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ON THE INFLUENCE OF VARIABLE LARGE-SCALE WIND SYSTEMS ON THE HEAT BALANCE IN THE ACTIVE LAYER OF THE OCEAN

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F. A. Berson
Extended Forecast Branch
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SUMMARY

Month to month variations in sea surface temperature anomaly are examined for a middle-latitude zone of the North Pacific Ocean during September 1959 through August 1960. Analysis on Hovmöller diagrams reveals the existence of slowly progressive wave patterns with a wave length about 5,000 km and a typical period about 5 months. To understand the heat-exchange processes associated with these fluctuations of sea-surface temperature, the anomalies of storage and of meridional flux-divergence (advection) of heat are evaluated in ten-degree longitude sections. A simplified model for the homogenous layer is adopted and the advection anomaly is calculated from meridional (Sverdrup) drift induced by surface wind stress anomaly and the normal south-to-north temperature gradient.

Next the excess or deficit of energy flux at the ocean surface, i.e. with respect to the normal value, is deduced from the imbalance of advection and storage anomalies in the mixed layer. The results indicate that in central longitudes and during autumn and winter when anomalies of net radiation at the surface appear to be rather small in comparison, a large proportion of the anomalous evaporation and turbulent heat conduction is generally tapped from the accumulation of heat due to the effect of wind stress anomalies on advection. To this extent the investigation lends support to the conclusion by Namias that energy feedback from the ocean is operative in the maintenance of certain

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systems of anomalous large-scale circulations in the troposphere.

A theoretical treatment of the thermal wave patterns is then given with a view to determining their period as a function of wave length and of amplitude of windstress anomaly. For temperature amplitudes of the magnitude indicated in the Hovmöller diagram (1° to 2° C on the average) the theoretically deduced period agrees with the observed period indicated in both temperature and wind-stress data only if the wind dependent effect of surface energy exchange is taken into consideration. It is shown that this effect can be incorporated into the thermal wave equation for a closed system (no energy exchange with the atmosphere) by factorizing the wind stress amplitude.

Allowing for the wind stress (forcing) field and the geostrophic-motion (internal) term, calculated periods are too large by a factor of five. The effect of surface energy exchange from anomalous sea-level wind is to shorten always the period of thermal waves with respect to those in the closed system.

The theory also shows that resonance with the forcing field occurs for retrogressive waves only and at periods exceeding a year, i.e. at wave lengths and wind-stress amplitudes of the observed magnitudes.

The study concludes with a proposal for a numerical method for computing, from an initial state, sea surface temperature anomaly changes when the wind stresses are given. Some aspects of the dynamics of feed-back loops are discussed.

1. Introduction[1]

Long-range forecasting is hampered by, among other things, a lack of better understanding and by the difficulties in the way of assessing quantitatively the mutual influences of large-scale circulations in the atmosphere and the distribution of heat energy in the ocean. This is particularly true for the slow anomalous changes in the wind systems and in the thermal state of the ocean that remain after one has eliminated the normal seasonal changes which are on the whole better understood.

The present investigation is mainly concerned with the ocean boundary or mixed layer in its thermal response to the anomalous changes of the large-scale wind systems in middle latitudes. The normal seasonal response will be re-examined, as it is the permanent background for the complex changes that interact with the seasonal processes.

In a sense there are mainly two effects of air flowing over the ocean on energy exchange between hydrosphere and atmosphere. The more immediate of these is the effect of variation in wind strength on evaporation and turbulent heat conduction from the ocean. The influence of seasonally varying zonal wind systems is well known in this respect. Wind speed is of course an important factor in the comparatively rapid transformations of continental air masses moving out over water when the sea surface temperature differs very much from the air temperature a few feet above the sea. Such transformations generally take place in the westernmost

1. A more exact derivation of equations (8) and (11) in sections 3 and 10 can be found in a paper to be published in the Journal of Geophysical Research.

parts of the ocean, i.e. in middle latitudes. The associated energy exchanges will not be the objective of the investigations in the present report.

The other effect on energy exchange stems from variable wind stress giving rise to change in intensity of vertical mixing and in horizontal water transport and through it to change in heat flux divergence in the thermally active top layer of the ocean. Studies of surface temperature fluctuations induced by wind stress variations treat mainly effects in the immediate vicinity of or some distance from relatively narrow and rapid surface currents. As these occur mainly in markedly stratified baroclinic surface layers, small mass transport divergence can account for temperature changes of considerable magnitude. Surface cooling from upwelling is an example of such temperature effects.

The present investigation is mainly concerned with the dynamic causes of surface temperature anomalies in the wider open expanses of the major oceans. The problem is this: What are the evolution and longevity of anomalies that would arise from a response of heat advection in, and heat energy exchange at the surface of the top layer of the ocean to slowly variable wind stress, or more precisely, to non-seasonal fluctuations on a spatial scale commensurate with monthly averages?

2. Review of earlier work and scope of the present study.

The problem has been treated by, among others, Neuman and Pandolfo (1958) in an empirical study for a large part of the North Atlantic. Mechanistically based on Ekman's stationary drift current solution, their investigation gives evidence of wind induced, predominantly secular-scale changes in sea-surface temperature. The changes were attributed to variations mainly in the atmospheric zonal circulation at sea level to the extent of their being deducible from parameters of the large scale pressure distribution. The approach and some of the results are similar to those of J. Bjerknes (1959).

Using in part the data applied in the aforementioned paper by Neuman and Pandolfo, Pierson (1958) investigated the spectra and cross spectra of temperature anomaly[1] and of sea-level pressure gradient in two neighboring rectangular regions of the Northeastern Atlantic. He found from the lag relationships between atmospheric pressure gradient and ocean temperature that the part of the system containing statistically predictable fluctuations is in the longer periods, i.e. greater than a year, while the shorter period fluctuations "are only weakly connected from one part of the system to the other".

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[1] Anomaly in this reference is used in its oceanographic meaning, viz. departure from average over longitude.

However - and this qualification is worth stressing in the context of the present study - "there is some connection between temperature anomaly fluctuations in the two areas at the shorter period of two to three months."

It is not clear whether Pierson's conclusions are in agreement with results obtained by Roden (1960) from statistical analyses of sea level and temperature anomaly time series at selected stations along the Pacific east coast and elsewhere, but it seems that they might be.

Investigations of the above kind would indicate a significant effect of anomalous atmospheric circulations on the distribution of heat energy in the surface layers of the ocean. As to energy flux anomalies from the sea to the air, Arkhipova (1960) investigated for the North Atlantic the year-to-year fluctuations of monthly surface heat balance using data for the meteorological elements and cloud observations at weathership stations. Horizontal heat flux divergence in the active layer of the ocean was deduced as a residual between surface energy balance and heat storage in the layer which was estimated from hydrographic data. A connection of heat balance with the year-to-year variations of the wind-driven and thermo-haline circulations in the active layer was however not examined explicitly.

Studying synoptically the evolution of anomalous planetary-scale circulation patterns in the lower troposphere Namias (1959,

1960a, b) considered the implications of both effects, viz., of anomalous windstress-induced heat advection at the ocean surface and the transfer of heat energy to the atmosphere. Pattern-wise, the mutual influence is manifest in the behaviour, on one hand, of the pressure systems determining a class of motions highly relevant to foreshadowing of monthly and seasonal temperature and precipitation anomalies; and on the other, in the evolution of the sea surface temperature anomalies in the open expanses of the ocean. Namias invokes some kind of seasonal-scale interaction between baroclinic perturbations in the atmosphere and the circulation in the thermally active top layer of the ocean to account for certain phase shifts in the planetary wave patterns. In the prediction of the tracks of the disturbances and their associated fronts such phase shifts are critical, and are therefore a major consideration in seasonal forecasting. Applying the monthly sea-level anomaly pressure field to Ekman's formula for wind driven surface transport in the ocean he showed that in principle it was possible to explain salient features in the large-scale sea surface temperature anomalies in the eastern North Pacific as a manifestation of energy feedback from atmosphere to ocean.

In terms of current concepts of the atmospheric energy cycle (cf. Winston and Krueger, 1961) such feedback process would require that anomalous transfer of heat energy from ocean to atmosphere caused rates of generation of available potential energy in excess of or less than

the average (normal) seasonal amount. In turn, these anomalies must be reflected in a fractional excess or deficit of available potential energy for release in large scale cyclonic and anticyclonic development which then finally can effect, through the action of surface wind stresses, the distribution of temperature in the thermally active layer of the ocean. The latter process tends to restore, at least during certain favorable phases of seasonal change, a pre-existing pattern of sea surface temperature in the manner demonstrated by Namias (1957, 1959) synoptically for the Pacific Ocean.

In the present study an attempt is made to determine from data, as well as theoretically, the two major effects on the heat balance in the ocean from variable large-scale surface wind systems from which normal seasonal changes have been eliminated. The effects are attributed to the power of the wind systems:

(i) to induce through the action of surface stresses quasi-geostrophic EKMAN or SVERDRUP drift current anomalies and thereby to orderly redistribute heat in an oceanic region with marked meridional gradients of sea surface temperature;

(ii) to control aerodynamically the rate of evaporation and eddy conduction of sensible heat to or from the ocean.

A more detailed examination is made of the first effect only (sections 6, 9, and in other places). A background study of the similar influences of the normal seasonal change of windstresses on heat

balance (section 5) was deemed best to precede the treatment of anomalous changes.

The second of the mentioned effects is being treated mainly in its broad implications for the computation of anomalous vertical flux of heat energy from the ocean (section 7). Quantitative consideration, however, is given to this effect toward the close of the report dealing with the theoretical deductions of free and forced "thermal waves," their characteristic periods, occurrence of resonance with the forcing system of windstresses, and the application of the theory to data (sections 10 and 11).

For ease of style the qualifying terms "anomaly", "anomalous", etc. will be mainly omitted in the remainder of this section.

A quantitative assessment of heat balance in the layer of the ocean effected by the windstresses requires one also to determine the storage of heat in this layer whose depth will be time dependent. (Section 6 and Appendix I). Evaluation of heat advection and storage are the necessary steps for determining heating and cooling in the ocean by the thermodynamic energy equation method. Because (anomalous) flux of heat across the bottom of the mixed layer will be neglected as very small in comparison to advection; and since heat sources and sinks are practically non-existent <u>within</u> the ocean except in a generally very shallow surface layer absorbing short wave radiation, the vertical flux of heat at the surface ("surface heat balance") will be determined

as the sum of advection and storage. The methods for computing the latter quantities can be questioned in certain respects and this calls for a critical appraisal of the values obtained for surface heat balance and of their distribution in time and in longitude (sections 8 and 9).

So that unnecessary repetition of terminology will be avoided as much as possible, the formulation of the equations for thermal energy balance in the mixed layer, volume transport, heat advection, and storage as well as specifications of the adopted layer model (section 3) will best precede the discussion of the chosen regions, the data, their source and treatment (section 4).

The report concludes with a very brief discussion (section 12) of a proposal for a numerical computation of the evolution of sea surface temperature anomaly patterns from an initial state. This envisages hind-casting by means of a known forcing wind field, and prediction allowing for feedback of energy from the lower troposphere when the kinetic energy of the wind field is part of the computational output, with time step iterations of the order of days to weeks.

3. The equations of thermal energy balance anomaly in the mixed layer.

(a) Notations:

$T(z)$ Ocean temperature as a function of depth (z).

T_s Sea surface temperature (the subscript will be mainly omitted)

T_o Air temperature above the sea

e_s Saturation vapor pressure at sea surface temperature

r	Relative humidity
R	Bowen ratio
H	Eddy flux of sensible heat at sea level
E	Evaporation from the sea
R_0	Net radiation at the sea surface
V	Current velocity (components u, v)
W_0	Wind velocity above sea
$\mathcal{F}$	Volume transport through depth of mixed layer
τ	Surface wind stress
γ^2	Drag coefficient
D	Mixed layer depth
$\varepsilon \equiv \frac{\partial \rho}{\partial T}$	The negative of the thermal expansion coefficient of sea water at a salinity of 35‰
a, b	Constant and coefficient of regression for $\log D$ on T_s

Subscripts:

N	Long-term (normal) monthly average
x, y	on stress τ denote eastward, northward components
y, t	on the function Φ denote partial derivatives with respect to latitude, time
$*$	Departure from a longitude average
a	referring to the air, as in c_{pa}, ρ_a

Superscripts:

Horizontal bar: monthly average (mainly omitted except when required for clarity)

Prime: Departure from normal monthly value (D.N.).

Other symbols have their usual meaning unless explained in text.

(b) The basic equation for the mixed layer.

The thermal energy balance equation for the mixed layer may be written as follows:

$$R_0 - H - LE = \rho c_p \left(\frac{\partial (D T_s)}{\partial t} + \vec{\mathcal{F}} \cdot \nabla T_s \right) . \qquad (1)$$

The assumptions underlying this equation are permissible in the case of its being applied to monthly mean balance anomaly, i.e., after subtracting out the long-term (normal) monthly balance. They are:

(1) Energy changes caused by non-seasonal accumulation or depletion of salinity and by precipitation cooling are negligible.

(2) Heat transfer across the bottom of the mixed layer (secondary thermocline) is also negligible. The transfer is of importance in the long term annual heat balance of the oceans and possibly in the normal seasonal processes. In regard to these it has also been found (Bryan, 1962) that a toroidal meridional heat transport of the species $\int v(z)\, T(z)\, dz$ and its convergence contribute substantially to the heat balance of oceans on a global scale. The toroidal term has been omitted in equation (1).

(3) The effect of wind stress (D.N.) on drift current penetrates to depths not exceeding the depth of the thermally mixed layer.

(4) An eddy term of the species $\overline{\tilde{\mathfrak{F}}''\cdot\nabla T_s''}$ is disregarded so that in equation (1) the original advection term $\overline{\tilde{\mathfrak{F}}\cdot\nabla T_s}$ has been set equal to $\overline{\tilde{\mathfrak{F}}}\cdot\nabla\overline{T_s}$. (Here a double prime denotes departure of the daily from a monthly mean value).

The terms on the r.h.s. of equation (1) - or rather its D.N. equivalent - are evaluated from monthly and longitude-sector averages (c.f. section 4) of the meteorological and hydrographic elements. Consequently an effect on heat advection and storage from correlation in longitude between temperature-layer depth and current speed on the meso-scale is excluded. However, analyses of comprehensive data for the North Atlantic Ocean have established the existence of quasi-meridionally oriented bands of thermally differentiated water with significantly varying layer depth, having dimensions 20-60 Km in width, 500 Km or more in length. Surprisingly stable correlations with current velocity suggest long persistence of these bands (Perlroth, 1962) which is in agreement with strong indications of quasi-geostrophic balance prevailing in the long-term seasonal averages of the current component along the bands (Gibson, 1962). It is thus quite possible that these oceanic standing eddies on a meso-scale would form an integral part of the mechanism for secular-scale meridional heat transport in the oceans. But it is unlikely in view of their nature that they bear significantly upon the evolution of the large-scale <u>anomalies</u> of heat advection.

It will be understood without mathematical language that all terms in equation (1) are taken as averages over a specified latitude zone. Some details of computation procedure are given in section 4. It remains here to mention in this connection an additional slight simplification of the layer model (see Fig. 1): The spatial variation of layer depth within this zone may be regarded as sufficiently small for approximating the integral $(y_2 - y_1)^{-1} \int_{y_1}^{y_2} \int_D^0 (V \cdot \nabla T_s)\, dz\, dy$ by the integral $\int_{D_m}^0 (V \cdot \nabla T_s)\, dz$ where D_m is the mean layer depth over the width of the zone $(y_2 - y_1)$ (Fig. 1). The subscript m will be omitted for brevity's sake. Similarly the quantities $H', D', T_s', \mathcal{F}'$, etc., will be zone-width averages and thus functions of time and longitude only.

(c) The meridional mass transport and heat advection (D.N.) terms.

On integrating over the width of the zone a meridional component of heat advection $(\propto \mathcal{F}_y \frac{\partial T}{\partial y})$ would be equal to a differential transport $(\propto \mathcal{F}_y T)$ between the northern and southern boundaries of the zone. In fact the normal temperature gradient ∇T_N to a sufficient approximation may be set equal to its meridional component. It is then permissible to neglect as small order terms $\mathcal{F}_{Ny} \frac{\partial T'}{\partial y}$, $\mathcal{F}_y' \frac{\partial T'}{\partial y}$, $\mathcal{F}_x' \frac{\partial T'}{\partial x}$ and retain merely the term $\mathcal{F}_y' \frac{\partial T_N}{\partial y} \equiv \mathcal{F}_y' \Phi_y$.

As we are interested in the response of temperature distribution to anomalies of monthly average wind stress, inertial current activity with characteristic periods of the order of a half pendulum day can be ignored. The meridional mass transport anomaly is then given by the Ekman quasi-geostrophic approximation:

$$\rho \mathcal{F}_y' = \int_D^0 \rho v_g\, dz - \frac{1}{f} \tau_x' \quad , \tag{2}$$

The quantity on the l. h. s. may also be deduced from the filtered (vorticity) equation following Sverdrup (1957). The integrated meridional volume transport from the surface to the bottom of the ocean and from coast to coast is proportional to the (vertical component of) wind stress vorticity providing bottom stresses can be neglected. This is certainly permissible in regard to monthly stress anomalies which in middle latitudes are many times greater at the surface than at the bottom. However, on account of mass continuity the total integrated meridional mass transport must vanish (except for a third order residual). Therefore, at any latitude in a wide ocean basin, the transport per unit length at any particular longitude is given by $\beta^{-1}(\nabla \times \tau_*)$; and the transport anomaly accordingly by $\beta^{-1}(\nabla \times \tau_*')$. Thus for (2) we have the alternative equation:

$$\rho \tilde{\mathcal{F}}_y' = \frac{1}{\beta}\left(\nabla \times \tau_*'\right) - \text{SUBLAYER TRANSPORT}. \tag{3}$$

The layer depth being variable, internal Rossby-type waves on the thermocline with periods of the order of a year are included ab initio. It has been shown by Veronis (1956) that the fraction of kinetic energy imparted to the ocean by windstress that is absorbed by non-geostrophic motions in the ocean is negligibly small for periods much longer than a year. Although the mentioned internal waves are largely geostrophic, meridional heat advection anomaly associated with these waves must be negligibly small owing to the small horizontal temperature gradients associated with the slope of the thermocline away from fast boundary

currents. The term for heat advection D.N. may then be written as:

$$c_p \left(\rho \, \tilde{\mathcal{F}} \cdot \nabla T \right)' = c_p \left[\int_D^0 \rho \, v_g' \, dz - \frac{1}{f} \tau_x' \right] \Phi_y \tag{4a}$$

or on account of eq. (3)

$$= \frac{c_p}{\beta} \left(\nabla \times \tau_*' \right) \Phi_y \quad . \tag{4b}$$

since the meridional temperature gradient in the sublayer is of smaller order of magnitude then the gradient in the mixed layer (Φ_y).

A geostrophic term in equation (4a) requires the zonal pressure gradient D.N. integral to be finite. Since the current anomaly has been assumed to vanish at $z = D$, zonal sea level gradients must be established. Ignoring the effect of mean monthly atmospheric pressure D.N. - it can not be ignored in individual storms - the sea level anomaly $\eta' \left(\equiv z_0 - z_{0N} \right)$ can be expressed as a function of T'; for we may write:

$$\int_D^0 v_g' \, dz = \frac{g D}{f} \frac{\partial z_\alpha'}{\partial x} \tag{5}$$

where z_α is the steric sea level anomaly defined by

$$z_\alpha = g^{-1} \int_{p_s}^{p_D} \alpha \, dp$$

α being the specific volume anomaly in the oceanographic terminology. Consistent with the basic assumption (1), section 3b, it is permissible to regard specific volume anomaly as a function only of temperature anomaly, whence:

$$\int_D^0 v_g' \, dz = \frac{g D^2 \varepsilon}{f \rho} \frac{\partial T'}{\partial x} \tag{5a}$$

where ε is the coefficient of thermal expansion (see list in section 3a).

(d) Heat storage D.N.

At northern middle latitudes the seasonal variation of z_N is known to be smaller than the seasonal variation of D_N by a factor of 100. (cf. Patullo, Munk, Revelle, and Strong, 1955). This makes it possible to eliminate the anomaly D' as dependent variable from the analytical expression for heat storage anomaly. In the present mixed layer model D' may be expressed as a function of T' or τ'. Analysis of available data for the North Pacific (see Section 4) has shown that the seasonal variation of layer depth can be represented by a regression

$$\log D_N = a + b T_N \equiv b \Phi \quad . \tag{6}$$

A similar regression can be deduced for the seasonal dependency of D_N on τ_N . Here the temperature regressions are being applied. Details are given in Appendix I. It should be emphasized that a theoretical relationship rather than an empirical regression could serve the purpose of the following calculations just as well or better. There could be no difference, qualitatively, in the results.

It is now reasonable to assume that D' is related to T' in a similar manner as layer depth to temperature in their seasonal interdependence; for the variations of the large-scale T' field are on a time scale comparable to the seasonal cycle. For simplicity equation (6) will be considered valid for the D.N. values and thus we are in a position to express heat storage

anomaly as a function of T' as dependent variable and of the parameters

T_N, $\frac{\partial T_N}{\partial t}$, (or Φ, Φ_t), a, and b.

The result is:

$$\frac{\partial (DT)'}{\partial t} = e^{b\Phi}(b\Phi - a + 1)\left[e^{bT'}\left(\frac{\partial T'}{\partial t} + \Phi_t\right) - \Phi_t\right]. \qquad (7)$$

(e) "Tendency" equation for sea surface temperature anomaly.

Substituting equation (5) in (4a) and subsequently equations (4a) and (7) in the D.N. equivalent of equation (1) one finds upon solving for $\frac{\partial T'}{\partial t}$:

$$\frac{\partial T'}{\partial t} = e^{-bT'}\left[\Phi_t + \frac{C}{f\rho}\left(\tau_x' - G\frac{\partial T'}{\partial x}\right) - \frac{C}{c_p \Phi_y} B'\right] - \Phi_t \qquad (8)$$

where $C = \frac{\Phi_y}{(b\Phi - a + 1)e^{b\Phi}}$, $G = g\varepsilon e^{2b\Phi}$,

$B' = H' + LE' - R_o'$.

As here defined, B' represents the heat balance (or energy flux) anomaly at the ocean surface counted as positive when there is an excess of net energy transfer from ocean to atmosphere over and above the "normal" value. The second term in the square bracket can be replaced by $C\rho^{-1}\beta^{-1}(\nabla \times \tau_*')$ according to equation (4b).

4. The region, the data, and their treatment.

The relevant feature of normal seasonal cycle of layer heat balance and the evolution of the departures from normal are examined for ten-degree broad zones centered on 40° N in the Atlantic and Pacific oceans. The selection of this zone within the middle latitude belt is in part dictated by data availability but also on the grounds that it mostly contains the

largest poleward gradient of sea surface temperature, and so satisfies the simplifying assumption in section 3c.

This temperature gradient is however larger in the western than in the eastern part of the ocean. Moreover, the gradient of time-averaged temperature in individual months and thus also the D.N. patterns would be sensitive to slight shifts in the eastward extensions of the major western boundary currents, i.e., the Kuroshio and Gulf streams. Yet this effect is largely eliminated in the D.N. temperature longitude-time section (see figure 2) on account of the averaging over the width of the zone; and will be of no consequence for the calculations of meridional volume transport and heat advection as these are performed on a sufficiently coarse grid. Furthermore, the wind stress components represent themselves spatial averages over five-degree latitude longitude squares.

For the most part, the variables and parameters are monthly averages, though in some instances, as e.g. for T' in the Western Pacific, these were formed from consecutive ten-day means. In other instances, as e.g. for the normal stresses in the Pacific, monthly averages were obtained by judicious interpolation of seasonal averages.

In the following are listed the kind of data, their source and the essential steps in computational procedure are outlined. Letters in bracket refer to the appended DATA SOURCE.

SEA SURFACE TEMPERATURE

Normals: These are monthly climatological means [A]. Meridional gradients are evaluated from grid points at 37.5° and 42.5°N.

D.N.: The departures are examined for the Pacific zone only and for the period August 1959 through September 1960. For longitudes 180° and eastward the "raw" data are taken from [B], west of 180° from [C] in combination with [A].

SURFACE WINDSTRESS

Normals: Zonal and meridional components for the Atlantic are taken from [D] and [E], for the Pacific from [E]. Average of zonal components obtained from values at 37.5° and 42.5°N. Values of both zonal and meridional stress components at these latitudes and at longitudes five degrees apart are used in the computation of stress vorticity. Coriolis parameter and β parameter are set equal to their respective values at 40°N.

D.N.: The anomalies of the zonal component of wind stress are evaluated from Normals in [E] and from the meridional EKMAN transports for individual monthly periods tabulated for the Pacific by Fofonoff [G]. To determine stress vorticity D.N., the vorticity of the normal stress vector is calculated by the finite difference method; while the individual monthly values were again taken from [G] but in this case from the meridional Sverdrup transports. Adjustment for mass continuity is performed by deriving the departures $\nabla \times (\tau')_* \equiv \nabla \times \tau_*' \equiv (\nabla \times \tau_*)'$,

the order in the steps of computation thus being arbitrary.

It should be noted that the stress vorticity field in [G] is based on the monthly mean geostrophic windfield derived from historical weather maps while the normal (seasonal) vorticity field has been here derived from wind stresses based on climatological wind roses. In view of the fact that the heat advection D.N. term, $\beta^{-1}(\nabla \times \tau_x')\Phi_y$, is proportional by definition to the covariance of $(\nabla \times \tau')$ and Φ_y with respect to longitude (cf. section 3), any difference between the method in [E] of computing τ_N and the method in [G] of computing τ will be of little consequence for the advection D.N. terms unless the discrepancy in stress vorticity D.N. due to that difference alone were correlated with Φ_y. Because this seems highly unlikely there is some assurance that the computed D.N. advection values are the best possible estimate of the true values.

A check on the effect of a certain degree of arbitrariness in the assumptions underlying the original computations of normal wind-stress in [E] on the derived distributinn of vorticity with longitude would be afforded by comparing with other sources such as [D] or [F]. However, for the Pacific ocean Hidaka's seasonal means in [D] are identical with the mid-season (monthly) mean stresses assessed by Scripps Oceanographic Institution [E]. Priestley [F] computed only zonal stresses. From the grid point values, kindly made available by him, the present author calculated the correlation with respect to longitude,

between zonal shearing vorticity $\frac{\partial W_x'}{\partial y}$ derived from [D] and [E] with that derived from [F]. The correlation coefficient turned out to be no larger than 0.41 (in April) and as low as 0.35 (in July). However, it should be remembered that small differences in the angle of the stress vector τ with the x-axis (west-east direction) might readily entail large differences in $\frac{\partial W_x'}{\partial y}$.

SEA LEVEL

The "geostrophic contribution" to the annual oscillation of meridional heat advection is assessed for middle latitudes of the Atlantic (see section 5). As already discussed in section 3(c) equation (5) the geostrophic volume transport is proportional to the gradient of steric sea level. Data for the annual variations of the computed steric and actually recorded sea level are taken from [H].

LAYER DEPTH

Normals: In connection with the regression equation (6) in section 3 (d), the seasonal cycles of D and T_s were examined from data contained in [I] and [J]. The slope of the seasonal regression during spring and summer was also compared with the diurnal cycle in [J]. Lumby [M] investigated the depth of the wind-produced homogeneous layer in the oceans applying the formula by Rossby and Montgomery (1935). Examination of "observed" layer depth for checking on the derived values was made by him only for the eastern North Atlantic. However, for the purpose of the present study computed values were

taken off the isopleths for the Pacific Ocean. The seasonal regressions are given in Appendix I.

SEA-AIR TEMPERATURE DIFFERENCE (see Section 7)

Some climatological atlases contain maps showing the distribution of this quantity separately for months or seasons. These maps are derived by means of graphical subtraction of the separate fields, a method that gives an all too crude picture of the seasonal cycle of the difference field.

For the seasonal regression of mean monthly air temperature just above the sea on 1000-700 mb atmospheric layer thickness (cf. Appendix II), data sources [K] and [L] were used. Gridpoint values of thickness anomalies were extracted from tabulated data sheets filed in the Extended Forecast Branch [N].

5. Differential heating and the annual cycle of meridional heat advection.

An examination of the annual cycle of heat advection in the ocean provides a means for estimating the relative magnitudes of the forcing stress term and the internal geostrophic term in equation (4b). The latter (D.N.) quantity is very difficult to compute from data.

The magnitude of the geostrophic term plays a part in the problem of response time to wind stress activity and is critical in the occurrence of resonance (cf.Section 10). It has therefore also a bearing on energy feedback. In the following we shall examine the magnitude of differences, with respect to longitude, in amplitude and phase of the annual cycle of

heat content in the layer of seasonal penetration. The effect of such differential heating is to set up large-scale zonal pressure gradients and corresponding geostrophic meridional mass transport. At the bottom of the layer of seasonal penetration the annual variation of that transport will become imperceptible and in analogy to equation (5) the mean geostrophic volume transport through the depth of the layer can be determined from the steric sea-level gradient. In this case a prime will denote departure from the annual mean (D.A.) and D the depth of the layer of seasonal penetration.

Similarly one may apply equation (4b) to compute directly the total meridional advection D.A. The computed values of the latter and the Ekman advection terms are likely to have systematic errors due to the errors in the wind stresses. To resolve this difficulty it is necessary to go back to the complete thermal energy balance equation, equation (1), but re-written to be applicable to D.A.

Average D.A. of surface energy balance B and of the seasonal heat storage in the considered zone are estimated from station values calculated by Patullo (1957) who investigated the annual cycle of these quantities from observations and measurements at weather ship and oceanographic locations in the North Atlantic and North Pacific oceans. Comparison will be also made with results of an investigation of seasonal heat storage in the Atlantic by Bryan and Schroeder (1959) and of the earlier cited study by Arkhipova.

Calculations of the wind drift advection (Ekman, Sverdrup) terms are made for the Atlantic in accordance with the procedure outlined in section 4. For the calculation of the geostrophic advection term, departures from the annual mean of steric sea level at suitably located oceanographic stations and weather ships are applied (DATA SOURCE [H]).

The results of these computations, and of collations of results derived by other authors, are set out in Table 1. It will be first noted that the use of one or the other of two sets of derived wind stresses for means (row 2) and vorticities (row 1) makes little difference on the whole. It is true that both Hidaka's [D] and the Scripps [E] assessments for the Atlantic are based on the same values of drag coefficient (6 and 20.10^{-3} for low and high wind speeds respectively, the threshold wind speed being 6.6 m sec^{-1} in both cases). But the working up of components from wind frequencies is not altogether standard procedure.

Rows 3 and 4 give the heat advection by geostrophic transport expressed once as difference of advection by Sverdrup and Ekman drifts, and computed subsequently from steric sea level gradients. There is satisfactory qualitative agreement. In view of the comparable although somewhat less satisfactory agreement of heat advection deduced from surface heat balance and layer storage according to Patullo's analysis (row 9) with heat advection computed directly (row 1), the relative magnitudes of geostrophic and Ekman-drift induced heat advection in the seasonal cycle can now be assessed.

The ratio of amplitude of computed geostrophic into total meridional heat advection (rows 4 and 1) is about 0.2 to 0.3. No comparison should be made in this respect between rows 3 and 1 because the geostrophic drift values in row 3 are computationally dependent on the Sverdrup drift values in row 1.

There is a feature in the phase of the computed geostrophic advection cycle (row 4) worth mentioning. The maximum falling into the late spring period is quite marked. It is seen from the rows farther below that the layer heat content attains its minimum in mid-spring at which time net radiative absorption of heat in the top layer is still negligible (row 5). The sharp warming during late spring is therefore mainly advective (row 9) and owing to the phase difference in layer heat content between western and eastern parts of the ocean (cf. Patullo et. al., 1955; Patullo, 1957) ensuing zonal pressure gradients favor comparatively large geostrophic transports in late spring.

Attention might be drawn too to the excellent agreement between cumulative heat storage values (from late winter to late summer) obtained by Patullo and by Bryan and Schroeder (1959), respectively. Hydrographic data used by Arkhipova (1960) were clearly not adequate.

6. Monthly anomalies of heat storage and advection in North Pacific middle latitudes, September 1959 through August 1960.

(a) Qualitative considerations.

This particular period has been selected because it coincides with

one treated in a wider Extended Forecast Branch project aiming at the synoptic characterization of the Earth's surface thermal properties on a hemispheric scale. However, no attempt is made here to relate explicitly planetary-scale circulation patterns and their concommitant weather systems (temperatures and precipitation anomalies) to the oceanic anomalies.

The sea surface temperature D.N. field in a Hovmöller diagram as described in Section 4, is shown in Fig. 2. The patterns during autumn and winter are surprisingly definite and suggest the occurrence in Mid- and East Pacific of slowly progressive thermal waves with lengths of the order of 5,000 km. and period of the order of six months. Somewhat shorter and more rapid disturbances are indicated in the western portion of the diagram, i.e., to eastward of the Kuroshio current.

In the eastern North Pacific during spring and summer, wave features are weakly developed and suggest retrogression rather than progression of thermal systems.

In Figure 3 the negative of the zonal wind stress anomaly is shown in the same way but for autumn and winter only. (Multiplying it by $c_p f^{-1} |\Phi_y|$ the corresponding induced field of oceanic warm advection D.N. is obtained).

If surface heat exchange and radiation processes were negligible, warm advection and heat storage in the mixed layer would be equal to one another. Therefore, even in such hypothetical conditions, the fields in Figure 2 and Figure 3 would not be correlated. However, were one to omit

for a first approximation layer depth variations, the temperature anomaly "wave" and monthly storage "wave" would be 90° out of phase, with equal amplitudes. The T' and τ_x' patterns in Figures 2 and 3 have one distinct feature in common. This is their time and longitude differentiation. To this extent their resemblance lends qualitative support to the hypothesis forwarded by Namias (1959) concerning the dominance of wind-stress effect in the generation of the observed large-scale sea surface temperature anomalies (see section 2). Namias calculated surface advection D.N. using Ekman's empirical surface drift formula. This implies the addition of a zonal advection term $u_s' \frac{\partial T_s}{\partial x}$ to the primary term $v_s' \frac{\partial T_s}{\partial y}$. The total surface advection is routinely evaluated in the Extended Forecast Branch for the eastern North Pacific. For the period in question calculations were extended in the present study to 150°E. Although in principle the patterns of surface advection should be closer to the true ones than the patterns resulting from the omission of the zonal advection term, no purpose would be served here to pursue this question; for in considering the heat balance of the active layer one cannot apply Ekman's empirical surface drift formula.

A qualitative appraisal of the effect of surface wind anomalies on redistribution of heat in the open ocean expanses is afforded by comparing heat storage D.N. with total (Sverdrup species) advection D.N. The quantities were calculated from equations (4b) and (7) from the data and by the method explained in section 4. They are shown in the diagrams Figures 4 and 5.

A study of these patterns leads to the following observations:

(i) The range of variations, in both time and longitude, of storage is considerably less than the range of variations in advection.

(ii) Progressive waves during autumn and winter and retrogressive or stagnant patterns during spring or summer are evident in both storage and advection.

(iii) There is no indication of any but small contemporary or lag correlation between these quantities. Any such correlation could be readily attributed to autocorrelations in the time series of each of them similar to most geophysical time series. It would be futile to furnish here numbers since interpretation would be of a low (not a high) correlation coefficient.

In view of the validity of equations (1) or (8), the point under (iii) above leads to the conclusion that anomalies in surface heat balance (B') must be of the same order of magnitude as either of the remaining components of the heat balance of the mixed layer, i.e., storage and advection. Such "constitution" of heat balance is of course a necessary although not sufficient condition of energy feedback processes mentioned in the Introduction.

The field of B' can be assessed from the thermal energy equation by determining values of the sum of storage D.N. and layer advection at points in the λ, t domain. However, it is known (Budyko, 1958) that the normal monthly or seasonal budget of surface energy flux over much of the North Pacific area during spring and summer is a delicate balance

between evaporation and turbulent heat conduction on one hand and net radiation on the other. It is thus represented by a comparatively small residual between two large energy fluxes. At somewhat lower than the average latitude considered here (40°N) such delicate balance obtains during much of the year. It has been also demonstrated by Fritz (1958) that the heat energy added to the atmosphere (except for the usually small eddy conduction effect) is through much of the year a small residual of large terms when it is evaluated from the heat balance of the system Earth-atmosphere and the storage of heat in the ocean.

If for some reason the field of B' as calculated by the thermodynamic energy equation method was also represented by small residuals of storage and advection anomalies the accuracy of the calculated B' would be much in doubt. This problem will be treated from two different angles in the following subsection and in section 7.

(b) Annual range of heat advection and storage (Comparison between Normals and D.N.).

The observation mentioned in point (i) above requires to be examined in some detail. We shall first compare the ranges (through the period under investigation) of the calculated anomalies of heat storage and advection in Figures 4 and 5 with the annual ranges of long-term (normal) storage and advection. Normal meridional heat advection $c_p \beta^{-1}(\nabla \times \tau_{*N})$ has been computed as part of the D.N. ($\propto \nabla \times (\tau_* - \tau_{*N}) = \nabla \times \tau_*'$). Storage D.N. has been directly calculated from T_s' values with the aid of normal seasonal regressions of layer depth on sea surface temperature

(equation (6) and Appendix I.)

In order to see whether the calculated annual ranges of normal advection at various longitudes in the selected Pacific zone are realistic, they will be compared with values derived independently from annual ranges of: (a) normal heat storage based on bathythermograph data, and of (b) normal surface heat balance assessed from meteorological and hydrographic observations. A comprehensive investigation of seasonal variations of heat storage and surface balance in the North Pacific has been made in the earlier cited Ph. D. thesis by Juno Patullo (1957) who based it on data from weathership locations and oceanographic stations.

Table 2 has been compiled from Patullo's monthly values and first harmonics of the annual march. Values to the west of 150°E have not been included on account of their being unrepresentative of large-scale conditions in the considered zone.

In Table 3 are set out annual ranges of normal meridional heat advection at longitudes as indicated. The values in the fourth column are deduced from Table 2 and those in the following columns are computed from the annual wind stress vorticity and meridional gradient of sea surface temperatures. The next column gives the range of heat advection anomaly in the period under study and this can be gleaned also from Figure 5.

The first of the two columns under "heat storage" in Table 3 contains the corresponding zone-width averages of the normal values in column 4 of Table 2; while the last column in Table 3 gives the range of storage anomalies during the period under study (calculated from equation (7)). This range can also be seen in the diagram Figure 4.

It will be seen from Table 3 that there is good agreement between the two sets of independently assessed values for the normal advection range. In either case a marked decrease of the annual range occurs from the western to the central Pacific. It is obviously linked with the strongly contrasting zonal distribution of storage and surface balance as will be readily seen from Table 2. The former quantity falls off sharply east of 165°E especially in the lower latitudes whereas the latter remains more or less unaltered.

If we now turn our attention to columns 6 and 8 of Table 3 we may amplify the statement under (i) in the preceding subsection. Along the entire Pacific zone (except at the western and eastern extremities) the values calculated for advection D.N. undergo in the considered period far larger changes than the storage D.N. values, the ratio between them being 0.48 from the averages and 0.45 as average of the ratios. From columns 5 (or 4) and 7 it is apparent that with respect to D.N. the normal seasonal variations are reversed, the annual range of advection being much smaller than the annual range of storage. Forming the inverse of the above ratios they turn out to be 0.48 (or .45) and 0.41 (or 0.39) respectively.

It is not permissible to generalize on this striking result since the data sampling underlying the calculations of advection and storage D.N. is rather limited. Nevertheless it is felt that it will qualitatively stand up to testing on larger independent sampling.

These heat budget considerations are significant for atmosphere-ocean interchange processes. Referring back to Table 3, a measure of the range in the anomaly of surface heat balance, i.e. of net energy exchange at the ocean surface, is afforded by the difference between the values in columns 6 and 8. However, this range will be much less uniformly distributed with respect to longitude than the normal seasonal range (column 8 of Table 2). Again, these marked variations with longitude in the anomaly of energy exchange (B') are attributed mainly to the variations of the anomaly in heat advection rather than in storage. (No such distinction can be made in regard to the normal seasonal range since as was pointed out it is rather uniform in the considered longitudes. In fact the standard deviation of B_N' among the seven stations listed in Table 2 (column 8) is 12% of its mean value.)

A close study along these lines of anomalous processes in the mixed layer of the ocean in middle latitudes demands comprehensive and homogenous observational data. Although it could well stop short of spectrum analysis, it must be concerned with the phases of the dominant wave-like perturbations.

7. Anomaly of surface energy exchange and sea-air temperature difference.

Magnitude, phase and spatial distribution of non-seasonal (D.N.) variations in the large scale heat balance at the sea surface are the prime considerations in atmospheric feedback of energy of the kind mentioned in the introduction. Positive energy feedback must be invoked in order to account for the longevity of concommitant anomalies in the planetary circulation patterns of the atmopshere such as described by Namias (1957, 1959).

As mentioned in the preceding section surface heat balance anomaly (B') can be determined by the thermodynamic energy equation as the sum of layer advection D.N. and layer storage D.N. It has been shown earlier that with respect to the maximum range surface balance anomaly is not represented by a small residual of storage and advection anomalies. One may now relate B' to physical and aerodynamic parameters controlling anomalous (upward) flux of sensible heat and latent heat. These parameters are obviously D.N. of windspeed, W', of sea-air temperature difference, $T_s' - T_o'$; and of the difference of saturation vapour pressure of the air at sea temperature and at its actual temperature. A simplification is introduced by ignoring anomalies in Bowen ratio. It is not pretended here that such assumption is a priori justified. The approach is therefore a heuristic one.

Ignoring these anomalies in Bowen ratio the temperature parameter will control the anomalous flux of the sum of sensible and latent heat except for the aerodynamic wind effect. For the present purpose we shall not examine this effect as it is independent of the radiation anomaly.

However, absorption of short-wave radiation in excess or deficit of normal amounts - in a generally very shallow surface layer whose thickness can be ignored in comparison to the mixed layer depth - will affect the sea temperature anomaly T_s'. Thus by relating B' to $T_s'-T_o'$ it will be possible to infer on the relative magnitude of radiative and non-radiative parts in the surface energy flux.

Mean monthly values of surface air temperature anomaly T_o' are derived from 1000-700 mb layer thickness anomaly. The method and seasonal regressions used in this procedure are given in Appendix II (for data source see section 4).

In the scatter diagrams Figure 6 and 7 surface heat balance anomaly is shown as a function of sea-air temperature difference anomaly for the periods September 1959 through February 1960 and March to June 1960 respectively. It will be seen that B' changes linearly with $T_s'-T_o'$ although there is of course a wide scatter but this relationship is reversed as between the two periods. During autumn and winter B' increased with increasing $T_s'-T_o'$ in accordance with the heat transfer law. Within the limits of the scatter of points in figure 6 this relationship suggests that both anomalous absorption of short-wave radiation and

anomalous "nocturnal" (long-wave) radiation are small effects in the total over-all anomaly of energy flux at the sea surface.

One would expect the opposite rôle of short-wave radiation in the anomalies of energy flux during spring and early summer when the normal upward fluxes of sensible and latent heat are small and even reversed in places (downward flux). Such reversal is indeed observed as already pointed out, for there is a decrease of B' which means an increase in anomaly of downward directed energy flux - with increasing $T_s' - T_o'$.

D.N. absorption of radiation would require a negative correlation of B' (through R_o') with $\frac{\partial T_s'}{\partial t}$. However, in analogy to the considerations of heat storage and advection in section 5 (a) this will also imply a correlation of B' with T_s'.

8. The magnitude of anomalies in large-scale surface heat energy exchange.

(a) In relation to synoptic regimes.

Little is known from the literature on the magnitude of the non-seasonal variations in the large-scale heat balance at the ocean surface. Recently Petterssen, Bradbury, and Pedersen (1961) made a synoptic study of the various components of energy fluxes at sea level over the Atlantic Ocean during the period November 1956 through March 1957. These fluxes were computed for various air streams and air masses associated with the major sea-level depressions separately for different phases of the storms, ranging from the open warm-sector cyclone

through occluded depressions to the storm centers that are surrounded on all sides by air masses of polar origin. In order to assess here the magnitude of the departure from normal of the flux of sensible and latent heat, and of the net heat balance including radiative flux, values at weathership locations were extracted from the isopleth maps. D.N. values were then obtained by subtracting long-term averages of the representative quantities. For the latter, two sets of data were used, viz. from tabulations of monthly long-term averages contained in the earlier cited study by Patullo (1957) and by forming ten-year (1948-57) averages of monthly means for individual years published by Arkhipova (1960). The results of this data collation are indicated in the top right corner of Figure 6.

Numbers denote the latitude of the appropriate weather ship locations. The D.N. referring to the quantity printed below the vertical lines are shown by short horizontal lines against the scale of the scatter diagram to the left. Vertical lines (for the heat balance) connect D.N. levels calculated from the different sets of Normals whose source is shown on the top. It will be seen that the difference between them is generally much less than the average difference between the characteristic regimes.

It should be permissible to compare the range of (positive) D.N. of monthly averages for the Pacific with those of the Atlantic composite

storm samples in the same latitude zone. Although in both cases the data sampling is rather limited, the following conclusions can be drawn from such a comparison for the autumn and winter period:

(i) Anomalous (greater than normal) upward flux of energy at sea level on a monthly scale as computed here from storage and advection is comparable in magnitude with the average excess of upward flux in the warm sector regions of depressions (where it is smallest), but it is considerably less than excess in deep polar air streams.

(ii) Deficit of flux on the monthly scale has the same order of magnitude as excess.

(iii) Judging from the proportion of energy flux anomaly by anomalous evaporation and heat conduction combined, in the total energy transfer (surface balance) the effect of anomalous net radiation at sea level on the monthly surface heat balance is rather small although apparently not negligible. (Compare the vertical ranges in the right part of Figure 6).

Turning to Figure 7 for the spring to early summer period there is a suggestion of a slightly reduced range in the computed heat balance anomalies. No comparison can be made in this case with synoptic regimes.

The question also arises as to how large is the range of departures from normal in surface energy transfer in comparison to the average

seasonal level of normal transfer. The dashed horizontal lines in the scatter diagrams in Figures 6 and 7 indicate the lowest and highest values of normal transfer from among those obtained at weather ship locations in Table 2. These were obtained from individual monthly values given by Patullo for the periods September through February and March through June respectively. Evidently the range of anomalies is of the same order of magnitude as the normals, viz. 10 units (10^3 cal $cm^{-2}month^{-1}$). However, the normal range is smaller than the anomaly range.

(b) In relation to heating in the lower atmosphere.

It is of considerable interest to reconsider the findings in (a) above in the light of normal and D.N. of heating in the lower troposphere. Autumn-winter period. Clapp (1961) investigated the distribution over the Northern Hemisphere of normal winter heating between sea level and 500 mb, applying both the thermodynamic energy equation and heat balance methods. For the Pacific zone under consideration one finds from the chart for the distribution determined by the first method (Figure 1 of Clapp's paper) values ranging from -3 to +15 in the here adopted units; positive heating occurring mainly west of 180° long. The heat balance method (figure 5 of the cited paper) gives a corresponding range -1 to +11 units. It is evident from other maps in Clapp's paper that over oceanic heat source regions long-wave radiation cooling is only a small fraction of total heating in the layer; and since the vertical heat flux at 500 mb must also be rather small in comparison we note a satisfactory

measure of agreement between Patullo's Normals and the values of surface heat balance implied in the distribution given by Clapp. It should be mentioned that he also gives separately the distribution of precipitation heating and sensible heat transfer from the oceans but it does not seem justified to make here a more detailed comparison.

Spring Period. Aubert and Winston (1951) evaluated the normal "horizontal heating" by the thermodynamic energy equation method in the layer sea-level to 10,000 ft. This heating component accounted for most of the total heating. We need not here consider their autumn and winter distribution over the Pacific. The normal heating during April in the considered zone is predominantly negative (except near the west coast of the American continent) and the zone-width average varies from +0.5 to -3 in the above units. The range of normal surface heat balance in the same region is -1 to +5 (dashed horizontal lines in Fig. 7). These figures would imply that in spring the ocean gains more heat than the lower atmosphere loses which is consistent with the circumstance that the atmosphere in addition loses by long wave radiation whereas the ocean in addition gains by net radiation at the surface. [1]

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[1] It is however significant that heat storage in the ocean in spring exceeds heat gain by energy exchange at the surface, the residual being provided by advection due to ocean current activity. This is clearly brought out in the seasonal variations of the relevant quantities in the Atlantic middle latitude zone. Referring back to Table 1 the average rate of monthly heat storage (east of 65°W) amounts to 10 units but the surface heat balance is only 4 units (putting the annual mean to approximately zero), the difference being accounted for by oceanic warm advection. The computed total advection (row 1) in April-May amounts to about 4 units of which a substantial part-at this time of year-appears to be produced by geostrophic transport (row 4).

Anomalies. Aubert and Winston for two individual months (January and July 1945) mapped the D.N. heating patterns. These show that significant differences in magnitude and location of the regions of heating and cooling occurred. In both months the deviations were generally as large as the magnitudes of the normal heating itself. The January map did not cover the whole Pacific, but over the North Atlantic the departures were in magnitude as large or larger than the normals and the "chart brings into sharp focus the considerable abnormality of the heating areas over eastern North America and the North Atlantic".

It should be finally mentioned that the above mentioned study of year-to-year variations of monthly surface heat balance and oceanic heat storage in the North Atlantic made by Arkhipova (1960) is by its very nature highly relevant to the present investigation. However, since it is for the Atlantic only whereas the D.N. field is computed here for a limited zone in the Pacific during one year only (although on a continuous rather than station basis) no comparison will be attempted.

9. The influence of windstress anomaly through layer heat advection on surface energy exchange.

The examination in section 6 of variability (range) of anomalous heat storage and wind induced advection in the mixed layer of the ocean demonstrated what was believed to be a striking feature in the overall layer heat balance, viz. that advection D.N. undergoes twice as large variation as storage in opposition to the normal seasonal process for

which the reverse is found to occur. At that stage of the investigation it was not clear whether any reliance could be placed on the magnitude of the advection anomalies derived as they were entirely from the greatly simplified formula for meridional volume transport anomaly (eq. 4b). Furthermore, the computation of wind stress vorticity anomalies is based on finite differences in a somewhat arbritrary grid, the climatological and individual monthly windstresses being derived by different methods. However, the considerations in the preceding paragraphs have strengthened the belief that computed values of both heat advection and storage can be taken as giving more information than merely on their order of magnitude. It should be thus meaningful to examine in some more detail the ratio of heat advection anomaly into surface balance anomaly B':

$$K = \frac{A'}{A' + S'} = \frac{A'}{B'}$$

where A stands for advection and S for storage. This question is not a formal one but has important bearings on the time of response of oceanic temperatures to the non-seasonal changes in the large-scale patterns and intensity of the surface windfield (Section 10).

The distribution of K and the standard deviation are shown in Figure 8. A priori one might have anticipated positive and negative values of K to occur with equal frequency. It is noteworthy that among a total of 81 computed values of each, advection and storage D.N., all but three ratios K were positive. It is therefore meaningful to consider

the average value for the entire sample of both K and the ratio $\frac{|A'|}{|S'|}$. These ratios turned out to be 0.72 and 2.6 respectively. This means that the sign of the surface energy flux anomaly almost always and everywhere in the region is determined by the sign of the oceanic (wind induced) heat advection anomaly; and that in magnitude heat advection exceeds storage by a factor of 2.6.

These results may be interpreted as follows. In middle latitudes of the North Pacific a comparatively small proportion of horizontal heat flux convergence or divergence (warm or cold oceanic advection) induced by drift currents due to monthly mean wind stress anomalies appears to go into non-seasonal warming or cooling of the mixed layer. A larger proportion of the heat gain or loss shows up as excess or deficit in heat exchange between ocean and atmosphere ($H'+LE' \gtrless 0$). In winter when the radiation anomaly R_o' is a small item in the total heat budget that proportion for all practical purposes can be equated to all of the remainder of the horizontal heating or cooling in the top layer of the ocean.

Examining the distribution of the ratio K and its standard deviations, Figure 8 indicates $K \geq 1$ in the central longitudes of the zone. Since in these longitudes $\sigma(K)$ is the smallest - as one might expect for an area less subjected to the disturbing effects of continental boundary regimes - the larger values of K should also be significant. From the definition of K it follows then that in the open expanse of the ocean there is a prevailing tendency for partial compensation between advection and storage anomalies

and a corresponding reduction of surface energy exchange anomaly. Accordingly the decrease of K to values less than unity in the western and eastern portions of the zone would indicate an absence there of a tendency for such compensation. If heat storage D.N. were in the average approximately equal to advection D.N. the ratio K' would tend to infinity. The patterns of sea surface temperature anomaly and of surface wind stress anomaly (Figures 2 and 3) would then tend to be highly correlated and no effective non-seasonal feedback processes from the atmosphere could be expected since energy exchange intensity at the surface would tend to approach the normal level ($B'=0$). These considerations do not of course imply that a search for contemporary correlations between wind stress and temperature anomaly patterns would not be useful.

10. Characteristic periods of temperature anomaly waves in the thermally closed system ($B'=0$).

(a) Progressive and retrogressive waves.

To simplify the problem we shall first examine the relationship between the characteristic time and length parameters of wind induced disturbances in the mixed layer ignoring altogether heat energy exchange at the sea surface. The disturbances in the field of meridional volume transport and of layer heat content are taken to be represented by harmonic waves progressing or retrogressing zonally without change of shape. The basic equation relating the local temperature change to Ekman transport, geostrophic transport and layer depth had been derived in section 3e (equation (8)).

We assume that the forcing anomalous zonal wind stress is given as:

$$\tau_x' = \delta \tau_x' \cos(\mu x - \nu t) . \tag{9}$$

The mass transport anomaly will then be given according to equation (2) Section 3c introducing the quantity G from equation (8)

$$\begin{aligned} \int_D^0 v' \rho \, dz &\equiv \int_D^0 (v_\tau' + v_g') \rho \, dz \\ &= -f^{-1} \left[\delta \tau_x' \cos(\mu x - \nu t) + \mu G \, \delta T' \sin(\mu x - \nu t + \alpha) \right] . \end{aligned} \tag{10}$$

Here δ denotes amplitude, μ wave number, ν frequency, and α the angle by which the T' wave lags on the forcing wave τ_x'. Since by definition the wave of geostrophic velocity v_g' lags 90° on the T' wave, the phase lag α must be either +90° or -90° providing waves in the total meridional velocity (v') are simple harmonic and move without distortion. Imposing this constraint on the kinematical properties of the disturbance the following relationships can be derived by linearizing equation (8), i.e. by setting $e^{-bT'} = 1$.

$$\nu_I = -\frac{|C|}{f} \left(\frac{\delta \tau_x'}{\delta T'} + G\mu \right) \tag{11a}$$

$$\nu_{II} = \frac{|C|}{f} \left(\frac{\delta \tau_x'}{\delta T'} - G\mu \right) . \tag{11b}$$

On account of the form of the cyclic function in (9) waves of solution I are retrogressive whereas waves of solution II are progressive if $\delta \tau_x' > \delta T' G \mu$ but retrogressive if $\delta \tau_x' < \delta T' G \mu$.

We may represent diagramatically the wave period t' as a function of wave length L. In figures 9 and 10 the period in days (or months, years) is shown as function of wave length in kilometers for various values of the amplitude of the temperature anomaly $\delta T'$ in degrees Centigrade, the amplitude of the wind stress anomaly being taken in either case as 1 dyne cm^{-2} and the meridional te mperature gradient of the basic (undisturbed) field as 10°C in ten degrees of latitude. These values are of the order of magnitude observed, cf. Table 6.

Retrogressive wave disturbances are shown in Figure 9, the resonance asymptote ($\delta T' \to \infty$) separating solution I from solution II .

The progressive wave disturbances of solution II are shown in Figure 10.

In the typical range of observed wavelengths of the forcing stress field, i.e. 3000 to 8000 Km and of temperature anomaly amplitudes 1° to 2°C, solution I retrogression waves have periods from six months to about one year while no solution II retrogression waves would occur. The progressive waves of solution II are in the eight months to two years range.

With the above values of wind stress amplitude and meridional temperature gradient the period for resonance varies from about three and a half years at wavelength 3000 Km to nine years at wavelength 8000 Km. (Both these points are outside the top of Fig. 9). Even longer periods would obtain for smaller stress amplitudes and/or less steep meridional temperature gradients.

The inverse relationship between period and wind stress amplitude is shown in Table 4 where L is chosen as 10,000 Km and the meridional temperature as 5°C/1000 Km. Table 4 refers to the progressive waves but similar tables can be set up for retrogressive waves.

The zonal windstress anomaly τ_x' is determined by $\gamma^2 \rho_a \left(W^2 \sin\theta\right)'$ where γ^2 is the drag coefficient and θ the angle of wind direction clockwise with NORTH. Breaking down the W^2 term into its components it may be expressed as function of $W_N\sqrt{\sin\theta_N}$ and $W\sqrt{\sin|\theta|}$ since $\left(W^2 \sin\theta\right)' = W^2 \sin\theta - W_N^2 \sin\theta_N$.

In Table 5 values of τ_x' are set out for a wide range of the speed-direction functions with $\rho_a \gamma^2$ assumed to have the same value as adopted by Hidaka [D] and Saur [E] in computations of wind stress over the ocean (i.e., for $W_o = 6$ m sec^{-1}).

In order that one may guage the magnitude of the zonal stress anomaly amplitude in terms of the relevant function $\sqrt{\left(W_o^2 \sin|\theta|\right)'}$, the values of this function are inserted in Table 4 and should be compared with those in the top (heading) row and left (leading) column of Table 5.

(b) Free waves and resonance.

It is to be expected that the resonance asymptote in Figure 9 would bear closely upon the period-wavelength relationship for certain among the possible normal modes that have been derived for a baroclinic stratified ocean. The oscillatory baroclinic mode in a two-layer model is bound up with the internal surface (thermocline). In the present mixed-layer model

the time variations only of the layer depth were taken into account in connection with the determination of heat storage and these were absorbed in the time variations of layer temperature (equation 7).

However, the assumption of vanishing meridional current velocity anomaly at the bottom of the mixed layer ($v' = 0$ at $z = D$) made it possible to determine the geostrophic meridional volume transport from equation (5) or (5a) in close analogy to integrating the thermal wind equation in the atmosphere. In this way the effect of oscillations at the internal boundary in the two-layer model is implicitly allowed for in the oscillations of the steric sea level (eq.(5)). Consequently the present model may be regarded as quasi-baroclinic.

The appropriate class of waves to compare with the resonance solution are therefore the free non-dispersive baroclinic waves. Veronis and Stommel (1956) derived the normal modes from the vorticity equation for a two-layer model allowing for surface friction. The free waves which are inherent in the normal model equations are obtained by ignoring the surface stresses. This is equivalent in the case of thermal waves to letting the temperature anomaly amplitude go to infinity as will be readily seen from the frequency equations (11).

Veronis and Stommel gave three approximate roots* of the cubic frequency equation of which the retrogressive barotropic and baroclinic Rossby waves represent the basic modes to respectively external and

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* For wavelengths $< 12{,}000$ Km the error introduced by the approximation is only 2 percent.

internal geostrophic response of the ocean to a forcing system of wind stresses. The period-wavelength relationships of these waves taken from Figure 1 of the paper by Veronis and Stommel are indicated in Figures 9 and 10. Quantitatively these relations are valid for the specific parametric values of the two-layer model which in their case were: thickness of top layer 500 mb, of bottom layer 3,500; density factor 2.10^{-3}; latitude about 40° ($f \approx 10^{-4} sec^{-1}$).

The barotropic free-wave relationship in Figure 10 is here of no further interest for obvious reasons. It will be seen that the resonant asymptote in Figure 9 is parallel to the straight part of the long baroclinic Rossby wave relation. This represents the non-dispersive long baroclinic waves whereas for wavelengths less than 500 Km the baroclinic Rossby waves take on the character of isostatic waves with increasing energy dispersion. Again, these will be not relevant to the large-scale systems considered.

Both the particular resonance asymptote in Figure 9 and the baroclinic non-dispersive wave relationship, part of which is shown in Figure 9, are derived in a manner to make them coincide by a suitable choice of the parameter values in either model. That they are solutions of closely similar dynamic systems is evident from the fact that in either case the waves move from east to west. The following circumstances however must be pointed out.

The Rossby waves are derived from the vorticity equation. The thermal waves in this present study are derived from the primitive equations of motion in conjunction with the thermal energy equation which as far as this section is concerned simply states the conservation of potential temperature. The dynamics of the waves are closely similar as we have seen. Yet according to Charney (1955) the β vorticity term is of particular importance for the behaviour of the waves. According to Charney's conclusions the variations of Coriolis parameter with latitude permits self-propagating barotropic modes; and the impressed force systems (transient wind systems with sufficiently large period) may approach resonance with the free baroclinic oscillating mode. Furthermore Charney points out that only retrogressive systems are capable of resonance. To this extent the present analysis is consistent with the dynamic concepts. On the other hand, the statement regarding the pre-requisite of a β effect for resonance does not appear to be consistent with the results obtained here.

It is not the place here to delve deeper into this problem but it is felt that the physical meaning of the β term in the vorticity theorem could well be ambigous in certain applications of that equation (Berson, 1960)*.

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* The author had the benefit to discuss some aspects of this problem with Prof. Phillips, who pointed out that there might be no inconsistency or contradiction of this nature. The free oscillation mentioned in connection with Charney's conclusions are bound up with the internal boundary of the two-layer model. In the single mixed-layer model adopted here, the disturbance of density or temperature is a function of time and longitude but not of depth. Therefore that free mode is excluded from the solution, but a resonant case is provided by the temperature-geostrophic current relationship referred to earlier in this subsection.

(c) Comparison with observed periods.

In constructing the diagram in Figure 9 the zonal wind stress amplitude was assumed to be 1 dyne cm^{-2}. A glance at Figure 3 suffices to recognize that the amplitude of the zonal wind stress anomaly would be likely considerably smaller in reality; whence from equation (11) the period of long retrogressive waves ($>$ 5000 Km) would be larger than the period read off the relevant isopleths of temperature amplitude in Figure 9. The resonant period too, for long waves, could be of the order of several years.

Average wave length, temperature and zonal stress amplitudes have been estimated from Figures 2 and 3 for various longitudes, and over-all mean values were formed. These values are set out in Table 6 together with the average value of the climatological sea-surface (meridional) temperature gradient. These averages were introduced in eq. (11) to determine the theoretical value of the average period. An estimate was also made of the average period that actually occurred during the autumn and winter of 1959-60. This is also shown in Table 6. The results of these computations can be summarized as follows:

Whereas there is broad agreement in orders of magnitude the theoretical wave period in comparison to the "observed" is 3 to 4 times too long in the case of progressive waves. The estimated speed of propagation of the "observed waves" for the whole spell in Table 6 is 1150 Km/month (eastward) as against 200 Km/month from equation (11b) and -350 Km/month (westward) from equation (11a).

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The nature of the discrepancy between the magnitudes of period indicated by observations and calculated from the theoretical solutions on the basis of independent observations, suggests that the neglected effect of surface heat exchange will depend mainly on an aerodynamic wind effect (W'); for this will be equivalent to an increase in the windstress amplitude, the effect of which is to reduce the period (see Table 4).

11. Response to variable wind when surface energy exchange is not disregarded.

(a) Factorizing of windstress amplitude.

Ignoring anomalies in net radiation at the sea surface as small in comparison to eddy heat conduction and evaporation anomalies - a procedure that appears to be justified at least during autumn and winter (cf. section 7) - the equation for surface energy balance anomaly due to variable surface wind speed may be written as:

$$B' = W_0' (1-r)(1+R) L K e_s \equiv W_0' J \quad . \tag{12}$$

The particular form of the energy transfer equation underlying (12) is that given by Kraus and Rooth (1961) who quote a value of $1.5 \cdot 10^{-12}$ for the transfer coefficient K.* (For other symbols see NOTATION, section 2b).

* If e_s is expressed in millibars the value of K is $1.5 \cdot 10^{-9}$ as actually given by Kraus and Rooth.

Substituting equation (12) in (7) and carrying subsequently the B' term through the derivations of the frequency equation (11), it is readily seen that the form of the equations (11) will remain unchanged.

The zonal stress amplitude is merely multiplied by a "factor" providing the amplitude of linear wind speed can be expressed as a function of windstress amplitude. It is possible to do this as indicated below:

From the classic power law for the windstress it follows:

$$\tau_x' = \gamma^2 \rho_a \left(W^2 \sin\theta - W_N^2 \sin\theta_N \right)$$

$$= \gamma^2 \rho_a \left\{ \left[W_N^2 + 2 W_N W' - (W')^2 \right] \sin\theta - W_N^2 \sin\theta_N \right\} .$$

In the last term on the right side to a first approximation $\sin\theta_N \simeq \sin\theta$, whence

$$\tau_x' = \gamma^2 \rho_a W' \left(2 W_{Nx} + W_x' \right) .$$

Assuming now that generally in the zone $W_x' \ll 2 W_{Nx}$ we have

$$W_0' = \frac{\tau_x'}{2 \gamma^2 \rho_a W_{N_x}} . \tag{13}$$

The frequency equations (11) after substituting from (12) and (13) in (8) take the form:

$$\nu_{\mathrm{III}}' = - \frac{|C|}{f} \left[\frac{\delta \tau_x'}{\delta T'} \left(1 + \frac{f}{|\Phi_y|} \frac{J}{2 c_p \gamma^2 \rho_a W_{Nx}} \right) + G\mu \right] \tag{14a}$$

$$\nu_{IV}' = \frac{|C|}{f}\left[\frac{\delta\tau_x'}{\delta T'}\left(1 + \frac{f}{|\Phi_y|}\frac{J}{2c_p\gamma^2\rho_a W_{Nx}}\right) - G\mu\right]. \tag{14b}$$

With a view to recognizing more clearly the effect of including the energy transfer term on modifying the wave period, equation (14a) and (14b) may be rewritten for the period t'. It is sufficient to give here the period from equation (14a) only:

$$t_{III}' = -\frac{f}{|C|}\left[\frac{\delta T'}{\delta\tau_x'\left(1 + \frac{f}{|\Phi_y|}\frac{J}{2\gamma^2 c_p \rho_a W_{Nx}}\right) + \frac{\delta T' 2\pi G}{L}}\right]. \tag{14a'}$$

The quantity following 1 in the factor of $\delta\tau_x'$ will be positive. It should be mentioned that the values of Bowen ratio R, relative humidity r, and saturation water vapour pressure at sea surface temperature e_s are equated to their seasonal zone-width averages because we have limited the analysis to the effect of wind anomaly only. However, as long as the variable wind influence is the dominating one that factor will always be positive. Consequently, the periods of progressive and retrogressive waves will be shortened. The reduction of period is the larger the smaller r but the larger R and the saturation vapour pressure at sea surface temperature, i.e., the higher the normal sea surface temperature.

To get an estimate of the magnitude of the reduction in the period, the climatological meridional sea surface temperature gradient Φ_y and average zone-width sea surface temperature are taken at their values

obtained in course of the previous deductions ($T_N = 14°C$ whence $e_s = 16 mb$). W_{Nx} is taken as $5 \ m \ sec^{-1}$ which value approximates fairly closely both, the average surface zonal wind component in the zone computed from manuscript data made available to the author by Dr. Priestley (F) and the average middle latitude surface zonal index in the Western Hemisphere. The following perhaps somewhat arbitrarily chosen values are taken for relative humidity and Bowen ratio: 1-r= 0.3, 1+R= 1.1 . With these values and the value of K mentioned earlier, the reduction factor turns out to be 5. . The corresponding modified periods t'_{III} and t'_{IV} are shown in Table 6.

Two conclusions may be drawn from a consideration of these re-evaluations of the characteristic wave periods:

(i) The magnitude of the theoretically deduced period is now much closer to the " observed " period of about 5 months.

(ii) Retrogressive and progressive wave solutions have now very much the same period.

In regard to conclusion (ii) it should be recalled that conclusive evidence for one or the other kind of waves is not present in the data. To the extent that progressive waves (during autumn and winter) are more likely to occur (Figures 3, 4, and 5), this circumstance would preclude the occurrence of resonance at those periods. However, near resonance periods might be required to account for the persistence of temperature anomalies;

for these tend to be destroyed by diffusion processes of a kind that have been disregarded in the present study. In contrast to earlier conclusions in this investigation, this result could be looked upon as a negative one. On the other hand, it might well support Pierson's conclusions from the statistical treatment of the data referred to in the Introduction.

12. Numerical computations (outline of a proposed scheme).

(a) Generalized "tendency equation".

The above method of analysis of the behavior of sea surface temperature anomaly and windstress patterns was adapted to estimating their average characteristic length and time scale parameters, the magnitude of their variance, etc. It gave also some insight into the magnitude of the effects of dynamic and thermal interactions at the ocean-atmosphere boundary. A deeper understanding could be achieved by numerical computations of the pattern evolution from an initial state, basing them essentially on the "tendency equation", equation (8).

However, it is then no longer necessary to neglect advection terms other than $\tilde{\mathcal{F}}_y' \Phi_y$. Yet among the remaining advection terms i.e., $\tilde{\mathcal{F}}_x' \Phi_x$, $\tilde{\mathcal{F}}_y \frac{\partial T'}{\partial y}$, and $\tilde{\mathcal{F}}_x' \frac{\partial T'}{\partial x}$ it seems justified to omit the last two terms as second order quantities. With the additional contraints $u'=0$ for $z=D$ and for $x=x_1$, $x=x_2$ (where x_1 x_2 are the longitudes of meridional coast lines) and with a change of variables

$$\begin{aligned} dt_* &= \Phi_t \, dt \\ \nabla_* &= \Phi_t \, \nabla_2 \end{aligned} \tag{15}$$

we obtain in analogy to eq. (8) the following differential equation in T' and W' as dependent variables:

$$e^{bT'}\left(\frac{\partial T'}{\partial t_*} + 1\right) = 1 - \frac{g\varepsilon e^{2b\Phi}}{f F_N}\left(\mathbb{K} \times \nabla_* T'\right)\cdot \nabla\Phi + \frac{W_*'}{F_N}\left(\frac{\gamma^2 \rho_a \Phi_y \widetilde{W}}{f} + \frac{J e_s}{c_p}\right) \tag{16}$$

where $F_N = (b - a + 1)\, e^{b\Phi}$,

$W_* = \frac{dl}{dt_*}$, dl being a streamline differential,

and $\widetilde{W} = W' + 2W_x + W^2 (W')^{-1} (\sin\theta)^{-1}$.

Here the subscript zero on all W terms has been throughout omitted. Equation (16) may be rearranged similarly to equation (8) so that $\frac{\partial T'}{\partial t_*}$ alone is on its left side. Furthermore, the expression for $\widetilde{W}$ can be simplified in accordance with the assumptions in section 11 (equation 13) or in some other manner. The third term on the right will however remain quadratic in the surface windspeed anomaly W'.

It should be pointed out that equation (16) can be applied only for such time of the year when a definite normal seasonal trend in sea-surface temperature does exist ($\Phi_t \neq 0$) but it can be brought back to its form with $\frac{\partial}{\partial t}$ and ∇_2 as differential operators.

(b) Hindcasting from a known forcing windstress field .

It is suggested that equation (16) could be tested on past data which amounts to hindcasting the temperature anomaly field from a known field

of the sea level wind anomaly. Initial and subsequent input values at gridpoints should be averages over half-monthly or ten-day spells. The equation is then solved numerically for T' $(t_0+\frac{1}{2}\delta t)$. For the time $t_0+\delta t$ a new set of T' is substituted in $e^{bT'}$ and $\nabla T'$ and the process is repeated with the known set of W' values at time $t_0+\delta t$. The iteration might be continued, say, 20 times in steps of half months or ten days, respectively. Such computations should be possible with a 1401 IBM computer. A ϕ, λ grid is appropriate providing the latitudinal width of the ocean considered does not exceed 20 degrees in middle latitudes. Averages of F_N and γ^2 have to be introduced.

Normal climatological average zone-values have to be assigned to the Bowen ratio and relative humidity. There is some doubt as to whether this is justified in regard to relative humidity and also whether a uniform transfer coefficient K would be permissible.

(c) Prediction allowing for energy feedback from the atmosphere.

The basis for arriving at a practical solution of the problem of prediction (in the sense of an operational research task only) will be here briefly outlined. It is essential to take into account the thermal energy balance in at least a considerable depth of the troposphere. We neglect atmospheric heat storage anomaly as quite small compared with heat storage in the active layer of the ocean. By substraction of the thermal energy equation for the atmospheric layer from the thermal energy equation for the active ocean layer a differential equation in W'

and T' similar to eq. (16) is obtained but containing in addition an advection term

$$- W' \nabla_2 \Phi \; \frac{c_{pa} \, \delta p}{g \, e^{b\Phi}} \left(\frac{W_g}{W} \right)_N$$

where δp is the pressure thickness of the atmospheric layer (for example 1000-500 mb) and W_g the sea-level geostrophic wind. To derive it application has been made of the thermal shear relationship according to which geostrophic advection of thickness or mean (virtual) layer-temperature in the direction of the average thermal wind vector must vanish identically. Advection by the frictional non-geostrophic component of the wind in the lowest 1,000 feet is being ignored which would be justified except in frontal zones.

Net radiation anomaly at the ocean surface has not been taken into account in equation (16) for reasons that were discussed in sections 7 and 8, and which were valid for the autumn-winter regimes. However, the corresponding radiation term in the modified thermal energy (difference) equation is a radiation anomaly

$$R'_{o1} = R'_1 - 2 R'_o$$

where R_1 is the net (upward) flux of radiant energy at the top of the layer with pressure thickness δp.

Aerodynamically controlled anomalies of evaporation from the ocean had been accounted for in eq. (16). Since precipitation anomaly would be a measure of excess or deficit in the amounts of conversion of latent heat in the atmospheric layer, a corresponding conversion term should appear

also in the thermal difference equation. However, since we had tacitly ignored anomalous advection of moisture (latent heat), precipitation and evaporation anomalies would cancel out. This would be in fact the case providing the distribution of rain bearing systems over the ocean were not sensibly correlated to the forcing field of wind stress anomaly - or at least much less so than the latter is correlated to heat transport in the active layer of the ocean. The thermal energy difference equation has then a form similar to (16) except that presently we have to substitute the Bowen ratio R in the place of (1 + R) and of course allow for the above mentioned advection and radiation terms.

In the modified equation (16) the W' field and radiation anomaly R'_{o1} are the input at the initial time. The mapping of the salient features of the R_{o1} field over the oceans would be facilitated by satellite traverses. The capability of spectrographic radiation sensors based on carbon absorption laws presently developed in the U.S. Weather Bureau and NASA and to be mounted on future satellites, in conjunction with measurements of the total outgoing beam on top of the atmosphere (R_{∞}) by present TIROS-type instrumentation should make it feasible to map the average R_{o1} field over periods as long as a month. The normal distribution of the relevant radiation components on a hemispheric scale is sufficiently well known to derive then the patterns of the anomaly R'_{o1}. Such patterns would be adequate for testing the modified equation since the radiation terms are not part of the specific energy feedback with which we are here

primarily concerned.

If one wants to examine the significance of energy feedback from the atmosphere from an initial set of input packages by a numerical iteration procedure it is necessary to find a function for the energy conversion in the atmosphere in terms of W'. Let E_k be kinetic energy per unit mass, then in the layer considered

$$\frac{1}{2} g^{-1} \int \left(W(z)^2\right)' dp = E_k' \quad .$$

Applying the concepts of Lorenz (see Winston and Krueger, 1961) for the generation of available potential energy $\mathcal{G}$:

$$\frac{\partial E_k}{\partial t} = g^{-1} \Gamma \int^{\delta p} \frac{COV(TQ)}{\sigma T^2} dp - \frac{1}{2} \frac{\partial}{\partial t} \int^{\delta p} \frac{VAR(T^2)}{\sigma T^2} dp$$

$$+ \text{DISSIPATION TERM}$$

$$= \mathcal{G} - \frac{\partial A}{\partial t} + \tau_s W_0$$

whence

$$E_k' = \int_{t_0}^{t_1} \left(\mathcal{G}' - \frac{\partial A'}{\partial t} \right) dt + \int_{t_0}^{t_1} \left(\tau_s W_0 \right)' dt \qquad (17)$$

where Γ is the adiabatic lapse rate, σ static stability, Q the effective atmospheric heat source, COV denotes covariance and VAR denotes variance.

The second term on the right side is being evaluated on a current basis in the Planetary Meteorology Section of the U.S. Weather Bureau's Meteorological Satellite Laboratory. The evaluation of the first term is being planned after the launching of NIMBUS Satellite. There would be no insurmountable difficulties in assessing the contributions of the generation and conversion terms by discrete but not too limited longitude sectors of latitude belts.

The wind speed anomaly W'_o at sea level is determinable from $W'(z)$ assuming a normal parametrical wind shear law.

A necessary condition for the occurrence of energy feedback of the kind here considered would then be given by the equality in order of magnitude of the ratios $\frac{\partial E'_{ko}}{\partial t} \Big/ B'$ and $\frac{\partial E'_k}{\partial t} \Big/ \left(G' - \frac{\partial A'}{\partial t} \right)$ where the surface energy flux anomaly B' is approximated in the manner discussed above. Complete criteria for positive feedback could probably be obtained only by iterative numerical procedures similar to the one outlined here.

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LEGENDS TO FIGURES

Figure 1 Simplified mixed layer model in latitude zone centered on average maximum baroclinity. Relative slopes of sea-level and thermocline are arbitrary. (For symbols see text).

Figure 2 Sea-surface temperature anomaly (°F.) Pacific Ocean. Lat. 35°-45°N.

Figure 3 The negative of the zonal wind stress anomaly (10^{-2} dyne cm^{-2}) over the Pacific, 35°-45°N.

Figure 4 Heat storage anomaly (10^3 cal cm^{-2}/ month) in the mixed layer of the Pacific Ocean, 35°-45°N. (Based on eq. 7).

Figure 5 Heat advection anomaly (10^3 cal cm^{-2}/ month) in the mixed layer of the Pacific Ocean, 35°-45°N (based on Sverdrup formula for meridional mass transport, eq. 4b).

Figure 6 Monthly surface heat balance anomaly (B') as function of sea-air temperature difference anomaly, Sept. 1959 - February 1960, Pacific Ocean 35°-45°N. West of 180° Long, 180° and East thereof; b, nearest to continental boundaries; F, February. (For dashed lines and inset on right see text).

Figure 7 As fig. 6 except for period March to June 1960. (J indicates June).

Figure 8 The ratio of layer heat advection anomaly into surface heat balance anomaly, K. Mean values over longitude ranges - - - - - 10° -----20° Sampling numbers on top. Lengths of vertical lines indicate standard deviation.

Figure 9 Showing the period of forced (retrogressive) thermal waves in mixed layer for different temperature amplitudes as a function of wave length, (equation 11a and 11b respectively). The amplitude of the wind stress anomaly is taken as 1.0 dyne cm^{-2}, the meridional temperature gradient as 10°C over 10° latitude and the central latitude is 40°.

Figure 10 As in fig. 9 except progressive waves (eq. 11a).

Figure 11 Mixed layer depth as function of sea-surface temperature. Long-term monthly averages except for station "Papa". In fig. 11a temperature is represented as departure from a mean for part of year as stated in legend on left. In fig. 11b the mean is for the period March through August.

Figure 12 Normal sea-level air temperature (T_o) as function of thickness, 700 mb height minus 1000 mb height (z), both shown as departure from the respective annual mean of the average in the belt 35°N to 45°N over the Pacific Ocean. For broken line see text. - A, west of 180° longitude; B, 180° and east thereof.

APPENDIX I

NORMAL SEASONAL REGRESSION OF LAYER DEPTH ON SEA-SURFACE TEMPERATURE

A simple relationship between the depth of the mixed layer and average monthly sea-surface temperature has not been established as valid in any oceanic region. Heating by short-wave absorption in the uppermost strata, cooling by net long-wave radiation from the surface, vertical mixing by eddy motion mainly due to wind stirring, and lateral eddy diffusion usually produce highly complex changes apart from diurnal variations. Over longer time interval much of the complexity is however smoothed away and a broad relationship seems to exist between changes in mean layer depth and in the mean wind velocity. Recently Kraus and Rooth (1961) showed that on theoretical grounds there should be a less obvious correlation with the mean wind than with the surface heat balance and gave some evidence for this conclusion from oceanographic and meteorological data at station Echo (35°N, 48°W) for 10-day means for the periods July through September 1958, 1959.

With a view to examining relationships between layer depth and sea-surface temperature on time and spatial scales commensurate with the present requirements, a search of oceanographic and hydrographic data was made but it was soon found that mass data on layer depth

are mainly security classified. Useful information is gleaned, however, from a declassified publication (DATA SOURCE [I]) and from the papers listed in [J] and [M].

Examination of the data therein shows that when the departures of monthly mean layer depth from its annual average is plotted as a function of the corresponding departures of sea-surface temperature, (10m on 1°C scaling) linear regressions are on the whole poor mainly on account of strong shifts between the autumn-winter and spring-summer position of points; thus fairly although regular open ellipses result when the points in the diagram are connected by a smooth line. It ought to be mentioned that correlation coefficients $r(D, T_s)$ have nevertheless quite respectable values, e.g. -0.85 for Station PAPA at (50°N, 145°W, [J]) and -0.74 for the "Bermuda Area," [I].

When sea-surface temperature is being replaced by the square of the sea-level wind speed - the wind data for the Pacific being retrieved from Hidaka's wind stresses [D] - the corresponding ellipses appear more irregular although the correlation coefficient is higher, viz. 0.94.

Useable regressions of the form eq. (6) in section 3 are obtained by subdividing the year into periods September to February, and March to June in accordance with the procedure in section 7.

In figure 11a and b layer depth is shown as a function of sea-surface temperature (departure from average in period) using the aforementioned data. In figure 11a the Pacific middle latitude and Bermuda Area plots give a satisfactory linear regression of log D on T_s. The parameters of eq. (6) have the values a = 2.32 b = -0.037 when the logarithm is taken to the basis 10 and D is expressed in meters.

A regression for Station PAPA would be steeper but actually this station is well to the north of the zone considered here.

The two seasonal means of layer depth estimated from the Pacific maps of Lumby [M] have been placed at the temperature values taken from [I]. They would indicate a similar regression but with a different constant.

The regression for spring and early summer is much steeper (figure 11b) the parameters having the values a = 3.46, b = -0.17.

APPENDIX II

NORMAL SEASONAL REGRESSION OF AIR TEMPERATURES ABOVE SEA SURFACE ON 1000 - 700 MB LAYER THICKNESS

The method of determining the sea - air temperature difference anomaly, $T_s' - T_o'$, as indicated in section 7, is based on examining the normal seasonal relationship between air temperatures T_{oN} and thickness of the 1000 - 700 mb layer, δz_N.

The assumption is than made that similarly as in the case of the mixed layer depth the anomalies T_o' can be determined from a regression equation linking T_{oN} with δz_N. (Anomalies of sea-surface temperatures T_s' are, as before, taken from [B] and [C]).

In the following subscripts N will be omitted for brevity's sake.

Monthly means of T_o and δz are read off at the relevant longitudes and at latitudes 37.5°, 40°, and 42.5°N. Zone-width averages are then formed and from the monthly means the annual averages are deduced (T_{om}, δz_m). The departure of the monthly from the annual averages are plotted as functions of each other in Figure 12 where each point represents the mean value over the longitude sections indicated on the diagrams.

It is seen that the annual range of both temperature and thickness is far larger in the western than eastern parts of the ocean.

The distribution of plots in figure 12a shows well defined linear regressions of air temperature T_o on thickness δ? as could be anticipated on grounds of convective mixing, especially during period of intense upward flux of sensible hear from the ocean surface. Since in the eastern portion of the ocean such fluxes are less developed than in the western portions, a comparatively large scatter of points obtains in figure 12b.

The broken lines in both diagrams indicate the regression that would hold valid in the case the temperature lapse rate in the 1000 - 700 mb layer were constant throughout and equal to the dry-adiabatic value. The deviation of the slope of the regression lines from the broken line is therefore dependent on both the departure from adiabatic lapse-rate and the month by month changes of lapse-rate (except in the July and February regression lines in figure 12a and figure 12b respectively).

A more detailed analysis of these relationships would require an examination of aerological sounding data.

	$c_p \times$ AREA** Mean of:	DEPARTURE FROM ANNUAL AVERAGE JAN	APR	JUL	OCT	ANNUAL MEAN	DATA SOURCE
Meridional Heat Flux Convergence* Determined by:							Windstresses from:
(1) Sverdrup Drift	$\beta^{-1}(\nabla \times \tau_*)\,\Phi_y$	-3.4	4.4	-0.4	-0.5	-1.6	Hidaka ([D])
		-4.2	3.3	0.7	0.1	-1.7	Scripps ([E])
(2) Ekman Drift	$f^{-1}\,\tau_x\,\Phi_y$	-1.2	0.3	0.5	0.3	-1.1	[D]
		-1.3	0.4	0.5	0.4	-1.5	[E]
(3) Geostrophic Drift: Difference (1)-(2)		-2.2	4.1	-0.9	-0.9	-0.5	[D]
		-2.9	2.9	0.2	-0.3	-0.2	[E]

		DEC JAN	FEB MAR	APR MAY	JUN JUL	AUG SEP	OCT NOV		
(4) Geostrophic Transport	$f^{-1} g \rho D \frac{\partial z_T}{\partial x}\,\Phi_y$	-0.8	0.7	1.5	0.4	-0.4	-1.2	0.1	Steric sea-level: Patullo et al. [H]
(5) Surface Heat Balance	$c_p^{-1}(R_o - H - LE)$	-9.6	-4.6	4.1	10.4	4.1	-4.6		
(6) Layer Heat Content	$\int_D^0 T \rho \, dz$	10.2	-19.2	-29.7	-10.2	19.2	29.7	(10.2)	J.Patullo (1956)
(7) Monthly Heat Storage in Layer	$\delta \int_D^0 T \rho \, dz$	-14.7	-5.5	9.8	14.7	5.5	-9.8		
(8) Storage February To August	$\delta_{max.} \int_D^0 T \rho \, dz$			53					J.Patullo (1957)
				56					K.Bryan (1959)
				37					Arkhipova (1960)
(9) Meridional Heat Flux Convergence: Difference (7)-(5)		-5.1	-0.9	5.7	4.3	1.4	-5.2		--

* = "Warm advection"

** Subscript N omitted in this column.

Table 1 - Normal seasonal variations of heat balance in the mixed layer of the Atlantic Ocean. Zone 35°N -45°N, 15°W -65°W. All quantities in units of 10^3 cal cm^{-2} /month, except row (6) where the unit is 10^3 cal cm^{-2}.

STATION	LAT.	LONG.	HEAT STORAGE				SURFACE BALANCE ($-B_N$)			
					FIRST HARMONIC				FIRST HARMONIC	
			Range	Phase	Double Ampl.	Phase	Range	Phase	Double Ampl.	Phase
1	2	3	4	5	6	7	8	9	10	11
"Victor II"	30-32	163-165E	39.3	JUNE	30.5	AUG-SEPT	15.5	JUN-JULY	15.1	JUN-JULY
"Victor I"	33-35	172-174E	16.8	MAY	14.4	JUL-AUG	17.4	JULY	17.1	idem
"38158"	37-39	157-159W	18.8	MAY	12.2	AUG	16.7	JULY	14.6	idem
"Sugar"	43-45	164-166E	39.6	JUNE	30.8	AUG-SEPT	19.0	JULY	18.8	idem
"Queen"	43-46	172-174W	22.6	AUG	18.8	JUL-AUG	17.4	JULY	17.2	JUNE
"40150"	40.5	150.5W	33.9	JULY	28.9	AUG	14.1	JULY	12.9	JUN-JULY
VANCOUVER ISL.	47.5	127.5W	18.6	JUNE	18.5	AUG	13.3	JUNE	12.2	JUN-JULY
AVERAGE FOR PACIFIC ZONE			27.1		22.1		16.2		15.4	---
(AVERAGE FOR ATLANTIC ZONE cf. Table 1)			---		29		---		20	---

Table 2 - Annual range and double amplitude of first harmonic for heat storage and surface heat balance at stations in the middle latitude zone of the North Pacific (Complied from data by Patullo (1957)). In the columns headed by "PHASE" the months of the maximum value are given. - Values in units of 10^3 cal cm^{-2}/month.

Stations (Table 2)	Approx. Longitude	MERIDIONAL HEAT ADVECTION NORMAL ANNUAL Table 2	Stress Vorticity	D.N. Stress Vorticity	HEAT STORAGE NORMAL ANNUAL Table 2	D.N. Eq. 7
31N 164E 44 165E	165E	22.2 (13.7)	23	32	32	10
34 173E 44 173W	180	2.3 (0.5)	6	21	19	12
38 158W 40 151W	155W	10.9 (6.8)	10	18	23	7
47 128W	130W	5.3 (6.3)	3	13	18	10
Average		10.2 (6.6)	11	21	23	10

Table 3 - Ranges of normal oceanic heat advection and storage, and of departures from normal during period September 1959 through August 1960. Latitude zone 35-45°N, Pacific Ocean. - Values in parenthesis in column 4 correspond to the first harmonics in Table 2 - Unit: 10^3 ly/month.

	Stress Anomaly Amplitude					
	0.5	1.0	2.0	3.0	4.0	dyne cm^{-2}
	$\sqrt{(W_0^2 \sin\lvert\theta\rvert)'}$					
	4.1	5.8	8.2	10.0	11.5	m sec^{-1}
Sea temperature anomaly amplitude T_s'(°C)						
0.5°	15.1	7.3	3.6	2.4	1.8	months
1.0	31.7	15.1	7.3	4.8	3.6	"
1.5	51	23.3	11.2	7.3	5.5	"
2.0	72	32	15.1	9.9	7.3	"
2.5	97	41	19.1	12.4	9.2	"

Table 4 - The period of thermal waves as function of forcing windstress and sea-surface temperature anomaly amplitude. (Progressive wave solution, eq. 11b). The meridional temperature gradient is here assumed as 5°C in 1,000 km and the central latitude as 40°.

$W_o\sqrt{\sin\lvert\theta\rvert}$ (m sec^{-1})	$W_{oN}\sqrt{\sin\theta_N}$ 5	10	(m sec^{-1}) 15
-10	-3.8	-6.0	-9.7
-5	-1.5	-3.8	-7.5
0	-0.8	-3.0	-6.8
5	0.0	-2.3	-6.0
10	2.3	0.0	-3.8
15	6.0	3.8	0.0

Table 5 - The zonal wind stress anomaly $\tau_x = (\rho_a \gamma^2 W_o^2 \sin\theta)'$ in dyne cm^{-2}, as function of the "observed" and normal wind speeds and directions. The drag coefficient γ^2 and the density ρ_a are taken as constant ($\rho_a \gamma^2 = 3 \cdot 10^{-6}$g cm^{-3}). Negative values in the column to the left indicate easterly winds.

PERIOD	Average From Longitude-Time Graphs							Computed Wave Periods (Months)			
	Φ_y °C/5°lat.	L (T') 1000 km	L (τ'_x) 1000 km	t' (T') months	t' (τ_x') months	$\delta\tau'_x$ dyne cm^{-2}	T' °C	t'_I retrog.	t'_{II} progr.	t'_{III} retrog.	t'_{IV} progr.
SEPT 59-March 60	5.5	5.4	6.7	5.2	5.2	0.45	1.4	16.7 (16.0	28.0 27.6)	3.3	3.7
DEC 59 March 60	5.3	6.1	6.4	5.4	5.4	0.45	1.5	19.7 (17.6	32.0 31.1)	3.8	4.1

Table 6 - Mean meridional temperature gradient and parameters of the wind stress anomaly and temperature anomaly waves. Left side: Estimated parameters averages over longitude - time domain from Figs. 2 and 3. Right side: Wave periods computed from equations (11) and (13) using the parameters on the left (except the estimated period t'). Values in parentheses computed from average parameters.

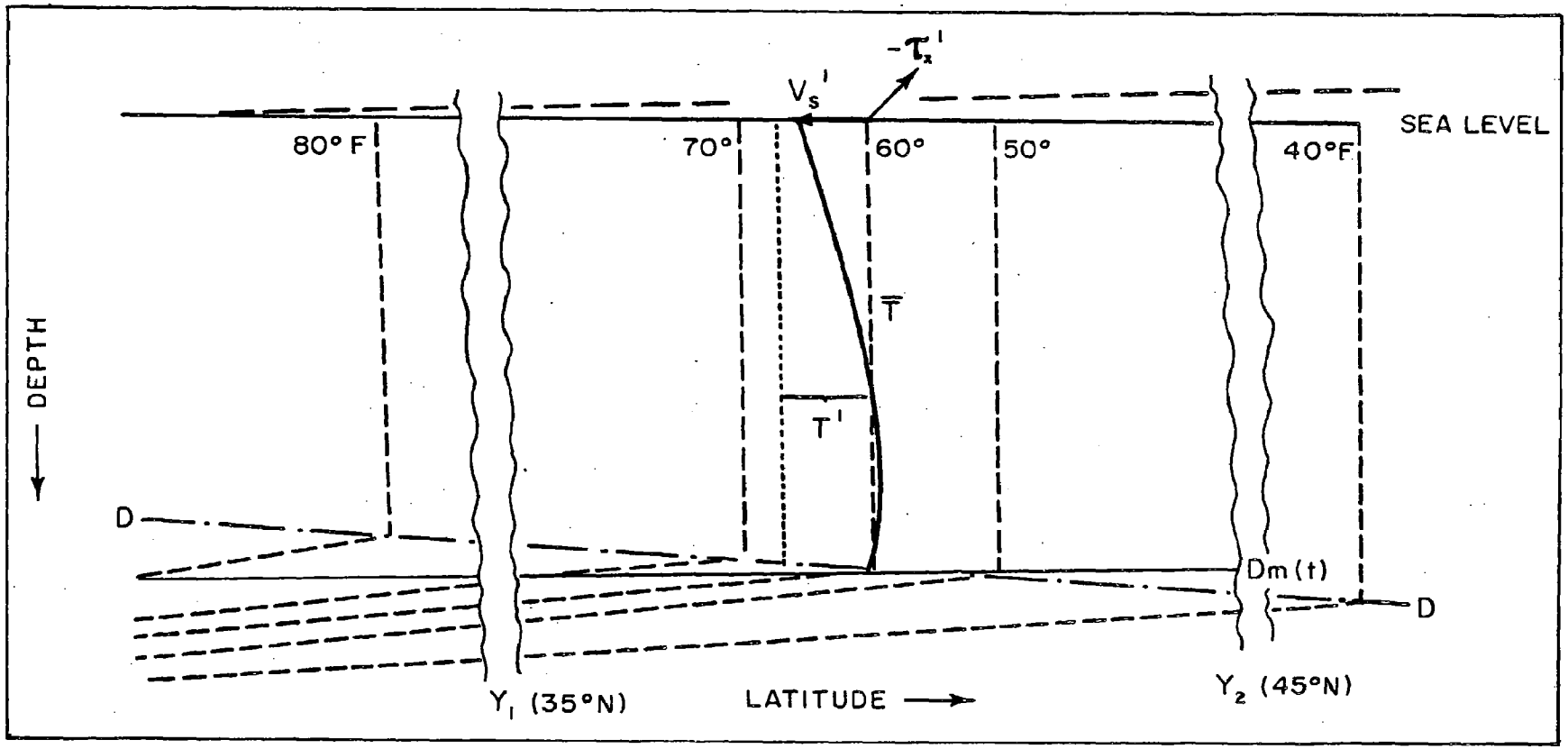

$-\tau_x'$
V_s'
SEA LEVEL
80° F
70°
60°
50°
40°F
$\bar{T}$
T'
DEPTH
D
Dm (t)
D
Y_1 (35°N)
LATITUDE
Y_2 (45°N)

Fig. 1

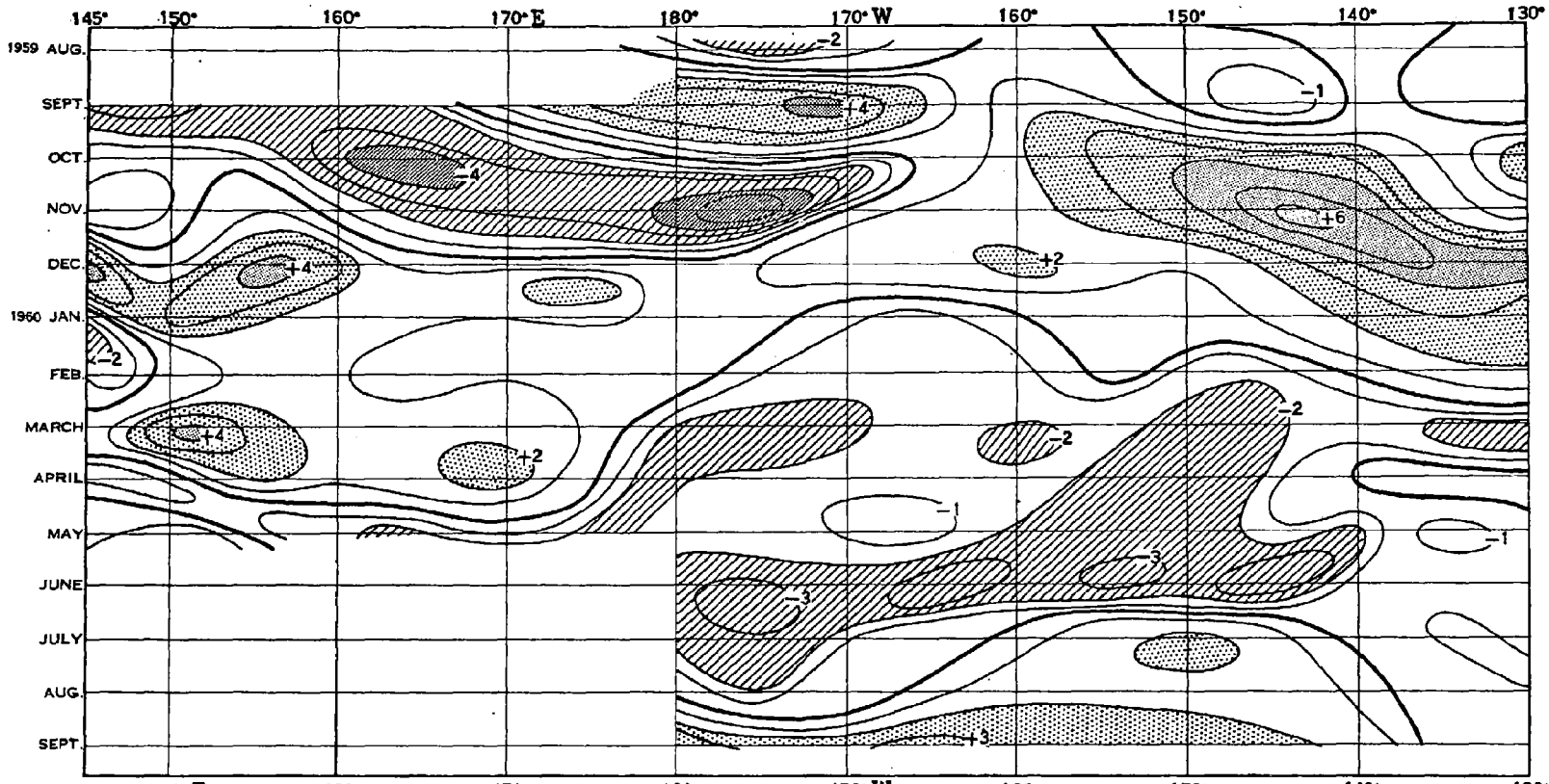

145°
150°
160°
170° E
180°
170° W
160°
150°
140°
130°
1959 AUG.
SEPT.
OCT.
NOV.
DEC.
1960 JAN.
FEB.
MARCH
APRIL
MAY
JUNE
JULY
AUG.
SEPT.
-2
-1
+4
-4
+6
+2
+4
-2
-2
+4
+2
-2
-1
-1
-3
-3
+3

Fig. 2

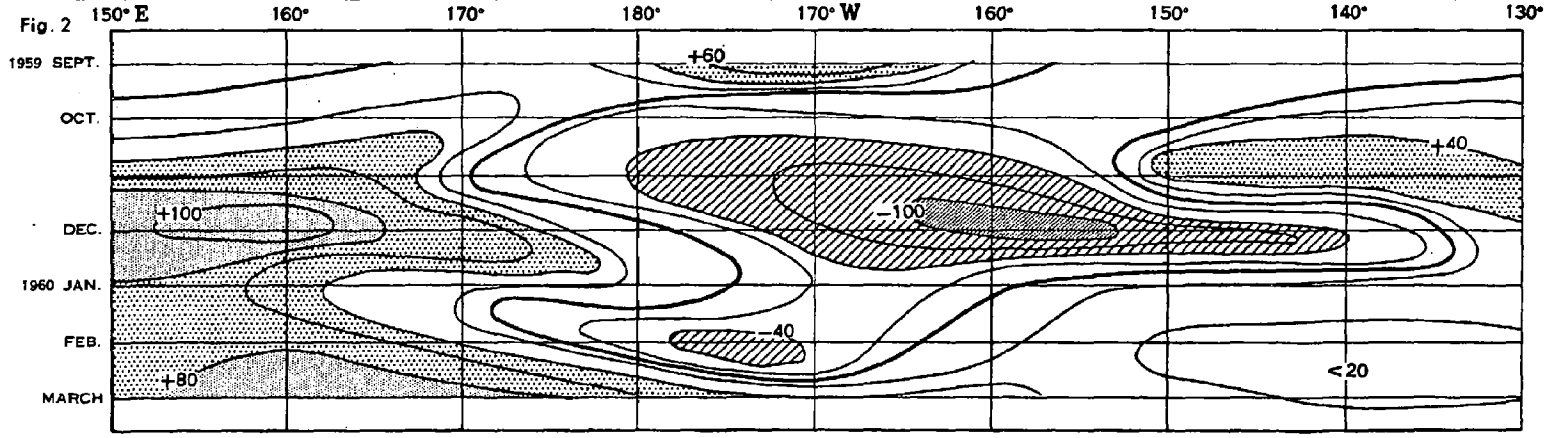

150° E
160°
170°
180°
170° W
160°
150°
140°
130°
1959 SEPT.
OCT.
DEC.
1960 JAN.
FEB.
MARCH
+60
+40
+100
-100
-40
+80
<20

Fig. 3

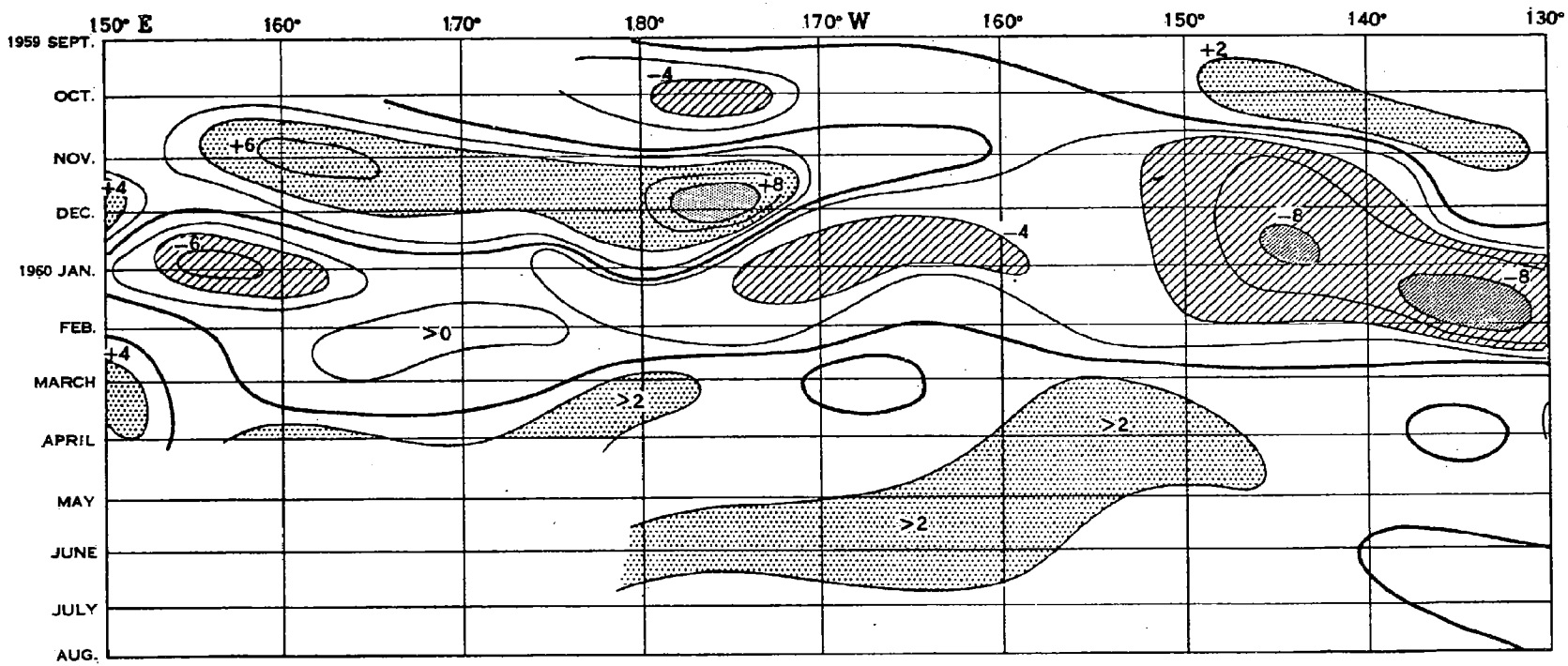

150° E
160°
170°
180°
170° W
160°
150°
140°
130°
1959 SEPT.
OCT.
NOV.
DEC.
1960 JAN.
FEB.
MARCH
APRIL
MAY
JUNE
JULY
AUG.
+2
-4
+6
+8
+4
-6
-4
-8
-8
>0
+4
>2
>2
>2

Fig. 4

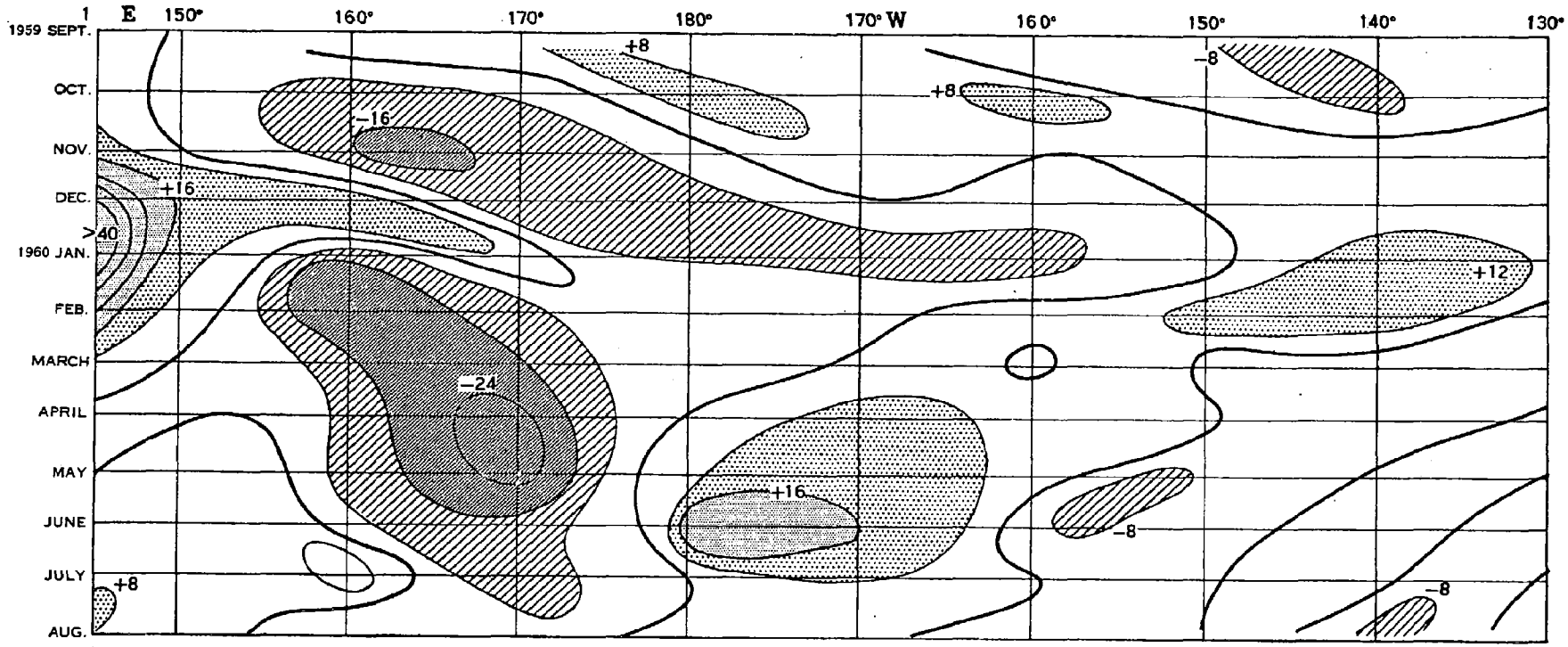
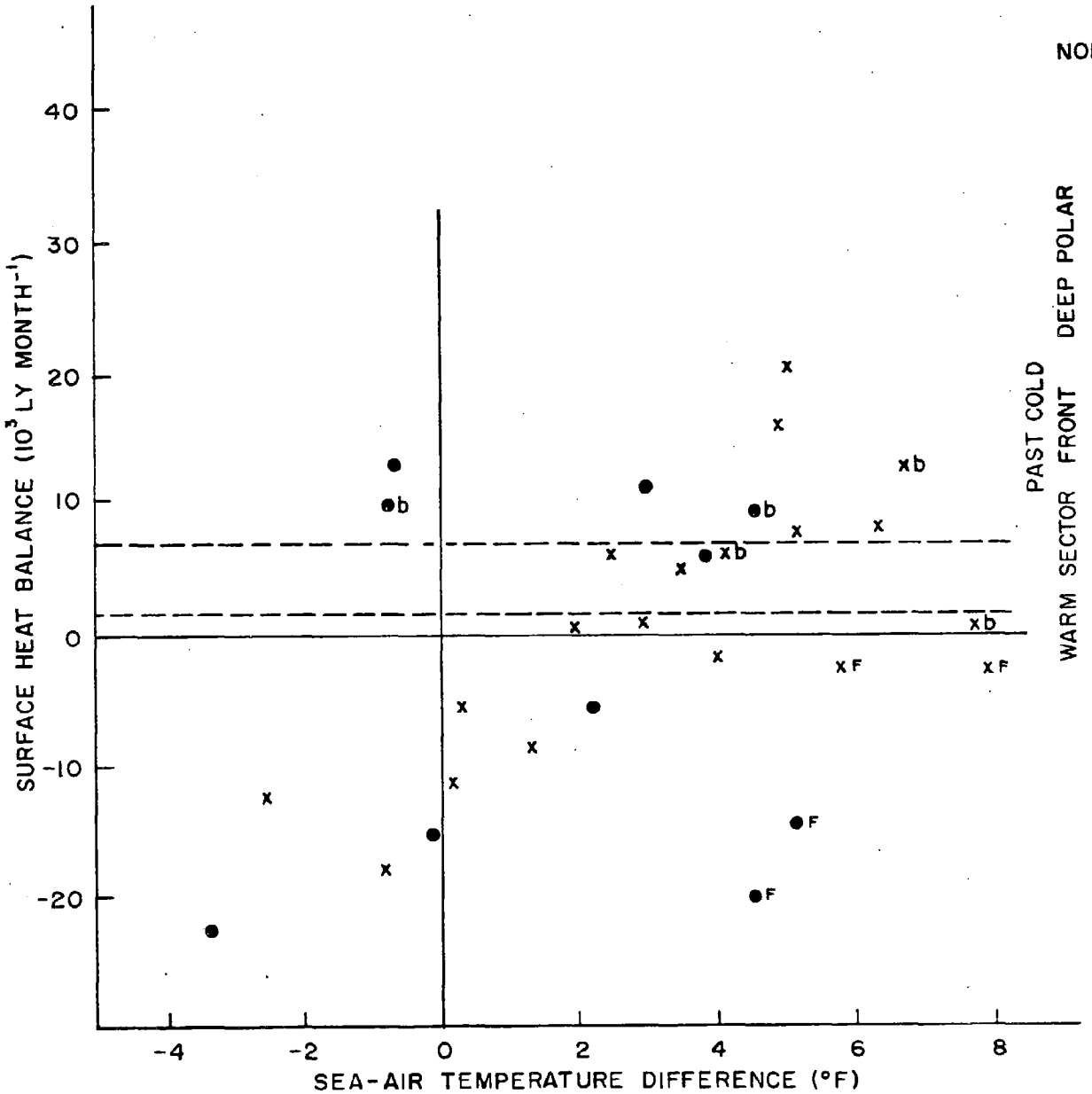

1 E
150°
160°
170°
180°
170° W
160°
150°
140°
130°
1959 SEPT.
OCT.
NOV.
DEC.
1960 JAN.
FEB.
MARCH
APRIL
MAY
JUNE
JULY
AUG.
+8
-8
+8
-16
+16
>40
+12
-24
+16
-8
+8
-8

Fig. 5


SURFACE HEAT BALANCE (10^3 LY MONTH^{-1})
40
30
20
10
0
-10
-20
-4
-2
0
2
4
6
8
SEA-AIR TEMPERATURE DIFFERENCE (°F)

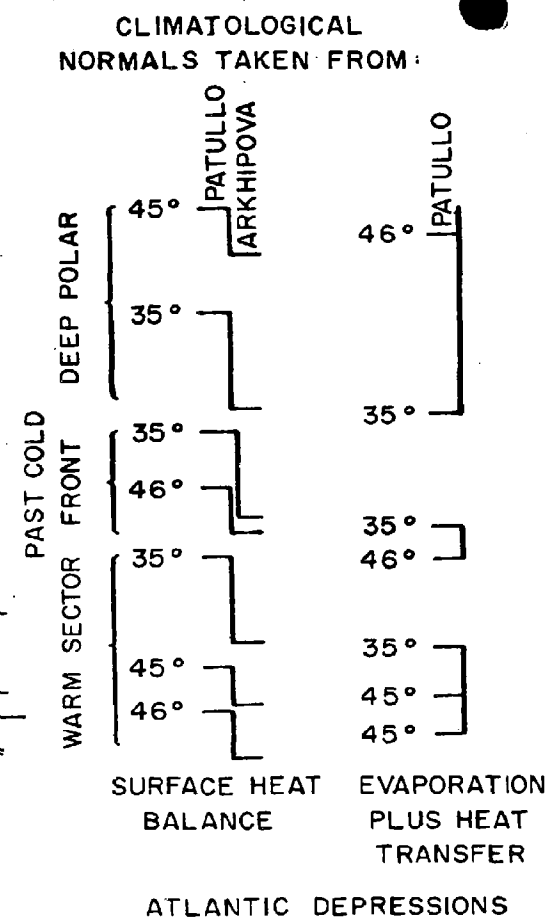

CLIMATOLOGICAL NORMALS TAKEN FROM:
PATULLO
ARKHIPOVA
PATULLO
DEEP POLAR
PAST COLD FRONT
WARM SECTOR
45°
35°
35°
46°
35°
46°
35°
45°
46°
46°
35°
35°
45°
45°
SURFACE HEAT BALANCE
EVAPORATION PLUS HEAT TRANSFER
ATLANTIC DEPRESSIONS

Fig.6

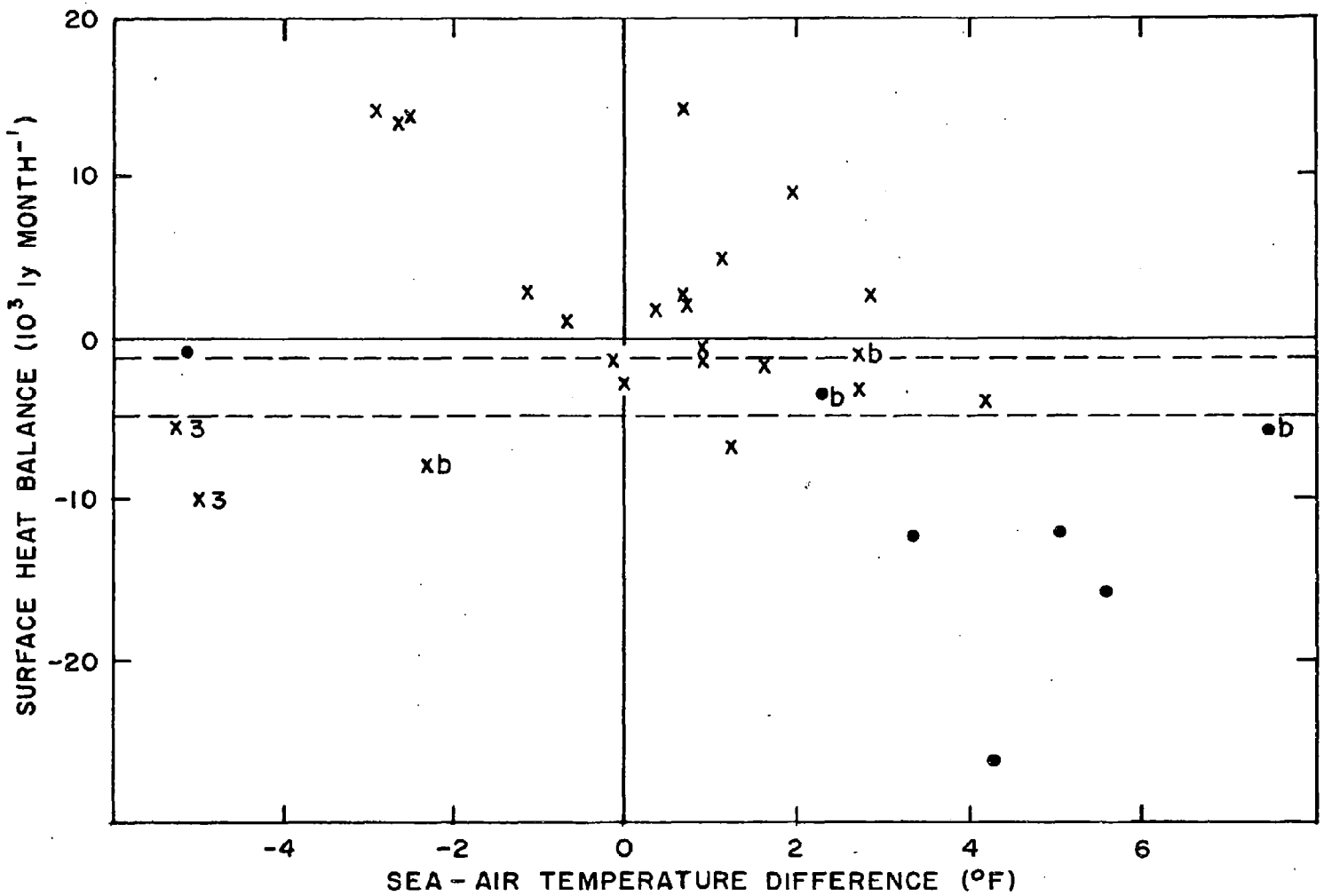
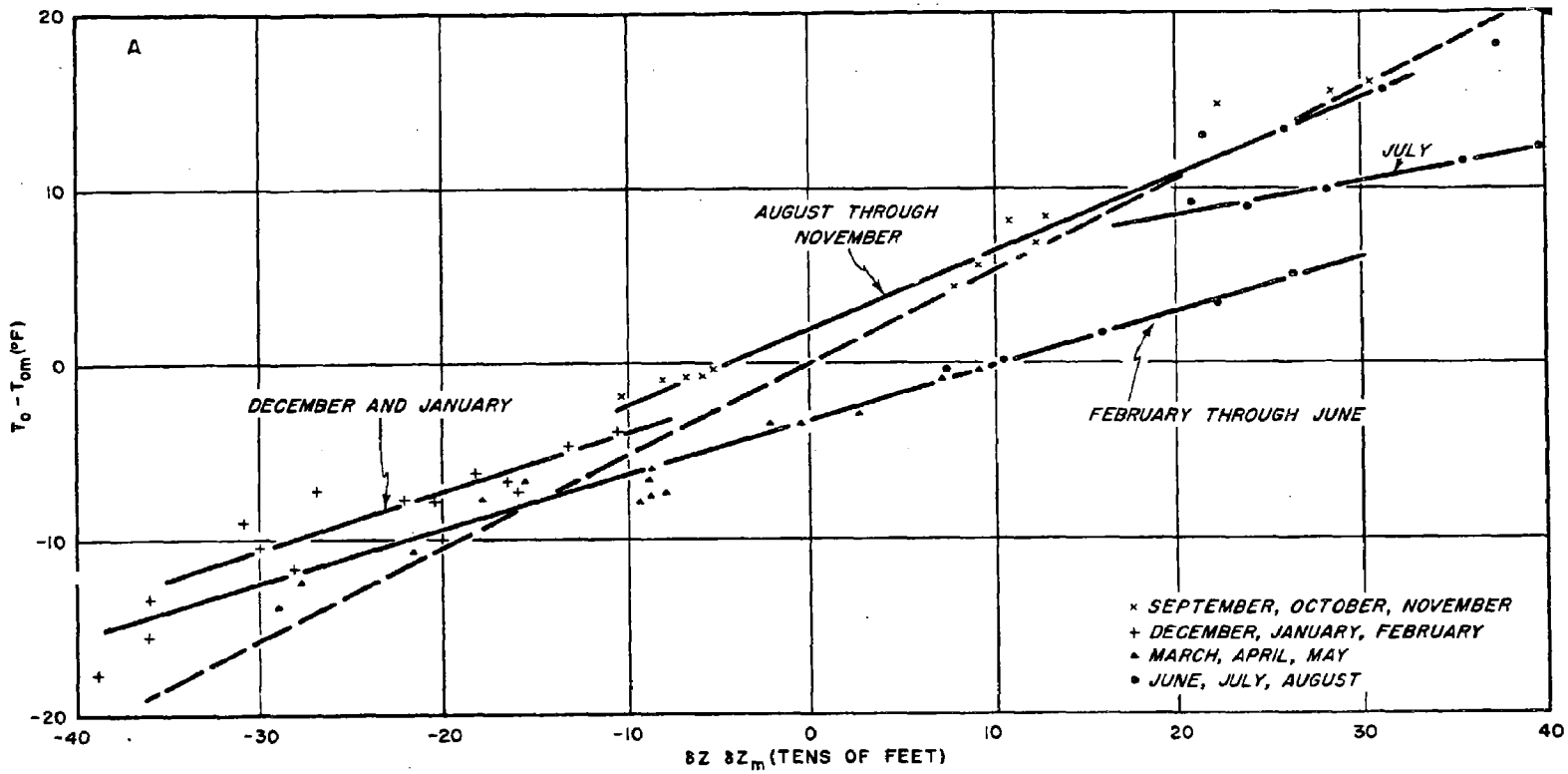

20
10
0
-10
-20
SURFACE HEAT BALANCE (10^3 ly MONTH^{-1})
-4
-2
0
2
4
6
SEA - AIR TEMPERATURE DIFFERENCE (°F)

Fig. 7



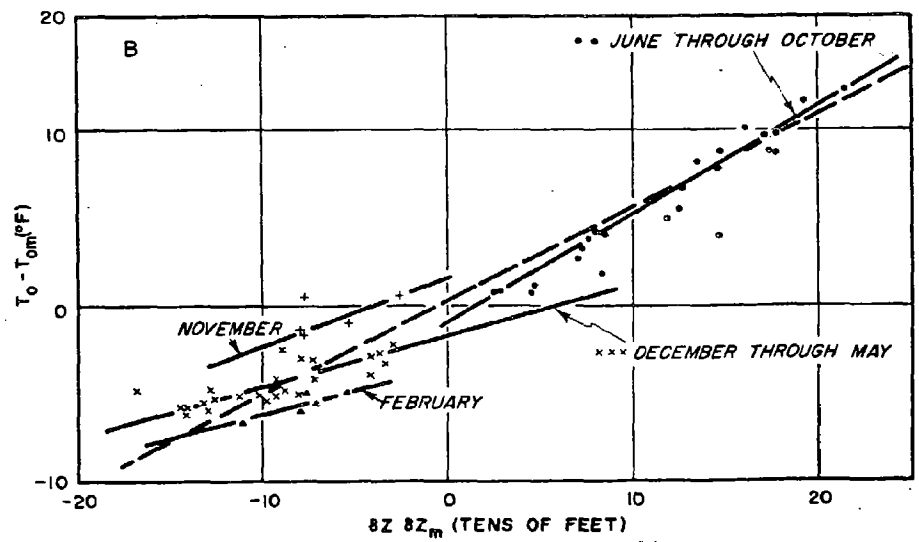


FIG. 12

USCOMM-WB-DC

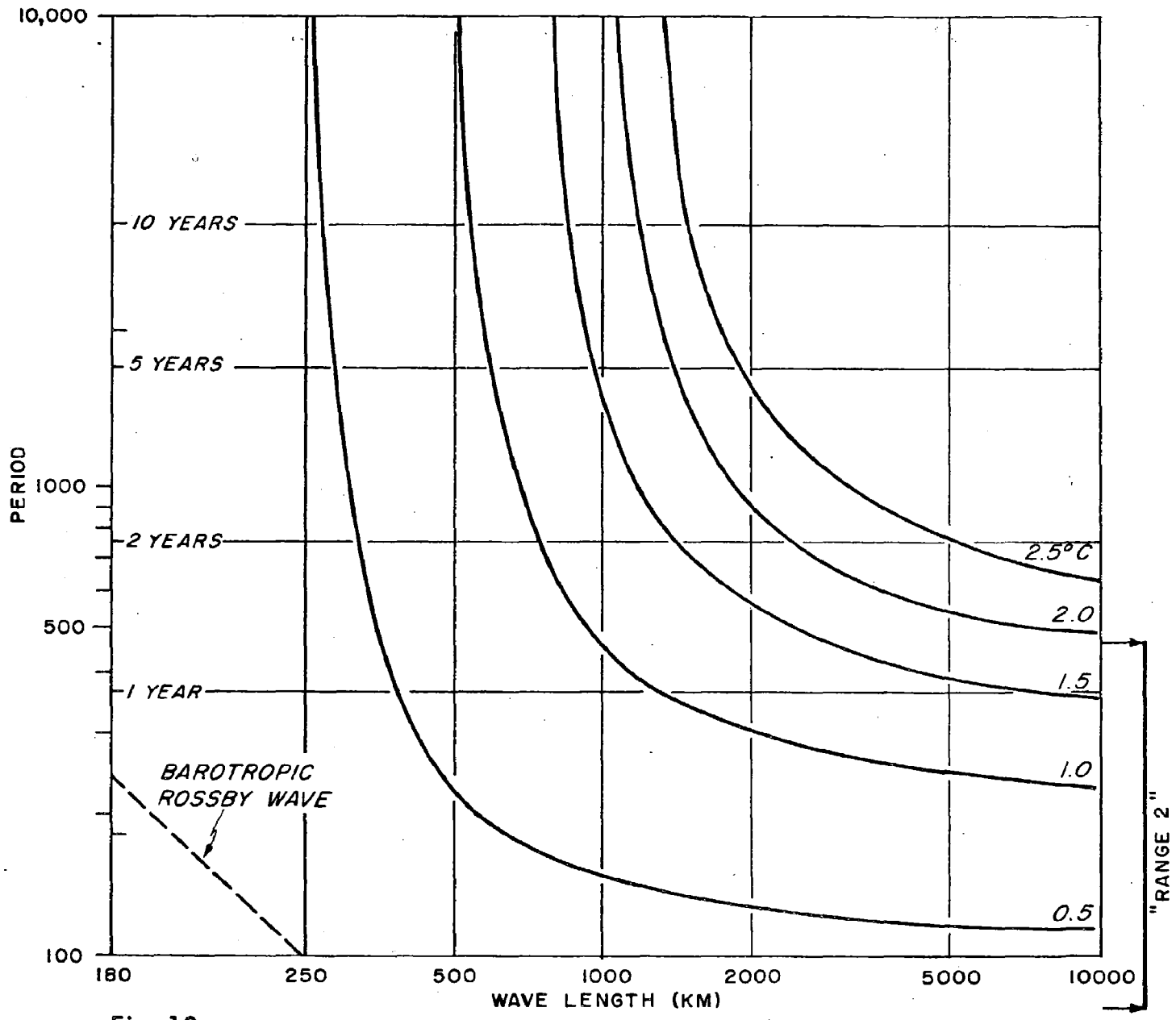
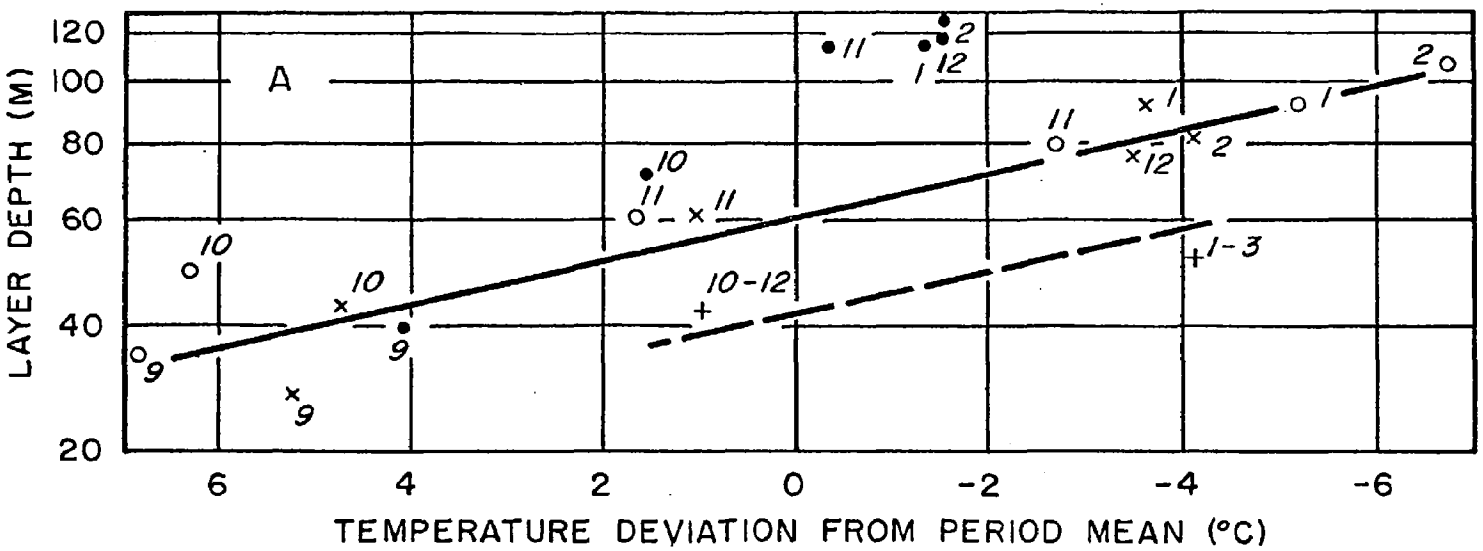

10,000
1000
500
100
PERIOD
10 YEARS
5 YEARS
2 YEARS
1 YEAR
BAROTROPIC
ROSSBY WAVE
2.5°C
2.0
1.5
1.0
0.5
"RANGE 2"
180
250
500
1000
2000
5000
10000
WAVE LENGTH (KM)

Fig. 10



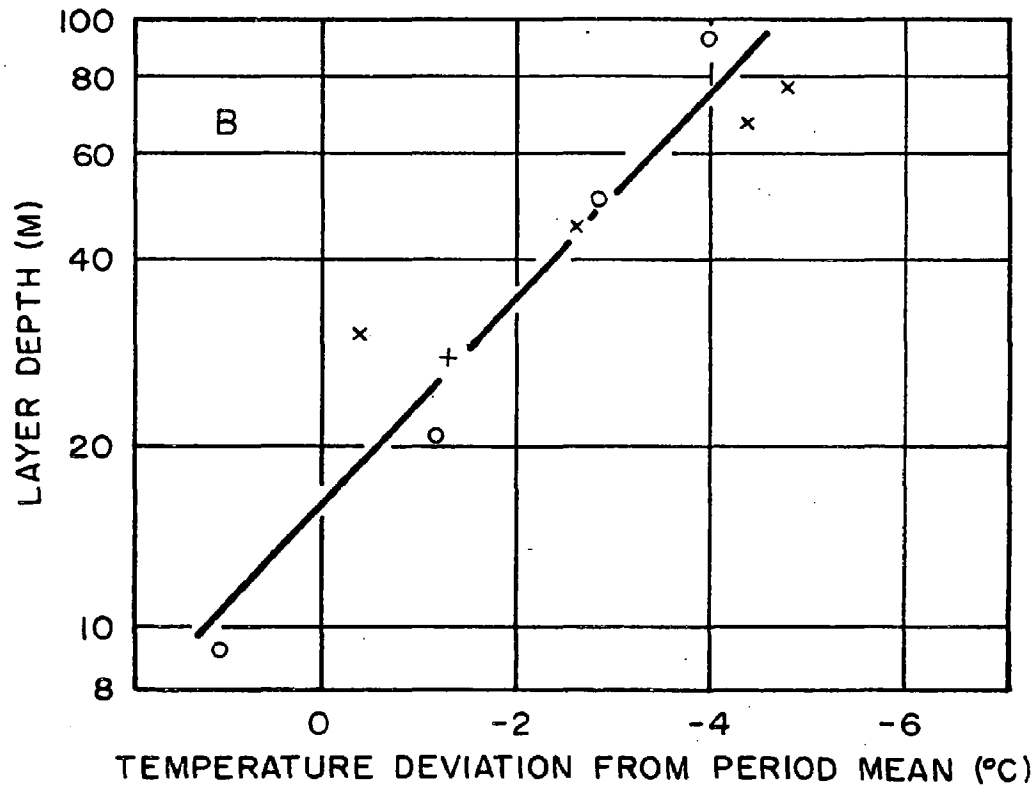


FIG. 11

x *MIDDLE LATITUDES OF NORTH PACIFIC*

\+ *ZONE 35°–45° OF NORTH PACIFIC*

o *BERMUDA AREA*

• *STATION "PAPA" 50° N 145° W*

NUMERALS INDICATE MONTHS

A. SEPT. THROUGH FEB.

B. MAR. THROUGH JUNE

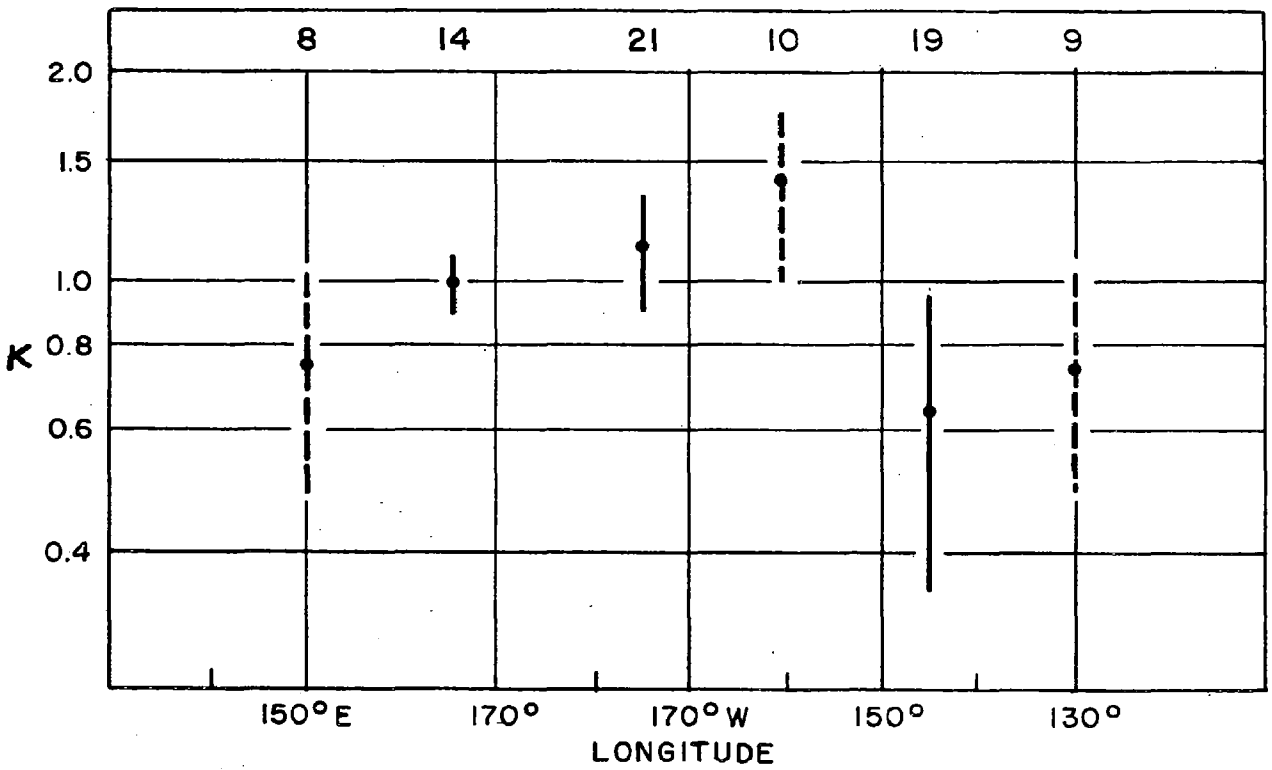


Fig. 8

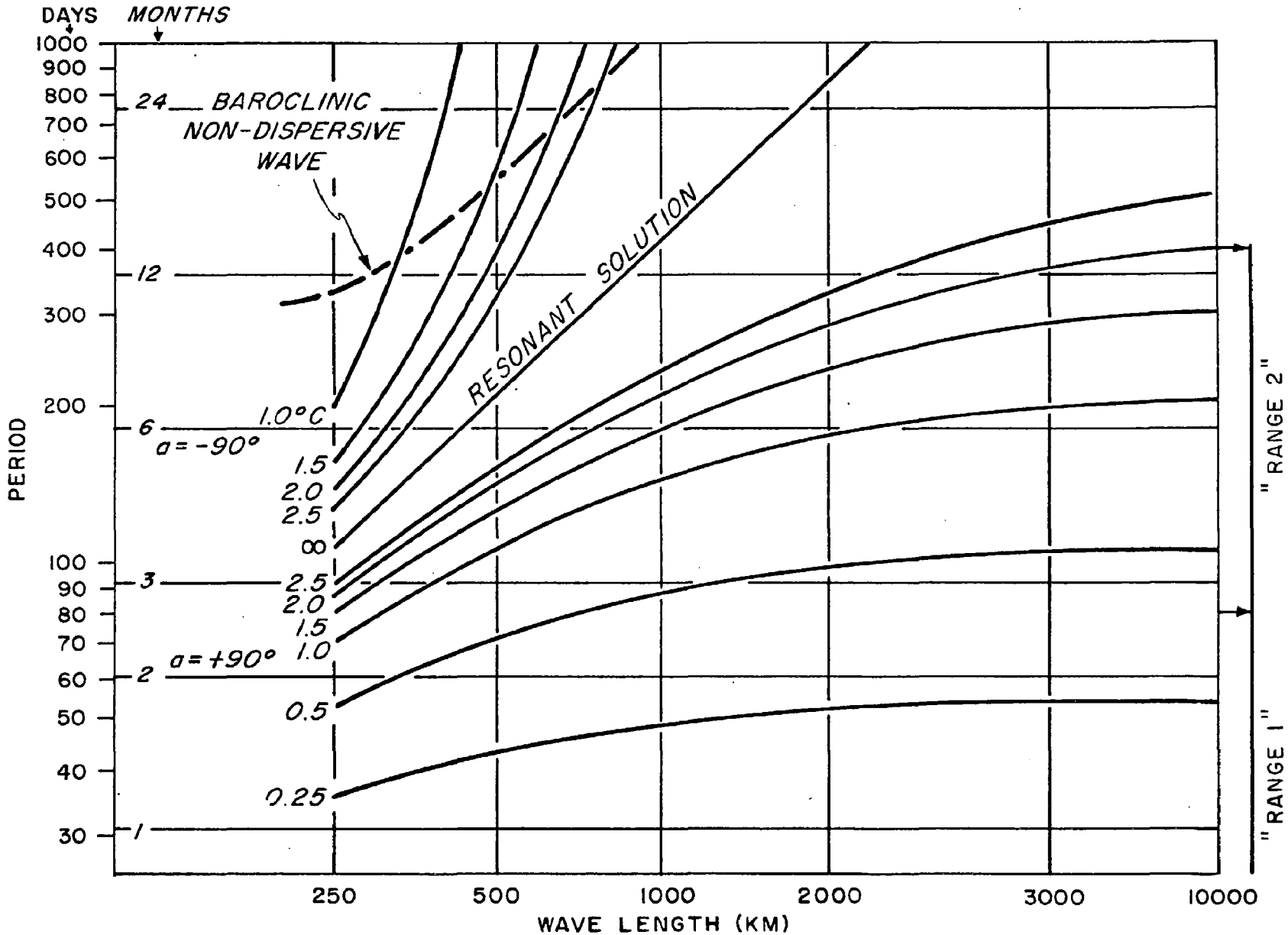

DAYS
MONTHS
PERIOD
1000
900
800
700
600
500
400
300
200
100
90
80
70
60
50
40
30
24
12
6
3
2
1
BAROCLINIC
NON-DISPERSIVE
WAVE
RESONANT SOLUTION
1.0°C
1.5
2.0
2.5
∞
2.5
2.0
1.5
1.0
0.5
0.25
a = -90°
a = +90°
250
500
1000
2000
3000
10000
WAVE LENGTH (KM)
" RANGE 2"
" RANGE 1"

Fig. 9