THE OCEANIC HEAT BUDGET NEAR BERMUDA

by

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S. B., University of Chicago
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The Oceanic Heat Budget Near Bermuda

Bruce Mitchell Gordon

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ABSTRACT

The net heat flux through the surface of the sea near Bermuda is calculated from meteorological data and compared with the observed change in heat storage in the surface layers of the sea determined from the semi-monthly series of Panulirus hydrographic stations (1954-1967). The mean calculated flux of 35 kg cal/cm^2/season cannot account for the observed seasonal change of 55 kg cal/cm^2, yielding a seasonal amplitude of the divergence of the heat transport of 350 g cal/cm^2/day an order of magnitude greater than the -25 g cal/cm^2/day mean. The greatest amplitude occurs at the June-July and December-January extremes of stability. These results are typical of comparison between seasonal heat balance calculation and observation. A qualitative relation between calculated and observed seasonal heat changes is found on adjusting the methods of estimation of the components of the heat exchange to local conditions, but accurate quantitative results require further basic study of radiative and turbulent exchange of energy between sea and atmosphere, under the full range of maritime conditions.

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Introduction

This paper is an examination of the energy exchange between ocean and atmosphere at one point in the region of the Atlantic known as the Sargasso Sea, the center of the current system associated with the North Atlantic subtropical gyre. The 'sea' is bounded by the Gulf Stream System, the North Atlantic Drift and (roughly) the Mid-Atlantic Ridge. The ocean-atmosphere system of the subtropical gyre is a natural physical unit. However, attention is usually concentrated on the clearly defined boundary current systems — the interior, lacking persistent circulation, has not received the same attention.

A unique opportunity for study of this area exists in the series of hydrographic stations occupied by the R. V. Panulirus of the Bermuda Biological Station for Research for the Woods Hole Oceanographic Institution. These stations have been occupied to about 2600 meters depth in the vicinity of 32°10' N 64°30' W southeast of Bermuda approximately twice per month since June 1954. This series provides a long record of variations in the heat storage of the surface layers of the ocean. Meteorological data are regularly collected at Kindley AFB (32°21'N 64°38'W), St. David's Island, Bermuda and Argus Island Texas Tower (31°57'N 65°10'W) Plantagenet Bank (about 55 meters depth) permitting calculation of the several components of the ocean-atmosphere heat exchange. These results are compared with the observed variations in heat storage in an effort both to evaluate the usefulness of these methods and to gain insight into the physical processes of energy exchange peculiar to this region.

As background for detailed discussion, some general properties of the region will be summarized. In the mean this is a region of atmospheric subsidence governed by the Bermuda-Azores high pressure system, a region of clear skies, light winds and near-neutral stability in the first tens of meters above the surface of the sea. These meteorological conditions, however, undergo seasonal and other, less regular variations. The Bermuda-Azores high circulation is frequently disturbed by frontal systems from the North American continent, particularly in winter. These synoptic scale disturbances carry important anomalies of atmospheric property. The general clockwise circulation in the atmosphere induces the subtropical convergence of the surface waters of the ocean. These warm saline surface waters accumulate in a lens to depths as great as a thousand meters. The water mass above the main thermocline is known as Atlantic Central Water with temperatures 8-19° C, salinity 35.10-36.70 ‰, below the depth of surface influence. That below is North Atlantic Deep Water with temperature 2.2-3.5° C, salinity 34.97-34.90 ‰. This discussion will be limited to the waters above the main thermocline.

The water mass above the main thermocline may be conveniently divided into two regions:

(1) that immediately governed by surface conditions with seasonally varying properties. The lower boundary of this region for most of the year is the seasonal thermocline of approximate depth thirty to one hundred fifty meters, as determined here by the depth of the maximum temperature gradient, depending on the season.

(2) that relatively homogeneous water mass of about 17-19° C and 36.5 ‰ salinity lying between the seasonal and main thermoclines from about 150 to 450 meters depth. This water mass is clearly marked by an inflection in the temperature and salinity profiles and maximum in the oxygen profile at about 300 meters depth. The stability and distribution of this 'eighteen degree' water have been discussed by Worthington (1959), Schroeder and Stommel (1959), and Istoshin (1961). North of Bermuda (32°N) water of these properties is observed at the surface in the winter season in a well-mixed layer extending to depths of several hundred meters. South of 32°N this water mass is not observed at the surface at any season, but the thickness and

oxygen content are observed to change seasonally. Regular winter renewal of this water mass at the surface of the sea is implied, but the stability of property observed suggests that effective seasonal variations are limited in depth. This 'eighteen degree' water is thus a 'basement' to the surface water with importance for the seasonal cycle of surface water properties.

Little is known of oceanic advection about Bermuda — neither the persistent patterns nor the high velocities of the boundary regions of the Gulf Stream system are observed in the interior. The difference between the observed march of local heat storage and the residual component of the heat balance is usually interpreted as the divergence of the local advective transport. Neglecting advection, the observed change in heat storage may be compared to the calculated net heat flux through the surface of the sea. Thus, the heat budget results are compared with oceanographic observation to check the validity of the simplifying assumptions used to approximate the components of the heat budget. In the absence of data on advection, correction must be tentative, but the variations of the anomaly may suggest the nature of the physical processes influencing the air-sea energy transfer.

Thus, two goals are set:

(1) to detail mean and variance of the several components of the oceanic heat budget characteristic of the interior of subtropical gyre circulations.

(2) by comparison with independent oceanographic data to investigate the validity of the simplifying assumptions commonly used in deriving approximate expressions for the components of the heat budget.

Chapter I

Nature of the Data

Panulirus Hydrographic Stations

The data collected by the R. V. Panulirus used in this paper consists of two hundred sixty-three stations taken from 6-7-54 to 6-27-67, with the exhibited distribution by months, at local times varying from about 10 A. M. to 4 P. M. but usually from 2-3 P.M. from no to three times per month, usually twice. Normally collected are temperature, salinity, oxygen, and phosphate at depths of approximately 1,10,25,50,100,150,200,250,300,350,400,450, 500, 600,700,800,900,1000,1200,1400,1600,1800,2000,2500 meters with some simple weather observations. Some of the original stations and sample depths of particular stations have been rejected due to inaccurate data or maximum depth less than 500 meters. Data is recorded as depth to the nearest meter, temperature to the nearest tenth degree centigrade, salinity to the nearest hundredth part per thousand, oxygen to the nearest hundredth milliliter per liter.

Schroeder and Stommel (1969) have determined that the Panulirus hydrographic time series is representative of monthly mean conditions off Bermuda. Steric sea levels determined from the series are compared with tide gage records at Bermuda. Strong correlations are observed both with mean (0.965) and departures from the mean (0.81) for the years 1954-1961.

Kindley A.F.B. Meteorological Data

The data collected at Kindley Air Force Base, Bermuda is supplied on magnetic tape by the National Weather Records Center, Asheville, North Carolina for the years 1949-1967, as described in Tape Reference Manual TDF-14 of the Center.

The following data have been used in calculating the components of the heat exchange:

Hourly observations	-	wind speed to .1 knot
		air temperature to .1° F
		relative humidity to .1%
		total observed sky cover to .1

Three hourly observations-sea level pressure to .1 mb

Insolation at Bermuda

Daily integrated values of observed insolation collected by Eppley pyranometer on the roof of the Bermuda Biological Research Station at St.George, Bermuda from 20 March 1959 to 15 March, 1962 and recorded to 1 g cal/cm^2/day have been provided by D. Menzel (W.H.O.I.) and J. Ryther (W.H.O.I.)

TABLE 1

PANULIRUS OBSERVATIONS FREQUENCIES

	I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII
1954						2	1	1	1	0	1	1
1955	0	2	1	1	2	1	2	2	2	0	2	1
1956	1	1	1	0	0	1	1	2	1	0	1	2
1957	1	0	0	0	2	2	2	2	1	2	2	1
1958	1	2	1	1	1	3	3	3	1	2	3	2
1959	4	3	0	3	3	2	2	2	1	2	2	2
1960	2	3	2	2	2	2	2	2	1	2	2	2
1961	2	2	1	1	2	2	2	2	2	2	2	1
1962	2	2	2	2	2	2	2	2	2	2	2	1
1963	2	2	2	2	1	2	2	2	2	1	2	1
1964	2	2	2	1	2	2	2	2	2	2	2	2
1965	1	2	2	2	2	2	2	2	1	1	1	2
1966	2	2	2	2	1	2	3	2	1	2	2	2
1967	2	3	2	0	3	1						
Total	22	25	18	17	23	26	26	26	18	18	24	20

Argus Island Texas Tower Observations

Argus Island Texas Tower daily weather logs from December, 1960 to present have been supplied by Alton B. Crumpler of the Naval Oceanographic Office. Of possible interest are the following:

air temperature to .1° F
dew point to .1° F
wind speed to 1 kt
sea level pressure to 1 mb
total sky cover to .1

collected at 0800, 1100, 1400, 1700, 2000 local time. Sea surface temperatures are collected by bucket thermometer at 1100 local time and recorded to the nearest .5° F beginning 26 May, 1963 and continuing to present.

Imperfect copy and evacuation for storms make data unavailable for certain periods. The number of unusable days for each month of observation is given in the following table:

Table 2

The number of days by month and year for which Argus Island sea surface temperature observations are not available

Month:	1	2	3	4	5	6	7	8	9	10	11	12
Year												
1963						1		7	4	5		
1964									19		1	2
1965	3							1	8	7		4
1966				5	6	2	15	14	11	16	18	7
1967	20	16	31	23					6	4	1	

A comparison of Argus Island and Panulirus sea surface temperature observations shows that Argus Island observations are generally representative of conditions at the site of the Panulirus soundings. As meteorological observations are collected at Argus Island only between 0800 and 2000, a comparison of mean conditions at Argus Island and Kindley A.F.B. is not possible. Therefore, it is not certain that Kindley observations are in the mean representative of oceanic conditions.

Chapter II

Method of Calculation of the Components of the Oceanic Heat Budget

General

These calculations are performed for a fixed column of water extending from the surface of the sea to 500 meters depth. Therefore, the exchanges to be considered are:

Advection of waters of different property through the bottom and side boundaries or,

the divergence of the advective heat flux designated by - - - - - - - - - - - - - - - - - - Q_V

Transfer of heat through the surface boundary, notably:

Absorption of direct and indirect (sky) radiation, or flux of insolation designated by - Q_I

Net long wave radiation to the atmosphere from the sea surface, or flux of back radiation designated by - - - - - - - - - - - - Q_B

Turbulent transfer of sensible heat designated by - Q_S

Turbulent transfer of latent heat designated by - Q_E

Other sources of heat, as mechanical and biological generation within the column, are small compared with solar radiation. Therefore, insolation is the only source of heat to be considered:

(1) $Q_I = Q_E + Q_S + Q_B + Q_{ST} + Q_v$

where Q_{ST} is the heat stored in the column. Of these Q_v is not directly calculable from the available data so that

(2) $Q_I = Q_E + Q_S + Q_B + \text{Residual}$

where Residual = $Q_v + Q_{ST}$.

Persistent change in annual mean properties of the surface layers is assumed small: the annual mean $\overline{Q}_{ST}$ is approximately zero, and

(3) $\overline{\text{Residual}} \simeq \overline{Q}_v$.

The Panulirus data gives the actual heat stored. Thus the seasonal variation of Q_v may be given as

(4) Q_v = Residual - Q_{ST} observed.

Insolation

Global surveys of the oceanic heat budget have used various latitude dependent approximations for the estimation of insolation. A number of these are reviewed by Budyko (1958). Most usual is the form

(5) $Q_I = Q_{IO}(f,t) * F(f,c)$

where Q_{IO} is the total possible radiation (clear sky), a function of season (t), and a latitude (f) dependent atmospheric transparency. F has some polynomial, usually linear or quadratic, dependence on the observed cloudiness (c) or sunshine hours. Variations in the nature and distribution of clouds are incorporated in latitude dependent coefficients, as for example from Budyko (1958)

(6) $Q_I = Q_{IO} * [1-k(f)*\bar{c}]$, the Savino-Angstrom formula where $\bar{c}$ is the mean cloud amount in tenths and k varies with latitude as indicated:

Table 3

The variation of the coefficient in the (linear) Savino-Angstrom relation for the estimation of insolation from total sky cover

f	0	5	10	15	20	25	30	35
k	.65	.66	.66	.67	.67	.68	.68	.68

f	40	45	50	55	60	65	70	75
k	.67	.66	.64	.62	.60	.55	.50	.45

on that of Berliand (1960)

(7) $Q_I = Q_{IO} * (1-a\bar{c}-b\bar{c}^{2})$

where b = 0.38 and a = a(f)

Table 4

f	0	5	10	15	20	25	30	35	40
a	138	.40	.40	.37	.37	.35	.36	.38	.38
f	45	50	55	60	65	70	75	80	85
a	.38	.40	.41	.36	.25	.18	.16	.15	.14

The variation of the linear coefficient in the (quadratic) Berliand relation for the estimation of insolation from total sky cover

These expressions are based on actinometric observations over limited area and duration assuming longitudinal constancy and regular seasonal variation in atmospheric transparency. The particular Q_{IO} used in these calculations as given by Berliand (1960) is determined by fitting a curve to insolation observed under cloudless skies at differing latitudes. The work of Berliand is employed by Budyko (1963) in his Atlas of the Heat Balance of the Earth, a global survey of heat exchange, and by Wyrtki (1965), Day (1968) and others.

Houghton (1954) has calculated clear sky insolation from insolation at the top of the atmosphere using the data of the Smithsonian Meteorological Tables and some simple assumptions about the variation of atmospheric transparency with vapor pressure, suspended particulates, etc.

Table 5

A comparison of the calculated clear sky insolation of Houghton (1954) with the observed clear sky insolation of Berliand (1960)

	Houghton	Berliand
March 21	557	630
June 21	683	780
Sept. 21	523	605
Dec. 21	308	375

The difference seems rather too large to be explained by greater transparency on clear days. It is assumed that the observed values of Berliand are more accurate.

Table 6

The clear sky insolation of Berliand by months

Month	1	2	3	4	5	6	7	8	9	10	11	12
QIO	410	509	613	703	763	780	771	716	628	530	430	378

Approximations using more detailed measures of cloudiness have been developed in an effort to improve the estimation of insolation, but suitably detailed data are not readily available at Bermuda.

It has been realized that these approximate methods are applicable only to mean conditions over large areas. As noted, several authors have applied these methods to global and other large scale surveys of heat exchange. This approach has several possible weaknesses: longitudinal constancy is assumed, but the observations used to determine the approximate formulae are made almost exclusively on land. The nature of the maritime air masses has been neglected, but the transmission of solar radiation is dependent on the nature of the air mass (see for instance Pochop, Shanklin, and Horner, 1968). In addition, variations in the distribution and type of clouds have not been considered. Thus Quinn and Burt (1968) note in the Western equatorial Pacific, where cirriform overcast is common, a consistent underestimation of insolation using standard approximations. An investigation of the nature of both the cloudiness distribution and the occurrence of weather disturbances permit an adequate (less than 10%) fit to observed insolation. Thus the best appraoch to the estimation of insolation at the surface is the study of local climate.

The correspondence between Berliand's insolation from clear skies and the observed daily insolation is not unreasonable. However, Berliand's observations are based on land stations. Observations made at Bermuda are preferable but the present record is not long enough to construct a fit for clear sky insolation. The zero sky cover intercepts of the best fits to normalized observed insolation are all less than unity — this may indicate that the atmosphere is more transparent on clear than on cloudy days, if Berlind's fit is representative of Bermuda conditions.

To examine the relation between insolation and total sky cover at Bermuda, the observed insolation is normalized by the clear sky insolation given by Berliand (1960). The normalized insolation is assumed a function of sky cover alone:

$$(8) \quad Q_I / Q_{IO} = f(\overline{c})$$

A least squares polynomial regression of normalized monthly insolation on monthly total sky cover for all 35 monthly means April, 1959 through February, 1962 yields

Improvement in terms of sum of the squares

(9) linear $Q_I/Q_{IO} = 1.048-0.654*\overline{c}$ .12276

(10) quadratic $Q_I/Q_{IO}=1.080-0.767*\overline{c}+.101*\overline{c}^2$ .0002

for all 1091 daily means from 20 March, 1959 through 14 March 1963

(11) linear $Q_I/Q_{IO}=1.046-0.649*\overline{c}$ 25.80

(12) quadratic $Q_I/Q_{IO}=0.870+0.078*\overline{c}-0.624*\overline{c}^2$ 1.358

(13) cubic $Q_I/Q_{IO}=0.950-0.492*\overline{c}+0.494*\overline{c}^2-0.648\overline{c}^3$ .089

The coefficient appropriate to the linear Savino-Angstrom relation (Budyko, 1955) for this latitude is -0.68.

The coefficients given by Berliand's quadratic relation (Berliand, 1960) are —— linear, -0.36; quadratic, -0.38.

The linear regression is in excellent agreement with that given by Budyko for both monthly and daily means, but it is obvious from an examination of the data both that a linear fit is not satisfactory over the full range of sky cover, although good in the limited range of monthly mean sky cover and that this simple method is inadequate for the estimation of insolation over periods as short as a day.

Year averages of insolation for 1960-1 for calculated and observed insolation are in good agreement.

Table 7

The observed year averages of insolation 1960-1 compared with the mean of the whole year regressions on months and days

	Observed	Linear fit monthly data	Cubic fit daily data	mean sky cover
1960	395	394	396	.602
1961	397	394	396	.612

but the seasonal march of insolation is not in such agreement, with underestimation in spring and early summer and overestimation in winter, indicating a change in the nature of sky cover from season to season. An examination of the fractional deviation of average observed insolation from the clear sky insolation of Berliand reveals a regular seasonal variation, suggesting that a seasonal fit for regression coefficients may improve the estimate. As there are but thirty-five monthly values of observed insolation, the year is divided into two parts only: April-September and October-March with the following result:

least squares regression of monthly insolation on monthly total sky cover

(14) April-September (18 months)

$$Q_I/Q_{IO} = 0.956-0.437*\overline{c}$$

(15) October-March (17 months)

$$= 0.844-0.387*\overline{c}$$

(There is not enough data to establish a realistic higher order relation.)

For least squares regression of daily insolation on daily total sky cover

April-September (549 observations)			Improvement in terms of sum of the squares
(16)	linear	$Q_I/Q_{IO} = 1.059-0.618*\overline{c}$	12.73
(17)	quadratic	$= 0.844+0.303*\overline{c}-0.805*\overline{c}^2$	1.18
(18)	cubic	$= 0.976-0.643*\overline{c}+1.093+\overline{c}^2-1.117*\overline{c}^3$	.128
October-March (542 observations)			
(19)	linear	$= 1.000-0.625*\overline{c}$	9.88
(20)	quadratic	$= 0.831+2.821*\overline{c}-0.543*\overline{c}^2$	0.48
(21)	cubic	$= 0.949-.801*\overline{c}+1.056*\overline{c}^2-0.913*\overline{c}^3$	0.08

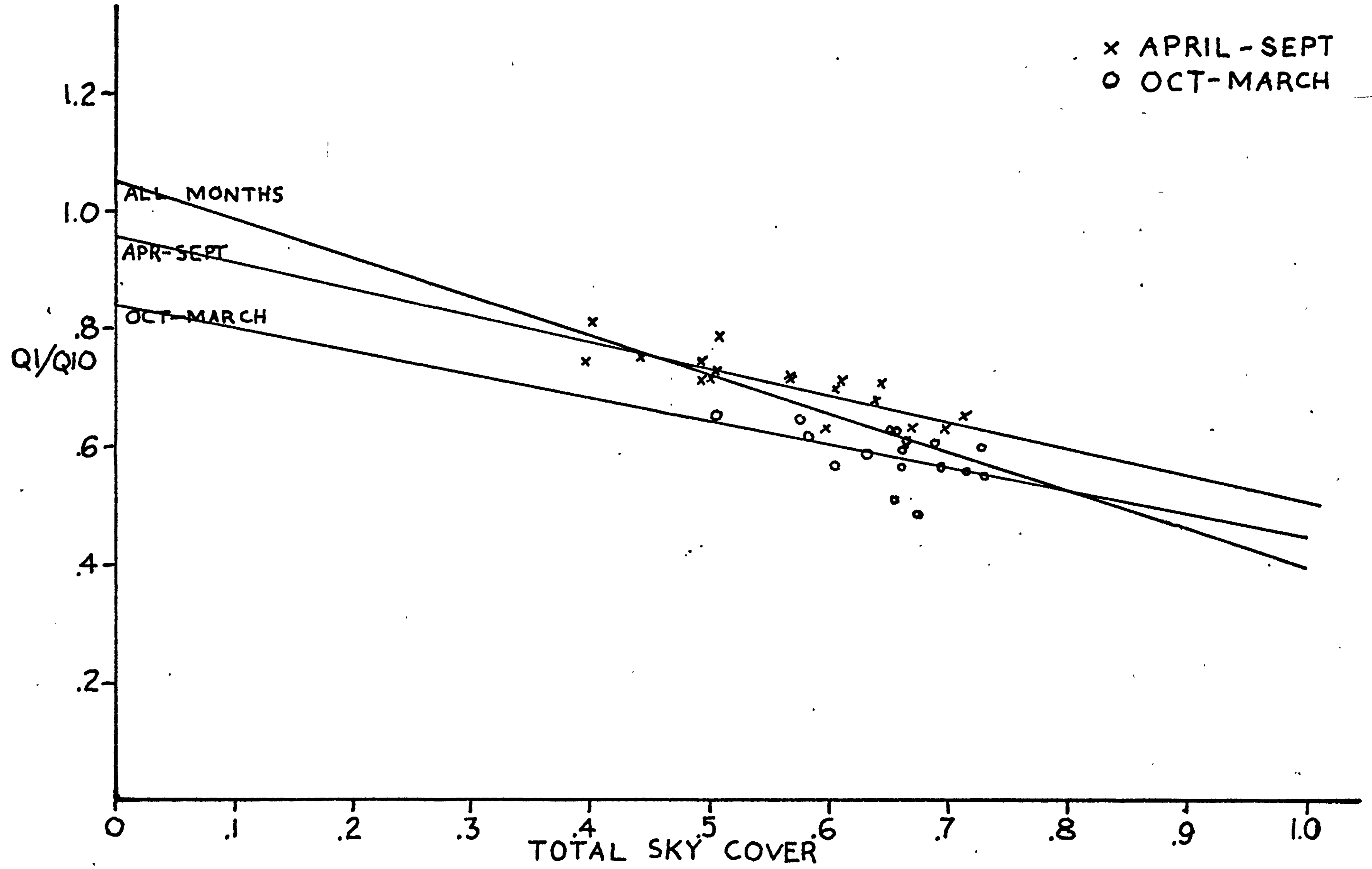


Figure 1 Regression of monthly insolation on total sky cover

The six month fits are each distinctly different from those of the whole year and from each other. The same reported sky cover yields a greater reduction of insolation in October-March than in April-September. These regressions deviate ten percent or less from the observed insolation for the monthly fit although the scatter about the day fit is still large. Monthly averages of insolation calculated from daily fit differ by only a few percent from monthly insolation calculated from monthly mean sky cover.

The calculated year averages of insolation are nearly unchanged:

Table 8

The observed year averages of insolation 1960-61 compared with the mean of the seasonal regressions on months and days

	Observed	Calculated from monthly fit
1960	395	398
1961	397	398

Some of the insolation at the surface of the sea is reflected rather than absorbed. The albedo of the sea, however, is small with daily mean values definitely less than ten percent at all times and probably but a few percent for most of the year. As this uncertain reflected component is small compared with the uncertainties of calculation it is neglected following Stommel, et al. (1968). Therefore, the energy absorbed by the sea is calculated using equations 14 and 15 for monthly values and 18 and 21 for daily values, without further modification.

<u>The Net Flux of Long Wave Radiation to the Atmosphere</u>

This component is the difference between the long wave radiation received and emitted at the surface of the sea. The radiation of the sea surface is approximately that of a black body. Therefore, following the Stefan-Boltzmann law

(22) $I = SaT^4$

where I is radiation emitted in g cal/cm^2/day

a Stefan-Boltzmann constant

T temperature of the sea surface in °K

S correction for deviation from black body behavior ($\sim$.954)

The radiation received by the sea surface is a function of the transparency of the atmosphere and especially water vapor content, sky cover, and temperature gradient.

A simple version of the theoretical relationship for net radiation from clear sky as established by Kindrat'ev and Berliand and taken from Budyko (1958) is

$$(23)\ QB_O = SaT_W{}^4\ (0.39-0.05\ EA^{\frac{1}{2}})$$

where EA is the vapor pressure of the atmosphere above the sea in mb.

Also from Budyko (1958) the influence of clouds may be included as

$$(24)\ QB = QB_O * (1-A*C^2)$$

with the coefficient C a function of latitude varying linearly from .50 at 0° Lat. to .82 at 75°N Lat.

Budyko suggests the addition of a correction for temperature stratification expressed as a function of air-sea temperature difference, such that the net long wave radiation flux to the atmosphere is represented approximately

$$(25)\ QB = SaT_W{}^4\ (0.39-0.05*EA^{\frac{1}{2}})(1-AC^2)$$

$$+4SaT_W{}^3\ (T_W-T_a).$$

There are no observations at Bermuda suitable for checking or improving these relationships.

The following expression from the Smithsonian Meteorological Tables is used to calculate the vapor pressure from the recorded relative humidity and air temperature

(26) EA = EASAT*RH

where

(27) EASAT=6.105*EXP(25.22*T/(T+273)-5.31
*LOG(T+273)/273))

is the saturation vapor pressure in mb at air temperature T in °C, and RH is the relative humidity.

Turbulent Fluxes of Latent and Sensible Heat

The most important steps in derivation of approximate expressions for these fluxes are summarized without detail for later reference.

It is assumed that turbulent exchange determines the distribution of property in the vertical to tens of meters above the surface of the sea. The structure of turbulence near boundaries has been intensively studied: the vertical flux of property (F_S) is alternatively formulated in terms of covariance or an eddy transfer coefficient (K_S) such that

$$(28) \quad F_S = \overline{(\rho' w')S'} = \rho \, K_S \frac{d\overline{S}}{dz}$$

where $\overline{S}$ is the distribution of the property in the mean, S' the fluctuation of property from the mean, ρ' and w' fluctuations of atmosphere density and vertical velocity.

In the atmosphere the intensity of turbulence is influenced by the gradient of potential temperature, stability decreasing and instability increasing the intensity, but the limitations of the available data require that the simplest finite difference approach be followed. Neglecting stability, the bulk aerodynamic method yields, following Deacon and Webb (1962), the following form for the flux of property F_S:

$$(29) \quad F_S = f_a \rho (S_w - S_a) U_a$$

where 'a' is some designated height at which the wind velocity U_a and the property S_a are measured and to which the as yet undetermined coefficient f_a is referred. S_w is the property at the surface of the sea.

It is assumed that the flux is constant — no accumulation of property occurs — in the first tens of meters above the sea, and shear turbulence dominated.

Following the arguments of Prandtl and others, assuming neutral stability and an aerodynamically rough boundary, the logarithmic profile of mean wind is derived from momentum transport considerations.

$$(30) \quad \bar{u} = (u^*/k)\ \mathrm{LOG}\left(\frac{z + z_o}{z_o}\right)$$

where $u^* = \tau/\rho$ = stress/air density

$k = 0.4$ approximately, a universal constant

z_o is a 'roughness' length characteristic of the surface — to be determined

z is the height above the boundary

such that the eddy transfer coefficient for momentum

$$(31)\ K_M = k(z + z_o)u^*$$

This formulation is reviewed by Deacon and Webb (1962), Roll (1965), and many others for observed wind profiles over the sea. Confirmation is obtained for the neutrally stratified atmosphere.

Using the bulk aerodynamic method the transfers of momentum, sensible heat, and latent heat (water vapor) may be presented as

(32) $\tau = C_a \rho u_a{}^2$

(33) $Q_S = H_a \rho C_p (T_w - T_a) u_a$

(34) $Q_E = D_a \rho L (q_w - q_a) u_a$

where ρ is the atmospheric density, $T_w - T_a$ the change in potential temperature from the sea surface to observation height 'a', approximately equal to the measured temperature difference, $q_w - q_a$ difference in specific humidity, L the latent heat of evaporation, C_p specific heat of air at constant pressure, C_a, H_a, D_a coefficients to be determined.

The logarithmic 'law of the wall' yields for the coefficient of drag

(35) $$C_a = (u^*/u_a)^2 = k^2 \text{ LOG } \left(\frac{10+z_o}{z_o}\right)^2$$

but the coefficients H_a and D_a remain undetermined. The roughness length z_o may be determined from the observed wind profile.

Following Malkus (1962) it is assumed

(36) $C_a = H_a = D_a$

that the mechanisms of turbulent exchange of momentum, sensible heat, and water vapor are the same for neutral stability, yielding similar profiles for the several properties above the surface of the sea. As little experimental evidence exists, this common assumption is perhaps one of convenience, although some supporting evidence of Charnock and Ellis is presented by Roll (1965).

The bulk aerodynamic equations 33 and 34 are assumed applicable if the lowest few meters of the air is "well-stirred, shear turbulence dominated, neutrally stable and barotropic" (Malkus, 1962, p. 108). The validity of these assumptions will be discussed later in terms of the results of the computation.

The assumption of equality of coefficients leaves the evaluation of the coefficient of drag. This topic has been extensively reviewed by Deacon and Webb (1962), Roll (1965) and most recently Wu (1969) each of whom have suggested functional relationships of the form

$$(37) \quad CD_a = CD_a(W) \qquad W \text{ the wind speed}$$

deduced from the historical record of drag observations in field and laboratory using both direct and indirect methods The scatter of the observations is large and the several suggestions differ significantly.

Originally computation was performed using the approximation of Wu (1969),

Table 9

The coefficient of drag as given by Wu (1969)

W	CD at ten meters observation height
$W \leq 1m/sec$	$CD_{10} = 1.25 \times 10^{-3}/u_{10}^{1/5}$
$1m/sec < W < 15m/sec$	$= 0.5 * u_{10}^{1/2} * 10^{-3}$
$W > 15m/sec$	$= 2.6 \times 10^{-3}$

(aerodynamically smooth boundary assumed for W < 1m/sec) who deduces a fundamental change in momentum transport to the sea at W = 15m/sec.

Recently, Ruggles (1969) has completed a detailed study of the wind field in the first ten meters of the atmosphere over the ocean in Buzzards Bay, Massachusetts during the summer of 1968. Using only clearly logarithmic profiles, he finds singularities in the drag coefficient at wind speeds of two, four and eight and one-half m/sec. with $CD \approx 1.6 \times .0^{-3}$ between singularities. The nature of the singularities is not yet well understood, but a modal structure is suggested by Ruggles. For these computations it is assumed that this recent work is the more reliable. A constant value of the coefficient of drag of 1.6×10^{-3} is assumed to be most representative of the range of Ruggles' observation from about one to fifteen meters/sec. Wu's approximation has been followed over the range of wind speeds not covered by Ruggles. The possible modal structure is neglected as undetermined.

The following expressions taken from the Smithsonian Meteorological Tables are used to calculate specific humidity from recorded pressure, relative humidity, and temperature:

(37) Q = [0.622*EA(T,RH)/(P-EA(T,RH)]/

[1+0.622*EA(T,RH)/(P-EA(T,RH)]

where EA is vapor pressure in mb (equation 27). T temperature in °C, RH relative humidity, and P the atmospheric pressure.

Chapter III

Calculations Performed on Panulirus Station Data

All subsequent calculations are performed on digitally coded station data.

(1) Interpolation

For machine use the observed data profiles are interpolated to every ten meters depth from zero to one thousand meters. A three point interpolation is performed such that the interpolation between any two points also uses the observation next above, except at the surface at which the observation next below is used. Thus this interpolation is similar to a smooth curve drawn between the points.

(2) Heat storage

The heat stored in the upper five hundred meters of the sea is computed for each station. Thus the lower limit of integration is well into the homogeneous stable 18 C water. Regular seasonal temperature variation is not observed at this depth.

The heat stored is computed numerically, performing tabular integration by numerical quadrature:

$$(38) \quad Q = \int_{-500}^{0} (T-10)*CP \, dz$$

where Q is heat content in $kgcal/cm^2$

z the depth below the surface in meters

T the temperature in °C

CP the specific heat of sea water at constant pressure

For convenience 10°C is selected as 'absolute zero' (no heat stored).

(3) Normalized heat flux

The observed heat stored is used to compute the normalized variations in heat storage. The difference in heat stored between two successive stations whose separation in time is forty days or less is divided by the separation in days and assigned to the middle of the interval as an estimate of the heat flux out of this oceanic region.

(4) Distribution of properties by depth

The interpolated data profiles are used to determine the depths of isolines of property to the nearest ten meters or the variation of property along a fixed depth or isoline of another property.

(5) <u>Depth and intensity of the maximum profile gradient</u>

The depth and intensity are estimated from the interpolated data profiles by assigning the depth to the middle of the interval of largest temperature change and the intensity to the change in °C/10 meters. These values are measures of the depth and sharpness of the thermocline, halocline, etc. The depth and intensity of the temperature gradient have been determined between 0-500 meters for the seasonal thermocline.

Ensemble averages and standard deviations of the above quantities are computed for each month of the year by averaging together all those values in a given month for the thirteen years of observed data.

Chapter IV

The State of the Sea and Atmosphere

The Seasonal Thermocline

The seasonal variation in property of the oceanic surface layers is particularly related to the depth and intensity of the seasonal thermocline. The mean seasonal thermocline is established in early May with a minimum depth of forty meters persisting through May, June, and July; the depth increases with increasing variance to approximately one hundred fifty meters in December and is inderterminate January to April. The intensity of the mean gradient increases with increasing variance from approximately 0.02°C/m in April to 0.2°C/m in August, decreases with decreasing variance to .04 °C/m in January, and remains approximately constant at .02°C/m from late January through April. Thus the homogeneity of the surface layers in winter (early January through late May) is indicated by the large variance of the depth of the maximum temperature gradient and the minimum value and variance of the intensity of the gradient.

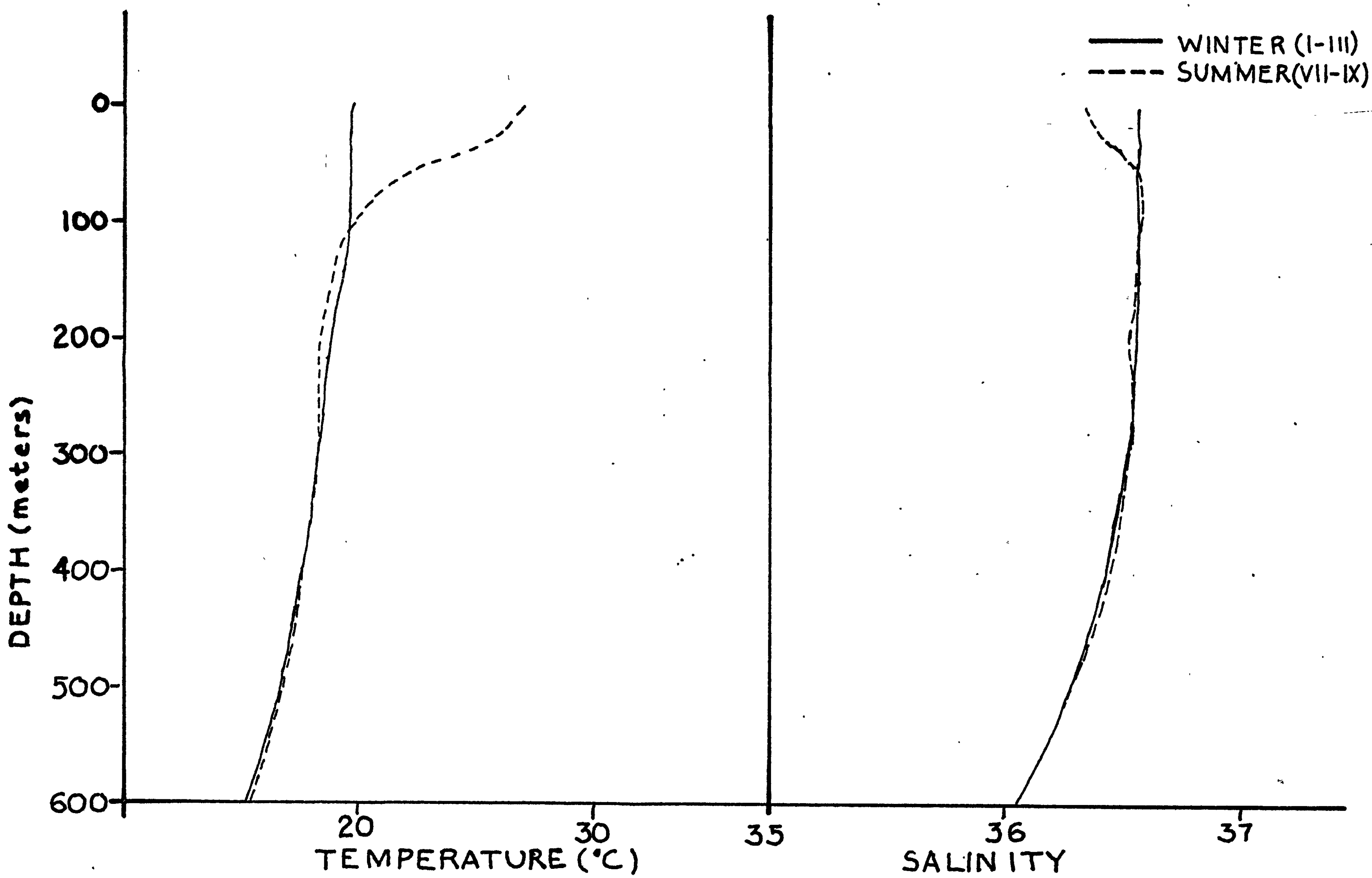


Figure 2 Seasonal depth profiles of salinity and temperature from Panulirus data (1955-1964)

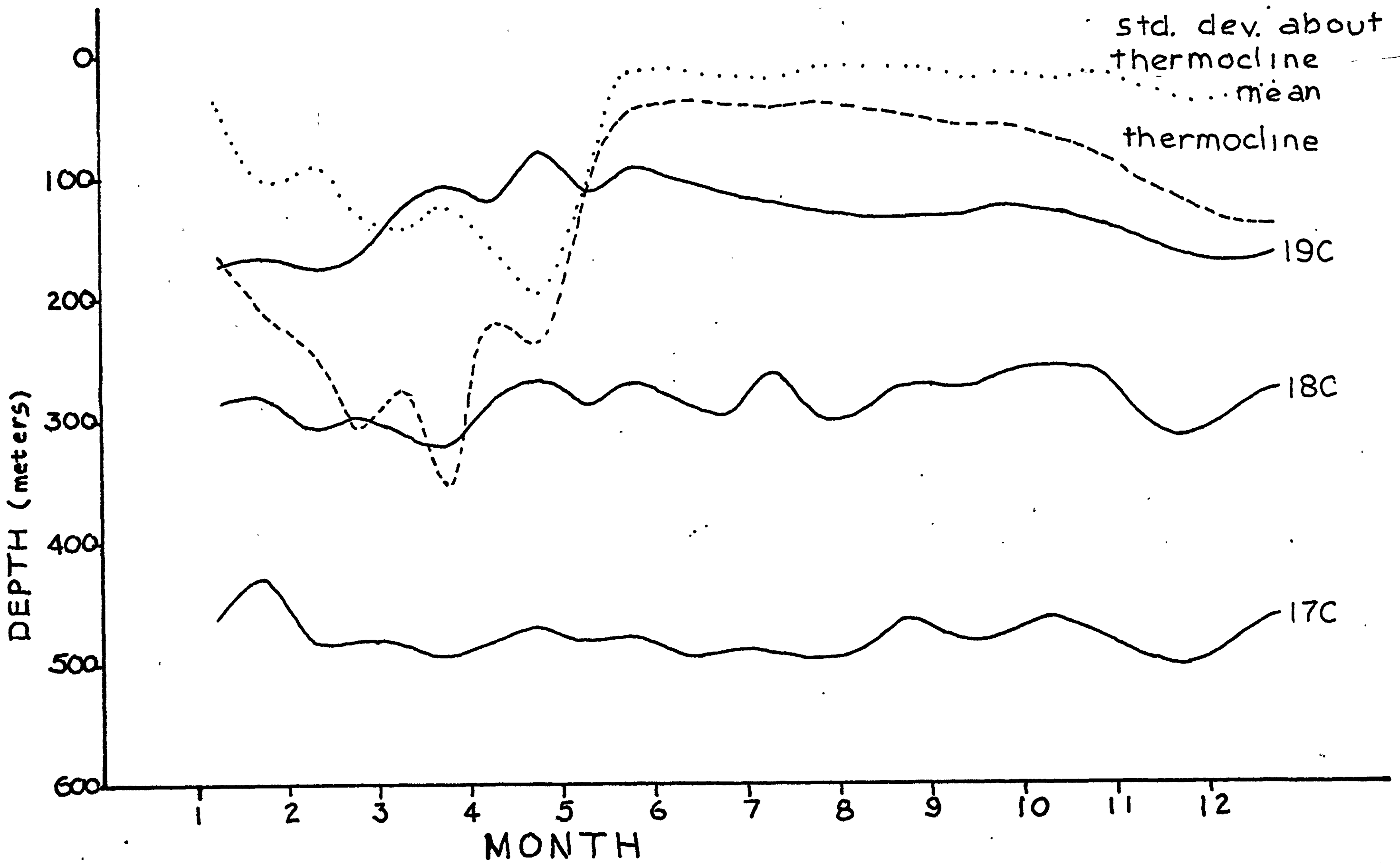


Figure 3 Depth of the seasonal thermocline

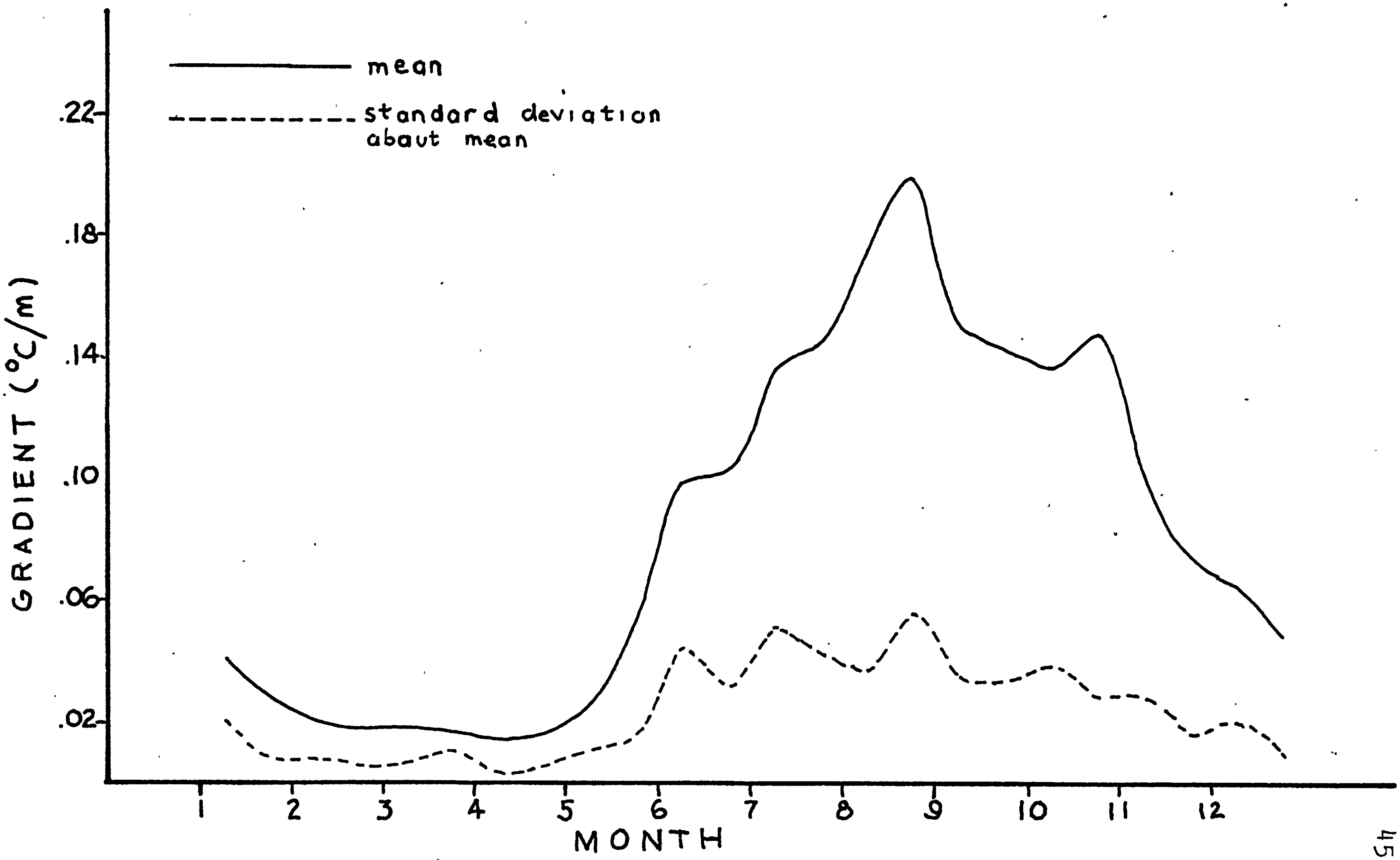


Figure 4 Intensity of the seasonal thermocline

Surface layers of the sea

The observed heat storage from zero to five hundred meters has a seasonal variation of magnitude about fifty-five kgcal/cm^2 in agreement with that deduced by Bryan and Schroeder (1960) from North Atlantic bathythermograph data. The observed spring increase in heat storage occurring May-July is related to the establishment of the seasonal thermocline in late May. The constant minimum depth of the thermocline during this period is associated with a constant rate of heating, a correspondence noted in the time-dependent thermocline model of Krause and Turner (1965). The abrupt decrease in December-January from the maximum storage of August-November to the nearly constant storage of February-April corresponds to the disappearance of the seasonal thermocline in late January when the 'surface' waters reach approximate homogeneity with the 18 C water beneath. There is little further change of surface property in the mean until the following spring. These differences in homogeneity are reflected in the march of sea surface temperature which displays a nearly constant minimum of 18-19° C February-April.and a maximum in August due to the correspondence of maximum heat storage, minimum depth of thermocline, and maximum intensity of thermocline in that month.

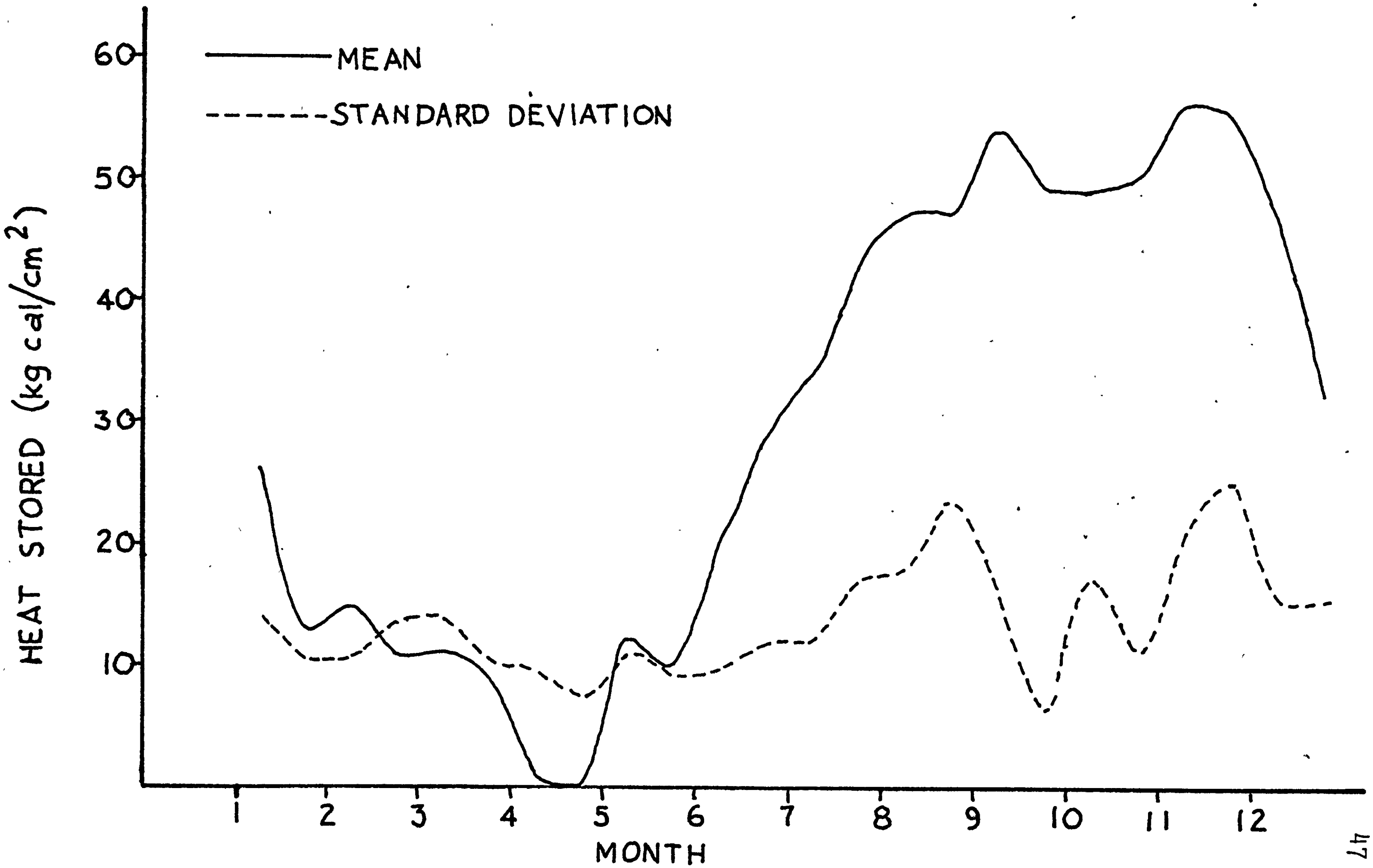


Figure 5 Seasonal march of heat storage.

The change in regime from that of summer-fall heating-cooling to winter-spring homogeneity is approximately indicated by the maximum temperature gradient depth contour passing the 19 C isotherm, the 'boundary' of the 18 c water mass. In some years the maximum gradient contour follows this isotherm for several stations indicating that convection is occurring throughout the surface layers, but homogeneity with the 18 C water has not yet been achieved. When the surface temperature is approximately 19° C convection penetrates the 18 C water, which apparently absorbs the heat losses occurring during the rest of the winter with little change in property — this heat loss is thus small compared with the heat stored in the whole 18 C water mass. However, there is a small net trend in heat stored and other parameters January-March, a reflection of the structure of the 'homogeneous' 18 C water.

This penetration of surface influence into the 18 C water is apparent in the mean seasonal variation of depth of the 19 C isotherm. The April minimum depth of about eighty meters corresponds to maximum winter convection and minimum heat storage; the depth increases May-November and is about constant December-February. The 18 and 17 C isotherms show no obvious seasonal variation.

Thus, convection does not occur through the entire depth of 18 C water, but some replacement seems to occur at this latitude. This interpretation is supported by both the depth profiles of temperature in the North Atlantic compiled by Schroeder (1965) and the 170 meter February mixed layer depth independently determined by D. M. Filippov, et al. (1968).

Seasonal march of atmospheric property

The variation of atmospheric property is governed particularly by the Bermuda-Azores high pressure system. Although the circulation may at any time of year be interrupted by frontal systems from the continent, these intrusions occur most frequently in winter. Thus spring-summer is characterized by maximum atmospheric pressure, maximum relative humidity (spring), minimum sky cover, minimum wind speed, increasing sea and air temperatures, and minimum sea-air differences, and fall-winter is characterized by maximum sky cover, maximum wind speed, minimum pressure, minimum sea and air temperatures and maximum air-sea property differences. A smaller summer variance of properties reflects the greater summer stability of the Bermuda-Azores circulation; a larger winter variance the more frequent frontal intrusions.

September to April the land cools more rapidly than the sea; frontal systems bring the colder continental air over the sea yielding air-sea temperature differences as much as $-3°$ C. Similarly May-August the air heats more rapidly than the sea and the temperature difference is slightly positive $< + .5°$ C. Thus the atmosphere above the sea may be expected to be most convectively unstable September to April and least unstable May-August if it is assumed that, for daily or longer means, the island of Bermuda has negligible influence on the meteorological data collected at Kindley A.F.B.

Chapter V

Heat Budget Calculations

Nature of the averaging procedure

The several components of the air-sea heat exchange are first calculated for every month from June 1954 to June 1967 from monthly means of the parameters: Kindley total sky cover, air temperature, specific humidity, vapor pressure, atmospheric pressure, and wind speed and SST as observed by the R. V. Panulirus on station. It is assumed that the mean of the Panulirus observations for any given month is representative of the sea surface temperature for that month. An average by months over the years 1954-1967 establishes the seasonal march of the several components. An average by years over months January-December displays year to year variations in the nature of the exchange.

The components of the exchange are then calculated for each day from daily mean parameters: total sky cover, air temperature, relative humidity, atmospheric pressure, wind speed, from observations at Kindley Field, and daily readings (11:00 A. M.) of sea surface temperature at Argus Island Texas Tower. These daily values are then averaged for each month of calculation May, 1963-December, 1967.

An average by months over the five years 1963-1967 and an average by years for months January-December are taken as above. An average of the results of monthly mean calculations with Argus data are also taken for each month 1963-1967 for comparison with Panulirus results.

The heat exchange

A summary of the seasonal march of the coefficients of heat exchange calculated from monthly and daily means is presented. The variation of the several components as presently calculated reflects the variation of air-sea property difference, wind speed, total sky cover, and insolation. Thus the radiation balance — the net radiative flux to the sea — displays a summer maximum in August displaced from the June maximum of clear sky insolation due to the march of total sky cover,.and a winter minimum in December. The turbulent transports of sensible and latent heat possess summer minima and winter maxima reflecting the seasonal march of wind speed and air-sea property difference.

The turbulent transport of sensible heat, although not negligible in winter makes the least contribution to the heat balance. The net transport of heat through the surface of the sea is thus principally governed by net radiation flux and turbulent transport of latent heat. The sensible heat transport is, however, important in fueling local adiabatic convection and may exercise influence on cloudiness and precipitation.

The two methods of calculation give differing results for the turbulent transports May-August when sea surface temperature is rapidly changing. This suggests that the monthly mean of the Panuliris surface temperature observations is not representative of the actual monthly temperature under such conditions. This data remains, however, the only available for years 1954-1962.

The average march by months of each component of the heat exchange is compared with the global survey results of Budyko (1963). The values of Budyko have been taken from his charts by linear interpolation. The residual of this budget calculation, representing the net heat flux through the surface of the sea, is a combination of local heat stored and the divergence of the advective heat transport. These calculations of residual: monthly calculations over the years 1954-1967 based on Panulirus station surface temperature observations and daily calculations over the years 1963-1967 based on Argus Island observations of surface temperature, together with the

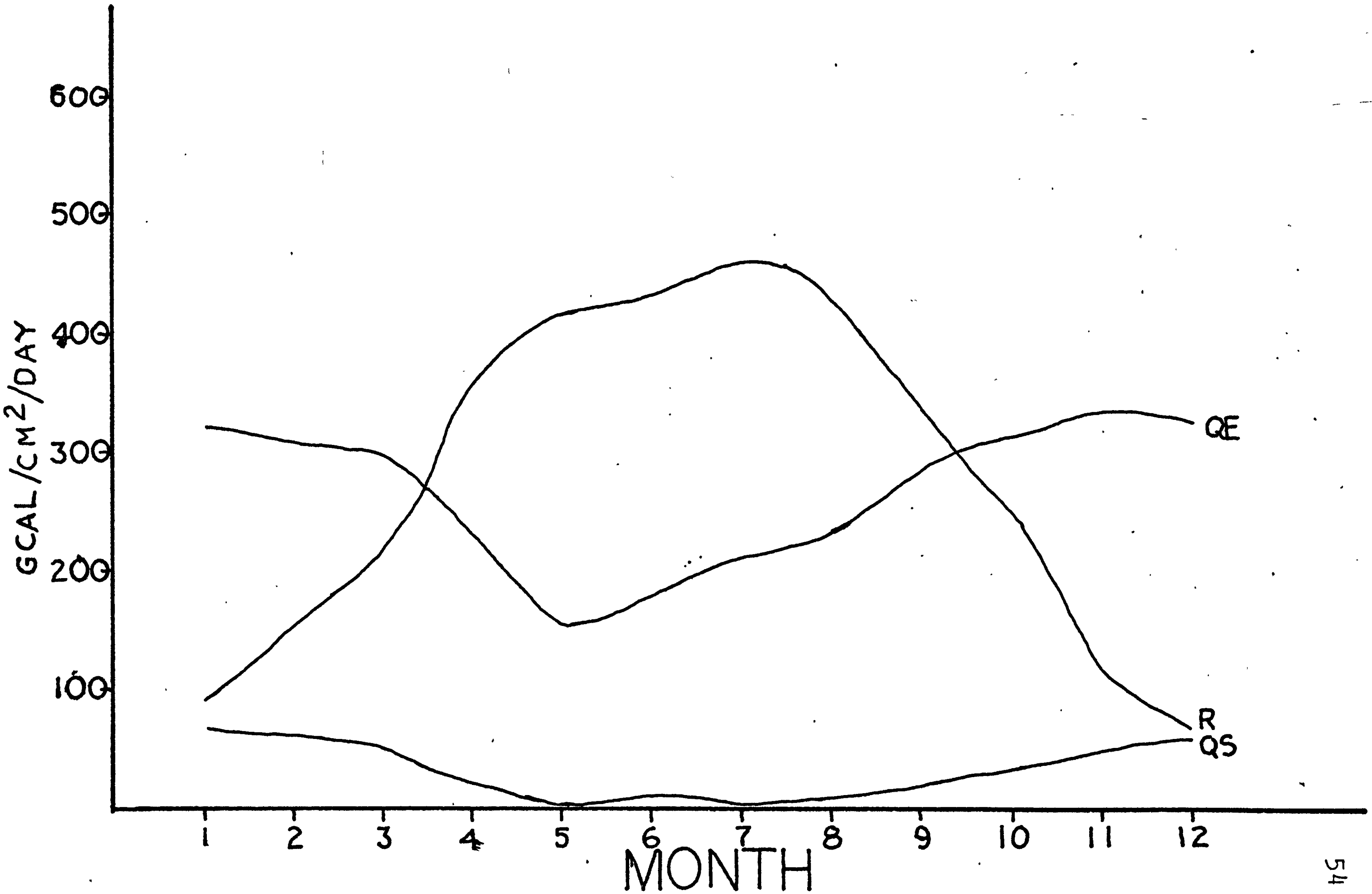


Figure 6 Summary of the heat exchange

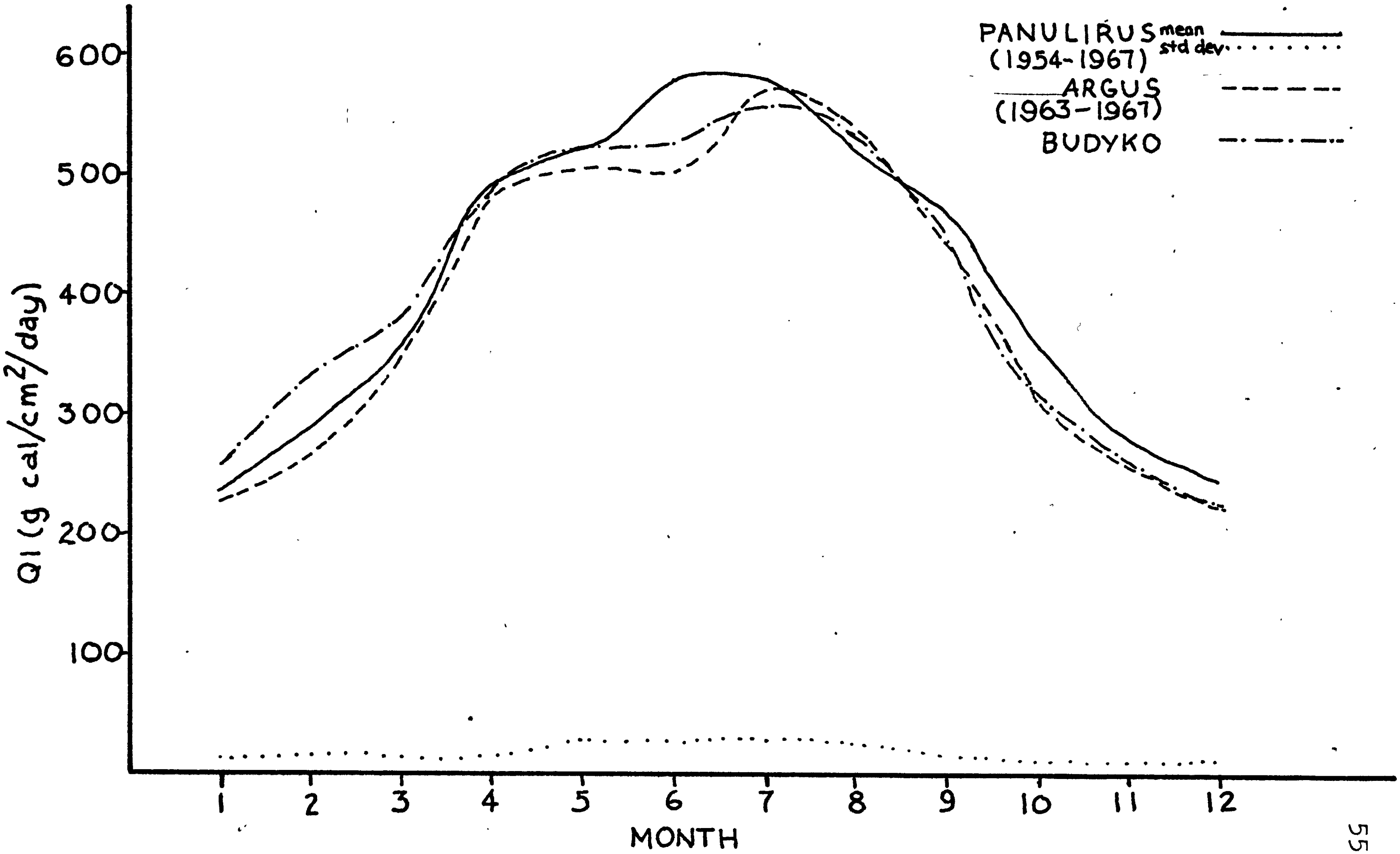


Figure 7 Insolation

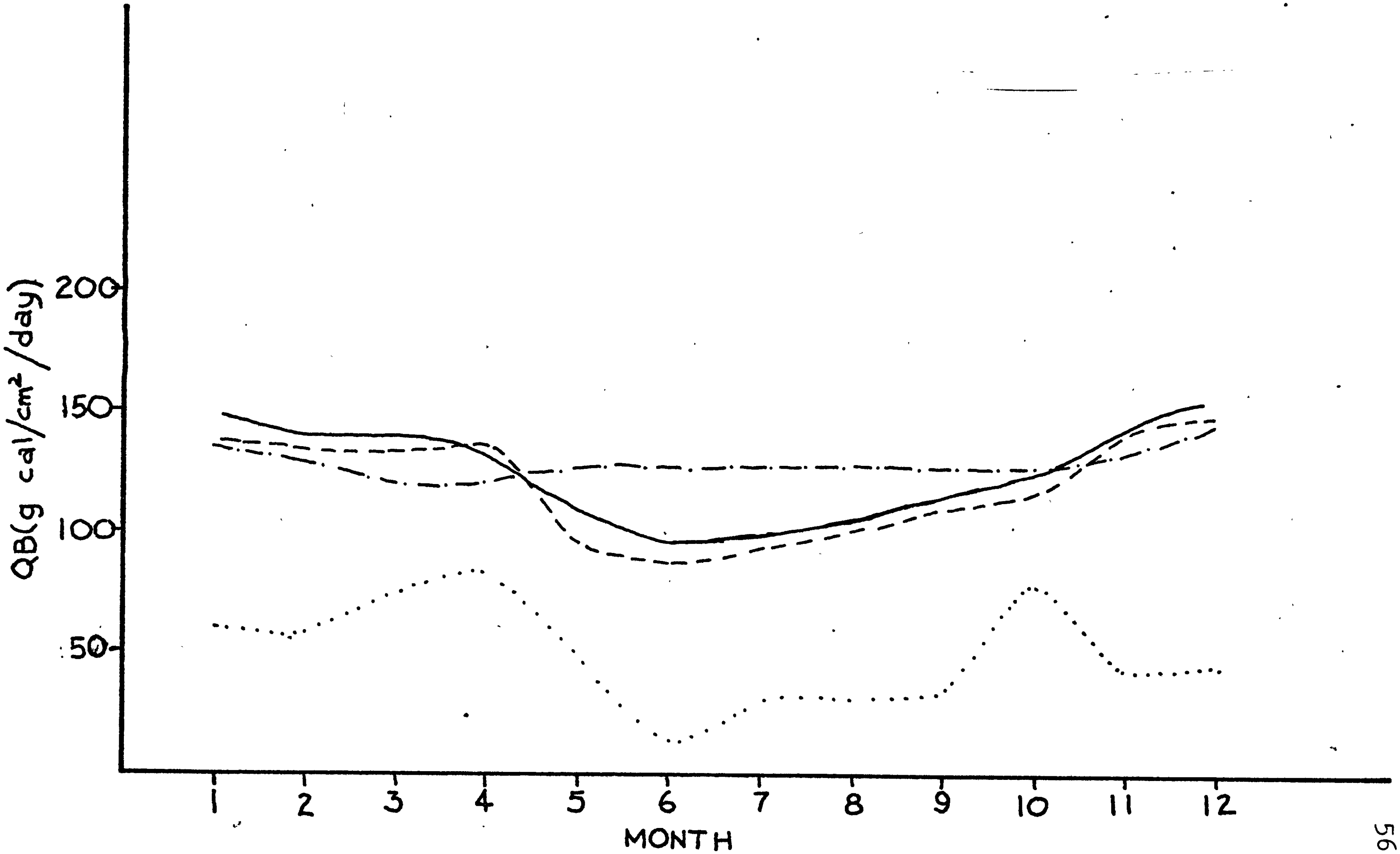


Figure 8 Flux of back radiation

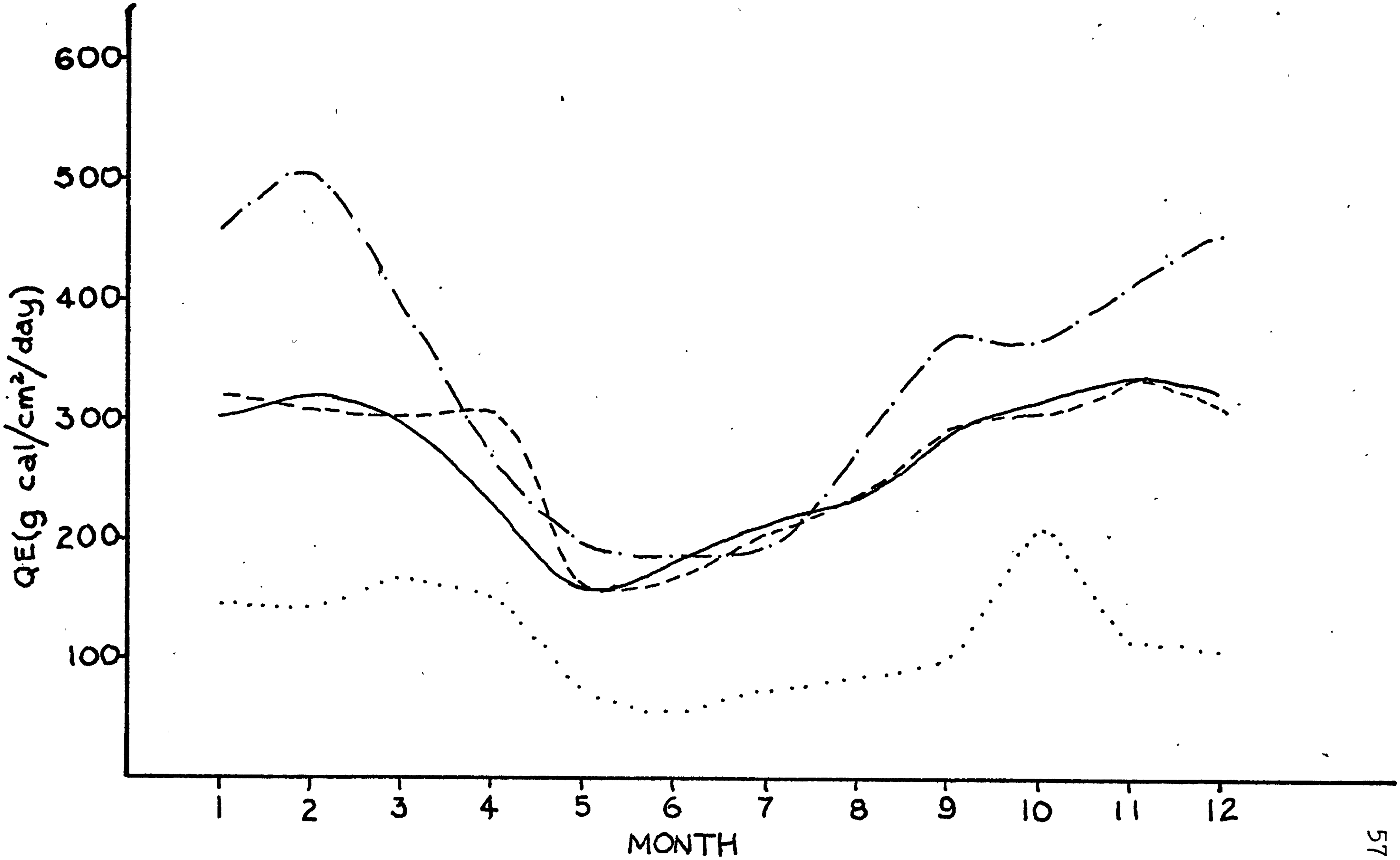


Figure 9 Flux of latent heat

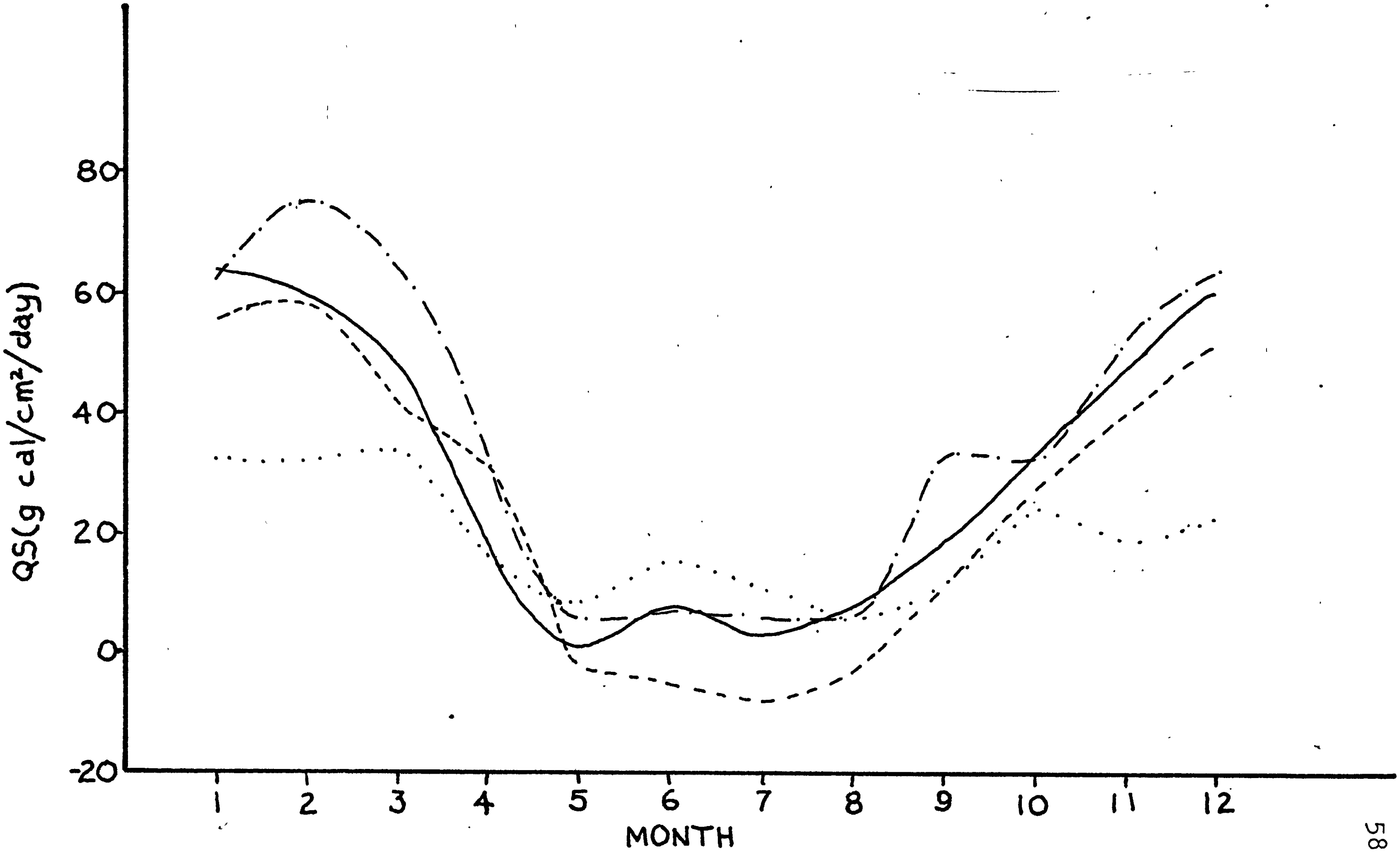


Figure 10 Flux of sensible heat

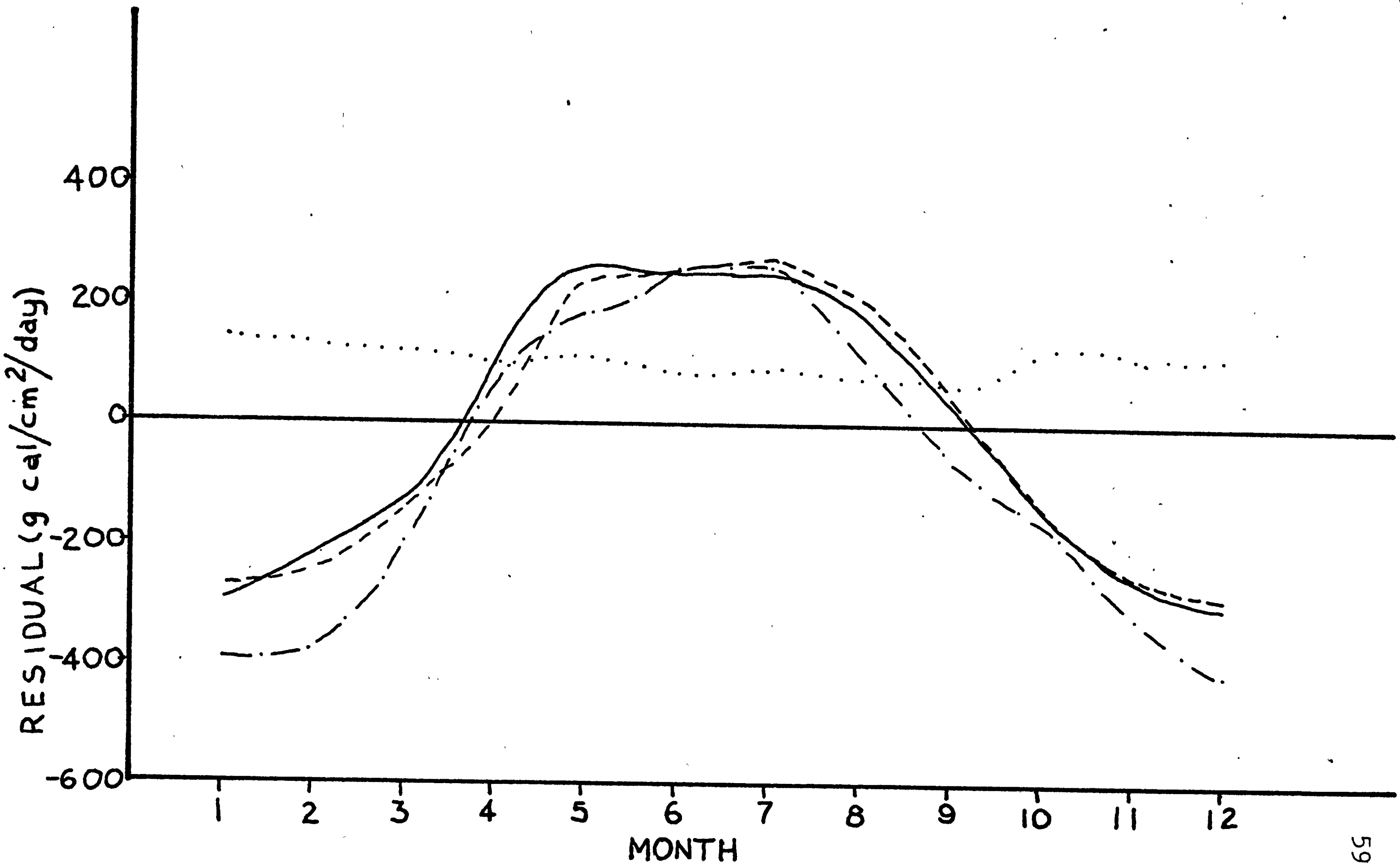


Figure 11 The residual of net heat flux through the surface of the sea

residual term given by Budyko are compared with the average by months of the observed change in heat content of the first five hundred meters of the sea as calculated from Panuliris data. The difference between the calculated net heat flux through the surface of the sea and the observed change in heat storage is considered separately as the divergence of the advective transport.

The calculations of Budyko

The method of data selection and processing used in compiling Budyko's Atlas of the Heat Balance of the Earth is not clear from the text. The estimation relations used differ for

i) insolation — Berliand's quadratic relation is used (equation 7)

ii) net flux of long wave ration to the atmosphere

$$(39) \quad QB = S\sigma T_w^4 \, (11.7-0.23*EA)(1-C\bar{c})+4S\sigma T_w{}^2(T_w-T_a)$$

where C(30°N) = 0.63. This relation differs from (25) in that the vapor pressure dependence is linear rather than square root, the sky cover dependence is linear rather than quadratic.

iii) turbulent fluxes of latent and sensible heat — this relation differs only in that the coefficient of drag is taken equal 2.1×10^{-3}, said to be deduced from closure of the heat balance problem for the Pacific Ocean.

The insolation of Budyko is smaller in summer (June-July), larger in winter (September-March) as the coefficients of the Berliand estimates are not seasonally dependent. The net flux of long wave radiation to the atmosphere of Budyko is of the same order but does not have the same seasonal variation as the present result. However, this small change may have been lost in interpolating from the given contours of Budyko, particularly as the back radiation is not directly given, but must be calculated as the difference between the given radiation balance and insolation. The net result of these differences in radiation budget is a winter estimate of the radiation balance about thirty g cal/cm^2/day greater than the present calculation.

The turbulent transfers of Budyko exceed the present values by a considerable margin in winter (December-March). The reason for this difference is not apparent from the methods of calculation — it is only known that Budyko's coefficient of drag, a constant of approximately

2.1×10^{-3} for all wind speeds, should result in transfers larger in magnitude everywhere, if the calculations are based on similarly processed data.

The residual of the calculation deviates most greatly (100 g cal/cm^2/day) in winter due principally to the difference in turbulent transfer of latent heat at that time.

Without more information on the methods of data selection and processing, no certain reason for these differences can be given. However, the relative magnitudes of the components of the heat exchange are similar and the seasonal variations qualitatively similar.

The Calculations of Jacobs

Jacobs (1951) has calculated the turbulent transfers of latent and sensible heat using ship data, a constant coefficient of drag numerically similar to the present one, but only one ship observation per day. The results are of doubtful value as both the data and the data handling techniques may be questioned. The values below are interpolated linearly from Jacobs' contours for the North Atlantic.

Table 10

The turbulent heat transfers of Jacobs(1951) compared with the transports here calculated and those of Budyko (1963)

	Jacobs		Panulirus based average 1954-67		Argus based average 1963-67		Budyko	
	Q_E	Q_S	Q_E	Q_S	Q_E	Q_S	Q_E	Q_S
Dec-Feb	327	60-70	314	61	321	58	468	67
Mar-May	171	18	224	22	248	25	286	35
June-Aug	263	0 to -1	207	6	208	-4	219	6
Sept-Nov	276	20	309	33	327	30	380	40

The turbulent transfer of sensible heat given by Jacobs is in good agreement, but the latent heat transfer has a spring, rather than summer minimum. This anomaly may be the result of poor ship humidity data.

Garstang (1965) has noted related diurnal cycles in the parameters on which these calculations depend, particularly air temperature, sky cover, wind, pressure, humidity. Some semi-diurnal variations attributed to the atmospheric tide are also present in his records. The earlier practice of Jacobs of selecting data at some fixed GMT for calculation thus may introduce a regional bias in the results.

Year averages

The year averages of the components of the heat flux are calculated by the several methods for those years in which data is available for every month of the year. Also included is the average of observed insolation from the years 1960 and 1961. A difference between daily and monthly based latent, sensible, and back radiation flux is present, probably a result of unrepresentative sea surface temperature sampling. Therefore, it is difficult to identify long-term trends in components of the heat exchange — in particular, the increase in the residual 1964-1966 may be spurious.

Seasonal heat storage

The seasonal march of the residual component of the heat budget is calculated by months from daily and monthly means of data. The several calculations of the heat flux through the surface of the sea deviate particularly from the observed storage change in June and July and in January (by approximately 350 g cal/cm^2/day) at the maximum and minimum observed storage change.

Integration of the calculated mean residuals over the positive and negative parts separately gives a measure

TABLE 11

YEAR AVERAGES of MONTHLY VALUES of HEAT EXCHANGE

	Insolation		Back Radiation		Latent Heat		Sensible Heat		~~Bowen~~ Ratio	Residual	
	mean	σ	mean	σ	mean	σ	mean	σ		mean	σ
1954	401	116									
55	395	119									
56	399	121									
57	399	117									
58	404	140	127	17	259	96	39	35	.150	-21	273
59	399	135									
60	398	118	130	26	246	73	32	26	.130	-10	224
61	398	125	126	24	256	112	31	30	.121	-15	272
62	381	117	115	23	271	103	32	32	.118	-38	250
63	392	128	118	19	267	63	27	29	.101	-19	216
	(390)	(132)									
64	304	117	123	25	281	83	31	28	.110	-51	233
	(376)	(117)	(119)	(21)	(262)	(81)	(28)	(25)	(.107)	(-31)	(232)
65	403	131	137	20	307	57	33	23	.107	-74	207
	(408)	(138)	(129)	(20)	(280)	(61)	(26)	(22)	(.093)	(-27)	(211)
66	385	124	130	20	300	66	31	26	.103	-69	216
	(391)	(130)	(117)	(26)	(273)	(77)	(21)	(32)	(.077)	(-29)	(242)

TABLE 11 (cont.)

YEAR AVERAGES OF MONTHLY VALUES OF HEAT EXCHANGE

	Insolation		Back Radiation		Latent Heat		Sensible Heat		Bowen Ratio	Residual	
	mean	σ	mean	σ	mean	σ	mean	σ		mean	σ
1967	395 (381)	115 (117)									
Mean exchange	395 (389)	122 (125)	125 (119)	19 (20)	263 (267)	58 (58)	31 (25)	23 (24)	.118 (.094)	-24 (-26)	217 (214)

σ: Standard deviation about the mean

Note: ( ) indicates daily calculation using Argus SST observations of the seasonal variation of heat storage, neglecting advection

Table 12

Seasonal heat change

	SUMMER	WINTER	SUMMER+ WINTER ($=Q_V$)	AVERAGE MAGNITUDE
1954-67 Panulirus based average	30.8 kgcal/cm²	-40.2	-9.4	36
1963-67 Argus based average	29.2	-39.6	-10.4	34
Budyko	23.8	-56.2	-32.4	40
observed change in heat stored	55	-55	0	55

Although the standard deviation about the mean of the observed march of change in heat storage is much larger than the mean, the mean displays a clear seasonal variation. However, the significance, if any, of variations near the

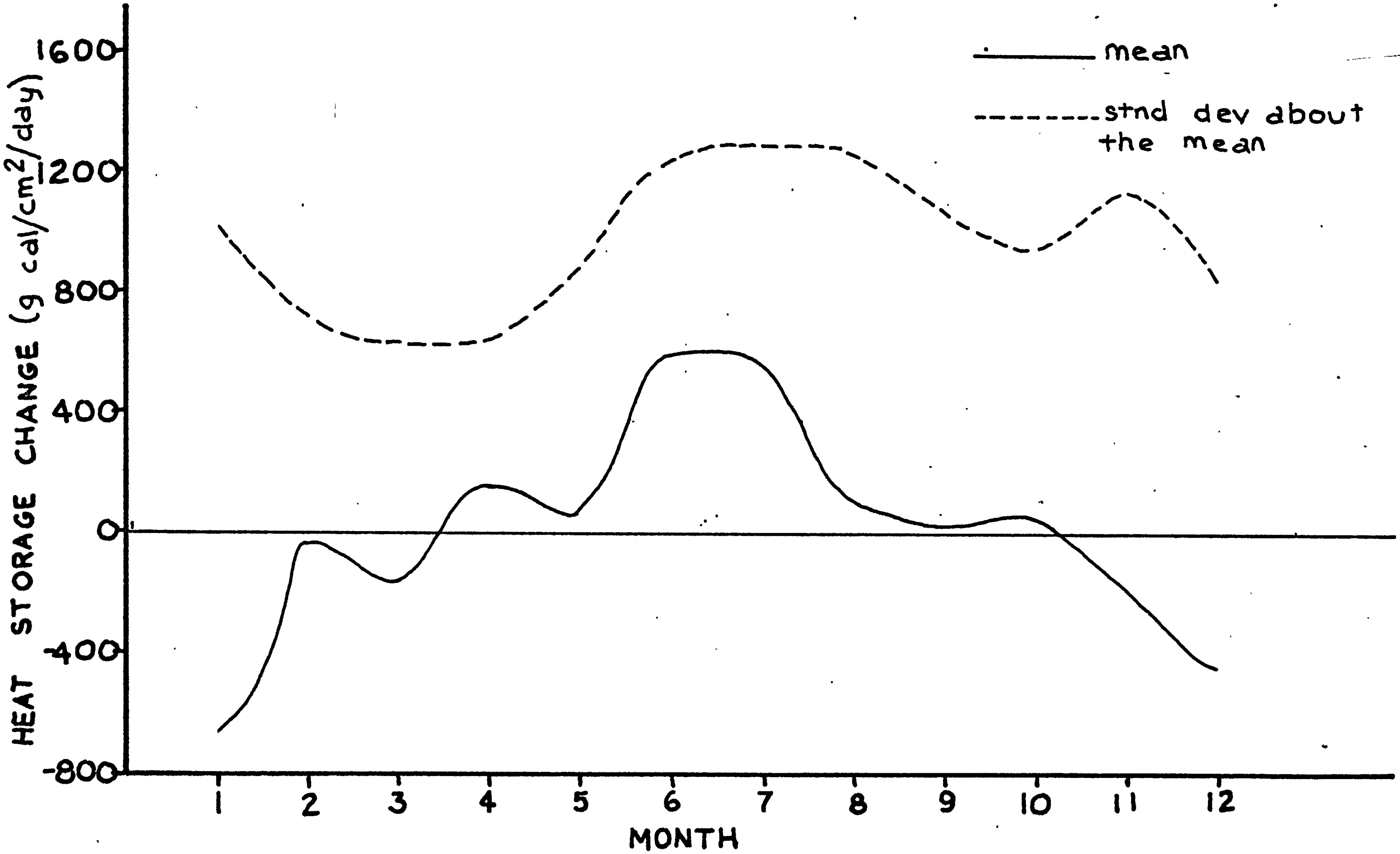


Figure 12 Observed change in heat storage

zero crossings is not clear. The amplitude of the calculated seasonal variation in local heat storage is notably smaller than the observed seasonal variation. Both Bryan and Schroeder (1960) from combined bathythermograph data of the North Atlantic 20°N-65°N and Pattullo (1957) for a series of stations in the North Atlantic and North Pacific have found that the seasonal heat storage change exceeds estimated local heating.

Divergence of advective heat transport

If the net heat stored over the year is assumed negligible, the year average of the residual is the year average of the divergence of advective heat transport, the negative values indicating a net convergence of heat over the year.

Table 13

Yearly mean advective heat transport

BUDYKO	-90 g cal/cm^2/day
PANULIRUS BASED	-24
ARGUS BASED	-26

If the usual practice is followed the difference between calculated local heat flux through the surface of the sea and observed change in storage is assigned to the divergence of the advective transport of heat by surface currents, implying that the divergence of heat transport is 700 g cal/cm^2/day greater in January than in June. Thus divergence of the advective transport must account for approximately forty percent of the observed variation in heat storage at the Panulirus site, out of phase with the observed heating cycle (convergence in summer and divergence in winter). The amplitude of this seasonal variation in divergence is an order of magnitude greater than the net transport. Is it reasonable that seasonal changes of this magnitude should occur? If Bryan and Schroeder (1960) are correct this situation is typical of the North Atlantic. Perhaps the abrupt changes in heat storage occurring in June-July and December-January suggest that some basic property of the heat exchange has been neglected.

Climatological averaging assumption

The averaging by months of meteorological variables before computation of the components of the heat balance will introduce error if the variables are correlated as

$$\overline{F(x_1, x_2, x_3, \ldots)} \text{ does not necessarily equal } F(\bar{x}_1, \bar{x}_2, \bar{x}_3)$$

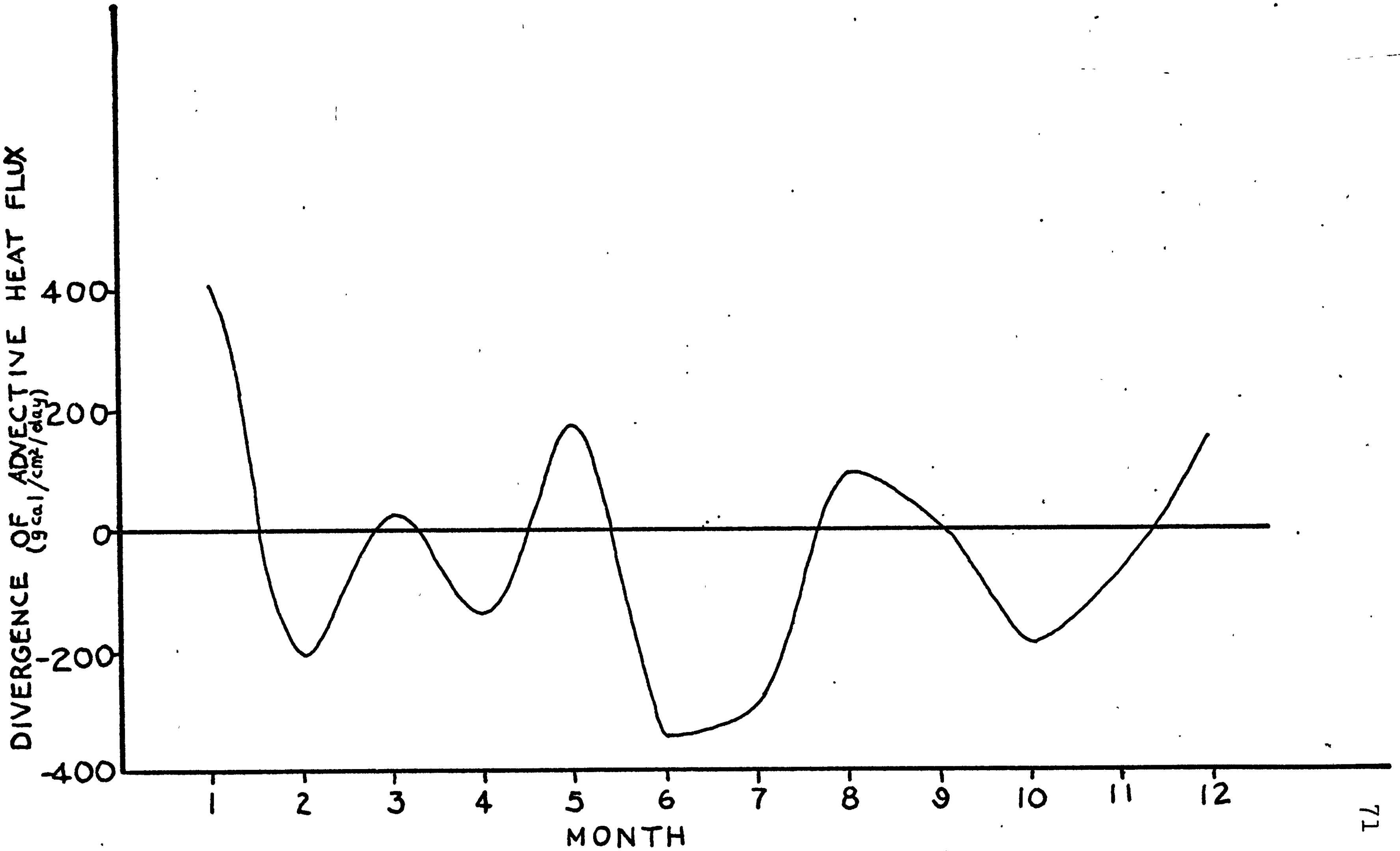


Figure 13 Divergence of the advective heat transport

Negative correlations between air-sea property differences and wind speed have been observed by Malkus (1962), Day (1967) and others to introduce reductions of the order of ten percent in calculations based on long term mean data compared with calculations based on hourly data.

Garstang (1965) and Petterssen (1962)find a particularly strong dependence of the turbulent transport of latent and sensible heat on the occurrence of synoptic scale phenomena in the atmosphere. Daily values of several times the monthly mean are observed in frontal systems moving off the continent. with integrated transports twice the expected mean. In the present calculations the latent heat transfer has been observed to exceed 1000 g cal/cm^2/day with mean wind speed of the order of thirty knots. Thus the climatological averaging procedure may reduce the turbulent transports, particularly during winter months when frontal passages are most frequent. The several components of the air-sea heat exchange, calculated in somewhat different ways, from daily and monthly means of the same data for the years 1964-1966 do not show the expected deviations: the deviation observed is most pronounced in summer and autumn with the monthly mean results usually greater than the daily mean results. These differences may be due to the different locations of collection.

For example, the Argus Island Texas Tower sea-surface temperature observations are taken in waters of depth of less than the depth of the seasonal thermocline at this season. The well-mixed layer extends to the bottom and the topography and local circulation may then influence the temperature variations. However, the Panulirus and Argus sea surface temperatures are generally in good agreement except for the summer of 1966, when there occurs a persistent difference of about 1°C. Also the island of Bermuda may influence the behavior of the mean daily meteorological parameters. For example a strong wind may change mean daily temperature and specific humidity over the island during the summer, reducing rather than increasing the air sea-property differences, as the air over the sea is likely more cool and moist than that over the island, at least during daylight hours. Unfortunately Argus Island does not collect data at night, so no comparison of daily mean sea and land observations can be made.

The larger value and variance of the turbulent fluxes during the winter season is likely due to the unsettled weather conditions accompanying the passage of frontal systems,but the wind air-sea property difference correlation expected is not apparent in these transports. The results of these calculations, therefore, probably should be assigned uncertainties greater than the expected ten percent reduction.

The neutral stability assumption

It has been assumed that the atmosphere immediately above the surface of the sea is neutrally stratified. However, the air-sea temperature difference varies from -3 to +.5°C during the course of the year. This variation results in a seasonal march of atmospheric stability reflected in the variation of the bulk Richardson number from -.0121 (December) to +.0018 (June), following Deacon and Webb (1962)

$$(41)\quad R_B = \frac{10g}{T}\, z\, \frac{T_a - T_W}{W^2} = \frac{g}{T}\, z\, \frac{G_H}{{G_M}^2}\, \frac{T_a - T_W}{W^2}$$

where it is assumed that the Montgomery coefficients of transfer for sensible heat and momentum $G_H = G_M = 0.1$

and

- T is the temperature of the sea in °K
- z height of observation above the surface of the sea
- T_a air temperature at that height in °C
- W wind speed at that height in m/sec
- T_W sea surface temperature

This variation may be partly due to influence of the island on the meteorological data.

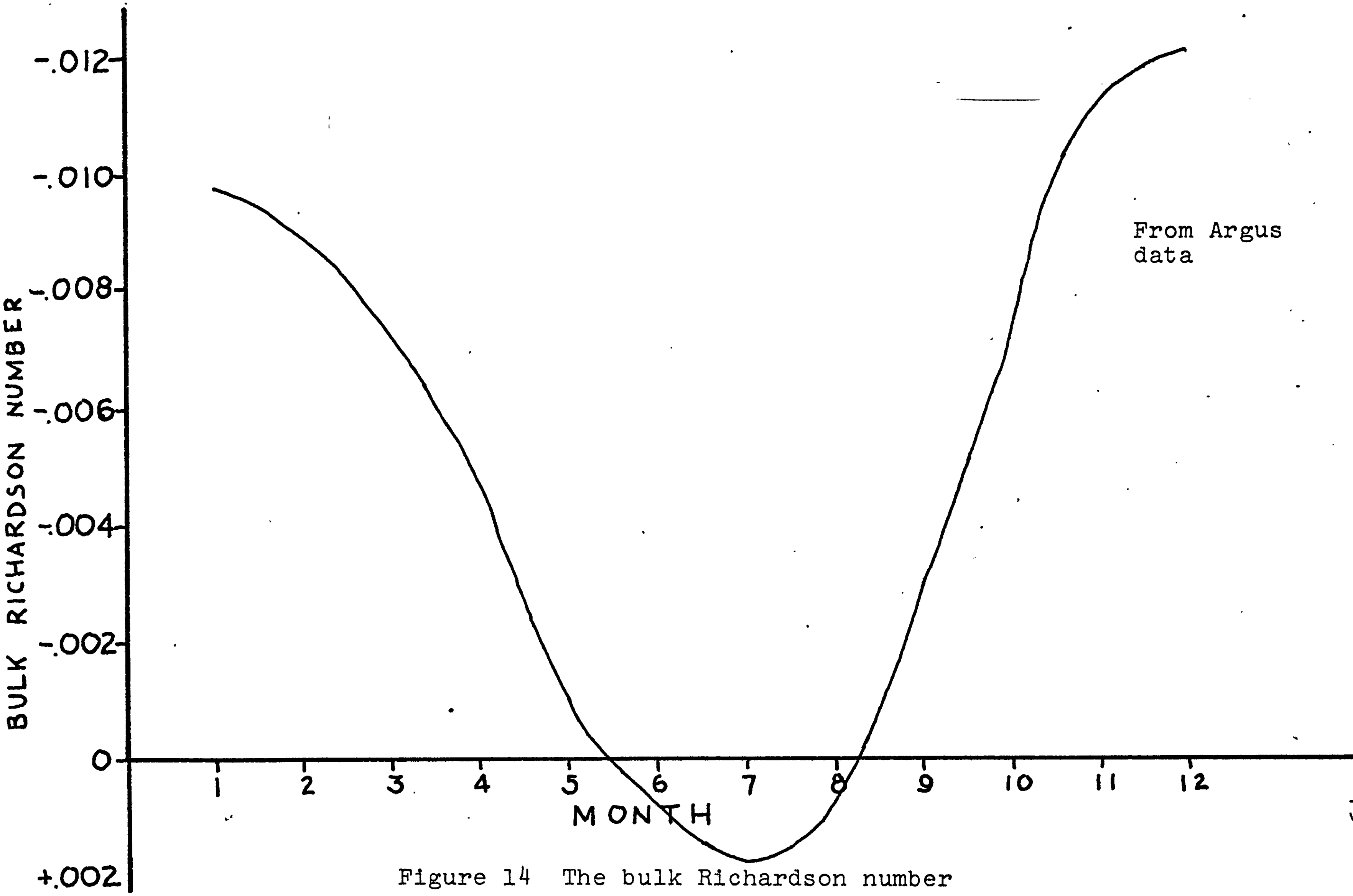


Figure 14 The bulk Richardson number

For near neutral conditions one may approximate the mean gradient of the wind over the sea as

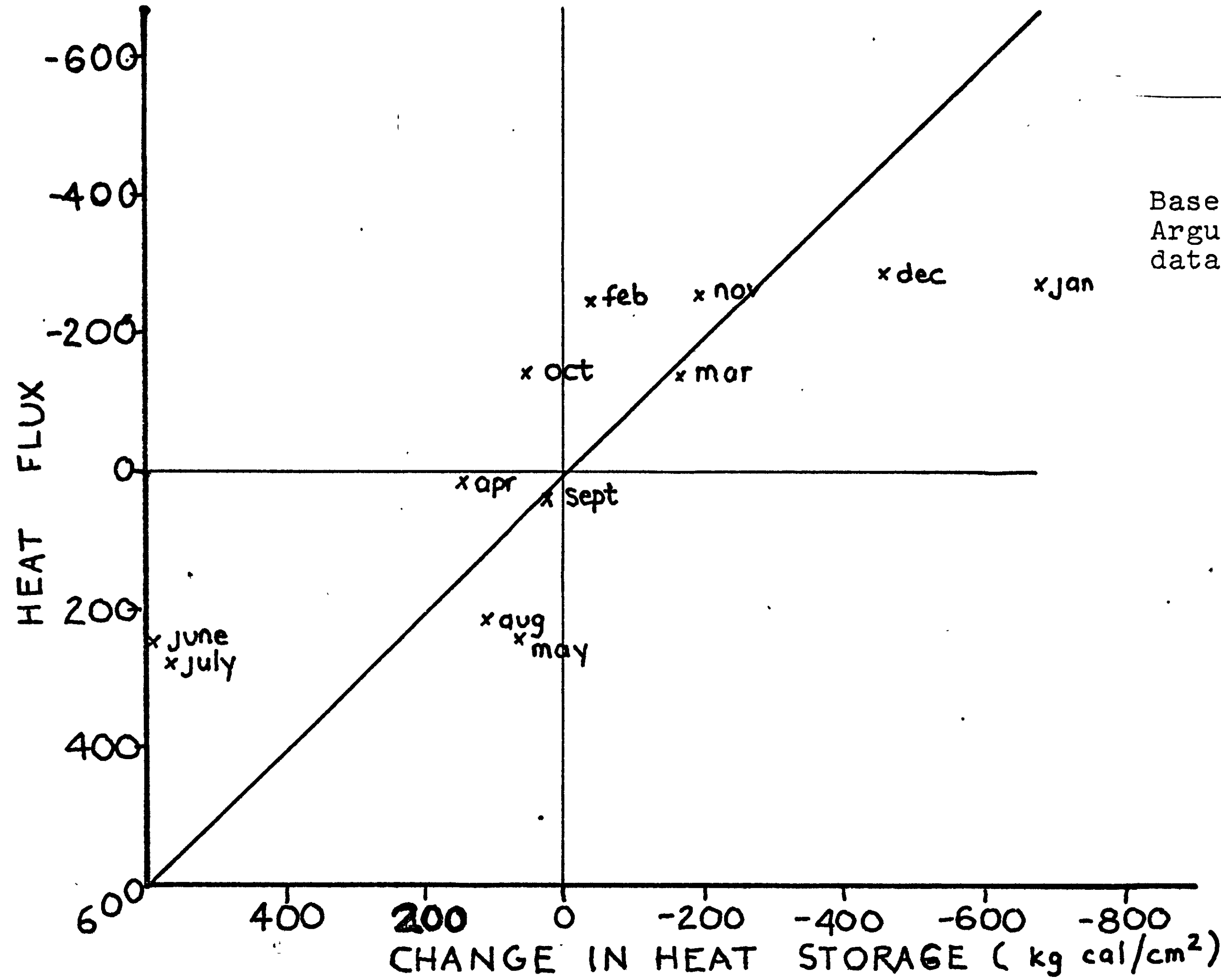
$$(42) \quad \frac{\partial u}{\partial z} = \frac{U^*}{K(z+z_o)} (1 + \text{constant}^* R_i) \text{ for } |R_i| \lesssim .03$$

from Roll (1965). Applying these considerations Garstang (1965) has introduced an approximate linear dependence into his coefficient of drag as

$$(43) \quad C_6^* = (C_6(w) - 4.2\ R_B)^*10^{-3}.$$

The present data is probably not adequate for stability corrections as it is not all collected at the deep ocean site of interest. However, the influence of stability on the turbulent fluxes near Bermuda seems unlikely to be negligible.

The largest deviation of calculated heat flux through the surface of the sea from observed change in heat storage is as noted a loss of approximately 350 g cal/cm^2/day in January and a similar gain in June and July, corresponding to the extremes of the bulk Richardson number. Therefore, it seems difficult to determine from present methods that final residual; the seasonal march of the divergence of the advective transport of heat, and the significance of



Based on
Argus SST
data

Figure 15 Calculated vs. observed monthly heat change the residual term of the heat budget calculations remains uncertain.

Advection near Bermuda

Some speculation on the patterns of advection near Bermuda may be made, however. The Panulirus observation site is in the region to which Wust, Defant and others assign the subtropical convergence of surface waters. Well-defined thermal fronts have been observed south of Bermuda. Therefore, it is possible that a regular seasonal variation of the heat transport occurs near Bermuda, but this will not explain the difference between observation and calculation also found by Bryan and Schroeder (1960) and Pattullo (1957) in widely differing oceanic regions. To determine the nature of the advective transport near Bermuda direct observation, rather than indirect heat balance calculation will probably be required.

Value of calculations of seasonal heat flux

Does this residual term of the heat budget calculation have any relation whatever to the observed variations in heat storage? To answer this question, the several calculations of the residual and the observed heat change are approximately integrated from monthly values to obtain a measure of the seasonal change in heat storage: a smooth curve is drawn between the monthly values by eye and the positive and negative portions are then integrated by planimeter.

The observed mean seasonal change in heat stored is greater than the integral of observed heat flux by approximately ten kg cal/cm^2/season. The integral of the heat flux, however, is less well determined than the heat storage as the interpolation among the monthly values is rather arbitrary. Therefore, the observed change will be used for comparison. The several calculations yield April as the month of the first heat gain in agreement with the observed variation, but there is no significant loss of heat stored before November, contrary to the calculated first loss in September. Also, the increase in depth of the seasonal thermocline accelerates and the intensity of the gradient decreases after the negative crossing of the observed heat flux in October, while the sea surface temperature maximum occurs around September 10 in the previous month — both changes in agreement with the predictions of the potential energy model of the time-dependent thermocline proposed by Krause and Turner (1965). The sea surface temperature minimum occurs around the tenth of April at the positive crossing of the observed heat flux, as expected, since the seasonal thermocline is absent at

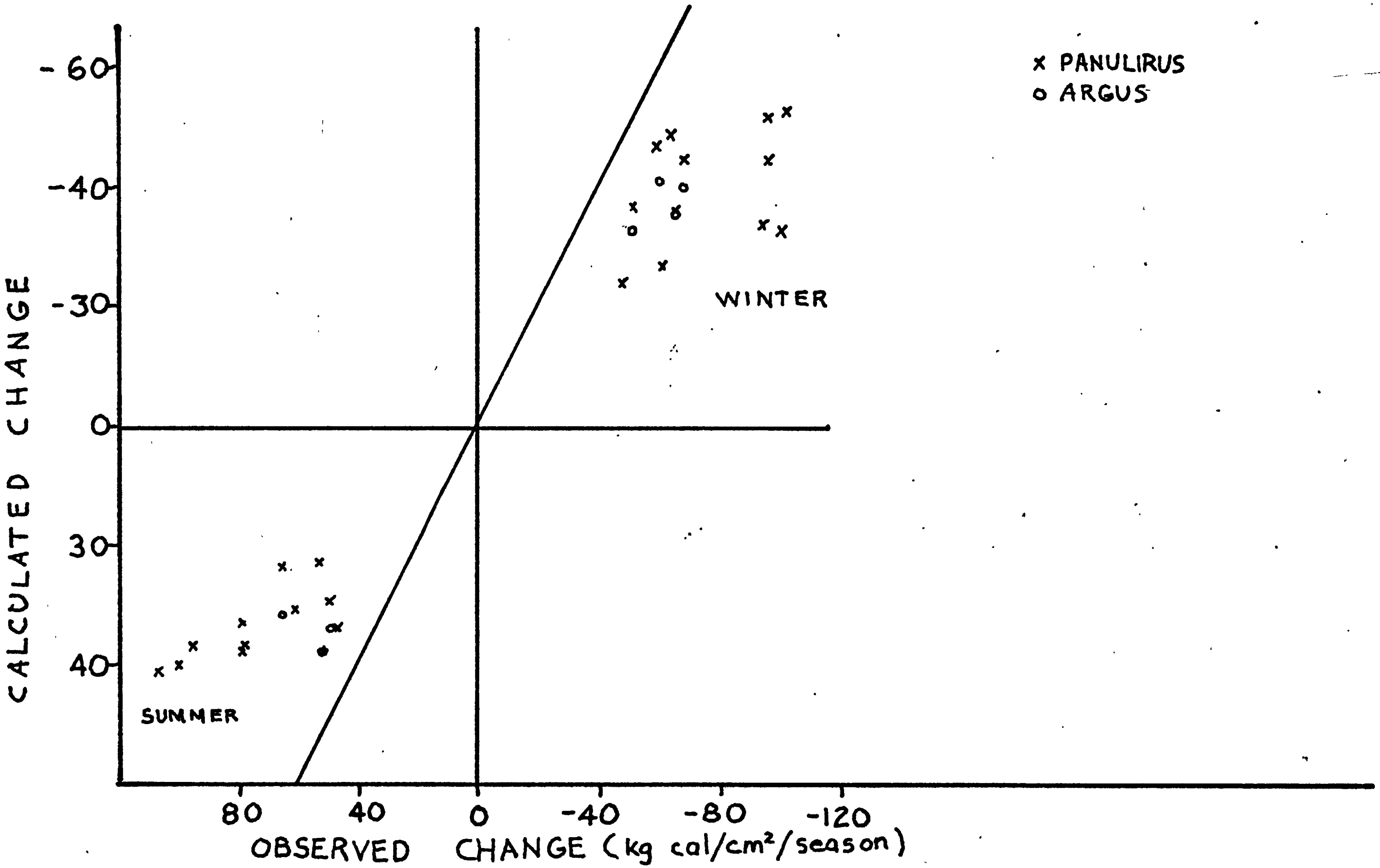


Figure 16 Calculated seasonal heat change vs. the difference of the observed extremes of heat stored

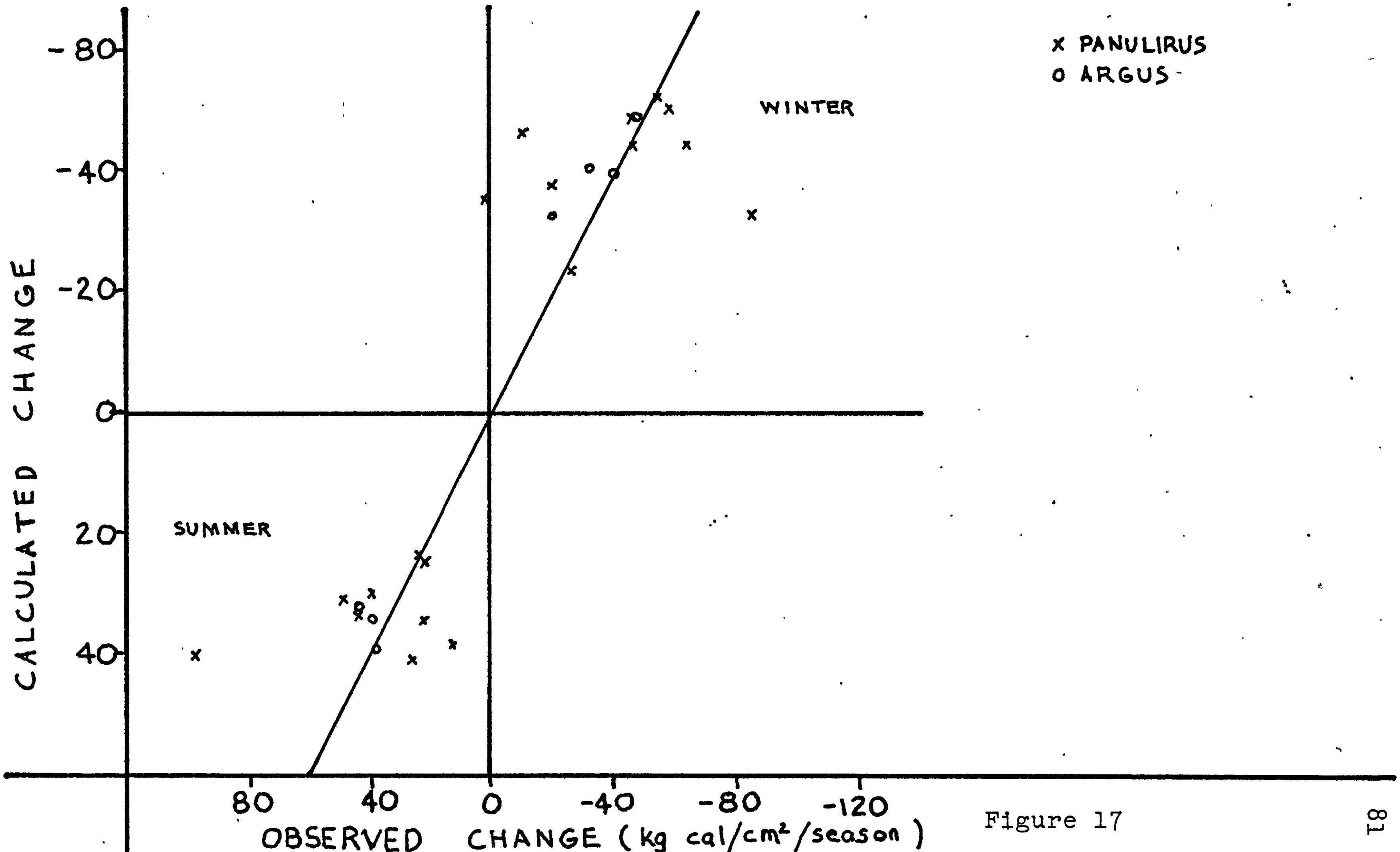


Figure 17

Calculated seasonal heat change vs. the observed difference in heat stored at the calculated start and finish of the heating cycle this time. The calculated residuals all display a net heat loss over the year, implying a not unreasonable net advective transport of heat into the region. Corroboration is not possible, however, without data on the circulation of the Sargasso Sea. Thus, the observed and calculated heat fluxes are in agreement on the initiation, but on neither the magnitude of seasonal variation nor termination of heating.

The variations in observed heat stored are irregular such that the calculated and observed occurrence of maximum and minimum heat storage may be separated by more than one month. There does seem to exist a relation among the measures of seasonal heat flux. Although the relative changes in amplitude often differ, the sense of the change is the same except for 1965. 1965, however, has a particularly irregular heat storage record. The commonly employed techniques of heat budget calculation, therefore, provide at least a qualitative measure of the variations of heat stored in the surface layers of the sea.

It should be realized that the various components of the ocean-atmosphere heat exchange are closely interrelated, and controlled by complex feedback mechanisms. Thus, an increase in insolation provides more energy to the ocean surface layer, which in turn may give up the energy in

TABLE 14

SEASONAL VARIATIONS IN HEATING DATES

Date of calculated initiation of heating	Date of heat storage minimum	Date of calculated termination of heating	Date of heat storage maximum
		10-1-54	
3-18-55	4-12-55	10-1-55	9-4-55 ?[1]
-	2-1-56 ?	10-6-56	8-27-56 ?
4-1-57	5-23-57 ?	9-11-57	11-20-57
4-2-58	4-16-58	10-1-58	10-1-58
3-9-59	4-1-59	10-5-59	7-22-59
4-4-60	5-25-60	9-17-60	7-27-60
3-20-61	4-12-61	9-21-61	11-24-61
4-8-62	4-14-62	9-21-62	8-15-62
4-7-63	4-14-63	9-4-63(9-18)[2]	10-1-63
4-5-64(4-1)	5-9-64	9-22-64(10-1)	9-17-64
4-14-65(4-14)	3-3-65	9-1-65(9-11)	10-4-65
4-1-66(4-1)	3-1-66	9-4-66(9-26)	10-4-66
4-10-67(5-4)	6-1-67		

[1] ? indicates result based on incomplete data sets

[2] ( ) indicates Argus based results

latent and sensible form to the atmosphere. The resulting warm moist surface air may rise to form clouds, reducing the insolation, or the turbulent transport during strong convective activity may be enhanced by an increase in wind speed, but inhibited by a decrease in sea-air property difference. In general, the system varies only slightly about an equilibrium imposed basically by the insolation at the top of the atmosphere and the physical properties of the earth as a planet. Only persistent trends in solar radiation or atmospheric transparency can make any significant change in the mean characteristics of the air-sea energy exchange.

Chapter VI

Conclusion

Implications

The following conclusions seem plausible

(1) The qualitative behavior of the mean of the several components of the heat flux is well established by the agreement of the several independent calculations discussed and is related to the heat storage in the surface layers of the sea as observed in the Panulirus hydrographic series.

(2) The absolute magnitude of the several components of heat exchange cannot be determined without further refinement of both the data collection and methods of estimation. In particular (a) the calculation of the local radiation balance for periods shorter than one month will require measured rather than estimated solar insolation. Estimation of insolation for longer periods should be based on local rather than global climate, with consideration of variations of atmospheric transparency, cloud type and distribution. Field study of the transmission properties of the various types of sky cover for both long and short wave radiation and systematic observation of insolation in ocean areas are vital.

(b) The turbulent exchange of heat is governed largely by short period synoptic disturbances; the instantaneous distribution of heat exchange may be quite unlike the mean. Therefore, to determine the local exchange, preserving the correlations of the variables, simultaneous hourly observations of the profiles of wind, temperature, and humidity at the site of interest are necessary.

(c) The simplifying assumptions used in deriving the bulk aerodynamic relationships are not always justified. The assumption of the neutrally stratified atmosphere does not appear realistic when the calculated residual is compared with the observed seasonal heat storage in the sea. However, turbulent transport in the stratified atmosphere is not well understood. Detailed field study of the profiles of atmospheric property above the sea under non-neutral conditions is indicated. Even for the neutrally stratified atmosphere, it is not certain that the mechanisms of transport of momentum, sensible heat, and water vapor are similar and as the coefficient of drag (momentum transfer) is not clearly formulated, uncertainty is inevitable with this method.

(d) To accurately determine the divergence of the advective transport of heat, synoptic observation of ocean advection, rather than further study of the heat balance, seems most useful. With this information the heat balance results may be more fruitfully compared with oceanographic observation.

That present calculations have some value is indicated by the agreement with large scale features of the circulation of ocean and atmosphere as noted by Malkus (1962). However, although present global surveys are still entirely dependent on limited ship data, improvement through detailed local studies is possible. This work, unfortunately, with its scattered data sources does not qualify, but can only suggest the direction of investigation. Until the methods used here are refined to apply to dynamically changing systems, the distribution of air-sea energy will remain conjectural and little progress will be made in the study of time-dependent phenomena. For instance, an accurate measurement of the air-sea interaction coupled with the theory of the seasonal thermocline proposed by Kraus and Turner (1965) could predict the depth of winter mixing in the sea. In this way the formation of water masses could be studied. Or the way in which tropical storms draw upon the stored heat of the sea to grow may be more accurately revealed. But to fully develop these techniques requires better knowledge of the behavior of the coupled systems in sea and atmosphere, such that the exchange relations may be mathematically formulated as conditions on the air-sea interface. Meanwhile, the numerical results of heat budget studies may continue to give hints of the nature of the physical processes of global and local air-sea interaction.

Further Study

As a basic investigation of the transfer phenomena it would be particularly interesting to observe at some fixed platform such as that used by Ruggles in Buzzards Bay, simultaneous profiles of the several variables in atmosphere and ocean together with direct measurements of the radiative and turbulent fluxes of heat for all seasons, under the natural variety of weather conditions. Shipboard observations, scattered in time and space, usually limited to 'good' weather conditions, and hampered by an unsteady platform are less useful for basic study. The results of such a study, if applied to such synoptic observations at sea that an instrumented buoy network might perform, may make possible for the first time an accurate quantitative measure of air-sea energy exchange. Prediction, however, is unlikely without a far more general study of the coupling of the (nonlinear) ocean and atmosphere circulations. That a program of such varied, simultaneous observations would be successful is uncertain, but observation of any of these components of the oceanic heat balance, separately or in combination is sure to yield useful information.

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