mixed layer may then be treated in the same way as the upper 2.5 metre layer in the continuous density model, with the difference that all the energy which will be absorbed in it according to Table 6 should be included in $a_{w(l)}$, where l represents the thickness of this mixed layer. The layers in and below the thermocline can be treated as in the continuous density model. Depending on whether cooling or heating has taken place in the mixed layer, its thickness will increase or decrease, if there has been no other factor (e.g. wave action) to increase the thickness. The change in the thickness depends on the original sharpness of the thermocline and can be determined either graphically (see Figure 15) or approximately with the following formula:

$$\Delta D_m = \frac{(D_{lm} - D_{om})\,(T_{ow(m)} - T_{w(m)})}{T_{ow(m)} - T_{dw}} \quad \text{(m)} \tag{43}$$

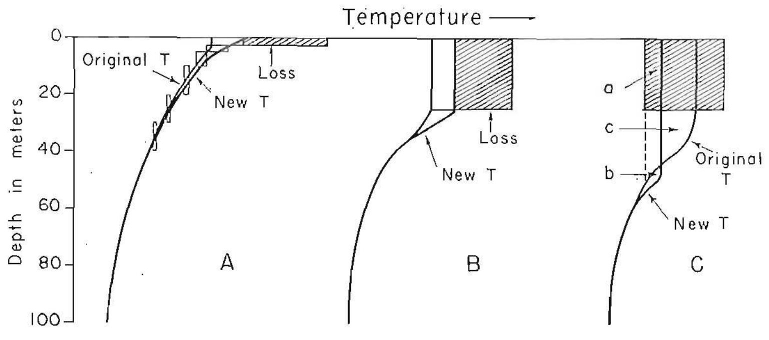

Figure 15. Graphical determination of the changes of temperature structure.

A — continuous density model
B — two-layer system, positive heat balance
C — two-layer system, negative heat balance

However, the surface layer above the thermocline may not always be considered as thoroughly mixed, because the factors which cause this mixing may change from day to day. Therefore a computation of the thickness of this layer, as determined by waves, currents and other factors, must usually be made before applying the methods outlined above. This is especially necessary when heating of the surface layer is taking place or the stability is increased by other factors (e.g. precipitation, runoff, etc.)

16.3. ESTIMATION OF THE DEPTH OF THE THERMOCLINE AND ITS VARIATIONS

Several different attempts have been made to determine the average depth of the thermocline (FREEMAN, 1954; MUNK and ANDERSON, 1948; LUMBY, 1955; and others). In quantitative attempts, the depth of the thermocline has been related to wind speed and latitude and by Russian researchers to waves. Unfortunately the Russian researchers do not give the formulas used. The other attempts are unsatisfactory and sometimes directly misleading (for example, the effect of convective stirring and the effects of divergences and convergences are not considered; the included »latitudinal effect» gives misleading values in high and low latitudes, etc.).

In the following consideration the problem of the determination of the thermocline depth is divided into: 1) prediction of the mean depth around which the thermocline oscillates; 2) prediction of the amplitude of the oscillation.

It is obvious that the average depth of the thermocline is directly related to mixing, which in this paper is divided into two, according to the cause: the first cause is convective stirring and the second is wave action. The final depth of the thermocline depends, in the case of a negative heat exchange, on which action reaches deeper, and, in the case of a positive exchange, only on the wave action. The possible influence of mixing by currents is dealt with separately, as well as the effects of upwelling and piling-up on the thermocline depth, especially at the convergences and divergences of surface currents.

16.3.1 *Convective stirring*

Only a few qualitative remarks can be found in the literature on convective stirring. MUNK and ANDERSON (1948) believed that convective stirring plays as important a part in the oceans as wind stirring. FOFONOFF (1956) concluded that the formation of Antarctic bottom water appears to be influenced and controlled by nonlinearities in the dependence of the density of sea water on temperature and salinity. This mechanism of water-mass formation and mixing can also be included in convective stirring. An attempt to consider convective stirring quantitatively is made below.

Convective stirring occurs when instability of the upper layer(s) is caused by evaporation and loss of heat to the atmosphere. This is a continuous process and at present it is difficult to derive an exact, applicable equation to describe it. However, using a 24 hour period, an attempt is made below to derive an approximate solution for the determination of the depth to which this convective stirring can arrive (by a continuous density model) and for the determination of the resulting changes of density in the stirred (mixed) layer.

When the heat balance and the resulting density for 24 hours for the upper 2.5 metre layer has been computed, and the density of this upper layer is greater than the density of the layer below, the computation of mixing should be continued layer by layer, taking the next layer (2.5 to 5 m) and its average density after the heat balance computations:

$$\bar{\varrho} = \frac{l_1\bar{\varrho}_1 + l_2\bar{\varrho}_2}{l_1 + l_2} \tag{44}$$

If the average density of these two layers is still higher than the density of the following layer (5 to 10 m) the computation should be continued as above. The average temperature of the layer mixed by convective stirring can be found with the following simple formula:

$$T_{w\,(l_1+l_2)} = \frac{l_1 T_{w(l_1)} + l_2 T_{w(l_2)}}{l_1 + l_2} \tag{45}$$

Another and probably more exact way for approximate calculation of convective stirring is by assuming that mixing reaches the depth where the density equals the density of the surface 2.5 metre layer caused by a change over the 24-hour period. For computation of the resulting density and temperature, the $\bar{\varrho}_2$ is taken as the average density of the layer between the surface 2.5 metre layer and the point where the density equals that of the surface layer after the change.

Molecular diffusion and conductivity can be omitted from these approximate calculations. Only on relatively few occasions can the upper layers be treated according to the procedures for the continuous density model. When warming up occurs, the main task is to determine the average depth of the thermocline which has been created, or is already present, and its fluctuations. This calculation is described in the following sections.

16.3.2 *Estimation of the average depth of the thermocline from wind and wave data*

Although wind is the direct force for mixing, the actual mixing in the water is caused by its effects — wind waves and wind currents (drift currents). The mixing by waves is mainly caused by the presence of waves of different length and height. In order to determine the thickness of a mixed layer caused by wave action, we must first determine the wave characteristics on the surface and then define a depth where the particle motion and accompanying mixing are negligible. This procedure is possible in the case of surface waves (short waves) which can be considered as trochoid waves and where the orbital paths of the particles are near circles. In long waves the paths are nearly horizontal and nearly constant with depth. However, these long waves are less important for vertical mixing (their effect being similar to that of the current), but they are of importance in causing the periodic fluctuations of thermocline depth.

In general, the height of wind waves depends on wind speed, its duration, length of fetch and the depth of the water. A summary of the existing formulas relating the waves to some or all of these factors is given by Bruns (1955). In addition, Roll (1952) and Fleagle (1956) have pointed out an important fact, generally known to seamen, to be considered in the prediction of waves, namely that significantly higher waves are generated by cold air rather than by warm air. According to Fleagle (1956), the difference amounts to roughly 10 per cent per degree Celsius of sea-air temperature difference.

Rapid determination of the wave height may be made (a) by using appropriate tables, and (b) by computation, using the relations of wave characteristics and wind data. Using the wave data from Darbyshire (1959) and the wave observations made on board USS »Rehobot» (U.S.

Navy Hydrographic Office 1955), including the temperature difference effect, mentioned above, and the effect of the fetch and the duration of the wind, based on the data from BURGESS (1950), NEUMANN (1955) and SVERDRUP, JOHNSON, FLEMING (1942), a simplified single formula is

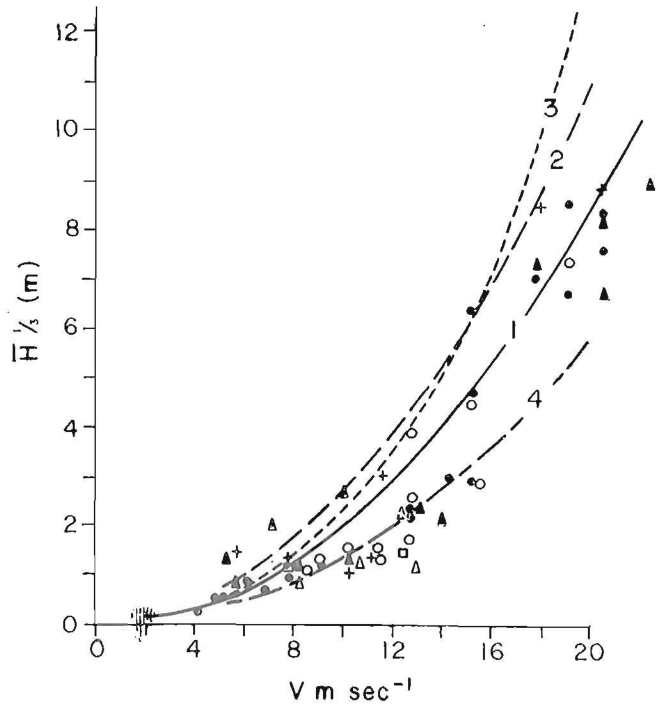

Figure 16. Significant heights of fully developed sea according to Darbyshire, Neumann, Sverdrup-Munk, and Formula (46).

Data for the curves (curves 2 to 4 from Neumann, Pierson 1957)

Curve 1 $H_{1/3} = \dfrac{0.0008\, V^2\, [50 + (T_w - T_a)]}{1 + \left(\dfrac{5\,V}{F}\right)\left(1 + \dfrac{V}{3t_h}\right)}$

$T_w - T_a = 0$; $F = 100$ km $t_h = 6$ hours

Curve 2 Sverdrup-Munk $H_{1/3} = 2.667 \times 10^{-2} V^2$

Curve 3 Neumann $H_{1/2} = 7.065 \times 10^{-3} V^{2.5}$

Curve 4 Darbyshire $H_{1/3} = 1.39 \times 10^{-2} V^2$

$(T_w - T_a)$: $\pm$ 0.5°C + + 1 to + 2°C ▲ > + 2°C

− − 1 to − 2°C △ < − 2°C

derived, by means of statistical approximations, for the estimation of the significant height of the waves:

$$H_{1/3} = \frac{0.0008\,V^2\,[50 + (T_w - T_a)]}{\left(1 + \dfrac{5V}{F}\right)\left(1 + \dfrac{V}{3t_h}\right)} \quad \text{(m)} \tag{46}$$

A comparison of Formula (46) with other existing wave forecasting formulas is made in Figure 16.

The depth of the water has not been taken into consideration. However, we have to consider it in the determination of the mixing depth caused by waves. The following formula can be used for checking the dependence of wave height on the depth of the water:

$$H_{max} = 0.5\sqrt{H_w} \quad \text{(m)} \tag{47}$$

Detailed instructions for the forecasting of waves are given in U.S. Hydrographic Office (1951) Publication No. 604. The Formula (46) is an approximation and useful for quick estimates. For the construction of a more exact formula, more data on the effects of temperature, fetch and the duration of the wind are needed. In the present work we are not interested in the decay of waves. The advection caused by the waves is mentioned at the end of Chapter 17.3 and the possible propagation of the depressions of thermocline, caused by heavy seas is described in Section 16.3.3.

The next step is to determine the depth at which the mixing caused by wave action is negligible. Using the relations for trochoid waves for computing the velocities of water particles at various wave heights and at various depths*, the depths given by NEUMANN (1955), where the total wave energy has decreased to 5 per cent of its value at the sea surface, and assuming that the mixing by waves is negligible at approximately a depth where the diameter of orbital baths is smaller than 10 cm., we can arrive at a tentative formula for determining this depth from the wave height:

$$D_m = 12.5\,H_{1/3} \quad \text{(m)} \tag{48}$$

* For relating the significant height to »corresponding» wave length, the following empirical relation, derived by the author, has been used: $L_{1/3} = 50\sqrt{H_{1/3}}$ (m).

This relation, which is not valid for seasonal data, is shown in Figure (17) together with a few available actual data from Weather Ship M and some data from LUMBY (1955). The last mentioned data are monthly averages and therefore give a greater depth of the mixed layer, because the depth was determined by heavy winds in the near past and not by average winds. It is therefore obvious that for the estimation of the thermocline depth the heavy wind and/or wave data for a few days back should be used, because the establishment of a new thermocline may take considerable time. Furthermore, the relation between waves and thermocline depth can usually not be used when convective stirring occurs (during autumn and winter). Furthermore, at and close to convergences and divergences and other areas where upwelling or piling-up of waters occurs, the approximate relation given with Formula (48) is not valid. These cases are described in a later chapter. (The available monthly data show that the average thermocline depth (m) is *circ.* 8 times the average speed of the wind (m sec^{-1}) in the same locality and month, in those months and areas when and where no convective stirring occurs.)

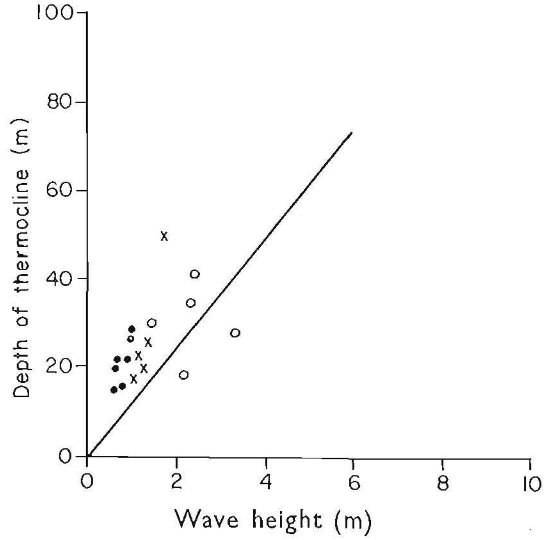

Figure 17. Relation between wave height and depth of the mixed layer.

○ Weather ship M, June 1957

× Mardsen square 185, April to August } Monthly averages of the mixed

• Northern North Sea, April to August } layer depth from Lumby (1955).

16.3.3 *Fluctuations of the depth of the thermocline*

SCHULE (1952) classified five main types of variations of thermocline depth: 1) seasonal; 2) lunar (monthly and bimonthly); 3) semidiurnal and diurnal, tidal; 4) short period; and 5) random variations due to mixing. He found that the third type had the largest single effect. He, furthermore, found that the task of predicting these fluctuations is still very difficult, mainly because of the scarcity of data. As the tidal forces may be the main cause of these fluctuations, some prediction in coastal areas can be made, assuming that the fluctuations are proportional to the tidal height. However, great uncertainties are also involved in this new method. The difficulties of determining tidal currents in offshore areas are mentioned in Chapter 17.4.

According to DIETRICH and KALLE (1957), GROEN (1948) has shown from the theory of short internal waves that there is a lower boundary of the period of internal waves which depends on the density gradient. As the fluctuation period of the short internal waves is usually rather short, this fluctuation can be included in the actual thickness of the thermocline in the 24-hour predictions. The great short period variations of thermocline depth are caused by internal tidal waves (semidiurnal and diurnal). The average amplitude of these variations is 2.5 to 5 metres. In coastal areas, REID (1956) has found internal waves with heights of 10 metres and greater. DEFANT (1950) has analysed the internal tide waves in the open sea and concluded that the displacement at the two layer boundary is always greater than at the surface and this displacement depends on density differences. SCHULE (1952), in cooperation with the U.S. Coast and Geodetic Survey, attempted to analyse the semidiurnal and diurnal constituents of these fluctuations. However, these attempts failed because no continuous registrations of the fluctuations existed and the available 6 BT data per day did not allow exact determination of the ridges and troughs of the internal waves. SEIWELL (1937) attempted a harmonic analysis of the hydrographic data of a four-day station in the North Atlantic. He found that the 24 lunar hour vertical oscillation was the most pronounced and the effects of the 12 lunar hour oscillations were about 50 per cent less.

An analysis was made, by the present author, of a graphical summary of North Pacific bathythermograph data (see LEIPPER 1954) in order to find the monthly and bimonthly fluctuations of the thermocline. An

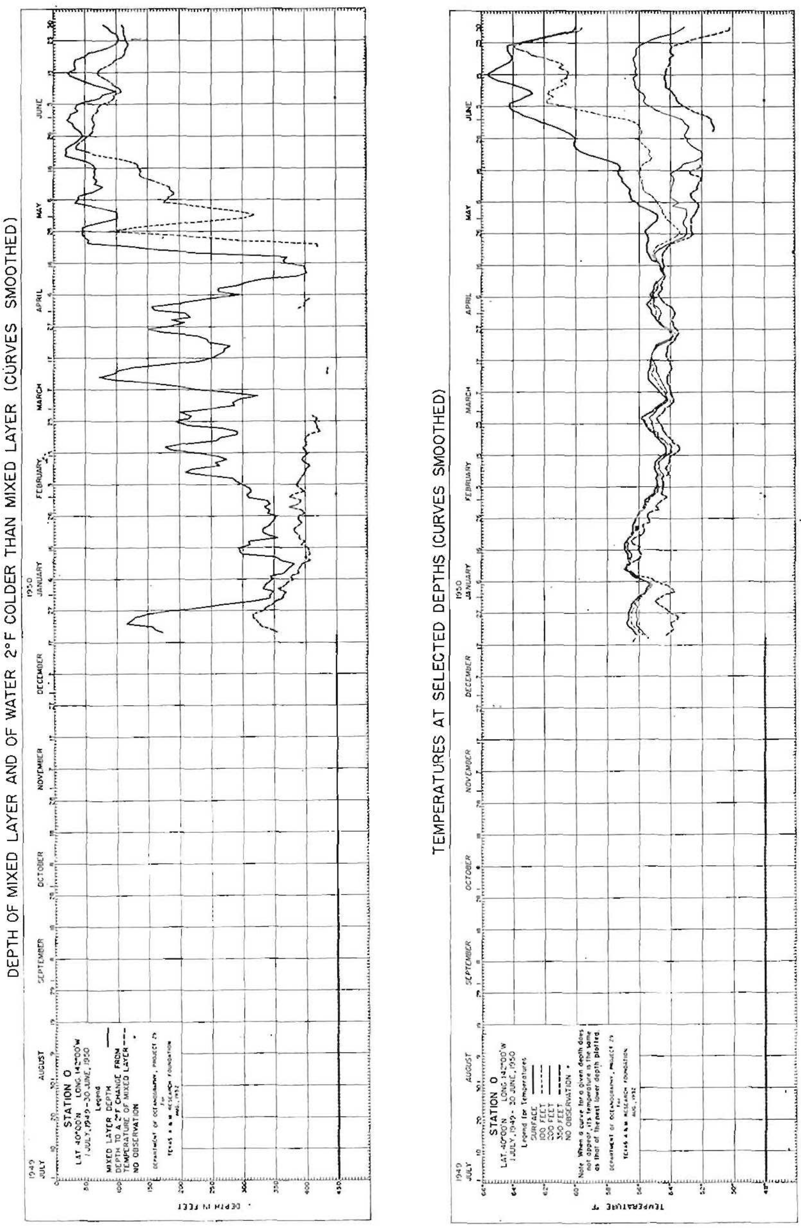

Figure 18. Example of fluctuations of thickness of mixed layer and temperature structure in the North Pacific (Leipper 1954).

example of such graphical data is given in Figure 18. The following observations were made from these data:

1) The fluctuations of the depth where temperature is 1°F different from the sea surface are much smaller than the fluctuations of the depth of the »mixed layer». (The »mixed layer depth» was defined in the graphical summary as the depth of a $\pm$ 0.3°F change from the temperature of the sea surface.)

2) The seasonal fluctuations of the depth of the thermocline are much greater than the monthly and other periodic fluctuations. It seems that the meteorological factors have the far greatest influence on the long term fluctuation of the thermocline depth. A preliminary examination gives an impression that the *circ.* fortnightly great fluctuations of thermocline depth are caused by heavy seas and atmospheric pressure changes in limited localities and that these fluctuations move with approximately the same speed as fronts and are being damped down over certain distances.

3) When the thermocline descends deeper (at the end of September and the beginning of October) and is sharp, the periodic fluctuations are relatively small. During the winter and early spring, when the thermocline is less sharp, the fluctuations within depth are greatest, but the changes of temperature with time in a given depth are small.

4) There are considerable variations in the average depth of the thermocline from year to year in the same locality and season.

Table 7. *Long-period fluctuations of the depth of mixed layer in the North Pacific*

Latitude N	28°	30°	31°	33°	40°	49°	50°
Number of fluctuations recorded	14	14	8	12	10	10	9
Range of periods in days	10 to 19	10 to 22	11 to 15	7 to 19	9 to 23	7 to 15	9 to 19
Average period in days	13.9	15.6	13.8	13.8	15.4	11.5	12.8

An attempt was also made to determine bimonthly and/or monthly fluctuations of the thermocline depth. The same difficulties were encountered in more exact analyses as have been described by SCHULE (1952). A summary of the average periods of fluctuation is given in Table 7. The overall average period is 13.95 days. It is difficult to relate this period to internal tides because of the difficulties of determination of offshore tides. It is interesting to note that standing internal waves with a 14 day period have also been observed in Gullmarfjord. The observed long-period fluctuations in the North Pacific might result from various factors which are difficult to analyse (e.g. tides — Mf and MSf; meteorological factors — heat exchange, pressure changes and winds; and standing internal waves etc.). Fortnightly tides, caused by the constituents of Mf and MSf, might be the important cause of these long term fluctuations of thermocline depth. GROVES (1957) also found fortnightly variations of sea level over large areas, assuming the cause of these variations to be the tidal constituents. Unfortunately, these constituents have been computed only for very few tidal stations. Mf is a purely astronomical constituent, arising from the variable declination of the moon, and MSf results from the nonlinear relations between M_2 and S_2.

It is doubtful whether these fluctuations are due to tidal forces, and it is more likely that they are caused by combined meteorological effects: The difference (and change) of atmospheric pressure causes a wave rather than a current, because the pressure gradient is not balanced by the deflecting force of the earth's rotation. Assuming the thickness of the surface layer to be 50 m and the density difference 0.002, this wave would propagate *circ.* 100 cm sec^{-1}, which is nearly the speed of an atmospheric cold front. A very rough estimation of the speed of a gradient current caused by a pressure change of 10 mb gives *circ.* 140 cm sec^{-1}. If particle transport velocity of a surface wave is 1/5 of wave speed, as determined experimentally, the average velocity in upper 5 m of a 5 m high sea would be around 100 cm sec^{-1}. The above rough estimates give speeds close to each other. If it could be supposed that the main causes of long term fluctuations (*circ.* fortnightly) of thermocline depth are the meteorological factors and that the speed of the movement of these fluctuations is approximately the same as the speeds of fronts, pressure changes and wave transport, and the direction of movement is also determined by these factors, the fluctuations would be predictable. Future investigations must clarify these points.

16.4. SUMMARY OF CHAPTER 16

Earlier attempts to predict the temperature variations in a given locality have been reviewed and no satisfactory theoretical or empirical solution has been found to exist. In the attempt to find an empirical and/

or theoretical solution, the problem is initially divided, by the previous temperature structure, into prediction by the continuous density model and prediction by the two-layer system. For the computation of the changes of temperature at different depths by the continuous density model, two theoretical formulas are given. The graphical estimation of temperature changes is shown in a figure. For the two-layer system a formula for the estimation of the changes in thickness caused by the heat exchange is given. However, the average depth of the thermocline is often unknown. For the prediction of this depth two approaches have been developed: prediction of the depth caused 1) by convective stirring, and 2) by wave action. For the quantitative estimation of convective stirring, for which no formulas existed, a new procedure with two theoretical formulas has been developed. For the determination of the depth of thermocline from wave data a new empirical wave formula has been developed, where the effects of the fetch, the duration of the wind and the temperature differences between the sea surface and air are included. The depth of the thermocline is related on theoretical and empirical consideration to the wave height, and the limitations to the application of this formula are discussed. The periodic fluctuations of the thermocline and the factors affecting these fluctuations are discussed. It is suggested that the main causes of the long-term fluctuations (e.g. fortnightly) might be meteorological factors and that the fluctuations might move with approximately the same speed as atmospheric disturbances. At present no data are available for developing formulas for the prediction of these fluctuations.

17. CURRENTS AND TRANSPORT OF HEAT

In the prediction of temperature we are interested in currents under two aspects: (a) transport of heat and consequent local changes caused by advection, and (b) mixing by currents.

In predicting the changes of surface water temperature we are mainly concerned with the trajectories (paths) of limited water masses, in order to follow their possible local changes day by day.

Munk and Anderson (1948) showed that currents and temperature distribution are intimately related. Mathematically, they are linked with coefficients of eddy viscosity and conductivity, which are functions of temperature (and salinity) gradients and currents.

SUDA (1937—38) and FUKUOKA (1957) concluded that the causes of the irregular variations of temperature and salinity with time in the Kuroshio area were connected only with the advection of the current.

Monthly and seasonal surface current charts based upon dead reckoning are available for all oceans. However, they are not very reliable, especially for short periods. HELA (1954) found that if the current displacement determined by astronomical fix is less than 2 nautical miles per 24 hours, the given value in the atlas is entirely unreliable. In most cases, little weight can be given to displacements of 5 nautical miles or less per 24 hours. JAMES (1957) came to an analogous conclusion, namely, that the surface current charts can be considered reliable and accurate as mean values if there are at least 5 observations from 3 600 square nautical miles and the drift is > 5 miles (24 h.)$^{-1}$. Consequently, there is a need for synoptic or prognostic current charts (JAMES, 1957).

The current charts in the atlases can be used for a rough average guess of the currents in a certain month. For our present purpose we need a more accurate method for the prediction of the surface currents on a daily or weekly basis. In this prediction use can be made of the atlases for estimation of the »normal» situations, but the anomalies of these situations must be predicted.

The following problems regarding currents are dealt with below:

1) Heat transport by currents.
2) The method of separation of wind driven currents from permanent currents.
3) The method for the prediction of wind currents.
4) Influence of changes of atmospheric pressure and transport by waves on the surface currents.
5) The influence of tidal currents on temperature structure.
6) Mixing by currents.
7) The fluctuation of current boundaries (convergences and divergences of currents).

17.1. HEAT TRANSPORT BY CURRENTS

The temperature change caused by advection can easily be computed by the use of the following formula, if the current speed and direction has been ascertained:

$$\frac{dT}{dt} = W \frac{dT_w}{dx}$$

where $\frac{dT}{dt}$ is the local change of temperature, per unit of time, caused by advection, and $\frac{dT_w}{dx}$ is the change of the surface temperature of the current with horizontal distance. This term should include temperature changes by mixing and by heat exchange during transport. It can be determined from seasonal charts of surface temperature and/or from previous measurements in the localities under consideration. However, heat exchange computations should also be made in the regions of expected inflow into the hydroptic area when abnormally rapid cooling or warming-up of surface layers is expected, and appropriate corrections should be made to T_w.

In hydropsis it is seldom necessary to compute the volume transport except by certain accurate forecasting of the movement of divergences and convergences of currents. Conventional formulas are available for the computation of this mass transport. The volume transport in a non-homogeneous layer is:

$$T' = 10\,W \int_0^d (\varDelta D_A - \varDelta D_B)\, dz_w \tag{50}$$

and in a homogeneous layer:

$$T' = \tfrac{1}{2}\,Wg\,\frac{\varrho - \varrho_1}{\varrho}\,z_w^2 \tag{51}$$

In the northern hemisphere, the direction of the average current and mass transport is assumed to deviate 90° to the right in relation to the direction of the slope of the isobaric surface.

Our main interest in the hydropsis is to determine the direction and speed of the surface currents in order to be able to use these data in Formula (49) and at other times.

17.2. PERMANENT CURRENTS AND THE SEPARATION OF WIND CURRENTS FROM PERMANENT FLOW

The permanent gradient currents are caused by the existence of pressure gradients, resulting usually from density differences, or they are maintained by the great permanent and more or less constant wind systems. The conventional formulas were usually developed for the computation of currents at right angles to the hydrographic sections, utilising the difference of the dynamic heights at these stations.

These methods for computing relative currents, using the dynamic height difference, are much criticized. It is difficult to find a real »layer of no motion». A dynamic height difference may also exist as a result of the piling-up of waters due to wind action, and the current perpendicular to the section might be balanced by the same force. Offshore tides and other factors causing the fluctuations of pycnocline depth also interfere with the dynamic height.

For hydropsis the computation of geostrophic currents would be of importance only if several more or less synoptic hydrographical stations were available. This is rarely the case, and therefore in hydropsis there only remains the possibility of computing the permanent flow and superimposed wind currents and currents caused by atmospheric pressure changes and by waves from the meteorological data. However, if a large area is under consideration, it would in certain cases be possible to compute changes of dynamic heights which might be caused by heat exchange, the piling-up of waters by wind, etc. Corrections to the estimated permanent flow could then be made, using the formula of MONTGOMERY (1938). He found that the logical method of deducing oceanic flow patterns from temperature, salinity and oxygen observations, is to chart these properties for surfaces of constant potential density. His formula for the gradient flow in a σ_t surface is:

$$W = \frac{37}{\sin L} \quad \frac{\Delta H_g}{\Delta l_n} \tag{52}$$

where H_g (the geostrophic potential) is: $\varphi_a - a_a\, p_h$.

It is necessary to separate wind currents, which vary with the direction and velocity of the wind and can be predicted from wind data, from the permanent flow (gradient currents) which is considerably more stable in space and time.

Although one has to recognize HANSEN's (1956) complaint that, for the detection of non-stationary movements in the ocean, no theory exists at present that could stand a test with regard to its practical application, nevertheless, the practical approaches made by PALMÉN (1930 b) and HELA (1952) have given satisfactory results in some areas and merit further application.

PALMÉN (1930 b) and HELA (1952) took the currents associated with a certain wind velocity, grouped them according to the directions of the winds and currents and thus computed the resultant currents for each direction. In the case of pure wind-drift currents, the final resultant of the resulting currents connected with all the different directions of the wind must be zero.

It is generally assumed that a rough concept about the permanent flow can be obtained from the resultant surface currents which occur during calm weather. MANDELBAUM (1955) has used a method similar to that of PALMÉN and HELA for computing »residua lcurrents during calms».

It is obvious that the wind currents have a certain inertia and that the sea does not respond instantaneously to the changes of wind-speed and direction. PALMÉN (1930 b) has shown that in the Baltic, when eliminating the drift current from the actual current data, the rest, usually called the permanent flow, is also a function of the wind. For this reason he introduced the name »characteristic current», instead of »permanent flow».

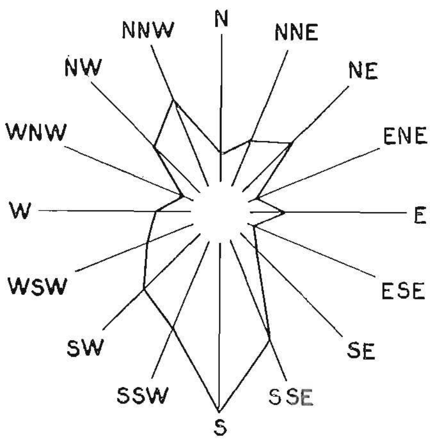

Figure 19. Directions of winds at »no current», Lightship »Storbrotten», 1953, 60°26′N, 19°13′E.

An examination of current data from lightship »Storbrotten» during 1953 (see LISITZIN 1955) shows that the currents during calms often indicate inertial wind currents from the previous winds and can come from all directions, showing the limitations of MANDELBAUM's method. By plotting wind directions at »no current» (Figure 19) a better idea of permanent flow can be obtained. From this figure it could be concluded that the »permanent flow» in this locality is from the North which is in agreement with LISITZIN's (1938) results. Plotting the N—S components of winds and currents at the same lightship (Figure 20) reveals that a

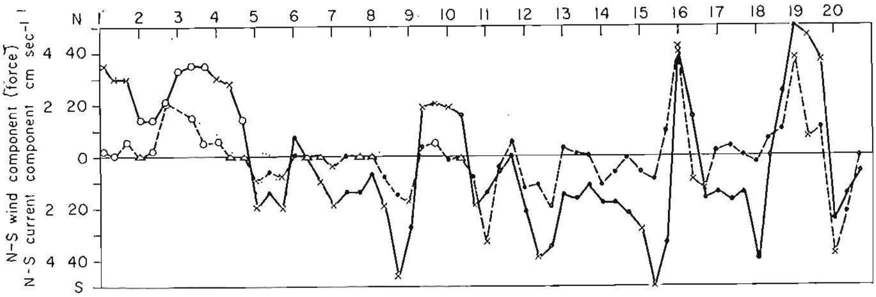

Figure 20. North-South components of winds and currents at Lightship »Storbrotten» during 1 to 21 January 1953.

— wind; — — — surface current;
△ no current; × S or N component only;
○ E component present; • W component present.

N—S seiche, normally expected to be present in narrow, long bays, is superimposed on this permanent flow from the north.

For the purpose of hydropsis, it would be necessary to compute the permanent flow for the area of hydropsis for each month, using the method of Palmén and Hela and available surface-current and average-wind charts. The »actual expected» wind current would be predicted from gradient wind data for the hydroptic period required, as outlined below.

17.3. RELATION BETWEEN WIND AND SURFACE CURRENT

Hughes (1956) found from the results of drift card experiments that the current velocity was 2 per cent of the velocity of the gradient wind. Scruton (1956) reported slightly higher values (the current speed being 2 to 3 per cent of the wind speed).

Hela (1952) found the following relation:

(53) $$W = 0.0137\ V \qquad (\text{cm sec}^{-1})$$

which is in good agreement with Hughes' values if we assume the surface wind to be 2/3 of the gradient wind. Lawford and Veley (1956) found that, while the expression connecting wind and surface-water movements was linear with winds up to about 3 Beauforts, above this strength it appeared to obey a different linear relationship or to become curved. A similar conclusion was also reached by Mandelbaum (1956). However, this idea was subsequently disregarded by Lawford and Veley, and also by Mandelbaum (1958), who assumed that the following equation best presents the relations between surface currents and winds:

(54) $$W = K\sqrt{V} \qquad (\text{cm sec}^{-1})$$

where K is 2.7, W in cm sec^{-1} and V in m sec^{-1}. It was pointed out already by WITTING in 1909 that the surface current speed is proportional to $\sqrt{V}$ (LISITZIN 1938).

Although Formulas (53) and (54) are derived on the basis of long-period (weekly and monthly) data, Formula (54) is best suited for the estimation of the speed of the wind currents during 24 hours and longer periods and also allows partly for the »rest currents» usually associated with low wind velocities (possible »inertial» wind currents caused by earlier stronger winds). For steady winds the Formula of HELA (53) is useful but tends to give too high values for winds above 15 m sec^{-1}.

PALMÉN (1930 a) found from theory, which was partly confirmed by observations, that the change of current speed with depth is proportional to the density gradient. There are no additional proper data available for quantitative re-evaluation of PALMÉN's finding. The examination of the data on the change of current speed with depth, given by STEVENSON (1958) and the actual measurements on Finnish lightships, show that the speed varies only slightly with depth over the continental shelf, the greatest changes occurring in the upper first metre. Formulas (53) and (54) can be considered as presenting the average wind current in the upper 5 metres.

There seems to be no appreciable latitudinal speed effect in the case of wind currents. However, the steadiness and duration of the wind should be accounted for in more accurate hindcasts, estimated with the use of accumulated data.

The direction and speed of the prevailing winds also considerably influence tidal currents, causing displacement of the centre of the current ellipse and its deformation (MANDELBAUM 1957). These modifications of tidal currents by wind vary with locality and can be allowed for in hydropsis, using experiences based on corresponding local measurements.

Besides knowing the speed of the wind current, we need to estimate also the direction of the current in relation to the direction of the wind. HUGHES (1956) found from the drift of the plastic envelopes that they moved approximately in the same direction as the gradient wind. However, the envelopes are exposed to the direct influence of the wind at the sea surface and might not represent the actual direction of wind currents in relation to the direction of the wind.

According to EKMAN's theory the deflection of the wind currents should be about 45° to the right of the wind direction in the northern hemisphere and 45° to the left in the southern hemisphere. HAY (1954) found that, at two different weather ships in the North Atlantic, the average deflection of the current was respectively 42° and 29° to the right of the wind direction. TAKEDA (1938) showed theoretically, on the basis of the mixing length theory of turbulence, that the deflection of the surface drift current from the wind is less than 45°. It has been shown by many other workers that EKMAN's theory is not valid in detail either on the deflection of the wind current or on the change of current speed with depth.

HELA (1952) showed that the stronger the frictional force, as compared relatively with the corresponding drift current, the smaller the angle of deflection *ceteris paribus*. He also found that, in the Gulf of Finland, the rotation of the characteristic current was *cum sole* with increasing wind velocity. This direction of rotation was confirmed by STOMMEL (1954). However, in shallow seas the direction of rotation of wind currents in the case of a change of wind direction might be influenced by the rotation of tidal currents, which may, on certain occasions, be *contra sole*.

BOWDEN (1953) found with the GEK that during a radical wind shift, when wind velocities ranged from 18 to 30 knots, the current direction changed and became steady within 3 hours. The deflection was 18° to the right of the wind direction. However, STEVENSON (1958) found that on the continental shelf the direction of the wind current in relation to the direction of the wind varied with locality.

To check the direction of the current in relation to the wind in coastal areas, a comparison of the directions of the wind and the surface currents was made from the data of the lightship »Storbrotten» (see LISITZIN 1955). The data plotted on Figures 21 and 22 reveal that in most cases the surface current is from the same direction as the wind or flows slightly to the right. Further analysis of the data showed that, with stronger winds the angle of deflection is small or the directions of the wind and current are approximately the same, thus confirming the conclusions of LISITZIN (1938) and HELA (1952). LISITZIN (1938) found an average deflection of the surface current to be 12° to the right of the wind at the lightship »Storbrotten»*

The observations on this lightship of directions of winds and currents were made for 32 directions. From the data it is obvious that the observer

* HELA (1952) found that in the Gulf of Finland the angle of deflection can be expressed with an empirical formula: $a = 34° - 7.5\sqrt{V}$.

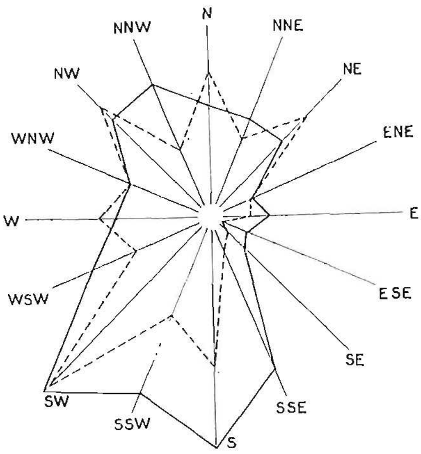

Figure 21. Directions of winds and currents at Lightship »Storbrotten», 1953.
——— wind – – – current

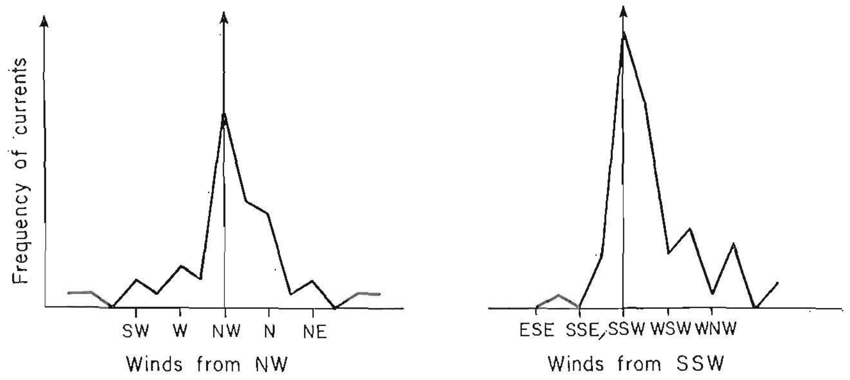

Figure 22. Directions of currents in relation to directions of winds at Lightship »Strobrotten», 1953.

in most cases used only 16 directions, probably due to fluctuations and other factors making accurate determination of direction difficult. However, as a very rough estimation of the deflection, 20° to the right can be given. HELA (1952) also found that the »characteristic current» was the sum of a real permanent flow and an additional current vector deflected about 20° to the right of the direction of the resultant wind.

Comparison was also made of the wind and current data of two lightships close to each other. (»Storbrotten» 60°26′N, 19°13′N and »Grundkallen» 60°34'N, 18°58'E). Three observations a day were made from the Finnish lightship »Storbrotten» and currents were measured at three depths. The observations from the Swedish lightship »Grundkallen» were made only once a day at 8 a.m. (probably Swedish Standard Time) and the currents were measured at two depths only (Anon., 1954). The first observation on the »Storbrotten», 7 a.m. Finnish Standard Time, and the Swedish observation are compared, there being, however, a time difference of 2 hours between these observations.

There were few differences in the wind directions at the two lightships, the average ranging from 10° to 20°. Considerable differences existed in the estimation of the wind speeds, which were generally 1 to 2 Beauforts higher at the »Grundkallen», but occasionally lower. This is obviously the result of incorrect estimates of the force.

The differences in the current directions measured by the two ships were sometimes up to and above 90°. The most remarkable were the differences in the current velocities, which were considerably lower at the »Grundkallen». These differences could be caused by the differences in the current measurement techniques and also partly by differences in bottom topography.

A comparison of the surface temperature and wind data at the »Grundkallen» showed a relation between the direction of the wind and the surface temperatures. A similar observation was made by Hela (1951) in another locality. This relation might be caused by the effects of advection, the piling-up of surface water or the upwelling of intermediate water.

From the discussions above, it becomes evident that the deflection of the wind currents varies with locality and should be ascertained from measurements. 20° to the right could be taken as a rough estimate of the deflection in medium latitudes if no data are available. The deflection most probably varies slightly with latitudes.

The sea level responds to the changes of atmospheric pressure, as will be described in Chapter 19, so that 1 millibar change of pressure corresponds roughly to 1 cm. change of sea level. This response can partly be dynamic, i.e. some water has to flow from a high pressure region into a low pressure region, in connection with an internal »pressure wave», discussed earlier. The pressure changes in a given locality are rather rapid and these changes will most probably also influence the direction and speed of the resultant surface currents.

The distribution of the surface pressure and the winds over the North Atlantic on March 3rd, 1957, is shown in Figure 23, and Figure 24 shows

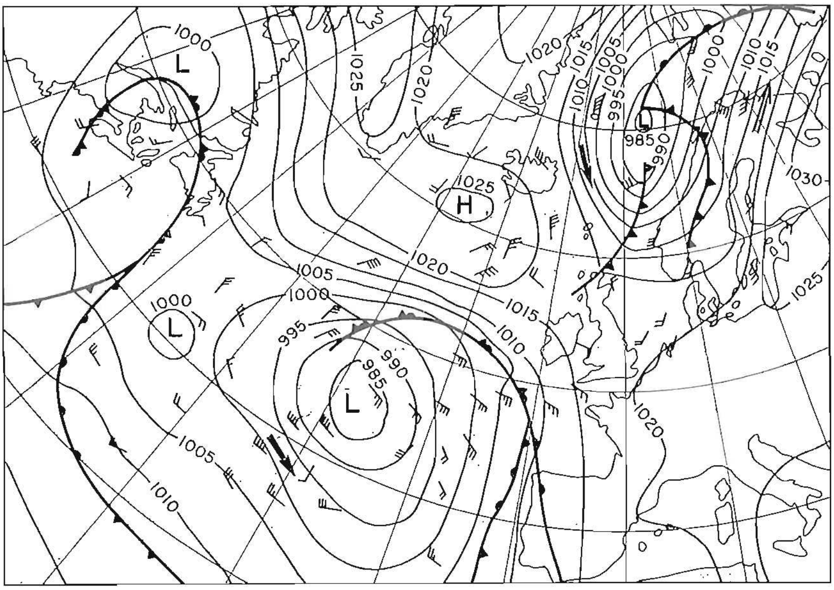

Figure 23. Barometric pressure (mb) and winds on 3 March 1957 in the North Atlantic.

the change of atmospheric pressure between the 2nd and 3rd of March 1957. The magnitude of the influence of the current component caused by pressure is unknown. However, some idea can be obtained by adapting the available approaches used in the dynamical computations of currents. A qualitative analysis of these figures (and other similar plottings not shown here) in respect to surface currents permits the following observations:

The surface current component caused by atmospheric pressure changes is directed approximately towards the movement of low pressure centres. Adding this current component to the wind currents, we find that the resultant surface current is approximately along the isobars, if no permanent flow is present. The zero changes of pressure coincide in the majority of cases with the fronts. There is also a transport of water particles by surface waves which is in general assumed to be with *circ.* 1/5 of the speed of the progression of the corresponding waves.

It seems that the gradient current caused by atmospheric pressure-changes and the water transport by waves may explain the short term fluctuations in the direction and speed of surface currents, the so-called

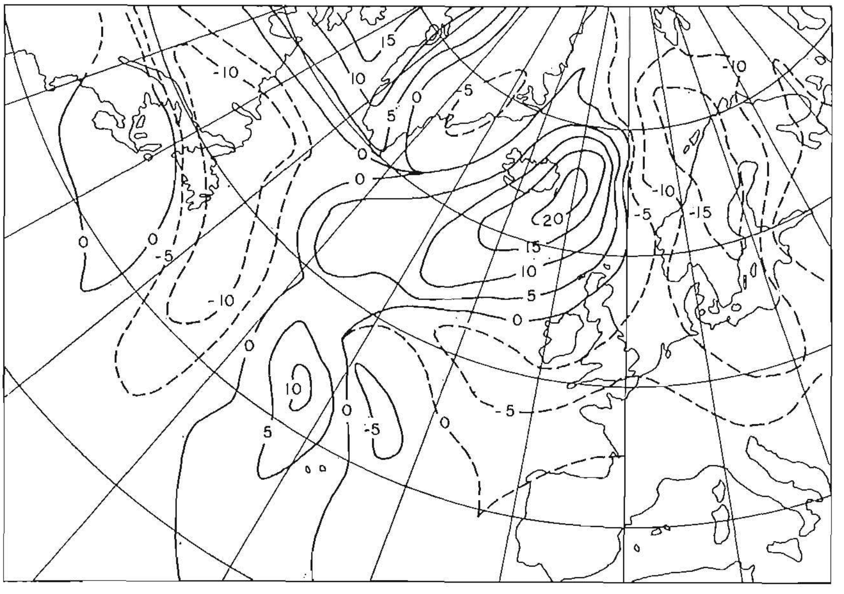

Figure 24. Change of barometric pressure (mb) between 2 and 3 March 1957 in the North Atlantic.

macroturbulence, and the cutting off eddies. Experimental verification of the influence of these current components would be difficult. One indirect possibility could be by means of prognostic current charts. A prognostic current chart could be prepared for every 12 hours on board (considering permanent flow, wind currents, transport by waves and gradient currents caused by atmospheric pressure). The correctness of this prognostic chart could be checked using the drift of the vessel under way. The other possibility is to use self-reporting (telemetring) drifting buoys.

17.4. TIDAL CURRENTS

Tidal currents usually predominate on the continental shelf. However in semi-closed basins they can be negligible (the Baltic and Mediterranean Seas, for example). Although our knowledge of the tides in coastal area is excellent, the offshore tides are difficult to measure and compute. RATTRAY (1957) tried to compute the tidal currents in offshore regions and obtained good results in regions where the frictional effect was negligible, using the data from adjacent shore stations. The results were poor for the longshore currents, where friction was strong.

Tidal currents are an important cause of mixing on the continental shelf. Large and almost periodic variations in the sea temperature, observed in shallow stratified water along the coast, may be caused by tidal stirring over an irregular bottom and by subsequent horizontal and vertical oscillating movements associated with the tides and by capsizing and cascading of internal waves in shallow water, especially on continental slope. DIETRICH (1954) found that the effect of the turbulence of the tidal stream increases with the approach to shallow water. He concluded also that the mixing layer close to the bottom depends more on the tidal stream, while near the surface it is caused by the wind effect.

The velocity of the tidal current remains nearly constant with depth, decreasing only close to the bottom, due to friction. This fact also explains DIETRICH's observations. The mixing by tidal currents occurs along the bottom, where it is limited to a relatively thin layer, and on the submarine slope where there is a change of depth. Mixing by tidal currents can also occur along the boundaries of different water masses. As tidal currents fluctuate back and forth, it is important to determine the horizontal distance from the slope, the shallowest depth or the current boundary where a thorough mixing is expected. The formula of HAYAMI, FUKUO and YODA (1957) can be used for the purpose.

They suggested that the length l_m travelled by a tidal current from slack to slack can be assumed to represent mixing length.

(55) $$l_m = W_t \frac{t_{at}}{2}$$

where $\frac{t_{a_t}}{2}$ is the half period of the tidal current.

This Formula (55) requires a knowledge of the average speed of the tidal stream, which must be ascertained by measurements in each locality or computed with the formula derived from FLEMING's (1938) work:

(56) $$W_{max} = \frac{2\pi x a_t}{TH_w}$$

The average velocity of the tidal current (W_t) is $\frac{2}{\pi}$ times W_{max}. Furthermore, directional corrections must be applied in most areas where the tidal currents form a pronounced ellipse.

For a rough approximation, one can assume for most areas that, in the horizontal distance of the tidal mixing length from the bottom of a slope where the thermocline would intercept the bottom contour, no sharp thermocline exists. It can also be assumed that this mixing length represents in certain conditions a mixing zone in current divergences and convergences.

17.5. CONVERGENCES AND DIVERGENCES OF CURRENTS

Mixing of water masses in a horizontal direction takes place at the convergences of currents with different water masses and with consequently different temperatures. The upwelling of deep water occurs at the divergences. These divergences and convergences are in no way stationary but fluctuate as to position and these fluctuations cause abrupt changes in the localities which they cover, and need to be predicted, because the sharp temperature boundaries act as barriers to the migration of marine animals.

The reason for the fluctuations of the current boundaries can be found in meteorological conditions. FUKUOKA (1955) concluded that variation of the oceanic polar front may be caused by zonal distribution of wind systems. MONTGOMERY (1938) concluded that the total horizontal convergence or divergence of the surface layer is given by convergence or divergence of drift current.

HELA (1954) gives a method for computing current divergences and convergences using available atlases of surface currents.

A strict mathematical treatment of the current boundaries is not possible at present. Tentatively, principles for the determination of the position of current boundaries and their changes are established below:

1) The average monthly or seasonal divergences and convergences can be computed from available surface current data, as done by HELA (1954);
2) Outside the regions of strong permanent currents, the divergences and convergences of surface currents are determined by the divergence and convergence of the winds. In order to account for the positions of these current boundaries and their changes with time, the velocities, directions and duration of the winds on both sides of the boundaries must be considered. When the winds are above normal on the warmer water side of a convergence or on the colder water side of a divergence, the warm water is expected to spread over the colder. If the opposite is the case with the winds, a more intensive mixing on the current boundary is expected;

3) When accounting for the changes of divergences and convergences along the coast and on the continental shelf, the piling-up action, mass transport and the depth of water must be taken into consideration, in addition to the winds;
4) The changes of the boundaries of permanent current are also influenced by the changes of total mass transport of either one of the currents. Normally, when the mass transport of one current increases, the boundary moves towards the current with less transport. The changes of the heat balances of the different currents might also cause a change in their positions. If the heat gain of the colder current increases above normal in relation to the warmer current, the convergence moves towards the colder current or, in case of current divergences, an increase of mass transport of the warmer current can be expected.
5) When the transport of a major current is small the current pattern changes easily. Overspread of a warm weak current over a cold current can then also be expected (TAKENOUTI, 1957).
6) The boundaries can be sharper towards the movement of the boundary especially if the boundary moves towards the direction of the colder side. Relatively little mixing is expected on these fast moving boundaries. The speed of the movement can be considerable. Speeds up to one knot have been reported (KNAUSS 1957).

A preliminary test of the influence of the winds on the movement of the current boundary between the Labrador Current and the North Atlantic Drift Current was made with the data from DINSMORE, MORSE and SOULE (1958) and from Wetterkarten des Seewetteramts, Hamburg (kindly put at the disposal of the author by Dr. RODEWALD). Figure 25 shows the 0°, 5° and 10°C surface isotherms SE of Newfoundland from the 20th to the 28th of February, 1957. Figures 26 and 27 show the same isotherms and average pressure distribution (only in Figure 26), and the resultant direction and relative speed of the surface winds in the same area from the 1st to the 15th of March and the 16th to the 31st of March, 1957. The heat gain of the water during March in this area is very small. Although a detailed analysis of the situation is not within the scope of this work, it can be easily seen that: average pressure distribution does not explain the changes of the positions of the isotherms, but some of the changing features can easily be explained on the basis of wind directions, relative speeds and resultant wind divergences and convergences. Only the surface winds deduced from the synoptic charts at noon have been used to plot the resultant direction and relative force in 2° squares. If a more detailed prediction of the situation is desired, more frequent wind data will be needed and should be plotted in 1° squares.

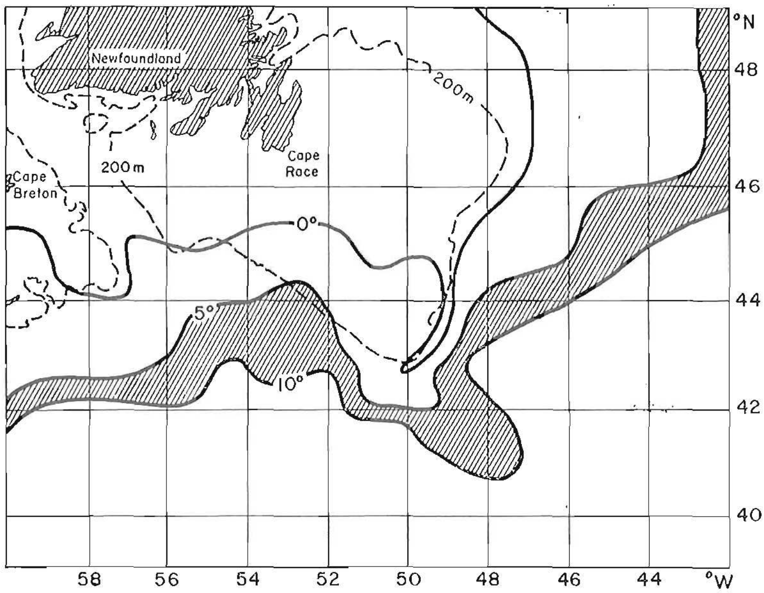

Figure 25. 0°, 5° and 10°C surface isotherms SE of Newfoundland from 20 to 28 February 1957. (Dinsmore, Morse and Soule 1958).

17.6. SUMMARY OF CHAPTER 17

The importance of the current as a factor affecting temperature changes in a given locality is discussed, and the necessity for the prediction of the actual surface currents for a day or week have been pointed out. For the computation of the temperature change caused by advection Formula (49) is given. This formula requires knowledge of the current speed and direction. In order to ascertain these parameters, the permanent and wind currents should be separated and the wind current predicted from the wind data. For this quantitative separation the method of PALMÉN and HELA is suggested. Another similar approach by MANDELBAUM has been found incorrect on the basis of the current data from the lightship »Storbrotten».

For the estimation of the speed of the wind current Formula (54) is suggested. The direction of rotation of wind currents is discussed and

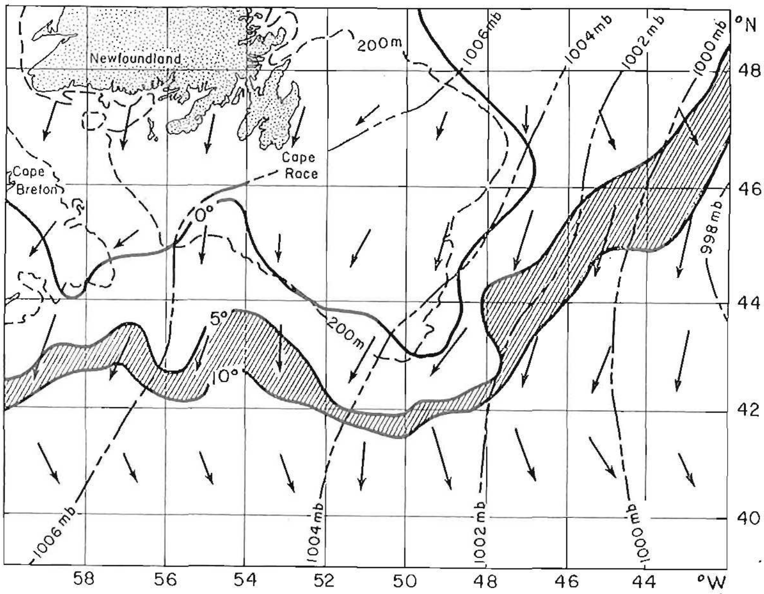

Figure 26. 0°, 5° and 10°C surface isotherms, average distribution of pressure and resultant directions and relative speeds of winds SE of Newfoundland from 1 to 15 March 1957.

found to be usually *cum sole* in the northern hemisphere. Local differences in the deflection of the current direction from the direction of the wind are discussed, and an average deflection of 20° to the right in medium latitudes in the northern hemisphere is suggested.

The influences of the changes of atmospheric pressure and wave transport on the surface current speed and direction are briefly discussed, and it is found probable that the change and distribution of the atmospheric pressure and wave transport cause the fluctuations of direction and speed of resultant surface currents. It is proposed to check the above hypothesis by the use of prognostic current charts.

The influence of tidal currents on the sea temperature is briefly discussed, and the difficulties for the computation of offshore tides are pointed out.

General rules for the estimation of the movement of convergences and divergences are established and the influence of winds on the movement of a current boundary is illustrated.

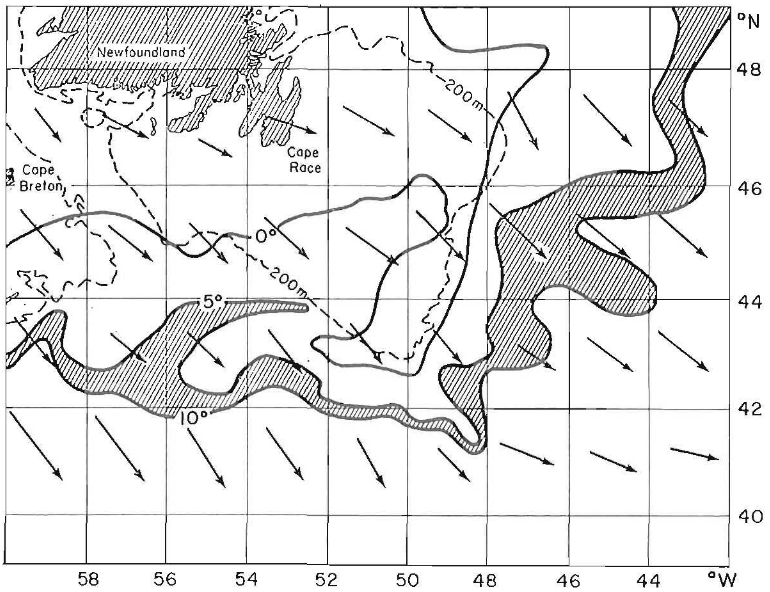

Figure 27. 0°, 5° and 10°C surface isotherms and resultant directions and relative speeds of winds SE of Newfoundland from 16 to 31 March 1957.

18. TURBULENT MIXING BY WATER MOVEMENT

In earlier chapters convective stirring and mixing by wave action have been described. This chapter deals with mixing by currents. In order to account for the temperature changes caused by the mixing of waters of different temperature one should be able to determine the mixing by currents quantitatively, using parameters contained in routine measurements.

The formulas for molecular diffusion and conductivity are not applicable in the upper layers of the ocean. Instead we have to deal with eddy diffusivity and eddy conductivity, which is usually called austausch. The vertical austausch coefficient A varies by more than three orders of magnitude with space and time in the oceans and is at present neither measurable nor predictable, although numerous attempts have been made to solve the problem. The problems of mixing and turbulence in the sea are discussed and summarized by DEFANT (1954).

JOSEPH and SENDER (1958) concluded that the existing expressions of the austausch were not suitable for the consideration of diffusion in the sea. They derived a useful formula for the computation of diffusion on a horizontal level from point source and for the computation of the concentration at any given distance from the point source at a given time.

The brief summary above shows that it is impossible to derive an exact and, under different conditions, universally applicable mathematical integration from the differential equations of motion and conductivity (and/or diffusivity). Also the necessary observations in time and space are not always available for computing the austausch or for predicting the mixing. Therefore our present task is to find, if possible, a simple consideration and expression of mixing in terms of the parameters which are contained in routine oceanographic and meteorological observations or can easily be derived from them (for example, stability, currents, waves etc.).

It is sometimes necessary to express quantitatively the vertical turbulent mixing by currents, for instance when there are two overlaying opposite currents or strong permanent currents with weak winds and low stability of water. Such situations are, however, not very common in the sea. Furthermore, in localities with two overlying strong currents, local factors considerably affect vertical turbulent mixing (for example, the shape of the coast and the bottom topography, superimposed tides, wind, etc.). Horizontal turbulent mixing due to current cannot be neglected if considerable horizontal gradients of conservative properties exist. This horizontal mixing is influenced by a) the speed of the current, b) the horizontal velocity gradient caused by changes in the depth of the current or its extension in a horizontal direction, and c) stability. These factors vary locally and seasonally. Therefore no satisfactory theoretical or 'universal' empirical solution can exist to account for the vertical mixing in the cases mentioned above. It is therefore suggested that the changes of properties caused by mixing by currents be determined empirically during different seasons and varying hydrographical conditions and the data so found can be used for correcting the temperature changes due to advection in the given locality. Experience shows that in stratified water the horizontal mixing by currents can be considerable, but in vertical direction negligible if the currents are continuous and bottom topography regular.

In certain »estuary conditions» the approximation by Keulegan (1949) is useful for a rough estimation. Keulegan (1949) showed that the amount of mixing U_m, defined as the volume of the heavier liquid that crosses unit area of the interface in unit time and having dimensions of velocity, could be related experimentally to the actual velocity of the surface layer W_o and the critical velocity of mixing W_c by:

$$U_m = 3.5 \times 10^{-4} (W_o - 1.15 W_c) \tag{57}$$

$W_c = 0.127$ in the case of REYNOLD's number < 450, and 0.178 in the case of REYNOLD's number > 450. This method has been applied by WALDICHUCK (1957).

Specific features of the distribution of properties in the currents must be borne in mind when making the corrections to the temperature changes, caused by advection; that is: 1) there is usually a tonguelike distribution of properties in a current, caused mainly by a) mixing which takes place along the edges of the current, and b) the horizontal current velocity in a gradient current, which is usually higher at its centre of horizontal extent; 2) formation of eddies occurring in the boundary regions of stronger currents as well as in coastal regions, caused by topography. These eddies are usually variable in space and time.

18.1. SUMMARY OF CHAPTER 18

The existing expression of austausch is found unsuitable for quantitative consideration of mixing in the sea. It is also not possible to derive a universally applicable empirical relation for this purpose, and the mixing by currents must be ascertained locally. Some characteristics of the turbulent mixing by currents are pointed out. For rough quantitative estimation of this mixing in »estuary conditions» the use of KEULEGAN's is Formula (57) is recommended.

19. SEA LEVEL AND TEMPERATURE

Among the procedures to correct the change of temperature of the sea in a given locality is the indirect way to estimate the influence of the sea level. The variations of sea level caused by piling-up of surface layers increase the temperature of a given locality and also increase the thickness of the thermosphere. The opposite is usually the case with a lowering of the sea level.

The unperiodic variations of sea level are caused by atmospheric pressure (UNOKI, 1950) and by winds. HELLAND-HANSEN and NANSEN (1920) found a correspondence between water level and temperature on the European coast. This finding was confirmed for the North Pacific by

FLEMING (1955). LISITZIN (1957) found that the slope of the water surface due to wind stress can be determined from the formula:

$$\Delta H = \frac{2.98\ V^2}{H_w} \tag{58}$$

LISITZIN (1958) tested a similar formula for the computation of sea level changes due to winds in the Gulf of Finland.

There are relatively few data on the influence of the barometric pressure on the sea level. HAMON (1957) and LISITZIN (1957 b) have tentatively assumed the following relation:

$$\Delta h_s = 13\ \Delta P \tag{59}$$

where Δh_s is the change of sea level in mm and ΔP the change of atmospheric pressure in mm Hg. It is obvious that the response of the sea level to atmospheric pressure is not simultaneous, but that a certain time lag must occur.

The sea level change caused by changes of atmospheric pressure and by advective factors (transport by waves and wind currents) is most pronounced along the coasts. This level change is related to the wind speed, cosinus of the angle between wind direction and the coastline, depth of the water, length of the fetch and duration of the wind as well as the atmospheric pressure change.

The influence of air pressure change and the piling-up by wind on the depth of the thermocline in offshore waters has been little investigated. On the basis of density differences, it can be assumed that the influence on the depth of the thermocline is considerably greater than on the change of sea level. Tentatively, the change of depth of the thermocline caused by piling-up and pressure changes could be assumed to be *circ.* 10 times the sea level changes by a pronounced pycnocline, and more if the pycnocline is less developed. The duration of the winds must also be taken into consideration in these estimates.

The piling-up action of the winds causes part of the relatively thin warmer surface layer to move approximately in the direction of the wind and causes temperature changes in the regions of piling-up. It is impossible to account for these changes by mathematical formulas, and necessary approximate empirical corrections should be made whereby the seasonal peculiarities of the upper surface layer, the state of mixing, etc., would be taken into consideration. A preliminary analysis of the wind and water temperature data from the Swedish lightship »Hävringe»

(58°33'N, 17°31'E) substantiates the conclusions and recommendations established above. Noticeable in these data is also the rise of the thermocline and the decrease of the surface temperature during winds when a »blowing away» of the surface layer is expected.

19.1. SUMMARY OF CHAPTER 19

The surface temperature and its changes within the depth may be also caused by the variations of the sea level, which in turn are caused by winds and atmospheric pressure. The formulas from LISITZIN (1957) are suggested for use in the estimation of the variations of the sea level. Tentatively, it is suggested that the effect of the sea level causes a tenfold effect in the depth of thermocline; for example, a 20 cm. rise of sea level will lower the thermocline *circ.* 2 metres.

PART III

TEMPERATURE HINDCASTS AND FORECASTS AND SOURCES OF ERROR

20. HYDROPTIC AREA, PERIOD AND SELECTION OF DATA

As in meteorology, so also in oceanography the forecaster's skill and his knowledge of local conditions play an important role in the success of the forecast. In this Chapter, some general advice is given on the methods of selection of data and on the necessity to consider local conditions carefully. In the following Chapters, a few examples illustrate the application of these methods, and the difficulties and sources of error are discussed.

20.1. HYDROPTIC AREA

The hydroptic area for a single computation should not be selected to be greater than 5° square because of the changes of meteorological and oceanographic elements with space. Several heat exchange computations must be made for forecasts of larger areas. For the forecasting of the temperature changes in a given locality, the effects of heat advection by currents must be considered, as discussed in Part II. Therefore the possible directions of inflow — either the permanent flow or surface currents caused by a steady prevailing direction of the wind — into the selected hydroptic areas must be ascertained and separate heat exchange computations made for these adjacent areas from which advection is expected, in order to account for the changes of temperature occurring during the transport. In coastal areas the hydroptic area can often be defined by the local topographical features. The general features of atmospheric pressure and wind distribution in a larger area around the selected hydroptic area should be ascertained in order to account for the changes of sea level and for the piling-up action of the wind, which cause secondary changes in the depth of the thermocline.

20.2. HYDROPTIC PERIOD AND SELECTION OF DATA

The hydropsis is usually calculated for a week; therefore the possible day-to-day sea surface and air temperature variations due to advection must be estimated from an extended (for example, five day) weather forecast. General seasonal trends can be used to some extent instead of long-range weather forecasts. These trends can usually be derived from data in hydrographic and maritime meteorological atlases. As day-to-day meteorological data become available during the hydroptic period, a new calculation for each day can be made to correct the original forecast. These corrections are especially necessary when no actual weekly hydrographic stations or sea surface temperature observations are made in the hydroptic area. Therefore these »back corrections» can ascertain, to a certain degree, that for the start of the next hydroptic period the starting conditions in the sea are close to real conditions.

During cloudy days (and generally during autumn and winter), when the diurnal variations of the meteorological elements over the sea are small, it would be sufficient to make one general 24-hour calculation, taking the average of the values of meteorological parameters reported at 00^{00} and 12^{00} local hours. During clear, fair days (and, in general, during late winter, spring, summer and early autumn) it is advisable to make separate calculations for night and day conditions if a higher accuracy of temperature is desired. In certain conditions, the downward convective transport of heat and the condensation of water vapour during the night might not be compensated for by using 24 hour average values. This is especially the case, if the days are shorter than 9 hours or longer than 15 hours. In this case, it is advisable to make separate calculations of the heat budget, using day and night average temperature and humidity values, assuming, for simplicity, that the length of the day is one-third or two-thirds respectively of the 24 hour period. The possible short distance variations of temperature and humidity must be considered, especially in coastal areas. The accuracy of the forecast depends to a great extent on the availability of accurate and proper meteorological data.

21. HINDCASTS OF CHANGES OF TEMPERATURE STRUCTURE IN TWO OCEANIC AND ONE COASTAL AREA

In this Chapter, some examples are given of computations of temperature forecasts in three different areas and different seasons, together with varying quality and quantity of available data for these computations. These examples might serve three purposes: (1) to illustrate the forecasting procedure in a variety of conditions; (2) to test possible accuracy and to discover sources of error; and (3) to show the possible future orientation of research and eventual partial re-organization of oceanographic observations and data handling.

There are difficulties in finding proper data and places for use in hindcasts, because past oceanographic and maritime meteorological data are usually incomplete for the purpose, or presented in such a manner that they cannot be used for present hindcast purposes. All the necessary data are usually not available in the same institution but are divided between oceanographic and meteorological institutes. The following computations are not made for an area but rather for a point from which time series of oceanographic data were available, because it was desired to compare hindcast and measured data. Most of the data and results of the computations are presented in tabular form and are easily comparable. The comments and discussions are therefore very brief. The general characteristics of the three selected points are listed below:

(1) 39°N, 153°E (Japanese Fixed Point E) in the North Pacific, SE of the convergence of Oyashio and the North Pacific Drift Current (the extension of Kuroshio). During winter, the convergence is closer to the Fixed Point than in summer. Cold eddies, cut off from the Oyashio, might reach the area. The sea surface temperature anomaly at the point is negative during the winter and positive during the summer.

(2) 66°N, 2°E (Weather Ship M) in the Norwegian Sea, SE of the Oceanic Polar Front. Cold eddies, cut off from the extension of the East Greenland Current, in the area NE of Iceland, are often carried into the area. The sea surface temperature anomaly is positive for most of the year.

(3) 58°33'N, 17°31'E (the lightship »Hävringe») in the Baltic Sea, NW of Gotland. The location is characteristic of a partly sheltered shallow coastal area.

21.1. NORTH PACIFIC (39°N, 153°E)

Meteorological and oceanographic data for this location have been published by the Japanese Meteorological Agency (Anon., 1952 a and b). Meteorological data were recorded 8 times a day. In Table 8, these data are averaged daily from noon to noon, to illustrate day-to-day variability.

In normal forecasting it is scarcely possible to make such detailed fore casts of meteorological conditions every day, and the best approach could be the estimation of average conditions for two to three days. Of interest in Table 9 is the seasonal variation of the evaporation and the convective transfer of heat, which are both determined by the wind speeds and the difference of the sea surface and air temperature and the water vapour pressure. Data on the last parameter are very sparse in daily weather reports. A possible method of improving the estimates of relative humidity for certain localities, especially in coastal areas, would be to ascertain the monthly average humidities for different wind directions (and/or different air masses).

In Table 10, the duration of the wind is estimated approximately for wave computation purposes, with the consideration of variations in direction and speed. The computed wave heights are compared with the observed wave heights. In the observed wave heights are given the higher values of the height ranges for every observed state of the sea code. These data can therefore be considered as a rough approximation. It is obviously desirable to replace in routine observations the state of the sea code with actual estimates of significant wave height.

The wind current is estimated for convenience to deflect *circ.* 22.5° to the right of the wind. The resultant current direction and speed is computed by simple dynamical addition (using the graphical method) of the wind current and permanent flow.

The permanent flow is estimated for the particular locality from the data on the surface currents (U.S. Navy Hydrographic Office, 1944) and from the resultant direction and force of the wind (U.S. Department of Agriculture, Weather Bureau, 1938). These data do not allow an exact computation of the permanent flow and the data given in Table 10 must be considered as very rough estimates only. No data are available to the author which would permit the estimation of the current component caused by changes of atmospheric pressure. Likewise the possible advection caused by wave transport is not considered. The surface temperatur gradient is estimated from the monthly temperature charts (U.S. Navy Hydrographic Office, 1944). No heat exchange computations have been made in the present tests on the water masses carried into the area, because of the relatively short distance of travel and the short hindcast period. No estimates of the range of the fluctuation of the depth of the thermocline are made in the following hindcasts. The advectional changes in Table 10, Column 14, have been computed from wind currents only. The resultant current velocities and directions (Columns 10 and 11 in Table 10) serve only as examples of the interaction between the permanent flow and the wind currents. However, knowledge of these resultant currents is important in practice (for example, for navigation, fishing, etc.). Comparison of the temperature changes caused by advection

Table 8. *Heat exchange at 39°N, 153°E in the North Pacific*
(Meteorological data)

No.	Date	T_{db}	T_{wb}	$T_{db}-T_{wb}$	U_o	T_w	T_w-T_a	e_w	e_a	e_w-e_a	V	C	t_d	A_n
February, 1950														
1	16–17	4.5	2.4	2.1	84	11.5	7.0	13.3	7.1	5.2	6	10	654	37
2	17–18	8.8	7.7	1.1	93	11.4	2.6	13.2	10.6	2.6	17	10	656	37.5
3	18–19	4.3	2.8	1.5	88	11.0	6.7	12.9	7.3	5.6	19	7	658	38
4	19–20	4.9	3.3	1.6	88	11.0	6.1	12.9	7.6	5.3	10	7	660	38.5
5	20–21	6.9	5.0	1.9	88	10.8	3.9	12.7	8.7	4.0	16	6	662	39
6	21–22	2.5	0.8	1.7	85	10.3	7.8	12.3	6.2	6.1	10	5	664	39.5
June, 1950														
7	23–24	17.4	16.1	1.3	96	18.0	0.6	20.2	19.0	1.2	10	8	890	74.5
8	24–25	16.9	14.7	2.2	92	18.0	1.1	20.2	17.8	2.4	8	7	890	74.5
9	25–26	17.2	15.3	1.9	93	17.8	0.6	19.9	18.3	1.6	7	8	890	74.5
10	26–27	18.5	17.8	0.7	98	18.4	−0.1	20.7	20.8	−0.1	9	10	890	74.5
11	27–28	19.0	17.7	1.3	96	18.9	−0.1	21.4	21.1	0.3	3	8	890	74.5
12	28–29	19.5	19.2	0.3	99	19.3	−0.2	21.9	22.5	−0.6	7	10	890	74.5
13	29–30	19.3	18.7	0.6	98	19.0	−0.3	21.6	22.0	−0.4	12	10	890	74.5
October, 1950														
14	10–11	19.8	18.3	1.5	96	20.2	0.4	23.2	22.1	1.1	4	6	690	44
15	11–12	18.9	16.5	2.4	93	19.8	0.9	22.7	20.2	2.5	6	5	687	44
16	12–13	20.0	19.6	0.4	99	20.3	0.3	23.4	23.1	0.3	6	8	684	43.5
17	13–14	18.7	15.4	3.3	90	19.9	1.2	22.8	19.3	3.5	9	5	681	43
18	14–15	18.2	12.3	5.9	81	19.8	1.6	22.7	16.7	6.0	9	5	677	42.5
19	15–16	16.3	9.4	6.9	75	20.1	3.8	22.9	13.8	9.1	7	6	675	42.5
20	16–17	18.5	14.1	4.4	86	19.7	1.2	22.5	18.3	4.2	14	10	672	42

Remarks: 1) Notations and units used, see Chapter 4. (Units: all T in °C, U_o in %, all e in mb, V in m sec^{-1}, C in tenth of sky, t_d in min and A_n in °.)

2) The meteorological data presented are averages from noon to noon of available eight observations in 24 hours.

3) The height of anemometer is not specified in the original data but is assumed to be at »standard» height (*circ.* 8 m).

Table 9. *Heat exchange at 39°N, 153°E, in the North Pacific*
(Heat budget computation)

No.	Q_{os}	Q_s	Q_r	Q_{ob}	Q_b	E	Q_e and Q_c	Q_h	Q_l Residual heat
February 1950									
1	339	136	18	194	46	4.46	263	197	−383
2	334	138	18	186	44	4.08	241	159	−324
3	350	278	34	192	89	9.63	568	450	−863
4	356	283	34	192	89	5.46	322	245	−407
5	361	314	37	192	104	5.96	352	227	−406
6	367	339	39	196	121	6.28	371	314	−506
Total		1 488	180		493		2 117	1 592	−2 894
June 1950									
7	928	643	55	171	66	1.24	73	24	425
8	928	737	56	174	81	2.11	124	37	439
9	928	643	55	174	68	1.28	75	19	426
10	928	371	42	168	39	−0.07	−4	−3	296
11	928	643	55	170	66	0.15	9	−1	514
12	928	371	42	166	39	−0.22	−13	−4	307
13	928	371	42	167	39	−0.37	−22	−11	323
Total		3 779	347		398		242	61	2 731
October 1950									
14	425	370	42	167	90	0.63	37	9	192
15	423	391	43	171	106	1.80	105	25	112
16	417	289	35	164	64	0.22	13	8	169
17	410	379	42	173	107	3.33	195	45	−10
18	403	373	42	181	112	5.70	334	59	−174
19	401	349	40	187	101	7.28	427	119	−338
20	395	158	21	177	42	5.63	330	63	−298
Total		2 309	205		622		1 441	328	−347

Remarks: 1) Number in the first column refers to the corresponding entry in Table 8.
2) Residual heat (Q_l) is computed with the Formula (2):

$$Q_l = Q_s + Q_p - Q_r - Q_b - Q_e - Q_h$$

3) Notations and units used, see Chapter 4. (Units: all Q in g cal cm^{-2}; E in mm cm^{-2})

Table 10. *Hindcast of waves, currents and temperature changes caused by advection at 39°N, 153°E in the North Pacific*

Date	Wind velocity m sec^{-1}	Wind direction (from)	Length of fetch km	Duration of wind h	Wave height m		Computed currents				Surface temp. gradient in direction of wind current °C/100 km	Distance travelled by wind current km	Temp. change by advection °C
							Wind current		Resultant current				
					Computed	Observed *a*	Direction (from)	Velocity cm sec^{-1}	Direction (from) *b*	Velocity cm sec^{-1} *b*			
1	2	3	4	5	6	7	8	9	10	11	12	13	14
February 1950 16–17	6	NE		24	1.5	1.25	ENE	6.6	NNW	6.7	−0.7	5.7	−0.04
17–18	17	SE		24	8.9	9	SSE	11.1	SSW	7.2	+1.2	9.6	+0.12
18–19	19	NW		48	10.8	9	NNW	11.8	NW	18.7	−2.0	10.2	−0.20
19–20	10	NE	500	12	3.3	5	ENE	8.5	N	6.8	−0.7	7.5	−0.05
20–21	16	W		24	8.2	6	WNW	10.8	WNW	18.8	−1.8	9.3	−0.17
21–22	10	NW		24	3.7	2.5	NNW	8.5	NW	15.3	−2.0	7.5	−0.15
													−0.49
June 1950 23–26	8	E	500	24	2.3	0.3 to 1.5	ESE	7.6	ESE	8.2	+0.5	19.8	+0.10
27–30	8	S		24	2.3	1.3 to 2.5	SSW	7.6	SSW	14.5	+1.3	19.8	+0.26
													+0.36
October 1950 10–13	6	SE	500	24	1.3	0.5 to 1.3	SSE	6.6	SSE	6.0	+1.2	17.1	+0.20
14–17	10	W		48	3.5	1.3to 5	WNW	8.5	WNW	12.5	−1.5	22.5	−0.34
													−0.14

Remarks: a Upper limits of average wave heights of the corresponding observed state of the sea

b Velocity and direction of current as corrected for the effects of permanent flow, which is estimated as follows:

Month	Direction	Velocity cm. sec^{-1}
February	WNW	8
June	SW	7
October	NW	4

(e.g. Table 10, Column 14) and the changes caused by the heat exchange (e.g. Table 11, Columns 4 and 5) shows the great influence of the advective term in the temperature forecasts and hindcasts. Extremely few computations of permanent flow have been made in the oceans and seas and, furthermore, the existing seasonal data are represented in most cases, so that no such computation can be made with accuracy. For simplification, and because of the lack of proper data, horizontal mixing during transport is not considered in the following example. It is also assumed that the advective change is uniform from the surface to the thermocline.

Table 11. *Water temperature changes at 39°N, 153°E in the North Pacific from 16 to 22 February 1950*

Heat received: 1 488 g cal cm^{-2}, heat lost: 4 382 g cal cm^{-2}, $Q_1 = -$ 2 894 g cal cm^{-2}
Optical water mass 2
Depth of thermocline: 16 II 50 : 150 m; estimated for 22 II 50 : 125 m[a], 135 m[b]; measured: 100 m.

Measured 16. II 1950.			Forecast 22. II 1950.			Measured 22. II 1950.
Depth m	Temp. °C[c]	Cl°/oo	Temp. correction for absorption °C	Temp. correction for losses °C	Temp. corrected for Q_1 and advection °C	Temp. °C[c]
1	2	3	4	5	6	7
0.5	12.3	19.05			11.6	11.4
10	12.49	.05				11.57
20	(12.48)	.05	+0.12	−0.35		11.58
30						
40						
50	12.23	.04				11.52
75	12.24	.03				(11.50)
100	12.17	.03	(125)[d]		11.6	11.45
150	11.78	.00			11.5	11.42
200	10.84	18.87			10.8	(9.75)

Remarks: a Estimated from wave data
b Estimated with consideration of convective stirring
c Temperatures in parentheses are interpolated
d Thermocline depth used for computations

Table 12. *Water temperature changes at 39°N, 153°E in the North Pacific from 23 to 30 June 1950*

Heat received: 3 779 g cal cm^{-2}, heat lost: 1 048 g cal cm^{-2}, Q_l = 2 731 g cal cm^{-2}
Optical water mass 2
Depth of the thermocline: 23. VI 50: 27 m; estimated for 30. VI 50: 29 m[a]; measured: 24 m.

Measured 23. VI 1950.			Forecast 30. VI 1950.			Measured 30. VI 1950.
Depth m	Temp. °C[b]	Cl⁰/₀₀	Temp. correction for absorption °C	Temp. correction for losses °C	Temp. corrected for Q_l and advection °C	Temp. °C[b]
1	2	3	4	5	6	7
0.5	17.5	18.97			18.8	18.5
10	17.56	.96	+1.26			18.55
20	(17.56)	.96		−0.35	18.8	(18.5)
30			(30)[c]		18.5	(17.9)
40	(14.9)	(19.11)	+0.02		14.9	(15.3)
50	(13.8)	(.06)	+0.01		13.8	13.97
75	(12.55)	(.02)				12.82
100	(11.70)	(.00)				(11.75)
150	(10.15)	(18.95)				(10.5)
200	(8.90)	(.80)				(9.65)

Remarks: a Estimated from wave data
b Temperatures in parentheses are interpolated
c Thermocline depth used for computations

The measured temperature at the beginning of the hydroptic period is shown in Column 2 in Tables 11 to 13, and the measured temperature at the end of the period in Column 7. The depth of the thermocline is estimated from the wave data and, where necessary, corrected graphically for the effect of convective stirring. In estimating the temperature changes caused by the heat exchange, the specific heat as well as specific weight of sea water is assumed to be 1. Errors introduced with this simplification are very minor compared with other sources of error. Below the thermocline no changes are assumed and changes within the thermocline are estimated graphically.

Table 13. *Water temperature changes at 39°N, 153°E in the North Pacific, from 10 to 17 October 1950*

Heat received: 2 309 g cal cm^{-2}, heat lost: 2 656 g cal cm^{-2}, $Q_l = -347$ g cal cm^{-2}
Optical water mass 2
Depth of the thermocline: 10. X 50: 45 m; estimated for 17. X 50: 44 m[a]; measured: 38 m.

Measured 10. X 1950.			Forecast 17. X 1950.			Measured 17. X 1950.
Depth m	Temp. °C[b]	Cl‰	Temp. correction for absorption °C	Temp. correction for losses °C	Temp. corrected for Q_l and advection °C	Temp. °C[b]
1	2	3	4	5	6	7
0.5	20.7	18.96	+0.52 (44)[c]	−0.60	20.0	19.6
10	20.29	.93				19.76
20	(20.1)					(19.75)
30	(19.9)	(.96)				(19.65)
40	(19.65)				20.0	(19.1)
50	19.20	19.01			19.2	16.84
75	14.12	.02			14.1	14.42
100	13.01	.09			13.0	13.01
150	10.72	18.99			10.7	(11.3)
200	9.29	.95			9.3	(9.9)

Remarks: a Estimated from wave and barometric pressure data
b Temperatures in parentheses are interpolated
c Thermocline depth used for computation

The possible effects of barometric pressure and piling-up action on the depth of the thermocline in this offshore area have not been considered, partly because the effects are masked by the diurnal and long-period fluctuations of the thermocline depth, and partly because no proper data were available.

In addition to the above-described hindcasts for one week, three tests of hindcasts for 13 to 16 day periods were made for the same location. (The available data were for 13, 14 and 16 day periods). The meteorological elements were averaged for the periods (Table 14) and only one heat exchange computation was made for each period (Table 15).

Table 14. *Heat exchange at 39°N, 153°E in the North Pacific* (13 to 16 days average meteorological data)

No	Dates	T_{db}	T_{wb}	T_{db}-T_{wb}	U_o	T_w	T_w-T_a	e_w	e_a	e_w-e_a	V	C	t_d	A_n
1	22. II to 10. III 50	4.9	3.2	1.7	75	9.3	4.4	11.5	6.5	5.0	11.8	8	684	44
2	30. VI to 14. VII 50	21.2	20.25	0.9	90	20.8	−0.4	24.0	22.6	1.4	5.7	7.5	888	74
3	17. X to 30. X 50	15.0	10.9	4.1	60	18.4	3.4	20.7	10.2	10.5	8.3	8.6	648	39

Remarks: 1) Notations and units used, see Chapter 4. (Units: all T in °C, U_0 in %, all e in mb, V in m sec^{-1}, C in tenth of sky, t_d in min and A_n in °.)
2) The meteorological data presented are averaged from eight observations in 24 hours throughout the given periods.
3) The height of anemometer is not specified in the original data but is assumed to be at standard height (*circ.* 8 m).

Table 15. *Heat exchange at 39°N, 153°E in the North Pacific.* (13 to 16 days heat budget computations)

No.	Q_{os}	Q_s	Q_r	Q_{ob}	Q_b	E	Q_e	Q_h	Q_l Residual heat
1	6742	4672	406	3300	1280	93.6	5522	3212	−5748
2	12880	9621	837	2400	1022	13.7	808	−96	7050
3	4600	2843	247	2660	910	122.9	1551	7250	−7115

Remarks: 1) The number in the first column refers to the corresponding entry in Table 14.
2) Residual heat (Q_l) is computed with the Formula (2):

$$Q_l = Q_s - Q_r - Q_b - Q_e - Q_h$$

3) Notations and units used, see Chapter 4. (Units: all Q in g cal cm^{-2}; E in mm cm^{-2})

Table 16 shows the estimated permanent flow and the computed resultant wind currents for the periods in question. The relatively good correspondence both in direction and speed of these currents indicates that the estimated permanent flow in this locality might be a »characteristic» current, as found in the Baltic by PALMÉN (1930 b) and HELA (1952). If this is true, the result is remarkable in that, although so close to two strong permanent currents — the Kuroshio and Oyashio — the North Pacific Drift depends only on the resultant direction and

Table 16. *Estimation of the temperature changes caused by advection at 39°N, 153°E in the North Pacific during 13 to 16 day periods*

Period (dates)	Estimated resultant wind		Estimated permanent flow [a])		Computed resultant wind current			Surface temperature gradient in direction of wind current °C/100 km	Estimated temperature change by advection °C
	Direction (from)	Speed m sec^{-1}	Direction (from)	Speed cm sec^{-1}	Direction (from)	Speed cm sec^{-1}	Distance travelled km		
22. II to 10. III 50	WNW	8	WNW	8	NW	7.6	105	−1.8	−1.9
30. VI to 14. VII 50	S	4.5	SW	7	SSW	5.7	67	+1.3	+0.9
17. X to 30. X 50	N	3	NW	4	NNW	4.7	53	−1.4	−0.7

[a]) This estimated permanent flow serves only for comparison with the computed wind current and is not used for estimation of advectional temperature changes.

force of winds and is no longer a true gradient current. The temperature changes caused by advection (Tables 16 and 17) are also computed on the basis of resultant wind currents only.

The seasonally varying advection might also explain the relatively great annual amplitude of the surface temperature in this location (see Figure 28).

The hindcasts of these approximately two-week periods are summarized in Table 17. During the period 22nd Feb. to 10th Mar. 1950, the surface temperature data show relatively great short-term variations, which might be caused by cutting off eddies from Oyashio, and the movement of the weather ship. These factors certainly affect the accuracy of the hindcast in this location. By averaging the surface temperature from the 9th to the 11th of March, the average value is 7.3°C and the difference between this and the forecasted temperature is 1.45°C.

In the period from 30th June to 14th July 1950, a relatively rapid warming-up occurs. In previous days in this period there were stronger winds than on the one used for the estimation of the depth of the thermocline and therefore some warm water from the surface had been mixed with the water in and below the shallow thermocline. If this had been taken into account a better correspondence between the hindcast and the measured temperature would have occurred.

For the period of 17th Oct. to 30th Oct. 1950 two values for the thermocline depth are given: the depth computed from wave data is given in parenthesis above the measured one, which has been determined by convective stirring.

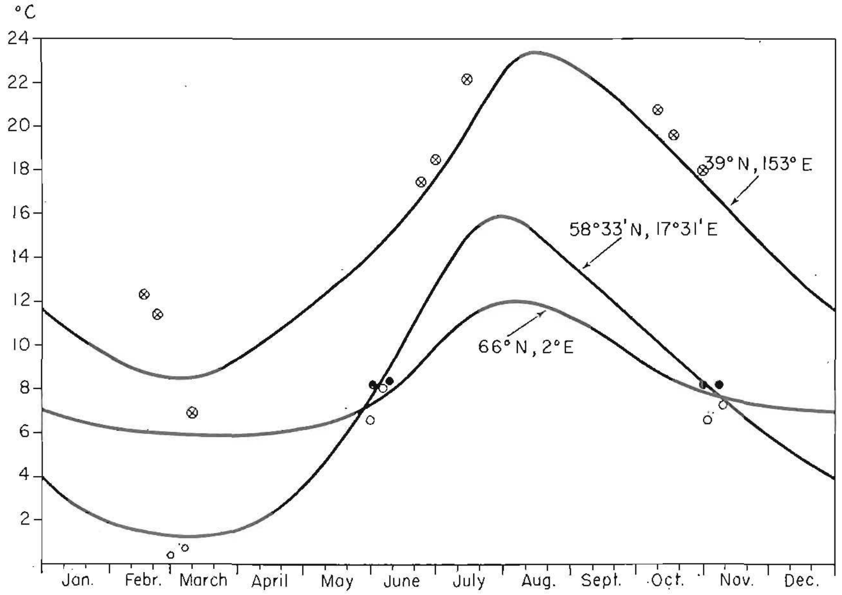

Figure 28. Annual variation of sea surface temperature at 39°N, 153°E; 58°33′N, 17°31′E and 66°N, 2°E.

21.2. NORWEGIAN SEA (66°N, 2°E)

For this area only daily BT curves (kindly supplied by the U.S. Navy Hydrographic Office) and daily synoptic weather maps (kindly supplied by Seewetteramt, Hamburg) were available. No permanent current is estimated for this point as the appropriate data were unavailable and also because the point is in the area of a great slow eddy in the Norwegian Sea. The position is in an area of a relatively great E—W surface temperature gradient. The surface temperature gradient is estimated from charts given by Schott (1942).

As seen from Table 22, a colder water mass from the East Greenland current moved into the area between the 5th and 7th of November, 1957, and caused relatively great differences between the hindcast and the actual temperatures. This change might also be caused by a slight change of the position of the lightship.

Table 17. *13 to 16 days temperature hindcasts at 39°N, 153°E in the North Pacific*

Time period (dates)	22. II to 10. III 50.	30. VI to 14. VII 50.	17. X to 30. X 50.
Initial average temperature of surface layers (°C)	11.5	18.5	19.7
Initial depth of thermocline (m)	100	24	38
Estimated average depth of thermocline (m) (Wind speeds, last 24 h: 13, 7 and 8.5 m respectively)	68	20	(32)* 60**
Total residual heat gained or lost (g cal cm^{-2} for the period)	−5748	7050	−7115
Temperature change caused by heat exchange (°C)	−0.85	+3.5	(−2.2)* −1.2**
Temperature change by advection (°C)	−1.9	+0.9	−0.7
Estimated final temperature (°C)	8.75	22.9	(16.8)* 17.8**
Measured final average temperature of surface layers (°C)	6.9	22.2	18.0
Measured final depth of thermocline (m)	60	10	60
Difference between measured and hindcast temperatures (°C)	−1.85	−0.7	(+1.2)* +0.2**
Total temperature change during the period (°C)	−4.65	+3.7	−1.7

Remarks: * Values calculated with estimated depth of thermocline.
** Values calculated with measured depth of thermocline.

21.3. BALTIC SEA (58°33′N, 17°31′E)

As seen from Tables 24 and 26, there is a relatively great difference between the hindcast and the actual temperatures for this area in the hindcast period in March. The heat loss by evaporation is also great in March, and it is possible that the estimate of relative humidity for this period is incorrect and causes the difference mentioned above.

In Table 25, the permanent flow has been computed from the wind and current observations made on the lightship »Hävringe» (Anon., 1958) by graphical subtraction of the computed wind currents of the resultant winds from the measured resultant currents. For the months of March and June, the current direction is from NE, as expected for this location. The current direction from WSW for November was, however, unexpected. It can, however, be the case, considering the prevailing direction

Table 18. *Heat exchange at 66°N, 2°E in the Norwegian Sea*
(Meteorological data)

No.	Date	T_{db}	U_o	T_w	T_w-T_a	e_w	e_a	e_w-e_a	V	C	t_d	A_n
June 1957												
1	2−5	8	80	8.2	0.2	10.7	8.6	2.1	8	7	1290	46.5
2	6−8	9	75	8.2	−0.7	10.7	8.6	2.1	10	5	1315	46.5
November 1957												
3	1−4	7.5	95	8.2	0.7	10.7	9.9	0.8	10	7	435	9
4	5−7	5	85	8.0	3.0	10.5	8.4	2.1	8	10	480	8

Remarks: 1) Notations and units used, see Chapter 4. (Units: all T in °C, U_o in %, all e in mb, V in m sec^{-1}, C in tenth of sky, t_d in min and A_n in °.)

2) The meteorological data presented are estimated from synoptic weather maps and grouped by 2 to 5 day periods. Wind speed converted from gradient wind to surface wind at 8 m level.

Table 19. *Heat exchange at 66°N, 2°E in the Norwegian Sea*
(Heat budget computation)

No.	Q_{os}	Q_s	Q_r	Q_{ob}	Q_b	E	Q_e	Q_h	Q_l Residual heat
June 1957									
1	2520	2000	168	612	284	5.55	330	207	1011
2	2568	2375	168	627	387	6.48	384	−63	1499
	Total	4375	336		671		714	144	2510
November 1957									
3	165	131	24	570	264	2.46	147	84	− 388
4	150	60	21	597	140	5.55	330	309	− 740
	Total	191	45		404		477	393	−1128

Remarks: 1) The number in the first column refers to the corresponding entry in Table 18

2) Residual heat (Q_l) is computed with the Formula (2):

$$Q_l = Q_s + Q_p - Q_r - Q_b - Q_e - Q_h$$

3) Notations and units used, see Chapter 4. (Units: all Q in g cal cm^{-2}; E in mm cm^{-2})

Table 20. *Hindcast of waves, wind currents and temperature changes caused by advection at 66°N, 2°E in the Norwegian Sea*

Date	Wind velocity m sec^{-1}	Wind direction (from)	Length of fetch km	Duration of wind h	Computed			Surface temp. gradient in direction of current °C/100 km	Distance travelled by current km	Temp. change by advection °C
					Wave height m	Current				
						Direction (from)	Velocity cm sec^{-1}			
1	2	3	4	5	6	7	8	9	10	11
June 1957										
2–5	8	W	500	24	2.2	WNW	7.6	−1.0	19.7	−0.20
6–8	10	N		72	3.6	NNE	8.5	−0.8	22.0	−0.18
										−0.38
November 1957										
1–4	7	NW	500	24	1.7	NNW	7.2	+0.6	18.7	+0.11
5–7	10	SE		24	3.5	SSE	8.5	−0.6	22.0	−0.13
										−0.02

of the winds during this season in the southern and central Baltic and the rise in the sea level in the same season. Winds in this season can also cause a large local eddy in the location under consideration. The estimated average velocities are rather approximate, considering the low accuracy in estimating wind force and current speeds on the lightships. Further, only three-year observations were used in the present computations of the permanent flow. Because of great local variations, it is difficult to estimate the horizontal temperature gradient in coastal areas if detailed data from actual surveys in the locality are not available, and consequently the value of the surface temperature gradient in the direction of the current (Table 25, column 9) is rather approximate (estimated from the data given by Bullig and Bintig 1954).

Considering the relative variability of the permanent flow and the relatively rapid changes of temperature in shallow coastal areas, the advective effect has been computed from the currents resulting from the permanent flow and the predicted wind currents.

The possible effect of atmospheric pressure and piling-up action on the depth of the thermocline for the March example is smaller than 50 cm., but for the June example this effect is estimated at *circ.* 1.7 metres.

Table 21. *Water temperature changes at 66°N, 2°E, in the Norwegian Sea from 2 to 8 June 1957*

Heat received: 4375 g cal cm^{-2}; heat lost: 1865 g cal cm^{-2}, $Q_l = 2510$ g cal cm^{-2}
Optical water mass 3
Depth of the thermocline: 2. VI 57: (30) m[a]; estimated for 8. VI 57: 45 m[b]; measured: (55) m[a]

Measured 2. VI 1957.		Forecast 8. VI 1957.			Measured 8. VI 1957.
Depth m	Temp. °C	Temp. correction for absorption °C	Temp. correction for losses °C	Temp. corrected for Q_l and advection °C	Temp. °C
1	2	3	4	5	6
0.5	8.22	+0.97 (45)[c] (0.5–40 m)	−0.41 (0.5–40 m)	8.4	8.39
10	8.22				8.33
20	8.22				8.33
30	8.22				8.33
40				8.4	
50	8.06			8.3	8.39
75	7.89			7.9	7.78
100	7.78			7.8	7.83
150	7.61			7.6	7.61
200	7.39			7.4	7.50

Remarks: a Thermocline not well developed
b Estimated from wave data
c Depth of the thermocline used for computation

22. FACTORS AFFECTING THE ACCURACY OF FORECASTS OF THE TEMPERATURE IN THE SEA

Forecasts of conditions in the sea are in this respect similar to weather forecasts in that success depends greatly on the experience of the forecaster and on his knowledge of specific local conditions. The main difference between these two forecasts is that in hydropsis several conditions can be computed with empirical and/or theoretical formulas, using meteorological parameters.

The accuracy and possible errors of heat budget computations are

Table 22. *Water temperature changes at 66°N, 2° in the Norwegian Sea from 1 to 7 November 1957*

Heat received: 191 g cal cm^{-2}, heat lost: 1319 g cal cm^{-2}, $Q_l = -1128$ g cal cm^{-2}
Optical water mass 2
Depth of the thermocline: 1. XI 57: 120 m; estimated for 7. XI 57: 44 m[a]; 190 m[b]; measured: (85) m[c].

Measured 1. XI 1957.		Forecast 7. XI 1957.			Measured 5. XI 1957.	Measured 7. XI 1957.
Depth m	Temp °C	Temp. correction for absorption °C	Temp. correction for losses °C	Temp. corrected for Q_l and advection °C	Temp. °C	Temp. °C
1	2	3	4	5	6	7
0.5	8.22			7.9	8.22	7.61
2.5		+0.60				
5		+0.07				
10	8.22	+0.03			8.28	7.67
20	8.22	+0.01	−0.07		8.28	7.67
30	8.22				8.33	7.67
40	8.22				8.33	7.67
50	8.22				8.33	7.67
75	8.22				8.33	7.67
100	8.22				8.44	6.67
150	7.94		(190)[d]	7.9	8.28	5.00
200	7.39			7.65	7.67	3.39

Remarks: a Estimated from wave data
b Estimated with consideration of convective stirring
c Thermocline not well developed
d Depth of the thermocline used for computation

discussed in Part I. It cannot be too often emphasized that successful hydropsis also depends greatly on successful weather forecasts and/or on wise selection of meteorological parameters from existing weather forecasts and from climatological data. On the other hand, the application of exact knowledge of the interaction between the sea and the atmosphere will allow an improvement of accuracy in daily, and especially in long-range weather forecasts.

The differences between the hindcast and the measured temperatures are summarized in Table 29. The possible errors are briefly discussed in Chapter 21.

A simplified approach to forecasting water temperature changes for a week can be made if a measurement is available. Arbitrarily it is assumed that the difference between the actual measured temperature and the annual mean temperature for the period in question will be equalized in four weeks. The temperature trend $\frac{dT}{dt}$ is taken from the annual curve. The simple formula for this prediction would be:

$$T_w = T_{ow} + \frac{dT}{dt} + \frac{(T_t - T_{ow})t_e}{28} \tag{60}$$

The annual curves for the three localities in question are given in Figure 28. The few data do not allow judgment about this approximate method. However, this approach is not, either on physical or logical basis, comparable to the detailed approach used earlier throughout this

Table 23. *Heat exchange at 58°33′N, 17°31′E in the Baltic Sea* (Meteorological data)

No.	Dates	T_{db}	U_o	T_w	T_w-T_a	e_w	e_a	e_w-e_a	V	C	P	t_d	A_n
March 1957													
1	1–3	0	90	0.4	0.4	6.1	5.5	0.6	5	10	6*	655	25
2	4–7	–2.0	80	0.5	2.5	6.2	4.1	2.1	8	6	—	675	26
June 1957													
1	1–2	12	80	6.5	5.5	9.6	11.2	–1.7	5	5	—	1065	54
2	3–7	10	75	7.0	3.0	10.0	9.2	0.8	5	3	—	1085	54.5
November 1957													
1	3–6	8	95	6.6	–1.4	9.7	10.2	–0.5	8	8	—	525	16.5
2	7–9	7	80	6.5	–0.5	9.6	10.0	–0.4	10	10	—	505	15.5

Remarks: 1) Notations and units used, see Chapter 4. (Units: all T in°C in %, all e in mb, V in m sec^{-1}, C in tenth of sky, t_d in min and A_n in °.)

2) The meteorological data presented are estimated from synoptic weather maps and grouped by 2 to 5 day periods. Wind speed converted from gradient wind to surface wind at 8 m level.

3) * Snow, equivalent to 6 mm of rain.

Table 24. *Heat exchange at 58°33'N, 17°31'E in the Baltic Sea*
(Heat budget computation)

No.	Q_{os}	Q_s	Q_r	Q_{ob}	Q_b	E	Q_e	Q_h	Q_p	Q_l Residual heat
March 1957										
1	485	183	58	418	98	0.78	46	20	−48	−87
2	983	855	120	868	470	7.39	440	343		−518
	Total	1038	178		568		486	363	−48	−605
June 1957										
3	805	745	56	207	128	−0.65	−39	140		460
4	4140	4070	280	1050	809	2.60	154	381		2446
	Total	4815	336		937		115	521		2906
November 1957										
5	363	252	48	579	225	−0.92	−55	−101		135
6	330	132	27	621	146	−0.92	−55	−45		59
	Total	384	75		371		−110	−146		194

Remarks: 1) The number in the first column refers to the corresponding entry in Table 23

2) Residual heat (Q_1) is computed with the Formula (2):

$$Q_l = Q_s + Q_p - Q_r - Q_b - Q_e - Q_h$$

3) Notations and units used, see Chapter 4. (Units: all Q in g cal cm^{-2}; E in mm cm^{-2})

study. The total temperature changes during the hindcast periods are shown in the last column of Table 29. These changes are relatively small. The relatively few available data do not allow statistical treatment. However, it can be mentioned that the average difference between the predicted and the measured surface temperatures, using the heat budget method, is *circ.* 36 per cent of the average temperature change during the 11 periods.

The estimation of wave height depends not only on wind speed, the main factor, but also on the accurate estimation of change of wind direction and of the winds in the vicinity. As the waves are subject to separate forecast, and several formulas and discussions of their accuracy are available, no further comments are given here.

Table 25. *Hindcast of waves, wind currents and temperature changes caused by advection at 58°33'N, 17°31'E in the Baltic Sea*

Date	Wind velocity m sec^{-1}	Wind direction (from)	Length of fetch km	Duration of wind h	Computed			Surface temp. gradient in direction of current °C/100 km	Distance travelled by current km	Temp. change by advection °C
					Wave height m	Current Direction (from)	Current Velocity cm sec^{-1}			
1	2	3	4	5	6	7	8	9	10	11
March 1957										
1–3	5	S	150	24	0.8	S	3.0	+0.2	5.2	+0.10
4–7	8	NW	25	48	0.7	N	9.4	−0.2	32.5	−0.65
										−0.55
June 1957										
1–2	5	NW	50	24	0.7	N	7.3	−0.05	6.3	−0.03
3–7	5	SW	100	24	0.8	WSW	3.7	+0.08	16.0	+0.13
										+0.10
November 1957										
3–6	8	S	150	24	1.8	SSW	8.8	+0.1	22.8	+0.23
7–9	10	SE	150	48	2.8	SSE	8.5	+0.1	22.1	+0.22
										+0.45

Remarks: a Speed and direction of the current corrected for the effects of permanent flow, which is estimated as follows:

Month	Direction	Velocity cm sec^{-1}
March	NE	3.5
June	NE	2.5
November	WSW	1.5

There has up to the present been very little investigation of the response of the thermocline to the waves. Table 30 presents the measured and estimated depths of the thermocline, from the examples given in the previous Chapter. The differences are to a great extent caused by the periodic fluctuations of the thermocline, discussed in Part II. These

Table 26. *Water temperature changes at 58°33′N, 17°31′E in the Baltic Sea from 1 to 7 March 1957*

Heat received: 1 038 g cal cm^{-2}, heat lost: 1 642 g cal cm^{-2}, $Q_l = -$ 605 g cal cm^{-2}
Optical water mass 4. Depth of water (60) m
Depth of the thermocline: 1. III 57: 20 m; estimated for 7. III 57: 10 m[a]; measured (15)[c] m.

Measured 1. III 1957.			Forecast 7. III 1957.			Measured 7. III 1957.
Depth m	Temp. °C	S ‰	Temp. correction for absorption °C	Temp. correction for losses °C	Temp. corrected for Q_l and advection °C	Temp. °C
1	2	3	4	5	6	7
0.5	0.4	6.93	+1.04		0	0.7
5	0.6	6.97		−1.64		0.7
10	0.6	7.01	(10)[b]		0	0.7
20	0.7	7.08	+0.01		0.5	0.9
30	1.2	7.09			0.8	1.0
40	1.2	7.09			0.8	1.0

Remarks: a Estimated from wave data
b Depth of the thermocline used for computations
c Thermocline not well developed

fluctuations also cause some mixing in the thermocline. A comparison of the measured data with the seasonal charts prepared by LUMBY (1955) indicates that the present method of estimation provides a much better and more accurate approach than the approach used by LUMBY.

The majority of the errors in the prediction of temperature changes are caused by inaccuracy in estimating the heat transport by currents. Very little data are available in such form as to allow separation of the permanent flow from the wind currents without time-consuming computations. It can be expected that in many areas the permanent flow will be a »characteristic current», which is a function of the prevailing winds. In addition, very little information is available on the variation of the wind current speed and direction with depth. EKMAN's theory can, however, on the basis of the few available measurements, be considered as invalid. There are also opposing permanent currents below the thermo-

Table 27. *Water temperature changes at 58°33'N, 17°31'E in the Baltic Sea from 1 to 7 June 1957*

Heat received: 4 815 g cal cm^{-2}, heat lost: 1 909 g cal cm^{-2}, Q_l = 2 906 g cal cm^{-2}. Optical water mass 4. Depth of water (60) m.

Depth of the thermocline: 1. VI 57: 8 m; estimated for 7. VI 57: 10 m[a]; 11.7 m[b] measured 12.5 m.

Measured 1. VI 1957.			Forecast 7. VI 1957.			Measured 7. VI 1957.
Depth m	Temp. °C	S ‰	Temp. correction for absorption °C	Temp. correction for losses °C	Temp. corrected for Q_l and advection °C	Temp. °C
1	2	3	4	5	6	7
0.5	6.6	6.41			9.2	8.1
5	6.6	6.43	+4.1	−1.6		8.1
10	6.3	6.42	(11.7)[c]		9.2	8.1
20	5.3	6.45	+0.04		5.4	4.8
30	4.6	6.55	+0.01		4.7	4.6
40	4.0	6.61			4.1	4.1

Remarks: a Estimated from wave data
b Estimated with consideration of changes in barometric pressure and piling-up action of wind
c Depth of the thermocline used for computations

cline, as yet unexplored. Furthermore, the deflection of the direction of the wind current from the wind direction is uncertain, because in earlier attempts to determine this deflection, separation of neither the permanent flow nor the gradient current component, caused by changes of atmospheric pressure, has been made. The mechanism of »eddying» and cutting off »water pockets» with different temperatures can most probably be ascribed to the action of the wind and variations of its direction and to the current component caused by the change of atmospheric pressure and transport by waves. If the dispersal and modification with time and distance of these eddies could be followed, some quantitative idea could be gained about horizontal mixing due to these eddies.

Finally, also to be considered are numerous specific local conditions in coastal areas; these local peculiarities must be ascertained through intensive surveys.

Table 28. *Water temperature changes at 58°33'N, 17°31'E in the Baltic Sea from 3 to 9 November 1957*

Heat received: 384 g cal cm^{-2}, heat lost: 190 g cal cm^{-2}, Q_1 = 194 g cal cm^{-2}
Optical water mass 4. Depth of water (60) m
Depth of the thermocline: 3. XI 57: 20 m; estimated for 9. XI 57: 35 m[a]; measured 23 m.

Measured 3. XI 1957.			Forecast 9. XI 1957.			Measured 9. XI 1957.
Depth m	Temp. °C	S $^0/_{00}$	Temp. correction for absorption °C	Temp. correction for losses °C	Temp. corrected for Q_l and advection °C	Temp. °C
1	2	3	4	5	6	7
0.5	6.6	7.08	+0.11 (35)[b] (0.5–30 m)	−0.05 (0.5–30 m)	7.1	7.3
5	6.6	7.08				7.3
10	6.6	7.08				7.3
20	6.5	7.08				7.3
30	5.0	7.39			7.1	5.6
40	3.4	8.00			4.0	3.4

Remarks: a Estimated from wave data
b Depth of the thermocline used for computations

23. AUTOEVALUATION AND NOTES FOR FUTURE INVESTIGATIONS NEEDED ON THE RESPONSE OF THE SEA TO ATMOSPHERIC CHANGES

A voluminous amount of literature has been reviewed (only part of which is shown in the bibliography) in an attempt to find theories, formulas and data applicable to the problems presented. Surprisingly few of the existing theories and formulas stand testing and are applicable in the actual conditions met with in nature.

A complete outline of the forecasting of various factors in the sea, with special emphasis on the forecasting of temperature, has been presented in all its complexity.

A classification of water masses has been made, using a few important and easily measurable properties. Considering the great local and seasonal

Table 29. *Comparison of accuracy of surface temperature hindcasts by the heat budget method and by the use of Formula (60)*

Position	Dates	Measured temperature minus hindcast temperature		Total temperature change during the period
		Heat budget method	With Formula (60)	
39°N, 153°E	16. II to 22. II 1950	−0.2	−0.1	−0.9
	23. VI to 30. VI 1950	−0.3	+0.2	+1.0
	10. X to 17. X 1950	−0.4	0	−1.1
	22. II to 10. III 1950	−1.85	−2.7	−4.6
	30. VI to 14. VII 1950	−0.7	+2.0	+3.7
	17. X to 30. X 1950	(+1.2)* +0.2	−0.2	−1.7
58°33′N, 17°31′E	1. III to 7. III 1957	+0.7	+0.2	+0.3
	1. VI to 7. VI 1957	−1.1	+0.2	+1.5
	3. XI to 9. XI 1957	+0.2	+1.0	+0.7
66°N, 2°E	2. VI to 8. VI 1957	−0.1	−0.2	+0.2
	1. XI to 7. XI 1957	−0.2	−0.3	−0.6

* Calculated with two different depths of the thermocline (see Table 17).

variations, this classification is sufficient for most practical and scientific problems.

In general, the formulas and procedures for computing heat exchange between the sea and the atmosphere are as well-established as present

Table 30. *Comparison between predicted and measured depths of the thermocline*

Position	Time period (dates)	Depth of thermocline (m)	
		Predicted	Measured
39°N, 153°E	16. II to 22. II 1950	135	100
	23. VI to 30. VI 1950	29	24
	10. X to 17. X 1950	44	38
	22. II to 10. III 1950	68	60
	30. VI to 14. VII 1950	20	10
	17. X to 30. X 1950	32	60
58°33′N, 17°31′E	1. III to 7. III 1957	10	(15)*
	1. VI to 7. VI 1957	11.7	12.5
	3. XI to 9. XI 1957	35	23
66°N, 2°E	2. VI to 8. VI 1957	45	(55)*
	1. XI to 7. XI 1957	190	(85)*

* Thermocline not well developed.

data allow and as is necessary, considering the short-period variations in meteorological conditions.

The approach to forecasting the insolation per day is a considerable improvement on the existing practice and the new formula for correcting the effect of cloudiness on insolation gives values far closer to the real conditions than earlier formulas. Although some quantitative data are available on the influence of different types of clouds on the insolation, the use of these relations is impracticable because of the great variability of clouds in short-term intervals and over short distances and because of the frequent lack of detailed data on the types and heights of clouds.

The improved quantitative relations between cloudiness and insolation make it possible to measure the average cloudiness instrumentally and also to predict low visibility from the measurements.

The existing disorder in the determination of reflected radiation has been partly sorted out. The new relation of albedo to insolation presented has been found adequate for the computation of daily albedo. A detailed paper is being prepared on this subject (Olson, MS).

For computing effective back radiation, the Lönnquist formula with additions from Möller has been adopted, which is considerably more accurate than the other formulas at present more commonly in use.

The revised evaporation formula of ROHWER has been found more accurate than others. In this connection, a graph is prepared for the conversion of the average wind speeds at various levels. Although there are slight variations in the speed relations at various levels in various stability conditions, it is often not possible to evaluate these differences, and a general relation is adequate for most purposes.

A new formula has been derived by physical reasoning for the transfer of sensible heat. As this formula is nearly identical to that derived from BOWEN's ratio and as the latter is considered adequately accurate, there is this indirect proof of the validity of the new formula.

The possibilities for estimation of the changes in the air masses moving over the ocean are illustrated by a small example. It is outside the scope of this paper to carry investigation on this subject further.

For convective transfer of heat to the sea and for the estimation of the condensation of vapour on the sea surface, formulas have been derived by physical reasoning. It is at present impossible to test the accuracy of these formulas.

Table 6 has been computed for the estimation of the absorption of insolation in different layers of different water masses. This table is useful for estimating temperature changes by a continuous density model and by a shallow thermocline.

There exists no general solution for the quantitative consideration of mixing in the sea. The problem of mixing is divided into two parts: mixing by wave-action and by convective stirring. A preliminary approach is given for the estimation of the depth which convective stirring can reach. The time unit for this consideration is still somewhat uncertain. Although, for practical purposes, a 24-hour period has been used in the present paper, it is more probable that only night conditions should be used in detailed study. For investigation of the problems of convective stirring, heat-exchange computations and BT measurements should be made at *circ.* 15-minute intervals daily in the autumn, in a location where there is no, or very slow, current.

Although notes were found in the Russian literature to the effect that wave height was used for the estimation of thermocline depth, no quantitative data or formulas were available. The present quantitative attempt is only a possible one, considering the available data, but presents, however, a great improvement over the existing procedures, in which wind data and latitudinal and seasonal effects were considered. The relatively great variations in depth of the thermocline caused by internal waves are pointed out. It would be possible to ascertain by actual measurement the range of these fluctuations in given locations and seasons. In areas where internal tides occur, frequent BT measurements (at 15-minute intervals) should be made over several days when the current speeds are low. A tide gauge station must be located near-shore, close to the site of these measurements, so that corrections for surface tides can be made. Attention should also be given to the possible relations between the atmospheric disturbances and their movements and the internal waves.

The simplified combined formula for wave forecast presents a slight improvement on the existing approaches. However, more data are necessary for improving the factors of wind duration and fetch in this formula. An investigation by the author on the influence of water temperature, salinity and the temperature differences between the sea surface and the air on the wave height is in progress. Field observations are also required on the coupling of waves from previous and present winds and on the decay of waves. Further field observations are also necessary for the perfection of the given relation between the wave height and the depth of the thermocline.

The accurate assessment of the heat transport by currents is probably one of the most difficult problems. Although methods are available for the separation of wind currents and permanent flow (PALMÉN, 1930 b; HELA, 1952), available climatological data are usually presented in such a form that to accomplish the task of computation one has to use huge amounts of original observations.

The existing relations between wind and current speeds have been critically reviewed. However, it has been shown that EKMAN's theory is not valid in detail, and there is very little actual data available on the change of current speed with depth. The question of the deflection of current direction has also not been finally solved, although a practical, more accurate deflection is suggested in this paper. Unfortunately, most workers in this field have not separated wind-current and permanent flow. Few observations are applicable for solving the problem (HELA, 1952, is one source). Further specific data on currents and winds are also urgently needed for improving forecasts of the movement of current eddies. Frequent current measurements should be conducted on lightships during and after the change of wind direction. Current measurements should be made at small depth-intervals close to the surface. Simple recording cumulative wind measuring apparatus should be used. Both cumulative direction and force should be recorded.

The effect of the changes of barometric pressure on the direction and speed of the surface currents has been very little considered by previous workers. It is felt by the present author that this effect must be considered in the prediction of surface currents and that it might also explain the often observed short-term fluctuations of speed and direction in the surface currents. The effects of gradient currents, caused by changing atmospheric pressure, on the resultant surface currents are discussed qualitatively, and a suggestion is made for the preparation of prognostic current charts which will make possible the empirical investigation of this current component. The »particle transport velocity» by waves also requires further laboratory and field investigations.

Mixing by currents can at present not be attacked quantitively. The following general procedure could be suggested for the investigation of mixing by currents. A series of parachute drogues should be set in one locality, with the parachutes at different depths. Their movement should be followed for several days and frequent (three-hourly) sampling and/or measurements of conservative properties should be made at the depth of the parachutes and above and below. The mixing, computed from these measurements, should be expressed as part per unit volume of the original water mass replaced by other water either per unit time or per unit distance of movement. This quantity should be related to: current velocity and its horizontal and/or vertical gradient, stability, and the gradient of the conservative property under consideration. Attention when planning these experiments should be given to the possibilities of avoiding such factors as mixing by waves and the influence of bottom configuration; for this, special local investigations should be conducted.

For the estimation of the movement of convergencies and divergencies only rules-of-thumb can be given at present, and no direct calculations are possible. However, detailed investigations on the movement of current boundaries, with simultaneous detailed analyses of wind situation, would enable the improvement and/or testing of the present rules-of-thumb.

There is nothing new to add on the excellent work of PALMÉN and LISITZIN on the estimation of sea-level changes. Special difficulties are, however, met with in offshore areas where we lack data on this subject. Therefore frequent BT measurements (at 15-minute intervals) should be made in offshore areas during winds, when piling-up or lowering of the sea level is expected in the given area. A mareograph station must be close to the site of these investigations. The current structure within the depth should also be ascertained at frequent intervals.

The hindcast tests in Part III mainly illustrate the application of and show the difficulties with the present data. The relative importance of different components and their local and seasonal changes can easily be seen from the tables. The necessity of good knowledge of local conditions is underlined.

The creation of procedures for weather forecasting was not the work of one man but of hundreds, during decades. The present paper cannot, therefore, be expected to establish complete hydroptic forecasting methods; it could be considered as a modest contribution to the new branch of oceanography.

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