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THESIS

VARIABILITY WITHIN THE OCEAN-ATMOSPHERIC SYSTEM
OVER THE NORTH PACIFIC

Submitted by

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ABSTRACT OF THESIS

VARIABILITY WITHIN THE OCEAN-ATMOSPHERIC SYSTEM OVER THE NORTH PACIFIC

Characteristics of the oceanic mixed layer over the North Pacific have been examined utilizing a number of statistical methods. Based on the analyses of twelve years of data, a quasi-meridional differentiation (QMD) in SST spectra across the North Pacific was observed. The SST spectra became increasingly "red" as an increasing function of latitude. A strong 21-26 day cycle in SST anomalies is discussed which may be a reflection of heat fluxes. These fluxes also vacillate significantly on this time-scale in conjunction with cycles observed in the atmospheric energy modes of available potential and kinetic energy.

Examination of an oceanic heat budget on a spatial and temporal basis suggest that the impact of latent and sensible heat fluxes upon Δ SST is partially a function of the magnitude of the heat fluxes as well as of the depth to which their effects are mixed. The heat budget analyses and the fitting of power spectra of SST anomalies over the North Pacific to a two-parameter oceanic model, suggest that SST behavior over the mid-oceanic regions of the North Pacific is dominated by the influence of latent and sensible heat fluxes. On the other hand, over the remainder of the North Pacific one could surmise that other processes, such as advection of heat within the ocean, the entrainment heat flux at the base of the mixed layer, and radiation are at least as important in determining the behavior of SST's.

By analyzing anomalous patterns of atmospheric thickness and SST's, it appears that the modification of air masses as they are advected over oceanic waters, as well as the stability of the lower atmosphere, are instrumental factors in determining the nature of large-scale air-sea heat exchange processes.

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I. Introduction

Numerous investigations within the last few years have focused a great deal of attention on understanding the interannual variability of the atmosphere's general circulation. To a large degree usage and production of natural resources are dependent upon the variation of weather from year to year. With the world's rapidly increasing population and demand for these resources, the area of climate variability, on all time scales, has become even a more popular and important research topic.

With the improvement of the historical records of sea surface temperatures (SST) within the past few years, we are slowly learning of the ocean's importance in influencing the global climate. No longer can we ignore the fact that the oceans, having a greater density and a high heat capacity, store most of the sun's radiant energy for later release to the atmosphere which is in contact with the oceans over 72 percent of the earth. Because the natural time scales of these two media vary from the order of days and weeks for the atmosphere to months and years for the ocean, scientists feel that changes in the atmosphere should be possible to predict from oceanic behavior. Assuming the atmosphere is sensitive to its bottom boundary, the oceans could act as an initiator of climate change. On the other hand, some scientists feel that the ocean's major role is to act as a low-pass filter on the high frequency fluctuations of the atmospheric field. This view suggests that the oceans act as a stabilizing agent on global climate. These examples of how the atmosphere and ocean can interact represent only the extreme choices within a broad spectrum of interactions operating on all time and space scales. To better understand what the dominant interactions

are within the ocean-atmosphere system and how and when their impact is felt, this study examines many of the characteristics and interactions of this complex, coupled system. In performing this investigation our analyses include a review of several statistical features of SST's, as well as a detailed examination of portions of the oceanic heat budget over the North Pacific.

The North Pacific was chosen as the area of interest for this study primarily because of the downstream barotropic response the climate of North America has to atmospheric circulations over this region. Another fascinating feature of this region, which has drawn the attention of numerous researchers, is that several atmospheric parameters exhibit the highest variability observed in the northern hemisphere over the central North Pacific. For example, maps of standard deviations of January surface pressures from 1899 to 1960 show a maximum value in the northern hemisphere of 10 mb located in the Gulf of Alaska (Namias, 1968). Likewise from a study conducted here at Colorado State University (Anonymous, 1979), a variation of nearly 20° of latitude is observed in the vicinity of 170°W in the mean position of the -30°C isotherm at 500 mb for three consecutive Januaries (Fig. 1.1). Yet another good illustration of this interannual variability is shown in Fig. 1.2, which examines four cold seasons (1958-1962) of sensible heat transport across the latitude 45°N . From this figure one can observe an interannual variability of nearly 100 percent in the magnitude of maximum transport between the longitudes 140°E - 120°W , which encompass the North Pacific, as well as a strong variability in the location of this maximum transport of sensible heat. Also in Fig. 1.2, one can surmise the location

Fig. 1.1 Monthly mean position of -30°C isotherm at 500 mb for three consecutive Januarys.

Fig. 1.2 Averages for four consecutive cold seasons (October-March) of contributions in each 5° longitude zone to poleward sensible heat transport across latitude 45°N. [From D.A. Haines and W.S. Winston, *Monthly Weather Review*, 91(7): 325 (1963).]

of two quasi-stationary long-wave troughs which are orographically induced by large mountain ranges. The more dominant of these two troughs, centered at 130°E , exists downstream from the Himalayan range, while the other is induced by the Rocky Mountains and is often found near 80°W , i.e. over the Eastern United States. The general circulation over the northern hemisphere is largely controlled by these two long wave orographically induced, quasi-stationary troughs. Between these two quasi-stationary long waves, which are asymmetrically located with respect to geographical longitude, other long-wave patterns (i.e. resonant Rossby waves) are observed which are far more oscillatory and irregular in nature. The strong variability in the atmosphere's circulation over the North Pacific, which has just been described, is a prime example of this oscillatory wave phenomenon. The existence and characteristics of these resonating Rossby waves over the North Pacific depend upon the mean speed of the circumpolar westerlies, and furthermore upon an outside forcing which exhibits a variable input into the system. In fact, many scientists have speculated that the vast reservoir of energy, which the Pacific ocean represents, may provide this source of variable forcing to the atmosphere which under certain conditions becomes sensitive to such inputs. Hopefully by examining the variability of the oceanic mixed layer over the North Pacific, fuller understanding of these oceanic inputs will be gained.

II. Data and Computational Procedures

A. The Data Sets

Accurate assessment of the variability of a coupled air-sea system requires an extensive data base for a number of parameters. Unfortunately the data bases pertinent to such a study are unreliable in their quality and very limited in extent prior to 1970. The computations in this paper were carried out with data furnished by the Fleet Numerical Weather Central (FWNC) and by the National Oceanographic Data Center (NODC). FWNC provided eight years (1970-1977) worth of twice daily (00Z and 12Z) values for the following parameters: air temperature, vapor pressures, and wind speed at ship-deck level (assumed to be 10m). Since SST's are one of the most easily measured environmental properties in the sea, their data base goes back several more years. Thus for a portion of the analyses in this study, twelve years (1966-1977) of twice daily values of SST's were used. The SST will be defined here as the average temperature of the surface mixed (turbulent) layer at a defined position and time. The ten-meter wind speeds were evaluated geostrophically from sea-level pressure data, using a quasi-geostrophic model described by Larsen (1975a, b). To better understand the oceans' response to seasonal changes in forcing, it was necessary to obtain expendable bathythermograph (XBT) data for the North Pacific region. These XBT data, obtained from NODC, were analyzed over the years (1970-1977) for which the oceanic heat budget was examined. The XBT data are concentrated along the shipping lanes, the location of which can be surmised from Fig. 2.1. It was inevitably due to a sparsity of XBT data over most oceanic regions, that the effects of a variable mixed layer on

Fig. 2.1 Location of weather observations from merchant ships during March, 1977, and outline of the 21 x 15 grid used exclusively in this study.

the heat budget could only be examined in detail over very limited areas.

The four fields of data from FNWC, which were used in the heat budget calculations, are stored on a 63 x 63 grid-point northern hemispheric polar stereographic grid. Figure 2.2 shows a 47 x 51 grid-point subset of this larger grid with the excess points located in the equatorial latitudes. Much interpolation is involved to produce such a network of data. One can note from Fig. 2.1 the sparsity of ship reports in infrequently navigated ocean areas. In such areas evaluation of SST's are made by comparison with nearby reports, climatology, and time continuity with previous analyses. Wolf et al. (1965) mentions in his discussions of these data bases that also present are instrumental, coding, and analysis errors. To eliminate as much bad data as possible, certain checks were placed upon the data bases before they were analyzed. For example, if the SST change at a grid point was greater than 6°C in a 12 hour period, the grids were visually inspected and a subjective decision on the reliability of the data was made, based on the overall SST patterns. Furthermore, the data were checked to see if they were within realistic limits. Due to mass storage capacities of the computer, the grid used in this investigation was limited to the area 20°N-55°N and 140°E-120°W (outlined in Fig. 2.1). For the sake of convenience the data points were interpolated to every 5° of longitude and 2.5° of latitude, using a four-point Bessel interpolation scheme made available by the computing facilities at NCAR. It was suggested that further smoothing of the actual data field by this interpolation scheme would alter the field's characteristics and thereby affect the results of the heat budget. To test this speculation, the standard

Fig. 2.2 National Meteorological Center 47 x 51 grid of the northern hemisphere.

deviations and means for the four data fields from FNWC previously mentioned were calculated at several points over the grid shown in Fig. 2.1 on a seasonal basis. The results showed that only for the SST field, and then only in regions of strong SST gradients, did the characteristics of the interpolated data field differ appreciably from the actual data field. In fact, the largest differences in the two fields were for a point at 35°N - 160°E , which is near the confluence region of the oceanic gyres. The interpolated SST data had a mean SST at this point that was systematically 0.24°C larger than the actual value, while the standard deviation was greater for the actual data by a factor of 0.026°C . Since the difference in the means of SST was systematic within $\pm 0.06^{\circ}\text{C}$, the parameter ΔSST (change in SST) used in the heat budget calculations should not be significantly affected by the interpolation scheme discussed above. As a consequence, we felt that the characteristics of the interpolated data field would not determinately alter the results of the oceanic heat budget as it was computed in this study.

As one might expect, due to smoothing and errors inherent within the data, the variables in the oceanic heat budget will not represent the exact magnitudes of the actual values. However, by examining a sufficiently large number of cases, the generalizations that are made by utilizing statistical methods take on greater significance. A number of investigators involved in the dynamics of ocean-climate variability, such as Jerome Namias and Elmar Reiter, have used statistical methods extensively to explore and explain the behavior of the coupled air-sea system. In this paper, attempts are made to substantiate statistical inferences with known physical principles, so that the significance

tests performed in this study should be regarded only as "references of reliability".

B. Oceanic Heat Budget

Over many parts of the world's oceans vertical exchange processes between the air and sea, as well as vertical mixing within the water column, are likely to affect the local conditions much more rapidly and effectively than horizontal advection and horizontal mixing. For many purposes one may treat the upper ocean layers as being statistically homogeneous along the horizontal. In other words, the horizontal derivatives can be omitted in a mathematical treatment with only changes in the vertical being considered. Since temperatures and salinities over the North Pacific tend, for the most part, to vary more along a vertical distance of a hundred meters than along a horizontal distance of a thousand kilometers, a one-dimensional treatment of this region can be useful. Probably the most important function of one-dimensional ocean models is the prediction of the mixed layer temperature given by Eq. 2.1.

$$\underbrace{\frac{\Delta SST}{\Delta t}}_A = - \underbrace{\frac{1}{\rho c_p h} (R_o + SH + LH - I_o e^{-vh})}_B - \underbrace{\frac{We \Delta T}{h}}_C \quad (2.1)$$

where

$\Delta SST \equiv$ temperature change of the mixed layer ($^{\circ}C$) during the time interval Δt

$\rho \equiv$ water density = 10^6 g m^{-3}

$c_p \equiv$ specific heat capacity of water = $1 \text{ cal g}^{-1} \text{ } ^{\circ}C^{-1}$

$h \equiv$ depth of mixed layer (m)

$R_0 \equiv$ surface flux of solar and terrestrial radiant energy
(watts m^{-2})

$SH \equiv$ surface flux of sensible heat (watts m^{-2})

$LH \equiv$ surface flux of latent heat (watts m^{-2})

$I_0 e^{-\nu h} \equiv$ penetrating solar radiation which passes into layers
below "h" (watts m^{-2})

$W_e \equiv$ rate at which water is entrained from below mixed layer
(ms^{-1})

$\Delta T \equiv$ quasi-discontinuity in temperature across bottom of
mixed layer ($^{\circ}C$)

This equation is equivalent to the one which Niiler and Kraus (1975) used in their one-dimensional oceanic model. The derivation of Eq. (2.1) (see Appendix 1) is based upon a similar methodology to which Lilly (1968) used to form his atmospheric model for a cloud-topped mixed layer under a strong inversion. Term A in the above equation represents the change in the mixed layer temperature (SST) during Δt . Term B is the surface heat flux minus the penetrating solar radiation which passes through the mixed layer. One should note that term A is negative (net cooling of the mixed layer) when the surface heat fluxes in term B are positive (heat flux directed out of mixed layer). Term C represents the heat flux at the bottom of the mixed layer due to entrainment through the bottom. The schematic sketch shown in Fig. 2.3 may help one to better understand the physical nature of the processes which can affect the mixed layer temperature.

The latent and sensible heat fluxes in Eq. 2.1 were calculated by using the bulk aerodynamic equations suggested by Friehe and Schmitt (1976). In their investigation, they examined several different bulk

Fig. 2.3 Schematic sketch of a simplified mixed layer model showing the underlying processes which can affect the mixed layer temperature.

formula parameterizations by least squares fitting a compilation of existing data. The final equations they proposed for parameterizing sensible and latent heat fluxes are given in Eqs. 2.2 and 2.3.

$$SH = \rho_a c_{p_a} (0.12 + 7.2 \times 10^{-2} (SST - AIRT) \times u_a) \quad (2.2)$$

where

$SH \equiv$ sensible heat flux (watts m^{-2})

$\rho_a \equiv$ air density = $1.16 \times 10^3 \text{ g m}^{-3}$

$c_{p_a} \equiv$ specific heat capacity of dry air = $0.24 \text{ cal g}^{-1} \text{ } ^\circ\text{K}^{-1}$

$SST \equiv$ sea surface temperature ($^\circ\text{K}$)

$AIRT \equiv$ air temperature at 10m ($^\circ\text{K}$)

$u_a \equiv$ wind speed at 10m (knots)

$$LH = L(0.62 u_a) \times \frac{ew}{(SST + 273.16)} - \frac{ea}{(AIRT + 273.16)} \quad (2.3)$$

where

$LH \equiv$ latent heat flux (watts m^{-2})

$L \equiv$ latent heat of vaporization = 597.3 cal g^{-1}

$ew \equiv$ saturation vapor pressure of sea surface (mb)

$ea \equiv$ water vapor pressure at 10m (mb)

The saturation vapor pressure needed in the above computation for latent heat flux is derived from the Clausius-Clapeyron equation,

$$ew = 6.11 e^{5415.23 \left(\frac{1}{273.16} - \frac{1}{SST} \right)}$$

A detailed discussion of the bulk aerodynamic equations is given by Hess (1959).

Previous studies (Tulley, 1964; Bathen, 1972; White and Walker, 1974) have discussed the strong seasonal changes in the depth of the mixed layer. In Tulley's paper, different models are examined for

various oceanic regions which show these seasonal changes in the depth of the thermocline (Fig. 2.4). One can note from this figure that the seasonal variability of the thermocline is especially large at high latitudes. Since the characteristics of the oceanic heat budget and of SST's depend strongly upon the vertical extent of the mixed layer, the author felt that more insight could be gained from this study by separating much of the analyses into seasons which exhibit different mixing layer depths. Hopefully our future research efforts will be more sophisticated in choosing seasons and regions over which the mixed layer varies, but for this study the only stratification of the analyses will be into the heating and cooling seasons of the ocean. These seasons were determined by examining the change between two successive six-day averaged SST's at 273 grid points within the region outlined in Fig. 2.1. The criterion to denote the start of the heating season is that greater than 50 percent of the SST changes at these points be positive (i.e. net warming over a 12-day period). A similar criterion is defined for the onset of the cooling season, except that net cooling had to be observed over the majority of the grid points over a 12-day period. By analyzing 12 years of SST data (1966-77) to determine the onset of the seasons as described above, average dates for these events were determined. Consequently, the average heating season for the North Pacific was found to be March (13-18) to September (3-8), whereas the average cooling season is September (9-14) to March (7-12). One must realize that these arbitrarily defined seasons only represent an average over the years and grid region considered, but in reality the duration of these seasons varies on a regional as well as an interannual basis.

Fig. 2.4 Three oceanic models of the mixed layer depicting their basic oceanographic structures and the processes affecting them. [After Tulley, 1964.]

ARCTIC MODEL

SUBARCTIC MODEL

SUBTROPIC MODEL

III. Characteristics of North Pacific SST

A. Statistical Properties of Δ SST and Related Fields.

Upon initial observation, SST variations over various oceanic regions appear to result from a complex set of feedback mechanisms within a coupled air-sea system. In Namias' (1968) words, "The physical sciences of the atmosphere and of the ocean are each extremely complex; the physical sciences of the combined atmospheric-ocean system are more than twice as complex". On the other hand, other investigators such as Tulley (1964) have suggested that many of the features which are observed in SST's are the result of a few relatively simple processes acting in a synergistic manner. Whatever be the case, this study does not attempt to do an in-depth investigation of the interactions between these processes. However, within the limitations of the oceanic data base described in the previous chapter, it does attempt to characterize on a regional basis a number of statistical aspects, as well as the dominant time scales of SST's over the North Pacific. In addition, speculations are made in this chapter as to what the dominant physical processes are which give rise to the characteristic SST features.

One of the major reasons for computing statistical parameters characterizing SST's over various oceanic regions is to provide data so that future studies can more adequately model a coupled air-sea system. The statistical parameters were calculated for forty points over the North Pacific grid which is outlined in Fig. 2.1. These points coincided with the intersection of every 10° of latitude and longitude.

To gain some insight into the heat budget parameter, Δ SST (SST change over a six-day period), its variance field was computed for the

three-year period (1975-77). The spatial pattern formed by the variances of Δ SST without seasonal bias (Fig. 3.1) corresponds well with that found by Davis (1976) in the variance of SST anomalies for the 28-year period (1947-74). The spatial distributions of variances of Δ SST, separated into their heating and cooling seasons, are shown in Figs. 3.2 and 3.3, respectively. Although these two figures exhibit similar spatial patterns irrespective of season, the magnitudes of the variances are larger in the heating season. This latter fact may be a result of the deeper mixed layer which characterizes the cooling season (see Fig. 3.4 which shows the depths of the mixed layer near the ends of the heating and cooling seasons). Thus, one can surmise that once a deep mixed layer is established in the cooling season, the changes in SST's due to heat fluxes into (or out of) the mixed layer are significantly reduced by the greater vertical dilution that occurs during the cooling season. To postulate an explanation for the spatial patterns in the variance fields shown in Figs. 3.1-3.3, other variables which influence Δ SST's (namely wind speed, air-sea temperature difference, air-sea moisture gradient) are examined.

The spatial distributions of the means and variances of these other three variables, which were calculated for the period (1975-77) using their six-day averaged values, are shown in Figs. 3.5-3.10. From Figs. 3.5 and 3.6, which represent the mean and variance fields of wind speed respectively, one can surmise the location of strong cyclone activity during these years from areas where large variances and strong mean winds are observed. The frequency of cyclonic centers over the North Pacific for this period would most likely resemble that which was observed during the period (1899-1940) shown in Fig. 3.11. Furthermore, a

Fig. 3.1 Spatial distribution of variances of ΔSST ($\times 10^2$) for the period 1975-77 without seasonal bias.

Fig. 3.2 Spatial distribution of variances of ΔSST ($\times 10^1$) for the cooling seasons for the period 1975-77.

Fig. 3.3 Same as Fig. 3.2, except for heating seasons.

Fig. 3.4 Mixed layer depth (meters) near end of heating and cooling seasons. [After Pattullo and Cochrane, 1951.]

Fig. 3.5 Spatial distribution of mean wind speeds (knots) for the period 1975-77.

Fig. 3.6 Spatial distribution of variances of wind speeds for the period 1975-77.

Fig. 3.7 Same as Fig. 3.5, except for air-sea temperature difference ($^{\circ}\text{C}$).

Fig. 3.8 Same as Fig. 3.6, except for air-sea temperature difference.

Fig. 3.9 Same as Fig. 3.5, except for air-sea moisture gradient (mb).

Fig. 3.10 Same as Fig. 3.6, except for air-sea moisture gradients.

Fig. 3.11 Percentage frequency of centers of cyclones during the summer and winter seasons for the period 1899-1940. [After Petterssen, 1956.]

study by Willebrand (1978) revealed that strongest fluctuations in the mean wind stress field existed over the northern ocean basin where traveling cyclones dominate the atmospheric variability. Although the statistical parameters computed from the wind speed field are informative as to the nature of baroclinic activity over the North Pacific, they alone offer little explanation for the observed spatial distribution in the variances of ΔSST . Figures 3.7-3.10, which represent the mean and variance fields of the air-to-sea gradients of temperature and moisture respectively, suggest a partial explanation for the observed spatial distribution of variances in Fig. 3.1. One should note that the air-sea temperature difference in the sensible heat transfer equation (Eq. 2.2) and the moisture gradient in the latent heat transfer equation (Eq. 2.3) were found by Seigel (1977) to be the dominant parameters which control these heat fluxes. Therefore, the similarities between the spatial patterns of the variances in Figs. 3.8 and 3.10 and that of Fig. 3.1 may be indicative of a relationship between latent and sensible heat fluxes and the variance of ΔSST . More specifically the similarities in the figures previously referred to are observed as a maximum in the variances off the east coast of the Asian continent, which then decrease to the east and away from the mid-latitudinal regions of the ocean. The relationship between the heat fluxes mentioned above and the entire spectra of SST's will be examined with greater detail later in the chapter. It is also of interest to note the effects of cold continental air masses being advected over the warmer waters of the northwest Pacific from Fig. 3.7., whereas from Fig. 3.9 one can deduce the existence of the subtropical ridge by the large moisture differences

across 20°N which results from the dry air within this region of strong subsidence.

A more likely hypothesis for explaining the observed spatial pattern in the variances of ΔSST can be obtained by examining the location of oceanic fronts. An extensive investigation of oceanic fronts over the North Pacific and the principles of their frontogenetic processes is described in a work by Roden (1975). He concludes that fronts are features of the upper ocean and occur near mass transport and energy flux convergence zones. By presenting the magnitudes of meridional SST gradients in Fig. 3.12, one can surmise the relative location of oceanic fronts along five longitudes across the North Pacific. One can note from this figure that the regions of largest variance in ΔSST (see Fig. 3.1) coincide with the location of the strongest oceanic temperature gradients. In other words, the large variances of ΔSST at several points over the North Pacific are probably a manifestation of the meanderings of the oceanic fronts through those points. These frontal meanderings are likely the result of perturbations in the mass transport and energy flux divergence zones due to the atmospheric variability manifested in the variance fields shown in Figs. 3.6, 3.8 and 3.10. The maximum of variances in ΔSST along 40°N and over the western portions of the North Pacific ocean is likely the outcome of the meanders of the subarctic temperature and salinity front and the Kuroshio Extension. On the other hand, the secondary maximum of variances of ΔSST in the vicinity of 20°N and $120\text{--}130^{\circ}\text{W}$ is likely the result of meanders in the California Current and in the Subtropical Front which tend to converge in this region as shown in Fig. 3.13.

Fig. 3.12 Time sections on a six-day basis of meridional gradients of SST at five different longitudes across the North Pacific (in $^{\circ}\text{C}$ per 5° latitude intervals) as a function of latitude.

Fig. 3.13 Schematic map of main North Pacific fronts. Arrows indicate prevailing current directions. Letters refer to the following fronts: A, Kuroshio front; B, Oyashio front; C, subarctic front; D, subtropical front; E, doldrum front.

B. Temporal Behavior of SST

Obtaining a proper perspective for the total spectrum of variations for both air and sea phenomena remains as a necessary step toward a complete understanding of air-sea interaction. The present section undertakes a portion of this formidable task through an investigation of the time dependent behavior of SST's. Since previous investigations of air-sea interactions have primarily concentrated on longer time scale phenomena (i.e. monthly or longer), this present study places a stronger emphasis on examining SST variability at synoptic time scales. Our investigation utilizes a variety of analysis methods on the following two features of the SST field: 1) the six-day change in SST at a point represented by ΔSST , and 2) SST anomalies generated from six-day filtered SST data. Discussion of the different analysis methods will be given throughout this section as they are used.

To gain an appreciation for the responsive nature of the surface temperature fields within the ocean, the temporal autocorrelations of ΔSST over the North Pacific are investigated. By noting the relationship between ΔSST and the various heat fluxes in Eq. 2.1, one could also perceive this investigation as an examination of the temporal structure of the variations in total heat flux into (and out of) the mixed layer. This view, of course, is true only to the extent that the approximations inherent in the derivation of Eq. 2.1 are valid. In performing this investigation, autocorrelations for ΔSST up to lag ten (60 days) were computed for the same forty points over the North Pacific as referred to earlier in this chapter. These computations were performed using data from the years 1970-77 as an entire set, but autocorrelations for

shorter periods within these eight years were also computed to examine the shorter term variability of the autocorrelation function.

The outstanding feature in the results of these computations is the predominance of a negative autocorrelation at lag one which is considerably larger in magnitude than at the other lags. This observation is quantified in Fig. 3.14 where the values of the autocorrelations up to lag ten, which were computed for the eight-year period (1970-77), are listed. The values in Fig. 3.14 represent the averages computed at each of these lags from the autocorrelations at the forty points described earlier in this chapter. These autocorrelations were averaged using the Z-transformation suggested by Panofsky and Brier (1958), which allows one to average a population of correlation coefficients that are not from a normal distribution. The Z-transformation is discussed in detail in the following chapter. The spatial distribution of these lag one autocorrelations, which is shown in Fig. 3.15, reveals a north to south increase in the magnitude of the negative autocorrelations with only a slight longitudinal dependence in this gradient. From Figs. 3.14 and 3.15 one might surmise that the total heat flux into the mixed layer (or ΔSST) exhibits a tendency to reverse itself from one six-day period to the next, and this tendency becomes increasingly pronounced at lower latitudes. In fact, a comparison of the time series of ΔSST 's from two oceanic regions (Fig. 3.16) shows that the fluctuations in the sign of ΔSST occur nearly twice as often in the region located at lower latitudes. These results, in particular the spatial pattern of lag one autocorrelations of ΔSST , take on additional significance when they are considered in view of the characteristics of the entire temporal spectra of SST's across the North Pacific.

Fig. 3.14 Forty-point average autocorrelations of Δ SST up to lag ten (60 days) for the period 1970-77.

Fig. 3.15 Spatial distribution of lag one autocorrelations of Δ SST for the period 1970-77.

Fig. 3.16 Time series of average ΔSST (six-day change in SST in °C) for two areas within the North Pacific.

The relationship between SST spectra and the lag one autocorrelation of Δ SST can best be understood with a brief review of the inherent properties of these statistical parameters. The value of the lag one autocorrelation of Δ SST represents the integrated area under a cosine-weighted power spectrum for Δ SST. This cosine weight, shown in Fig. 3.17, arises from the inverse transform of Eq. 3.1 which defines the relationship between the Δ SST power spectrum, $G_{\Delta\text{SST}}(f)$, and the autocorrelation function $R(\tau)$. This equation and its transform for the lag one autocorrelation function (Eq. 3.2) are as follows:

$$G_{\Delta\text{SST}}(f) = 2 \sum_{\tau} R(\tau) \cos(f\tau) \quad (3.1)$$

$$R(1) \propto \sum_f G_{\Delta\text{SST}}(f) \cos(f) \quad (3.2)$$

where

$\tau \equiv$ lag interval

$f \equiv$ frequency

From Fig. 3.17 one can see that the lag one autocorrelation function would weigh all frequencies of the Δ SST power spectrum lower than .25 cycles per six days (or 1 cycle per 24 days) in a positive sense, whereas all frequencies higher than this value would be weighed negatively. To complete this review of statistical parameters, one should now consider the relationship of $G_{\Delta\text{SST}}(f)$ to the power spectrum of SST's, $G_{\text{SST}}(f)$. By virtue of the difference operator Δ , which was applied to the original time series of SST's, $G_{\Delta\text{SST}}(f)$ actually represents a high-pass filtered spectrum of SST's. This operator is defined as:

$$\Delta(\text{SST}(t)) = \text{SST}(t) - \text{SST}(t-6)$$

where "t" is in days. The transfer function of Δ is given by Koopmans (1974) as:

Fig. 3.17 Cosine weight inherent within the lag one autocorrelation function.

$$H(\lambda) = 2 |\sin(\lambda/2)| e^{i(\pi-\lambda/2)}, \lambda > 0$$

where

$|H(\lambda)| = 2 |\sin(\lambda/2)|$ is the gain function

$\theta(\lambda) = (\pi-\lambda)/2$ is the phase function

$$\lambda = (2\pi x)/N \quad x = 0, 1, 2, 3, \dots, N/2$$

$N \equiv$ any large integer (e.g. $N=732$).

The alteration of the power spectrum $G_{SST}(f)$ can be represented at any given frequency by:

$$G_{\Delta SST}(f) = |H(f)|^2 G_{SST}(f)$$

This squared gain function, $|H(f)|^2$, is shown in Fig. 3.18.

Through recognition of the functions of the difference operator as a high-pass filter on the SST spectra and the cosine weighting inherent in the lag one autocorrelation of ΔSST , one can surmise that the values of the autocorrelations in Fig. 3.15 are a reflection of characteristics in the shape of SST power spectra across the North Pacific. The nature of this reflection becomes more evident when one considers the effects of the filter and weighting functions on a "red" noise versus a "white" noise spectrum. The color of noise in these spectra is analogous to the wave length characteristics in similar shades of light in the optical spectrum. In other words, a "white" noise spectrum has equal power (variance) in all frequency intervals, whereas a "red" noise spectrum is characterized by relatively greater power (variance) at lower frequencies than at the higher frequency range of the spectrum. The effects of high-pass filtering a "white" noise SST spectrum, as is done through application of the difference operator, would result in a "blue" noise spectrum of ΔSST . "Blue" noise implies greater power at higher frequencies. On the other hand, this type of high-pass filtering applied to a

Fig. 3.18 Squared gain function which results from difference filter $-\Delta$.

"red" noise SST spectrum would yield a relatively "white" Δ SST spectrum. By incorporating the cosine weighting effects on these filtered spectra, one can see that a "white" Δ SST spectrum would result in a lag one value for Δ SST equal to zero, since the positive contributions from frequencies lower than .25 cycles per six days would exactly cancel the negative contributions from higher frequencies. However, when cosine weighting a "blue" Δ SST spectrum, the value of the lag one autocorrelation of Δ SST would now be negative, for in this latter case the negative contribution from higher frequencies is far more dominant. In summary, it appears likely that the quasi-meridional gradient of negative lag one autocorrelations of Δ SST in Fig. 3.15 is indicative of an increasing "redness" in the spectra of SST's across the North Pacific with increasing latitude. To substantiate these findings, power spectra of SST's across the North Pacific were computed. Furthermore a spectral analysis scheme, as described in the following section, has the advantage of more precisely defining the temporal scales of variability which exist in these SST's.

C. Spectrum Analyses.

Spectrum analyses were performed on SST anomalies at the forty points over the North Pacific which were previously referred to. The anomalies were obtained from the time series of daily SST's which were smoothed using a six-day running mean filter. Ideally, a low-pass filter as such would leave the variance in time scales longer than six days unaltered, while the variance in time scales shorter than six days would hopefully be completely attenuated. This filter effectively reduces the spurious noise, which often characterizes SST's at very high frequencies due to sampling and data collection errors, yet still

permits the examination of SST anomalies on relatively short (i.e. synoptic) time scales. In reality, though, when such a filter is harmonically analyzed, the transfer function for the filter only approximates the discontinuous jump from 1 to 0 at the cutoff frequency. The power spectrum of the filtered time series will differ from that of the actual power spectrum above the cutoff frequency. This alteration in the power spectrum calculated from the filtered series can be represented at a given frequency by the product of the square of the gain (shown in Fig. 3.19) and the power spectrum calculated from the unfiltered time series. The transfer function for a six-day filter is given by:

$$H_j = \frac{1}{6} \sum_{m=-2}^{m=3} e^{-i \frac{2\pi j}{N} m} \quad j = 0, 1, 2 \dots \frac{N}{2}$$

where

$$H_j = |H_j| e^{i\theta_j}$$

$|H_j| \equiv$ gain of transfer function

$\theta \equiv$ phase of transfer function

$N \equiv$ integer (for the computation of Fig. 3.19, $N = 732$).

Figure 3.19 shows the gain and phase of the transfer function plotted against frequency. Based on the characteristics of the transfer function shown in this figure, one can determine that the alteration of the actual spectrum exceeds 60 percent attenuation at frequencies higher than the cutoff frequency. To eliminate the analysis of strongly attenuated frequency intervals and to reduce the amount of data handled, a subset of the original filtered time series was generated by resampling the original series at six-day intervals. Decimating the series in this manner resulted in a cutoff and Nyquist frequency which now coincide at

Fig. 3:19 Gain and phase of the transfer function for a six-day running mean filter.

PHASE OF TRANSFER FUNCTION FOR $N = 732$ GAIN OF TRANSFER FUNCTION FOR $N = 732$

0.5 cycles per six days with a spectral resolution in frequency space of 0.0014 cycles per six days. Before any power spectra are examined, it is useful to review the procedure by which they were calculated.

Spectrum analysis is a statistical method by which harmonic analysis schemes are applied to a time series. The analyzed spectrum shows the amounts of variance that are contributed to the total variance by oscillations at different frequencies. One should realize in using this statistical technique, that the true spectrum is a property of the infinite series, which can only be estimated from finite samples. The spectrum analysis method used in this study will reveal the dominant time scales of variability which show up consistently in the years over which the time series of the finite sample spans.

The procedural steps followed for computation of power spectral density estimates via fast fourier transforms are given below:

- 1) Removal of the mean from the data series, which eliminates the strong peak which might otherwise appear in the real fourier component at zero frequency leading to contamination of low frequency spectral estimates.
- 2) Detrend the data series, which eliminates low frequency components whose nature cannot be resolved within the length of the entire series.
- 3) Perform a cosine tapering over one-tenth at each end of the data series as discussed by Bendat and Piersol (1971). This tapering is done to suppress leakage in the power spectral estimates arising from a finite sample length.
- 4) Calculate the power spectral density estimates $\hat{G}(f)$,
 where $\hat{G}(f) = 2 \times (A^2 + B^2) N^{-1}$

$A \equiv$ real Fourier component

$B \equiv$ imaginary Fourier component

$N \equiv$ number of points in the data series.

- 5) Multiply these estimates from the previous step by $1/0.875$ to recover variance lost to cosine tapering.
- 6) Apply a smoothing to the spectrum by averaging over adjacent frequency estimates, so that a consistent estimate of the "true" spectrum can be obtained.

The resulting number of spectral degrees of freedom for a spectral estimate is given by Bendat and Piersol (1971) to be twice the number of estimates smoothed over. This number should in turn be reduced by 10% due to the cosine tapering (Julian, 1971) to yield DOF^* , the degrees of freedom for each spectral estimate.

Within this study there are several cases in which power spectra over the North Pacific are averaged together on a regional basis. Such a technique of ensemble averaging is useful for grouping similar spectra into categories so that characteristic variabilities can be differentiated between the groups. When averaging power spectra together in this manner, the effective number of degrees of freedom increases for the ensembled spectral estimates. This increase is given by the following equation:

$$DOF = (DOF^*) \times \overline{ns} \quad (3.3)$$

where $DOF^* \equiv$ number of degrees of freedom adjusted for smoothing and cosine tapering,

$\overline{ns} \equiv$ number of independent spectra over which the averaging is done.

Although the relationship in Eq. 3.3 assumes that the spectra which are averaged together are in fact independent, this criterion is likely

to be invalidated for the cases examined in this study, by the spatial coherence that exists within the data fields analyzed. Achieving a more accurate approximation to the actual number of degrees of freedom via knowledge of the spatial correlations in the various data fields presents itself as a considerable task. In fact, it was felt that this more rigorous approach would not yield enough of an improvement in the significance of the results, when compared to the effort and resources needed to generate them. Throughout this study DOF will, therefore, represent the upper limit for the actual degrees of freedom which exist for a spectral estimate. In all likelihood this value of DOF will be erroneously large due to the spatial dependence of the spectra estimates which are averaged together. It would follow that the confidence limits, which depend upon DOF, will be fallaciously narrow, giving the appearance of greater significance for some estimates than actually exist. If in fact the significance of a spectral estimate is ambiguous, yet necessary to elucidate a crucial point, further evidence will be given at that time to provide a firmer basis for any conclusions.

Before any power spectra are examined, it is useful to review how the confidence intervals were generated. These confidence intervals should express at a given confidence level how close to the values of the hypothetical spectrum one would expect to find sample estimates based on a given number of degrees of freedom. The $(1-\alpha)$ confidence interval, assuming that the hypothetical spectrum is in fact the true spectrum, $G(f)$, is given by the following expression:

$$\left[\frac{\chi^2_{n(1-\alpha/2)}}{n} G(f) < \hat{G}(f) \leq \frac{\chi^2_{n(\alpha/2)}}{n} G(f) \right]$$

where $n = \text{DOF}$

$\hat{G}(f) \equiv$ estimated spectrum

$G(f) \equiv$ hypothetical spectrum

$\chi^2_{n\alpha} \equiv$ the Chi Square value such that $\text{Prob}[\chi^2_n < \chi^2_{n\alpha}] = \alpha$.

In other words, given the validity of the assumptions made concerning the sampling distribution, one would expect the sample estimate $\hat{G}(f)$ to lie within this interval of the hypothetical value $(1-\alpha) \times 100\%$ of the time if the hypothetical spectrum is the true spectrum. One should note that this confidence interval depends upon $G(f)$. And since $G(f)$ is not independent of frequency, the width of the interval changes with frequency. For the reader's convenience the confidence limits are considered for the logarithm of $G(f)$, in which case the interval becomes independent of frequency. Thus by examining the spectra plotted versus frequency on semilog coordinates, the confidence interval once represented can be reused as the appropriate limits at any point on the hypothetical spectrum.

In this study the hypothetical spectrum, $G(f)$, is represented by a two parameter, aliased Markov spectrum which Reynolds (1977) gives as:

$$G(f) = \frac{2v}{1 + \alpha^2 - 2\alpha \cos(2\pi f)} \quad (3.4)$$

where $v \equiv$ variance of the "white" noise forcing

$\alpha \equiv$ autocorrelation of the first lag

$f \equiv$ frequency

In general Eq. 3.4 represents the spectrum of a discrete first-order regression process which is forced by "white" noise. When the lag one autocorrelation equals zero ($\alpha = 0$) in Eq. 3.4, this relationship becomes the spectrum for "white" noise. Since the variance of the "white"

noise forcing, v , was unavailable for this study, it was necessary to calculate $G(f)$ using the variance of the SST time series, σ_{SST} , and the lag one autocorrelation of this series, α . This allowance is possible because Reynolds (1977) has shown that the three parameters: σ_{SST} , α , and v are related in such a manner so that only two of them are independent, while the third parameter is uniquely determined by the other two. Frankignoul and Hasselmann (1977) have demonstrated that the hypothetical spectrum, $G(f)$, can represent a spectrum of SST anomalies generated for a model of a simplified atmospheric and oceanic system. The $G(f)$ inherent within their model reflects the integral response of SST anomalies which are driven by random "white" noise atmospheric forcing. Inclusive in their model is a negative stabilizing feedback to balance the atmospheric input and maintain a statistically stationary response. The derivation of $G(f)$ in Eq. 3.4 from Frankignoul and Hasselmann's model is shown in Appendix 2. The implications of their work on this study becomes more obvious later in this chapter when actual SST spectra from the North Pacific are considered with $G(f)$ as the null hypothesis (i.e. the model generated spectrum $G(f)$ fits the actual SST spectrum perfectly).

D. Results of Spectrum Analysis.

Earlier discussions in this chapter have shown that the spatial pattern of the lag one autocorrelation of ΔSST is in some way indicative of the shape of SST spectra across the North Pacific. More specifically, these results indicated that an increasing proportion of the variance in SST spectra exists at lower frequencies (relative to higher frequencies) as an increasing function of latitude over the North Pacific. Based on this premise, the power spectra of SST anomalies

analyzed for the period (1966-77) were partitioned in a quasi-zonal sense and then averaged to determine if characteristic variabilities could be identified among the ensemble-averaged spectra. This partitioning was achieved by attempting to rationally group the spectra so that those which were averaged together within a region would represent independent realizations of similar physical processes. The result of this regional grouping is shown in Fig. 3.20 and is based upon the following criteria. Area 1 in Fig. 3.20 represents the region influenced by the convergence of the subarctic and subtropical oceanic gyres. Due to the number and complexity of processes that are affecting Area 1, this region requires a study in itself. In view of this fact, Area 1 for the most part is eliminated from the discussions that follow on the averaged spectra from the five regions shown in Fig. 3.20. Area 2 is a region defined from climatology (see Fig. 3.21) by Tulley (1964) in which precipitation exceeds evaporation and the effect of mechanical mixing of the wind is important in determining the ocean's structure. On the other hand, Areas 3, 4, and 5 represent regions where evaporation exceeds precipitation and convective mixing is more important. The spectra within these last three regions were separated solely on a latitudinal basis, so that differences in the characteristic shape of the spectra might possibly be distinguishable as a function of latitude. A test for equivalence of spectral estimates (Bendat and Piersol, 1971) was performed on the spectra within each of the five areas. The results for Areas 1 and 3 showed that the majority of spectral estimates within these regions could be accepted at the 85% confidence level as being equivalent, which gave some credibility to the assumption that the spectra averaged together with these regions were in fact representing

Fig. 3.20 Regional partitioning of the North Pacific ocean for purposes of ensemble averaging SST spectra.

Fig. 3.21 Oceanographic climate regions. [After Tulley, 1964.]

similar physical processes. The results of the equivalence test for Areas 2, 4, and 5 showed that equivalence of spectral estimates could be assumed with a high degree of confidence at only a few frequency bands within these spectra. This result is probably due to the large longitudinal separation between many of the points in these latter three regions, in conjunction with the fact that some longitudinally dependent processes are acting upon these SST's. Nevertheless, in this initial approach to such an analysis, it was felt that a more elaborate definition of averaging areas than is presented in Fig. 3.20 would not be warranted.

The results of ensemble-averaging the forty power spectra within their respective regions are shown in Fig. 3.22. The smoothed curves in these diagrams represent the least squares fits to the logarithm of the SST anomaly spectra by the logarithm of the two-parameter aliased Markov spectra given by Eq. 3.4. The value of the α (lag one autocorrelation in Eq. 3.4), which is converged upon in the least squares fitting process, is a measure of the "redness" of the spectrum, and should not be confused with the interpretation of the lag one autocorrelation of Δ SST which was discussed earlier. A relatively large positive value for α obtained from Eq. 3.4 signifies greater persistence in the SST time series, which in turn indicates that the spectrum is "redder". By visually inspecting the ensemble-averaged spectra in Fig. 3.22 in conjunction with their α values, one can note that the spectra generally become increasingly "red" as function of latitude. These results are in agreement with those implied by the pattern of lag one autocorrelations of Δ SST in Fig. 3.15. The fact that region 2 is characterized by a slightly smaller α value (or "whiter" ensemble-averaged spectrum) than

Fig. 3.22 Ensemble-averaged power spectra with least squares fit (shown as the smooth curve) by the logarithm of a two parameter aliased Markov spectra. The α values in the upper right hand corner of these diagrams represents the lag one autocorrelation from Eq. 3.4 which is converged upon in the least squares fitting process.

region 3, which lies to its south and west, may be due in part to the relatively "white" SST spectra from the northeastern Pacific which were averaged into this former region. From all the evidence examined, it appears that a quasi-meridional differentiation (QMD) in the shape of SST spectra exists across the North Pacific, and in general, this differentiation is such that these spectra become "redder" as an increasing function of latitude.

Although a number of explanations can be offered for the existence of the QMD phenomenon discussed above, there are two hypotheses which are more readily apparent: 1) the QMD phenomenon is a function of the atmospheric forcing field, and 2) it results from the nature of the ocean's response to the atmosphere's forcing which in turn is a function of the ocean's structure. Upon examination of the effects of these two factors, one finds it difficult, if not impossible, to distinguish the individual contributions that they have in determining the shape of the SST spectra across the North Pacific. For example, the similarities in the spatial patterns of the lag one autocorrelations of Δ SST in Fig. 3.15 and the long-term mean in frequency of cyclone centers over the North Pacific in Fig. 3.11 may be indicative of the importance of atmospheric forcing upon the characteristics of SST spectra. This latter figure also shows that the atmospheric forcing exhibits a strong latitudinal dependence as manifested by the meridional gradient of baroclinic activity, and moreover this figure implies that the maximum intensity of atmospheric forcing over the North Pacific occurs between 50°N and 60°N . Before any contributions influencing the shape of SST spectra are directly attributed to the atmospheric forcing field, one should consider the effects of this forcing with respect to the second

hypothesis mentioned above. Past studies (Tulley, 1964; Bathen, 1972) have shown that the atmospheric forcing within its annual cycle has a strong effect upon the vertical structure of the ocean, which in turn effects the observed surface properties. In Tulley's paper, it is shown that the depth of the mixed layer represents a balance between the surface heat flux and the downward mixing process. With a simple oceanic model Tulley shows that in regions of strongest forcing the combined effects of convective and mechanical mixing within the ocean result in the deepest mixed layers. These effects can be observed by comparing the map of the annual average mixed layer depth (Fig. 3.23) to that of cyclone frequency in Fig. 3.11. Since vertical mixing within the ocean was described earlier in this chapter as a process which suppresses the net surface effects of heat fluxes as an increasing function of mixed layer depth, one could surmise from Fig. 3.11 that this suppression generally becomes stronger at higher latitudes. In other words, the low-pass filtering effects of the ocean on the atmospheric forcing signal increase as the mixed layer deepens which, on the average, is towards higher latitudes over the North Pacific. From this discussion it appears likely that the increasing "redness" in SST spectra as a function of latitude (i.e. the QMD phenomenon), may be accounted for, in part, by the synergistic effects of the atmospheric forcing field and the ocean's response to this forcing as a function of the ocean's vertical density structure.

From within the years 1970-77, autocorrelations for Δ SST were calculated for several shorter periods to examine fluctuations of this feature on a short-term basis. The results of these calculations revealed characteristics similar to those of the period 1970-77 (i.e.

Fig. 3.23 Annual average depth (meters) of the mixed layer. [After Bathen, 1972.]

autocorrelations at lag one were significantly negative) except in the years 1975-77. It was during this period that the forty-point average lag one autocorrelation, as shown in Fig. 3.24, was considerably smaller in magnitude in comparison to the longer period shown in Fig. 3.14. The spatial distribution of the lag one autocorrelations of Δ SST for the period (1975-77) is shown in Fig. 3.25. As in the other periods that were examined, Fig. 3.25 likewise reveals a meridional gradient in the values, except now positive autocorrelations appear at latitudes north of 35°N . To investigate further the characteristics of North Pacific SST's for the 1975-77 period, power spectra of SST anomalies were calculated for this period at the forty points previously referred to. These spectra were separated into two groups based on the sign of their lag one autocorrelation value in Fig. 3.25. The results of ensemble averaging the spectra within these two groups are shown in Figs. 3.26 and 3.27. As discussed earlier in this chapter, the α values included in these two figures are a reference to the "redness" of the spectrum. These results suggest that the spectra associated with positive lag one autocorrelations in Δ SST during the period 1975-77 were strongly "red" in their noise characteristics. The fact that these positive autocorrelations appear only sporadically in time and space in the years that were examined, suggests that an interannual variability in the regions of maximum atmospheric forcing and consequently in the oscillatory long-wave pattern (i.e. wave number 3-4) may be an important factor in determining the time scales of variability that an oceanic region exhibits.

Another interesting feature is noted by viewing the same ensemble-averaged spectra of SST anomalies as in Fig. 3.22, except now with their

Fig. 3.24 Same as Fig. 3.14, except for the period 1975-77.

Fig. 3.25 Same as Fig. 3.15, except for the period 1970-77.

Fig. 3.26 Ensemble-averaged spectrum for points containing a negative lag one autocorrelation in Fig. 3.25.

Fig. 3.27 Ensemble-averaged spectrum for points containing a positive lag one autocorrelation in Fig. 3.25.

"red noise" (null hypothesis) spectra. Figure 3.28 shows these spectra for the five areas specified earlier in Fig. 3.20. In the upper right-hand corner of these plots the 95% confidence interval is given along with the maximum number of degrees of freedom from which this interval was calculated. The "red noise" spectra, which are shown as the smooth curves in these plots, were obtained by using the area-averaged values of the lag one autocorrelation and the variance in Eq. 3.4. Here again the Z transformation was used to average these parameters. Upon inspection of the spectra in Fig. 3.28, one can observe in all the plots, except for Area 1, an excess of power above the null hypothesis for several spectral estimates in the frequency band .23-.29 (21-26 days per cycle). Table 3.1 summarizes for each of these areas the specific frequency band over which the excess of power mentioned above is most prominent. Included in Table 3.1 are the number of spectral estimates which were observed to be significant with an excess of power within the corresponding frequency band at both the 90 and 95 percent confidence levels, as well as the number of spectral estimates, given in parentheses, that one would expect to be significant in the given frequency band by random chance. The large differences between the number of spectral estimates which are significant and the number expected to be significant by random chance within these given frequency intervals should be regarded with some caution; for, in fact, these intervals were chosen because of their high significance. One can note from Table 3.1 that the significant cycles are centered around 23.4 days at high latitudes (40°N - 50°N) and around 22 days at lower latitudes (20°N - 30°N). The results presented here take on an additional significance by noting that similar periodicities (21-26 days) were

Fig. 3.28 Ensemble-averaged spectra with "red" noise null hypotheses (shown as the smooth curves). 95% confidence intervals along with the number of degrees of freedom used to calculate these intervals are shown in the upper right hand corner of these diagrams.

TABLE 3.1

Spectral analysis of SST anomalies.

Area	Frequency	<u>Period</u>	Center of Period (days)	Total Number of Spectral Estimates Within a Given Frequency Band	Number of Spectral Estimates Ob- served to be Significant at the Given Level Within the Given Fre- quency Band (number in parenthesis is that expected by random, chance).	
		Days			90% Level	95% Level
2	.231-.283	20.8-26.0	23.4	57	41 (6)	39 (3)
3	.232-.287	20.9-25.8	23.4	58	33 (6)	25 (3)
4	.251-.280	21.4-23.9	22.2	20	19 (2)	15 (1)
5	.266-.291	20.8-22.5	21.7	19	17 (2)	15 (1)

found by Seigel (1977) in cycles of oceanic heat fluxes and by McGuirk and Reiter (1976) in cycles of zonal and eddy available potential and kinetic energies of the atmosphere.

The results of McGuirk's paper showed the existence of a steady, persistent oscillation between zonal and eddy modes of hemispheric scale parameters (i.e. A_Z , A_E , K_Z , and K_E , where A_Z represents zonal available potential energy. K_E represents eddy kinetic energy, etc.). This vacillation, which appears to be fairly steady around 24 days, is significant at the 99.9% confidence level compared to the "red noise" spectrum. McGuirk and Reiter ascribe this vacillation in the atmospheric energy parameters to be a consequence of baroclinic energy transformations. This vacillation is also observed in the amplitudes, and sometimes even in the phase angles (i.e. positions) of planetary waves. This 24-day vacillation comprises the majority of the atmospheric energy variance at time-scales less than 100 days. Seigel takes these results a step further by trying to relate the 22-26 day cycles of the latent and sensible heat transfers, which her studies showed to be significant over the North Pacific and Atlantic, to the 24-day cycle in the atmospheric energy parameters.

In Seigel's investigations of latent and sensible heat transfers, average fluxes for the entire ocean north of 20°N latitude were spectrally analyzed. The results of Seigel's analysis, shown in Table 3.2, outline the important spectral peaks for the heat fluxes over the North Pacific. Since the eight-year records (1970-77) of daily latent and sensible heat fluxes over the North Pacific were available for our study, these two parameters were spectrally analyzed to allow for a comparison with Seigel's results. Since atmospheric forcing and

TABLE 3.2

Spectral analysis of heat fluxes (from Seigel, 1977).

Oct.-May	<u>Sensible Heat Flux</u>		<u>Latent Heat Flux</u>	
	Period (days)	Significance (% level)	Period (days)	Significance (% level)
1969-70	23	95	25	99
	7	99		
1970-71				
1971-72	21	90	21	90
	7	99	5.6	99
1972-73	7	99	8.7	90
1973-74			23	95
1969-73	21	95		
	7	99		

thermocline depths vacillate strongly on a seasonal basis over the North Pacific, it was felt that the results would be more informative by analyzing the latent and sensible heat fluxes on a seasonal basis as well. These analyses were carried out in a similar manner and over the same five areas as was done for the spectral analysis of SST anomalies. By utilizing daily values for the analysis of these parameters, the Nyquist frequency for these spectra is one cycle per two days with a bandwidth resolution of 0.0055 cycles per day. The results of this analysis, presented in Table 3.3, list the significant spectral peaks which are nearest to the 24-day cycle which Seigel found to be significant in her analyses of these parameters. Our results appear to be consistent with those found by Seigel, when one considers that the two analyses were done over slightly different time periods and utilized different techniques in obtaining their final results. From the results in Table 3.3, one would find it difficult to characterize the variability in heat fluxes over the North Pacific with a dominant scale of oscillation during the cooling season, whereas the heating season does exhibit a more consistent vacillation (around 3 weeks) in the heat fluxes over the North Pacific. Although Seigel's results indicate that events on the order of one week are more responsible for the variability in the heat fluxes than processes with oscillations around 21 to 25 days, she shows that these longer cycles are still of some significance with respect to air-sea interaction. Utilizing cross-spectrum analysis and a test of association applicable for time series containing serial correlation, Seigel examined with some detail the interrelationships between heat fluxes and energy parameters. In short, her results indicate that zonal available potential energy forces the heat fluxes which in

TABLE 3.3

Spectral analysis of heat fluxes.

Area	<u>Latent Heat Flux</u>		<u>Sensible Heat Flux</u>	
	Period (days)	Significance (% level)	Period (days)	Significance (% level)
<u>Cooling Season</u>				
2	20.3	95	20.3	90
3	18.3	95	36.6	90
4	30.5	90	30.5	95
5	20-23	95	36.6	95
<u>Heating Season</u>				
2	30.3	95	22-26	90
3	18.2-22.7	95	22.7	95
4	22.7	95	18.2-22.7	90
5	20.2-22.7	95	22.7	95
<u>Entire Cycle Without Seasonal Bias</u>				
2	22.0-23.0	95	22.5-24.3	95
3	20.0-21.2	95	21.4-24.5	90
4	27.5-30.4	95	21.3-22.6	95
5	20.0-22.5	95	20.8-21.4	90

turn influence the eddy modes of energy through vertical eddy heat transport. Taking these results a step further, it is plausible that the variability which is observed in SST anomalies at time scales around 21-26 days per cycle is a reflection of heat fluxes which vacillate at similar frequencies in conjunction with the atmospheric energy cycle described above. In light of these results, it appears likely that a significant time window for the occurrence of air-sea interaction is on the order of 21-26 days. Further implications of these results will have to await additional research with a likely emphasis being on cross-spectrum analysis of the parameters involved in the interaction process.

Another feature of the spectra in Fig. 3.28 which bears discussing is the implications that arise by determining in which region the hypothetical spectrum provides the best fit to the observed spectrum. To achieve this result the deviation, ε , between the logarithm of the observed spectrum and the logarithm of the hypothetical spectrum, $G(f)$, has to be minimized by a least squares fit. The logarithm of these spectra was used in calculating ε so that the statistical errors for each frequency would receive equal weight. If ε was less than some critical value, $\bar{\varepsilon}$, the model represented by the hypothetical spectrum was considered statistically significant at the 95% confidence level. This value $\bar{\varepsilon}$ was defined by Reynolds (1977) as:

$$\bar{\varepsilon} = 1.65\sigma + [\varepsilon]$$

where

$$[\varepsilon] = \frac{2}{V}(p - q)$$

$$\sigma^2 = [(\varepsilon - [\varepsilon])^2] = \frac{8}{V}(p - q)$$

$p \equiv$ number of spectral points

$q \equiv$ number of fitted parameters in the spectral model

$v \equiv$ number of spectral degrees of freedom.

For this significance test, the two-parameter model suggested by Frankignoul and Hasselmann in Eq. 3.4 was taken as the null hypothesis, this hypothesis being that the model perfectly fits the true spectrum. As mentioned earlier, the two parameters used in this study to generate the model's results are the variance and the lag one autocorrelation of the SST anomaly time series. Due to statistical finite-sampling errors the model spectrum will never fit the measured spectrum exactly, so that ε is never zero. Since $p - q$ is large, the probability distribution of ε can be approximated by a Gaussian model with a mean $[\varepsilon]$ and a variance σ^2 as indicated above. The results of this test for the five spectra in Fig. 3.28 are shown in Table 3.4.

One can note from this table that the model is never significant (i.e. ε never less than $\bar{\varepsilon}$), when DOF (the maximum possible number of degrees of freedom) is used to calculate $\bar{\varepsilon}$. Since in reality DOF is probably a gross overestimate of the actual v (as was discussed earlier in this chapter), a critical v was calculated which represents the number of degrees of freedom that are exactly needed in each region for the model to be accepted at the 95% confidence level. From these results and by visual inspection of the five spectra in Fig. 3.24, it appears that the two-parameter model provides the best fit for the observed spectrum in Area 3, whereas among Areas 2, 4 and 5, which have the same DOF, the model appears to fit the spectrum best in Area 4. These results resemble those found by Reynolds in which he used this same model over the North Pacific analyzing 28.5 years of monthly averaged SST anomalies from a NORPAX data set. The last row in Table

TABLE 3.4

Results of Spectrum Fitting.

	<u>Area</u>				
	1	2	3	4	5
Number of spectra assembled together within area	3	11	4	11	11
DOF (maximum possible degrees of freedom).	60	220	80	220	220
ν critical (degrees of freedom needed for spectrum to be confident at 95% level)	26	25	40	30	25
ϵ (observed square error)	31.7	32.2	20.6	27.2	32.2
$\bar{\epsilon}$ (square error calculated with $\nu = \text{DOF}$)	13.3	3.7	10.3	3.7	3.7
Fraction of points within each area that Reynolds found the two-parameter model to be significant at the 95% level	0/3	3/11	3/4	7/11	3/11

3.4 contains the fraction of points within each of these areas that Reynolds found the two-parameter model to be valid for. Reynolds concludes from his results that areas where the two-parameter model is not valid (statistically significant) can generally be related to strong oceanographic processes. These processes, which can significantly influence the SST fields, have been demonstrated by Clark (1972), Gill and Niiler (1973), and Elsberry and Garwood (1978) to be large-scale advection, local advection and upwelling due to eddies, mixed layer deepening by buoyant convection or wind actions and radiation flux. It is in the "non-significant" areas that oceanographic processes are apparently at least as important as the local atmospheric forcing in causing SST variability. In the "significant" areas, it appears that most of the features (variability) of mid-latitude (30°N - 40°N) SST anomalies can be explained by a simple model of "white" atmospheric forcing balanced by a stabilizing linear feedback. It is interesting to note that the results of the spectrum analyses of SST anomalies for our investigation and Reynold's study are for the most part analogous, even though they utilized different data sets and averaged the anomalies on different time scales. The implications of these results will be examined further in the next chapter when the spatial patterns of grid point correlations between ΔSST and heat fluxes are examined.

IV. Heat Budget Investigations

A. Applicability of Heat Budgets

Oceanic heat budgets have been used extensively by researchers investigating the function and importance of various air-sea interactions, the understanding of which is considered to be a crucial step in generating meaningful long-range forecasts of atmospheric and oceanic phenomena. One of the main intents of this present study is to examine a portion of the oceanic heat budget which past researchers have determined to be important. Whereas past studies have concentrated on selected geographical regions and time scales due to the complexity of possible interactions and the large quantities of data involved, our study improves upon the spatial and temporal resolution of the heat budget as well as its spatial coverage, while limiting the aspects of the heat budget (namely SST change, latent and sensible heat fluxes, and the variability of these parameters with respect to a transient thermocline) which are examined. The radiational effects of the heat budget are not examined here, due to serious limitations in their data bases over the North Pacific. Hopefully, the gradual improvement of satellite imagery over this oceanic region will allow future studies to examine with some detail the importance of radiation on the oceanic heat budget. Many previous oceanic heat budget studies (Clark, 1972; Husby and Seckel, 1975, etc.) have exclusively used monthly and seasonally averaged properties, and thereby have failed to capture the phenomena on time scales shorter than a month that effect the thermal characteristics of the ocean's surface layer. Such studies have demonstrated the important roles of radiation and advection on long-time and large-space scales. For example, Reiter (1978) investigating the possibility that

advective processes modulate long-term SST behavior in the North Pacific, has shown that changes in the advective ocean current transport could conceivably be derived from long-term anomalies in the trade-wind flow and in the intensity of the Hadley circulation. The choice of a 6-day time scale in our study allows for the large scale advective process of meandering ocean currents to be minimized. Furthermore this preference of time scales lends itself to a more detailed investigation of the temporal variability of local air-sea interaction and vertical entrainment of water from below the mixed layer. Basically this chapter inquires into two aspects of the heat budget: 1) when and where the observed SST changes over the North Pacific can best be related to the latent and sensible heat fluxes, and 2) how these processes are affected by a variable mixed layer.

The heat budget in this present study is analyzed over the expanse of the North Pacific outlined in Fig. 2.1. To facilitate a regional analysis of the heat budget, this larger grid was divided with criteria similar to those which were employed to group the power spectra in the previous chapter. In this present case though, the regions were further divided so that they would contain a minimum of 30 points. Partitioning the grid in this manner produced a finer regional resolution in the analyses, yet ensured the significance in the results of the statistical tests which were used to analyze the regional heat budgets. These regions, with the number of points they contain, are shown in Fig. 4.1.

From the prognostic mixed-layer temperature Eq. 2.1, one can note the five processes by which gains or losses of energy from an oceanic volume could be reflected in SST changes. By integrating Eq. 2.1 over a six-day period and utilizing daily values of latent and sensible heat

Fig. 4.1 Outline of areas used in the heat budget analysis. The number in parentheses indicates the number of points (represented with dots) within each area for which the heat budget was analyzed.

fluxes and SST's, the impact of these heat fluxes on the mixed layer temperature is examined on a regional and temporal basis. The strength of the impact is estimated by correlating the six-day change which is observed in the SST with the summation of the heat fluxes over that same period.

One can note from Eq. 2.1 that an increasing percentage of Δ SST is determined by the heat fluxes if they either increase in magnitude, or the depth to which their effects are mixed decreases. From a statistical viewpoint either of these two circumstances would signify that the random chance of an insignificant correlation between the heat fluxes and Δ SST would decrease various aspects of the heat budget. In fact, these two factors mentioned above will become apparent in the discussion that follows on the various aspects of the heat budget.

To ensure proper interpretation of the heat budget analyses, significance levels were calculated for the correlation coefficients with a method suggested by Marion Haurwitz (1979). This method, which requires no recognition of the number of degrees of freedom, was extremely useful in this particular analysis, since determination of the degrees of freedom is encumbered by the spatial correlations that exist in SST and heat flux fields. A description of this method by which significant levels were determined for each area in Fig. 4.1 on a seasonal basis is given in the following discussion. The heat fluxes (i.e. their six-day summation) and Δ SST's for all the six-day periods within the years 1970-77 were separated by area and season. The values for each area were then placed into two sets: one containing heat fluxes, the other Δ SST's. Each element within a set represents an array containing values for each grid point within that area. After randomizing the elements in

each set, they were paired up (one element from each set) and then correlated. To maintain a temporal independence between the elements being correlated, any random pairs which were from the same or adjacent six-day periods were not correlated due to serial correlations that exist within the time series of SST's and heat fluxes. This process of picking random pairs of elements from the two sets and correlating the corresponding arrays was continued until the relative distribution of correlation coefficients became "stable". Stable in this sense refers to an insignificant change (i.e. less than 0.1%) in the percentage of correlation coefficients that fall within each interval of the distribution when additional correlations are added to that distribution. The results, shown in Fig. 4.2 for each area on a seasonal basis, are presented in the form of a cumulative distribution. From these graphs one can determine a significance level for a correlation coefficient by observing the probability that by random chance a correlation between heat fluxes and Δ SST will yield a value that is less than or equal to that coefficient's value.

Closer inspection of the graphs in Fig. 4.2 show some obvious regional and seasonal differences in the significant levels of the heat budget correlations. To assist one in accounting for these differences Table 4.1 presents, on a regional and seasonal basis, the probability that by random chance a correlation coefficient will be less than or equal to -0.5. By examining the values from Table 4.1 with consideration to the average seasonal magnitudes of the heat fluxes (Fig. 4.3), one can see that the probability of a higher correlation occurring by random chance in a particular region is greater in the season characterized by larger heat fluxes (e.g. in the northern regions of the North

Fig. 4.2 Regional significance levels for the heat budget correlation (LH + SH versus Δ SST) presented in cumulative distributions, so that the probability of a given correlation coefficient occurring by random chance can be identified. Solid line represents cooling season; dashed line represents heating season.

Table 4.1

Probability (in percent) that by random chance a correlation coefficient will be less than or equal to -0.5.

		Area							
		1	2	3	4	5	6	7	8
LH vs Δ SST	Heating season	5.0	7.0	4.1	9.0	10.3	19.6	16.3	9.5
	Cooling season	8.8	6.7	6.5	8.7	6.7	7.2	7.3	5.8
SH vs Δ SST	Heating season	5.4	3.2	4.8	8.4	7.5	11.2	6.8	5.6
	Cooling season	8.1	6.2	5.6	9.5	9.9	10.2	10.5	5.4
LH+SH vs Δ SST	Heating season	5.3	6.1	4.5	9.4	10.1	18.5	14.9	5.9
	Cooling season	9.1	7.0	6.2	11.4	7.2	7.5	7.7	5.9

Fig. 4.3 Magnitude of the latent and sensible heat fluxes (watts/m^2) on a seasonal basis calculated for the period 1970-77, where $1.0 \text{ watt} \cdot \text{m}^{-2} = 2.06 \text{ calories} \cdot \text{cm}^{-2} \cdot \text{day}^{-1}$.

Fig. 4.3 (Continued)

Pacific grid, that is areas 1-4, where the heat fluxes during the cooling season are much larger in magnitude than those during the heating cycle). In the other areas (5-8), where the seasonal variability in the magnitude of the heat fluxes is much less, one can see from Table 4.1 that the probability of a high correlation occurring by chance is generally greater during the heating season when the mixed layer is relatively shallow (see Fig. 3.4, which shows the depth of mixing at the end of both seasons). In view of the above discussion, it would appear that the magnitude of heat fluxes and the depth of mixing are two of the more obvious factors which may be influencing the variability observed in the significance levels of the heat budget correlations on a regional and seasonal basis.

B. Examination of Heat Budget on a Temporal Basis

The impact of the latent and sensible heat fluxes upon Δ SST is examined on a temporal basis over the eight areas shown in Fig. 4.1 by computing correlation coefficients between these heat budget parameters at six-day intervals over the eight years (1970-77) in which data were available. An example of these computations is shown in time series format for area 1 and for the years 1970-72 (Fig. 4.4). Included in this figure are the 5 and 10 percent significance levels obtained from Fig. 4.2. The rapid fluctuations of the values and the "noisy" appearance of the time series in Fig. 4.4 are typical characteristics of the correlation coefficients for the other areas and years over which these computations were done (results not shown here). Aside from the fact that these features of the raw time series may be indicative of the complexity of the air-sea interactions that are affecting SST's, a

Fig. 4.4 Time series of correlations for LH + SH versus Δ SST on a six-day basis for area 1 (shown in Fig. 4.1). Included on this plot are the 5 and 10% probabilities that these correlations will occur by random chance.

presentation as such offers little further insight into the characteristics of the heat budget.

A more informative investigation of the heat budget on a temporal basis is achieved by examining the characteristics of the variability from an eight-year (1970-77) mean annual cycle of correlation coefficients. This mean annual cycle, being partitioned into sixty-one six-day periods, is represented in each of these periods by an average correlation. More specifically, an average correlation for one of these six-day periods is the arithmetic mean of the correlation coefficients associated with that particular six-day period for each of the eight years (1970-77) over which the heat budget calculations were done. In averaging correlations as is done in the above case, Panofsky and Brier (1958) have suggested that the population of coefficients should first be normalized. Therefore, throughout this study correlation coefficients are averaged by utilizing the transformation in Eq. 4.1 to normalize their populations.

$$Z = \frac{1}{2}[\ln(1 + r) - \ln(1-r)] \quad (4.1)$$

where $r \equiv$ correlation coefficient

$Z \equiv$ normalized value

After the population of coefficients has been normalized, their associated "Z - values" are averaged. This mean "Z-value" is then reconverted to an average correlation by solving Eq. 4.1 for r . The mean annual cycles of correlation coefficients for latent plus sensible heat fluxes versus Δ SST are shown in Fig. 4.5 for the eight areas previously referred to. The significance of these correlations, represented as the probability of their occurrence by random chance, is shown in Fig. 4.6.

Fig. 4.5 The eight-year (1970-77) average correlations for LH + SH versus ΔSST on a six-day basis for the eight areas defined in Fig. 4.1.

Fig. 4.6 The significance levels on a six-day basis corresponding to the correlations in Fig. 4.5.

SIGNIFICANCE LEVELS ON
A 6-DAY BASIS FOR AREA 1 SIGNIFICANCE LEVELS ON
A 6-DAY BASIS FOR AREA 2 SIGNIFICANCE LEVELS ON
A 6-DAY BASIS FOR AREA 3 SIGNIFICANCE LEVELS ON
A 6-DAY BASIS FOR AREA 4 SIGNIFICANCE LEVELS ON
A 6-DAY BASIS FOR AREA 5 SIGNIFICANCE LEVELS ON
A 6-DAY BASIS FOR AREA 6 SIGNIFICANCE LEVELS ON
A 6-DAY BASIS FOR AREA 7 SIGNIFICANCE LEVELS ON
A 6-DAY BASIS FOR AREA 8

The more significant correlations in these figures are characterized with a low probability of occurrence by random chance.

Examination of Figs. 4.5 and 4.6 reveals several interesting characteristics of the heat budget from within its annual cycle. In all the areas that were considered, the significance of the average correlations is nearly always greatest from midsummer to early fall, which coincides with the period when the mixed layer is normally the shallowest. One might speculate that during this period of maximum significance in the correlations, the effects of the heat fluxes in determining ΔSST 's are accentuated by the limited vertical mixing which the ocean's density structure permits. In the more northern areas (1-4) the period of maximum significance tends to occur later in the year, which may be indicative of the increasing magnitude of the heat fluxes over these regions during the early autumn season. Another interesting feature that appears in Fig. 4.6 is the occurrence in several areas of relative minima in the significance of the average correlations, bounded by peaks of strong significance. The fact that this feature exhibits temporal continuity between adjacent areas lends more credibility to the notion that some actual physical phenomenon is causing it. In areas 3 and 4 these relative minima of significance are observed in late June and early July, whereas more prominent minima exist around early September in areas 6 and 7. One could postulate that during these periods of minimum significance, when the magnitudes of the heat fluxes are likewise near their annual minima, other processes are exerting a dominant influence upon ΔSST 's. In fact, a study by Oort and Vonder Haar (1976) on the northern hemisphere's ocean-atmosphere heat balance based on satellite radiation data suggests that these minima of significance in

areas 3 and 4 are caused by radiation, since their results show that heat storage in the oceans at the latitudes occupied by these two areas (30-45°N) maximizes (at greater than $100 \text{ watts} \cdot \text{m}^{-2}$) in June and early July. Another result of Oort and Vonder Haar's study, which is of interest to note in light of this present analysis, is the northward transport of heat due to oceanic motions (Fig. 4.7). Since the bulk of the meridional heat transport at midlatitudes is accomplished by the western boundary currents, which traverse the region contained in area 3, it seems likely that the two periods characterized by minimum significance in April and November in area 3 are associated with the strong northward advection of heat indicated by Fig. 4.7 during these periods. Finally, from Figs. 4.5 and 4.6 one can observe that the period of least significance in these correlations occurs, as a general rule, from late winter to early spring, which is what one might expect, since during this period the mixed layer is characteristically deepest so that the impact of heat fluxes upon ΔSST is minimized by strong vertical dilution.

The sixty-one correlation coefficients appearing for each area in Fig. 4.5 are averaged together so that one eight-year average correlation can be assigned to each area. These correlations along with their corresponding significance levels are shown in Table 4.2. The results in this table reveal that in most regions the latent heat flux provides a higher correlation with ΔSST than sensible heat. The only notable exception to this behavior is observed in area 3, which is coincident with the region that exhibits the largest magnitudes in the sensible heat fluxes (see Fig. 4.3). These observations of the average regional correlations in Table 4.2 imply, as was suggested earlier in this

Fig. 4.7 Northward transport of heat due to oceanic motions computed as a residual term in the earth's heat balance. [After Oort and Vonder Haar, 1976.]

Table 4.2

Eight-year average (1970-77) correlations with corresponding significance (probability of occurrence by random chance) for regional heat budgets.

Average Correlation	1	2	3	Area 4	5	6	7	8
LH vs Δ SST	-.151	-.228	-.161	-.304	-.267	-.282	-.242	-.152
SH vs Δ SST	-.123	-.144	-.194	-.275	-.260	-.284	-.227	-.141
LH+SH vs Δ SST	-.149	-.216	-.181	-.309	-.282	-.299	-.260	-.165

Significance (in percent)
for above
correlations

	1	2	3	4	5	6	7	8
LH vs Δ SST	38.9	33.0	37.4	30.8	34.1	35.0	37.0	38.3
SH vs Δ SST	41.1	38.1	36.4	31.8	35.6	34.0	35.5	40.1
LH+SH vs Δ SST	39.2	33.5	36.4	30.5	34.0	34.2	35.7	36.9

chapter, that the impact which the latent and sensible heat fluxes have upon Δ SST is related to the magnitude attained by these heat fluxes. Finally, to obtain an areal perspective for the significance of the heat budget correlations (latent heat plus sensible heat versus Δ SST) over the North Pacific, the regional significance values in Table 4.2 are shown within the outline of their corresponding areas in Fig. 4.8. A more detailed investigation of the heat budget on a regional basis is given in the following section when the spatial patterns of grid point correlations are examined.

C. Examination of Heat Budget on a Spatial Basis

In this section the heat budget is analyzed on a grid point basis to facilitate a finer spatial resolution in examining its characteristics. A pattern of correlations is obtained for each season by correlating the Δ SST's at each grid point shown in Fig. 4.1 to the corresponding heat fluxes for all the six-day periods within an entire season. In this manner a total of eight heating and seven cooling seasons were analyzed from the years 1970-77. These seasonal patterns of correlations were then averaged so that the basic seasonal spatial patterns inherent within all these years could be examined. The resultant spatial patterns, shown in Figs. 4.9 and 4.10 for the heating and cooling seasons, respectively, were analyzed only for the more significant correlations (i.e. between -0.4 and -1.0). In a similar manner, the spatial pattern was computed with no seasonal bias for the entire period 1970-77 (Fig. 4.11). To obtain an estimate of the significance of a grid point correlation from one of these figures (i.e. Figs. 4.8-4.10), one should first locate that grid point within one of

Fig. 4.8 The eight-year (1970-77) average significance (probability of occurrence by random chance) corresponding to the heat budget correlations for latent plus sensible heat versus Δ SST plotted as a function of area.

Fig. 4.9 The eight-year (1970-77) average spatial pattern of grid point correlations for: SH versus Δ SST, LH versus Δ SST, and LH + SH versus Δ SST for the heating season.

Fig. 4.10 Same as Fig. 4.9, except for cooling season.

Fig. 4.11 Same as Fig. 4.9, except for entire cycle (no seasonal bias).

the regions shown in Fig. 4.1 and then refer to the graph of that region's significance levels in Fig. 4.2.

Several features in the spatial patterns of correlations from Figs. 4.9-4.11 further substantiate a number of characteristics of the heat budget and air-sea interaction that were mentioned earlier in this paper. For example, the predominance of higher correlations between Δ SST and the larger of the two heat fluxes, namely latent heat, can now be observed in a spatial distributions of Figs. 4.9 and 4.10. Furthermore, the regions of highest correlations (i.e. between -0.4 and -1.0) in Figs. 4.9-4.11 corroborate the results presented in conjunction with the temporal analysis of the heat budget shown in Fig. 4.8. These latter results indicate that the greatest significance in the heat budget's correlations of latent plus sensible heat versus Δ SST is in area 4, which coincides with the mid-oceanic regions of the North Pacific. These results, along with those discussed in the previous chapter, where ensemble-averaged SST power spectra for five areas over the North Pacific were fitted to "null hypothesis" spectra generated from a two-parameter oceanic model, suggest, as did a similar analyses by Reynolds (1977), that SST behavior over the mid-oceanic regions of the North Pacific is dominated by the influence of latent and sensible heat fluxes. Whereas in the regions that exhibit less significant correlations (i.e. greater than -0.4) in Fig. 4.9-4.11, such as along coast lines and in the vicinities of 20°N and 50°N latitudes, one can surmise from the results discussed above that other oceanic processes are at least as important in determining the behavior of SST's. Yet another interesting feature in Fig. 4.9-4.11 is the seasonal variability observed in the spatial patterns of correlations, which is manifested by

larger negative correlations over a greater areal extent in the heating season. Since the mixed layer depth also undergoes strong seasonal changes over much of the North Pacific, this feature lends support to observations made earlier in this chapter which indicate that the depth of vertical mixing has a notable effect upon SST behavior.

D. Interannual Variability of the Heat Budget

An examination of the spatial patterns of heat budget correlations for individual years reveals a significant component of interannual variability. An example of such variability in the spatial patterns of correlations can be seen between the two heating-cooling cycles (1975-76 and 1976-77) shown in Fig. 4.12. One should recall that a heating-cooling cycle was defined to begin on March 13 and extend to March 12 of the following year. The grid-point differences (see Fig. 4.13) between the two spatial patterns in Fig. 4.12 further emphasize the interannual variability of the heat budget correlations. Undoubtedly, some of the interannual differences in the grid-point correlations as shown in Fig. 4.13 are a function of random chance and cannot be related to any physical processes. On the other hand, it seems reasonable to speculate that the same factors which have been shown to affect the significance of heat budget correlations on a seasonal and regional basis (i.e. the depth of vertical mixing and the magnitude of the heat fluxes) could be of some consequence on an interannual basis as well. This speculation as it pertains to the latter factor mentioned above is investigated by examining the interannual differences in atmospheric forcing upon the ocean as manifested in the magnitudes of the latent and sensible heat fluxes for the same heating-cooling cycles as in Fig. 4.12 (i.e. 1975-76 and 1976-77). The interannual differences between the spatial patterns

Fig. 4.12 Spatial pattern of grid point correlations for LH + SH versus Δ SST for two entire cycles (heating and cooling seasons).

Fig. 4.13 Difference (1975-76 cycle minus 1976-77 cycle) between the two spatial patterns of grid point correlations shown in Fig. 4.12.

of heat fluxes previously referred to is shown in Fig. 4.14. One can observe some interesting similarities between the spatial patterns of interannual differences of correlations in Fig. 4.13 and heat fluxes in Fig. 4.14. By noting that the differences in these latter two figures were obtained by subtracting the spatial patterns of the 1976-77 cycle from those of the 1975-76 cycle, it generally appears that the more significant correlations between the latent plus sensible heat flux versus Δ SST exist in the year and over the region characterized by the larger average heat flux. For example, one can note from Figs. 4.13 and 4.14 that the region characterized by large interannual differences in the correlations near 140°W - 40°N (i.e. greater than -0.4 which is indicative of more significant correlations during the 1975-76 heating-cooling cycle) is also associated with considerably larger heat fluxes during the 1975-76 cycle over that region. Examples similar to the one cited above could be made for other regions from Figs. 4.13 and 4.14 and from other years during the period 1970-77, which would demonstrate that the interannual differences in heat budget correlations can be explained, to some extent, by the interannual variability in atmospheric forcing as manifested in the fluxes of latent and sensible heat. On the other hand, it was apparent that between many of the years during the period 1970-77 processes other than the one described above, which act upon or with the ocean, must be considered to explain the interannual component of variability which the heat budget correlations exhibit.

E. Effects of a Variable Mixing Depth on an Oceanic Heat Budget

Earlier in this chapter it was shown that the variability in the depth of vertical mixing within the ocean influences the heat budget

Fig. 4.14 Same as Fig. 4.13, except for the spatial patterns of heat fluxes LH + SH (watt · m⁻²).

correlations, especially on a seasonal basis. The characteristics of the seasonal change in the depth of mixed layer are such that heat fluxes in the cooling season, when the mixed layer is deepest, have only a fraction of the impact upon Δ SST's in comparison to heat fluxes in the warming season. On the other hand, recent studies (Clark, 1972; White and Walker, 1974) have indicated that the interannual fluctuations in the mixed layer depth can have a large effect in determining anomalous sea-surface temperature changes and anomalous heat storage within the mixed layer. In view of these results, the author has undertaken an investigation of the heat budget with respect to the variability in the mixed layer on a seasonal as well as interannual basis. Unfortunately, this aspect of the study alluded to here was still in progress at the time this report was written so that the following results will be only qualitative in nature.

Due to the sparsity of XBT data over much of the North Pacific, selected regions shown in Fig. 4.15 that lie across major shipping lanes were chosen for examining the heat budget as a function of mixed layer variability. Within these regions, mentioned above, the vertical thermal structure of the ocean as well as the heat budgets were analyzed on a six-day basis. The vertical temperature profiles for the four regions under consideration are shown in Figs. 4.16-4.19 for the years 1970-76. The temperature profiles presented in these figures were generated for each region by averaging all the XBT observations which were contained within each area on a six-day basis. With this method of compiling the data a number of periods were void of observations. To fill the data gaps the vertical temperature profile was inferred by interpolating between existing observations as a first guess and by

Fig. 4.15 Regions chosen for examination of XBT data.

Fig. 4.16 Time history of subsurface thermal structure for area A. The isotherms ($^{\circ}\text{C}$), which are labeled for every 2°C , are plotted as a function of depth (meters) and time (on a six-day basis).

Fig. 4.17 Same as Fig. 4.16, except for area B.

Fig. 4.13 Same as Fig. 4.16, except for area C.

Fig. 4.19 Same as Fig. 4.16, except for area D.

utilizing monthly mean temperature profiles from an oceanic atlas (Fleet Numerical Weather Central, 1971) when more than four consecutive six-day periods were missing. This latter means for inferring missing data (represented by dashed lines in Figs. 4.16-4.19) was used primarily in the higher latitude regions where rough seas often restricted shipping activities and thereby limited the acquisition of data. A final check of all interpolated values was made by visually reviewing the temporal plots for continuity. The mixed layer depth was then determined from the six-day averaged vertical temperature profiles in Figs. 4.16-4.19.

For the purposes of the following discussion, the mixed layer is defined as the layer contained between the surface and the depth at which either a vertically isothermal condition or a weak vertical thermal gradient changes to a more pronounced thermal gradient found in the main thermocline underlying the surface layer. The main thermocline separates the surface waters, which are mixed downward by wind action and convection, from the deeper waters which are unaffected by seasonal changes in atmospheric forcing. To determine the depth of the mixed layer, Bathen (1972) describes three methods: a visual inspection, and two statistical tests. Bathen comments in his paper that the two statistical methods are both necessary and useful only at high latitudes where the depth of mixing is not always visually obvious. Since the four areas investigated in the present study are from middle and low latitudes, visual inspection was used exclusively as the means for determining the depth of mixing. Despite the subjectiveness of the "visual inspection" method, in all but a few cases a pronounced thermocline, which clearly marked the vertical extent of mixing, virtually assured an accurate analysis. The mixed layer depths derived from the

temperature profiles in Figs. 4.16-19 are presented in a time series format along with the heat budget correlations on a six-day basis, in Figs. 4.20-23. The correlations presented in these figures were computed solely between the six-day summation of the latent heat flux and Δ SST, since the inclusion into the correlations of the sensible heat flux, which is often negligible (see Fig. 4.3) over the four areas outlined in Fig. 4.15, only makes the analysis appear noisier. By examining the vacillations in the two time series shown in these figures, some tentative conclusions can be drawn as to how a variable mixed layer affects the heat budgets over the four regions shown in Fig. 4.15.

The characteristics of the mixed layer in Fig. 4.18 are typical of the subarctic waters over which Region "C" is located. This figure reveals that the mixed layer in Region "C" generally varies between 20 and 100 meters on a seasonal basis, and in comparison to the other regions outlined in Fig. 4.15 it exhibits relatively little interannual variability in this seasonal range of depths. On the other hand, the onset of the cooling season, as depicted in Fig. 4.22 by a reforming of a shallow thermocline, varies considerably on an interannual basis. From the years in which sufficient data was available to determine the onset of the heating season, the results showed that a shallow thermocline began reforming as early as the middle of May (as in 1972) to as late as the end of June (as in 1973). After the mixed layer attains its minimum thickness around mid-summer, it begins to slowly erode away. The onset of the cooling season, manifested by rapid mixed layer deepening, also exhibits strong variability as indicated in Fig. 4.22. This rapid deepening of the mixed layer, which is a response to an intensification of the erosion processes of the thermocline with the advent of

Fig. 4.20 Time sections on a six-day basis of thermocline depths (meters) and LH versus ΔSST correlations for area 1'. Dashed lines on uppermost plot infer the use of interpolated or climatological data when real data was missing. Dotted lines: halocline depth.

Fig. 4.21 Same as Fig. 4.20, except for area 2'.

Fig. 4.22 Same as Fig. 4.20, except for area 3'.

Fig. 4.23 Same as Fig. 4.20, except for area 4'.

the stormier winter months, varies from early November (as in 1975) to late December (as in 1970). In Region "C" the mixed layer during the cooling season is normally limited by a permanent halocline (i.e. a salinity gradient). A halocline as such is a consequence of precipitation exceeding evaporation over a region. Since no salinity data were available for this present investigation, the halocline depths (denoted with a dotted line in Figs. 4.20-4.23) represent climatological values given by Tulley (1964).

Upon closer examination of Fig. 4.22, it appears that insignificant correlations between latent heat and ΔSST generally occur after the onset of the cooling season when the mixed layer deepens rapidly. Camp and Elsberry (1977) have shown that the heat flux at the base of the mixed layer can be ten times the magnitude of the surface heat flux at Ocean Weather Station Papa ($50^{\circ}N$, $145^{\circ}W$) during periods of rapid mixed layer deepening. In reference to Eq. 2.1, one can note that in a one-dimensional model the heat flux at the base of the mixed layer is accounted for the entrainment heat flux, " W_e ", which is equivalent to the rate of change in the mixed layer depth with time. Obviously a heat budget which ignores the entrainment heat flux will perform poorly when the mixed layer is rapidly deepening. In all likelihood, the insignificant correlations, which characterize the periods of rapid mixed layer deepening in Fig. 4.22, are indicative of a negligible impact of the latent heat flux upon ΔSST in comparison to the impact of the entrainment heat flux at the base of the mixed layer, which was not considered in the heat budget correlations.

The characteristics of the mixed layer in Regions "B" and "D" (Figs. 4.17 and 4.19, respectively) are typical of subtropical waters in

which evaporation is the dominant process. No permanent halocline exists in these regions since fresh water does not accumulate here as in the subarctic waters. A large latent heat flux over these regions removes heat, as well as fresh water, from the ocean's surface, leaving a high concentration of salt in the sea. This evaporation process acts to increase the density of the surface waters and promotes a salinity-driven circulation which generally dominates the mixing process in Regions "B" and "D". One can note from Fig. 4.21 that Region "B" is characterized by a relatively small seasonal transition in mixed layer depths (i.e. on the average around 70 meters) in comparison to the other three regions which were examined. Closer inspection of the heat budget correlations in Fig. 4.21 shows that less than 5% of the variance in ΔSST , on the average, can be explained by the influence of the latent heat flux over Region "B", and furthermore the temporal trends in the time series of these correlations have no detectable relationship to variations in the depth of the mixed layer either on a seasonal or interannual basis. These observations suggest that the characteristics of SST's in Region "B" are influenced more strongly by some processes other than a variable mixed layer and the fluxes of latent and sensible heat. However on an interannual basis, Region "B" exhibits the largest variability in the onset of the seasons as is observed in the reforming and rapid deepening of the mixed layer. On the other hand, the heat budget correlations for Region "D" shown in Fig. 4.23 not only explain more variance in ΔSST , but vacillations in the time series of these correlations also appear to be related to some extent to the fluctuations in the mixed layer depth. The most significant correlations in Fig. 4.23 predominate during periods when the mixed layer is generally

shallow, whereas the insignificant correlations commonly appear during two periods within the annual cycle. One of these periods is usually observed after the onset of the cooling season, and likely is associated with the entrainment heat flux at the base of the mixed layer as in Region "C". The other period characterized by insignificant correlations generally occurs when the depth of mixing is the greatest, and is probably related to a strong vertical dilution of the surface heat fluxes.

Categorizing the behavior of the mixed layer in Region "A" is more difficult, since on a yearly basis its features appear to vary between those observed in subarctic and subtropical waters. During four of the seven years shown in Fig. 4.16, a well-defined thermocline is present throughout the year as is characteristic over the subtropics. On the other hand, during the cooling seasons of 1973 and 1975 no thermocline is evident, which is a distinction of subarctic waters where the thermocline vanishes during the cooling season beneath a permanent halocline. One possible explanation for the absence of a notable thermocline during 1973 and 1975 might be that the atmosphere's circulation during these years was such that precipitation exceeded evaporation resulting in a halocline, prominent enough to limit the vertical extent of mixing. Although our data were insufficient to substantiate such a speculation, Fig. 4.20 nevertheless shows that the mixed layer in 1973 and 1975 deepened during the cooling season only to the extent of a halocline placed climatologically at 100 meters. Examination of Fig. 4.20 reveals a large interannual variability in the depth of the mixed layer in Region "A", which is especially evident between adjacent cooling seasons. The interesting aspect of this variability in the depth of the

mixed layer is its apparent relationship to the interannual variations observed in the heat budget correlations of Fig. 4.20. For example, less significant correlations in the heat budget occur during the cooling seasons that exhibit deeper mixing depths (as in '72 and '74), whereas shallower mixing depths during the cooling seasons (as in '71, '73, '75, and '76) are observed in conjunction with greater significance in the heat budget correlations. Furthermore, one can surmise from Fig. 4.20 that the predominance of high correlations is an indicator of the strong impact that the latent heat flux has upon Δ SST's in Region "A".

In summary, it would appear that by considering the variability in the depth of mixing on a seasonal as well as interannual basis, one could better determine the magnitude of impact that heat fluxes into the mixed layer would have upon Δ SST. In addition, the analyses presented in this last section suggests that the entrainment heat flux at the base of the mixed layer is a significant factor in determining SST characteristics, especially in the regions and during periods of rapid mixed layer deepening. Hopefully our continued research in these problem areas of air-sea interaction will disclose the true significance and applicability of the observations stated above.

V. Heat Exchange Feedbacks Within the Air-Sea System

Until now the investigations and analyses in this study have been carried out by viewing the ocean and its behavior as a passive agent to atmospheric forcing. Yet in the introduction of this paper concessions were made which suggested that the ocean could act as an initiator of climate fluctuations if the atmosphere were sensitive to inputs from the enormous oceanic energy reservoir at its bottom boundary. In view of this new emphasis on atmospheric sensitivity within the coupled air-sea system, this chapter examines the conditions surrounding the exchange of heat from the ocean to the underlying tropospheric air column.

One of the most striking features observed in the course of this paper is the exceptional resemblance which often occurs between anomalous fields of SST and atmospheric thickness patterns. A good example of such a resemblance is shown in Fig. 5.1, where the correlation coefficient between the six-day anomalous patterns of SST and 1000-700 mb thickness is 0.72. The anomalous thickness in Fig. 5.1 were computed by utilizing NMC height data over the 21 by 15 North Pacific grid outlined in Fig. 2.1. It is of interest to note that the average correlation between the six-day anomalous patterns of SST and 1000-700 mb thickness over the North Pacific grid was larger than 0.6 for the 10 six-day periods prior to that in Fig. 5.1 (i.e. February 18-23, 1977). The fact that anomalous fields of SST and atmospheric thickness are often highly correlated over extended periods of time would suggest that their resemblances are more than just random chance. By visually inspecting several concurrent anomaly patterns over a number of years, one can detect that strong SST patterns are more likely to be highly correlated with anomalous thickness patterns of the lower troposphere than weak SST

Fig. 5.1 Spatial patterns of anomalous SST ($^{\circ}\text{C}$) and 1000-700 mb thickness (geopotential meters) fields for the period Feb. 18 to Feb. 23, 1977.

anomaly patterns. In an earlier effort, Namias (1973), who first recognized the relationship between the strength of the SST anomaly pattern and the magnitude of its correlation with anomalous thickness, suggested that an increase in baroclinic activity, which often occurs between strong anomalies of the opposite sign, cooperates with the SST anomalies to destroy inversions and thereby reduce the stability of the lower troposphere over much of the region. In turn the reduction of stability would effectively increase the transport of heat from ocean to atmosphere. Unfortunately, in dealing with a complex, coupled air-sea system, it is impossible to distinguish from figures such as Fig. 5.1 whether the ocean is forcing the atmosphere or vice versa. Most likely, feedbacks are occurring in both directions as past studies have indicated (Reiter, 1978).

Since the magnitudes of SST anomalies appear to be related to the behavior of large-scale, air-sea interaction as suggested in the previous paragraph, an investigation of the areal distribution of absolute magnitudes which SST anomalies attain may lead to a better understanding of the coupled air-sea system. By utilizing six-day running mean filtered SST data for the period 1966-77, the spatial patterns of six-day averaged SST anomalies were computed over the North Pacific grid referred to earlier. The results are shown in Figs. 5.2 and 5.3 for the heating and cooling seasons, respectively. The largest SST anomalies in these figures occur in regions characterized by strong currents and oceanic fronts (see Figs. 3.12 and 3.13), which suggest that advection within the mixed layer is the most dramatic process affecting the magnitudes that SST anomalies can attain over the North Pacific. The advective effects appear strongest off the coast of Japan in the confluent

Fig. 5.2 Areal distribution of the absolute mean of six-day averaged SST anomalies for the heating season.

Fig. 5.3 Same as Fig. 5.2, except for cooling season.

region of the oceanic gyres and in the southeast North Pacific, where the California Current merges with the subtropical oceanic front. The large maxima in the magnitude of SST anomalies in Figs. 5.2 and 5.3, which lie over the region of the subarctic front (40°N), are most likely the result of anomalous wind stresses acting upon strong SST gradients.

Another oceanic feature, which appears to influence the magnitudes that SST anomalies can attain is the depth of the mixed layer. For example, the large SST anomalies during the heating season, which occupy a greater areal extent of the North Pacific than in the cooling season, may be an indication that the heat fluxes into the mixed layer and the corresponding temperature changes are more effectively concentrated in the surface waters by the shallower thermocline which characterizes the heating season (see Fig. 3.4). A similar rationale can be used to explain the relative minima in the magnitudes of SST anomalies along the West Pacific side of 50°N in Figs. 5.2 and 5.3, by noting in Fig. 3.23 that this region is characterized by an extremely deep mixed layer over most of the year. In other words, the deep mixed layer over this region allows the effects of the heat fluxes to be mixed to considerable depths, which in turn minimizes temperature changes in the surface waters. From the above discussion, it would appear that the variability in the depth of the mixed layer plays a prominent role in determining the amplitudes which SST anomalies can attain on a seasonal and regional basis. In addition, the results from a study by Clark (1972) indicate that the interannual variability in the mixed layer depth also can have a large effect in determining anomalous sea surface temperature changes.

Yet another process, which conceivably could have some bearing on the magnitude which SST anomalies attain, is the QMD phenomenon discussed earlier in Chapter 3. This phenomenon, which appeared in the autocorrelations of Δ SST and the power spectra of SST anomalies, suggests that SST spectra become increasingly "red" as an increasing function of latitude. In other words, the time span over which the ocean integrates a similar tendency of heating or cooling is probably shorter at lower latitudes, where higher frequency variability is more prevalent in SST's. The minima in the magnitudes of SST anomalies at 20°N latitude may be a reflection of this QMD phenomenon discussed above, which could limit the amplitudes of SST anomalies over lower latitudes. Certainly other processes are affecting the magnitudes that SST anomalies can attain, but hopefully the major features of Figs. 5.2 and 5.3 could be simulated by superimposing the effects of the three processes described above.

Although the absolute magnitudes attained by SST anomalies vary only slightly between seasons, the correlations between anomalies of atmospheric thickness patterns and SST patterns reveal a pronounced seasonal difference. An example of this seasonal variability can be detected in Fig. 5.4, which shows the eight-year (1970-77) time series of correlations between anomalous patterns of SST's and 1000-850 mb thicknesses over the North Pacific on a six-day basis. Table 5.1 demonstrates that 1000-850 mb thickness layer generally provides the highest correlations to anomalous oceanic temperature patterns in comparison to other height and thickness layers. The significance levels included in Fig. 5.4 were computed with a technique similar to that described in Chapter IV, where the heat budget correlations were evaluated. One can see from Fig. 5.5

Fig. 5.4 Time series on a six-day basis of correlations between anomalous SST and 1000-850 mb thickness fields over the North Pacific grid outlined in Fig. 2.1. The dashed lines represent the 5% significance level (probability of occurrence by random chance) on a seasonal basis.

Fig. 5.5 Significance levels for correlations between anomalous SST and 1000-850 mb thickness fields over the North Pacific presented in a cumulative distribution so that the probability of occurrence of a given correlation coefficient by random chance can be identified. Solid line represents cooling season; dashed lines represent heating seasons.

that the seasonal variation in significant levels for correlations between 1000-850mb thickness and SST is such that a given correlation coefficient during the cooling season will be less significant than if that same coefficient occurred during the heating season. In spite of this fact, only 14.1% of the correlations in Fig. 5.4 are significant at the 95% level during the heating seasons in comparison to 24.2% during the cooling seasons. Namias (1973), finding similar results over the eastern North Pacific, related the higher correlations in fall and winter to cold continental air masses being advected over relatively warmer water. Such events effectively reduce the static stability of the lower troposphere and thereby permit greater heat loss than with stable air, which is more prevalent in spring and summer when air-sea temperature differences (see Fig. 5.6) are smaller and sometimes reversed.

Although anomalous 500-mb height patterns generally exhibit little relationship to anomalous SST's as suggested in Table 5.1, it does appear that the ocean-to-air heat exchange process is related, at least in some years, to characteristics of the long-wave pattern as manifested at 500 mb. Such a relationship can be seen during the cooling seasons of 1970 and 1977 by categorizing the 500-mb patterns corresponding to each six-day period during the above seasons into groups based on the significance of their pattern correlations in Fig. 5.4. The 500-mb patterns are averaged for each cooling season for the following two groups: 1) for the group associated with correlations that are significant at greater than the 95% level (see Fig. 5.7), and 2) for the group associated with correlations significant at less than the 50% level (see Fig. 5.8). By comparing the height patterns in the latter

Table 5.1

Annual average correlations for anomalous SST fields vs anomalous height and thickness fields over the North Pacific.

Year	Anomalous SST vs Anomalous									
	1000 mb	850 mb	700 mb	500 mb	1000- 850 mb	1000- 700 mb	1000- 500 mb	850- 700 mb	850- 500 mb	700- 500 mb
1970	.109	.176	.213	.223	.282	.275	.246	.245	.220	.194
1971	.000	.088	.127	.142	.303	.245	.200	.180	.159	.139
1972	.045	.103	.138	.160	.253	.224	.206	.185	.180	.165
1973	.019	.081	.118	.130	.291	.239	.185	.184	.143	.107
1974	.051	.095	.115	.131	.17	.148	.154	.129	.140	.131
1975	.073	.098	.130	.138	.095	.141	.136	.171	.140	.111
1976	.163	.191	.225	.227	.149	.228	.201	.252	.196	.150
1977	.092	.171	.195	.210	.318	.289	.254	.217	.203	.187
1970-77	.069	.125	.158	.170	.231	.223	.198	.195	.172	.148

Fig. 5.6 Eight-year (1970-77) average air-sea temperature difference ($^{\circ}\text{C}$) on a seasonal basis.

Fig. 5.7 Average 500 mb heights ($\times 10^1$ gpm) for periods during the 1970 and 1977 cooling seasons when significant (less than a 5% probability of occurrence by random chance) correlations occurred between anomalous SST and 1000-850 mb thickness fields over the North Pacific.

Fig. 5.8 Same as Fig. 5.7, except for periods when insignificant (greater than a 50% probability of occurrence by random chance) correlations occurred.

two figures for that in the jet stream, which is located in the region with the strongest height gradient, is farther south and the 500-mb surface typically lower, one can see in the cases associated with the significant correlations in Fig. 5.7. In other words, the high ocean-to-air heat exchange may be related to the anomalous position of the jet stream, which would decrease the lower atmosphere's stability over a greater expanse of the North Pacific by virtue of an increase in the areal extent of cold-air advection associated with a more southerly storm track than normal.

The effects of stability on the large-scale heat exchange can also be observed in the spatial patterns of grid point correlations between anomalous SST's and atmospheric thicknesses (1000-850 mb) on a seasonal and regional basis in Fig. 5.9. The correlations in this figure, which represents an average for the period (1970-77) were computed in a fashion similar to those of Figs. 4.7-4.9 in which the spatial patterns of heat budget correlations were examined. The results in Fig. 5.9 east of 180° longitude are comparable to those found by Namias (1973), in which the effects of the Aleutian and Gulf of Alaska lows, as well as of the northeast trades, can be surmised from the spatial correlation patterns. For example, relatively high correlations in the Gulf of Alaska have resulted from the fact the heat from the northern waters is readily provided to the lower troposphere as a consequence of cold continental air masses being advected into the baroclinic zones of this region. On the other hand, the high correlations over the southern extremities of the grids in Fig. 5.9 are a result of the exceptionally strong latent heat flux from the ocean (see Fig. 4.3) associated with the dry northeast trades which flow over the warmer water of this region. Since the

Fig. 5.9 The eight-year (1970-77) average spatial pattern of grid point correlations for SST anomalies versus 1000-850 mb thickness anomalies (shown on a seasonal basis for heating and cooling seasons as well as for the entire cycle - no seasonal bias). For clarity, the more significant correlations (greater than 0.3) are highlighted.

air flow in these trade winds is cooler than the waters over which it is advected (see Fig. 5.6), the heat is readily furnished to the lower troposphere. Another interesting feature of Fig. 5.9 is the zone of minimum correlations off the west coast of the United States. Quite likely the cold California Current and the upwelling, which occurs off the west coast, act to stabilize the lower atmosphere over this region and thereby inhibit significant heat exchange. Furthermore, the high correlations between anomalous SST and thickness patterns as observed in Fig. 5.9 off the coast of Japan are a reflection of the cold continental air masses, which are forced over the warm Kuroshio current, as surmised from the large air-sea temperature differences in Fig. 5.6 over this region.

VI Conclusions

The future of long-range weather forecasting may lie in our improved understanding of the ocean-atmosphere system, and more specifically in the nature of the interactions which dominate the variability of this coupled and complex system. Our study has examined many of the characteristics of the oceanic mixed layer over the North Pacific utilizing a number of statistical methods to analyze the copious amounts of available data. Hopefully, characterizing features of the air-sea system with statistical parameters will allow future studies to model more adequately the ocean-atmosphere system and its variability.

The variance field of Δ SST (six-day change in SST at a grid point) was examined in conjunction with temporal plots of meridional SST gradients as well as the variance fields of other parameters (e.g. wind speed, air-sea temperature difference, air-sea moisture gradient) which presumably are related to Δ SST. The results of this examination showed that the meanders of oceanic fronts are most likely responsible for the observed spatial pattern in the variance of Δ SST over the North Pacific. These frontal meanders result from perturbations in the mass transport and energy flux divergence zones which, in turn, are due to the atmospheric variability manifest in the other variance fields mentioned above.

To gain an appreciation for the responsive nature of the surface temperature fields over the North Pacific, the temporal autocorrelations of Δ SST were investigated. A predominant feature which appeared in this analysis was the occurrence over the majority of the North Pacific of a negative autocorrelation in Δ SST at lag one. From all the years that were examined within the period (1970-77), a meridional gradient was

observed in lag-one autocorrelation values with the larger negative values occurring at lower latitude. By reviewing the functions of the difference operator, Δ , as a high-pass filter on the Δ SST spectra and the cosine weighting inherent in the lag-one autocorrelation function, the relationship between the value of the lag-one autocorrelation of Δ SST and the shape of the SST spectrum was discussed. Using the spatial pattern of the lag-one autocorrelations of Δ SST as an indicator of SST spectra shape over the North Pacific, it was surmised, and then verified by actually computing SST anomaly spectra that the spectra of SST's become increasing "red" with increasing latitude across the North Pacific. This quasi-meridional differentiation in the characteristics, or shape, of SST spectra across the North Pacific was referred to as the QMD phenomenon. Although the characteristics of the QMD phenomenon vary to some extent over shorter periods, its persistence leads us to hypothesize on the mechanism(s) within the coupled air-sea system that might be producing this phenomenon. Initial results indicate that the atmosphere's forcing field over the North Pacific, as well as the ocean's response to forcing, which is a function of the ocean's vertical density structure, may be acting in a synergistic manner to produce the QMD phenomenon.

Another interesting feature observed in the SST anomaly spectra was the excess of power relative to the red noise null hypothesis that several spectral estimates contain in the frequency band 0.23-0.29 cycles per six days (21-26 days per cycle). The significance of these results becomes apparent when one examines the similar periodicities (at 21-26 days per cycle) which were found by Seigel (1977) to be significant in the cycles of oceanic heat fluxes and by McGuirk and Reiter

(1976) in cycles of zonal and eddy available potential and kinetic energies. It was suggested that the variability of SST anomalies in the frequency band (21-26 days per cycle) is a reflection of heat fluxes which often vacillate significantly on this time scale in conjunction with the cycles observed in the atmospheric energy parameters.

One of the main intents of this study was to examine, with considerable detail, a portion of the oceanic heat budget over the North Pacific which past researchers have determined to be important. Although this investigation of the heat budget was limited in the parameters that it examined (e.g. SST change, latent and sensible heat fluxes, and the variability of the parameters with respect to fluctuations in the mixed layer depth), it was designed so that the spatial and temporal resolution of the heat budget, as well as its spatial coverage, could be affectively maximized. Examination of the heat budget on a spatial and temporal basis indicates that the impact of latent and sensible heat fluxes upon Δ SST is partially a function of the magnitude of these heat fluxes and the depth to which their effects are mixed. More specifically, the random chance of an insignificant correlation between latent plus sensible heat and Δ SST tends to decrease if either the heat fluxes increase in magnitude or the depth to which their effects are mixed decreases, since in either of these two circumstances the heat fluxes would be determining a larger percentage of Δ SST.

The results of the spatial analysis of the heat budget as well as the fitting of power spectra of SST anomalies over the North Pacific to a two-parameter model suggest that SST behavior over the mid-oceanic regions of the North Pacific is dominated by the influence of latent and sensible heat fluxes, whereas over the remainder of the North Pacific

one could surmise that other oceanic processes are at least as important in determining the behavior of SST's.

By evaluating the heat budget with consideration to the depth of vertical mixing, it appeared that the heat entrained at the base of mixed layer is a significant factor in determining SST characteristics, especially in the regions and during periods of rapid mixed layer deepening. Furthermore, it would appear that by considering the variability in the depth of mixing on a seasonal as well as interannual basis, one could better determine the magnitude of impact that heat fluxes into the mixed layer would have upon Δ SST.

In comparing anomalous patterns of atmospheric thickness and SST's, our findings were consistent with those discussed by Namias (1973), which indicated that SST patterns of large magnitudes are more apt to be highly correlated with anomalous thickness patterns of the lower troposphere than weak SST anomaly patterns. In addition, the regions characterized by the largest six-day averaged SST anomalies are located in the vicinity of strong currents and oceanic fronts, implying the importance of the advection within the mixed layer in generating large SST anomalies. Further evidence, though inconclusive, suggests that the magnitude which SST anomalies attain is also influenced by the variability in the depth of mixed layer as well as the QMD phenomenon.

Finally, it appears that the modification of air masses as they are advected over oceanic waters and the stability of the lower atmosphere are instrumental factors in determining the nature of large-scale heat exchange processes. For example, when the jet stream at 500 mb is located farther south over the North Pacific and the 500 mb heights are lower than average, there is a strong tendency for more significant

correlations to occur between anomalous lower tropospheric thickness and anomalous SST patterns, which could indicate that during some periods the high ocean-to-air heat exchange is related to a more southerly storm track which would effectively decrease the lower tropospheric stability over a greater expanse of the ocean by virtue of an increase in the areal extent of cold-air advection over a relatively warmer ocean.

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APPENDIX 1

From the conservation of sensible heat or enthalpy,

$$\rho c_p \frac{dT}{dt} = \dot{Q} \quad (1)$$

where

$$\rho \equiv \text{water density} = 10^6 \text{ gm}^{-3}$$

$$c_p \equiv \text{specific heat capacity of water} = 1 \text{ cal g}^{-1} \text{ } ^\circ\text{C}^{-1}$$

$$T \equiv \text{temperature (} ^\circ\text{C)}$$

$$\dot{Q} \equiv \text{heat source (cal m}^{-3} \text{ s}^{-1})$$

In Eq. 1, $\dot{Q}$ represents the absorption of solar radiation within the water column. The schematic diagram below shows the attenuation with depth in the penetrating component of solar radiation, I .

where sea level is represented by $Z = 0$. Within the finite layer, ΔZ , the heat absorbed would be:

$$\frac{I - (I - \Delta I)}{-\Delta Z} = - \frac{\Delta I}{\Delta Z} \quad (2)$$

and as the layer becomes infinitely small $\dot{Q} = -\frac{\partial I}{\partial Z}$. Therefore, if $\frac{\partial I}{\partial Z}$ is negative as shown in the diagram above, a flux convergence (i.e. a source of heat) will exist in the infinite layer. It follows that

$$\frac{dT}{dt} = - \frac{1}{\rho c_p} \frac{\partial I}{\partial Z}$$

$$\text{or} \quad \frac{\partial T}{\partial t} + u \frac{\partial T}{\partial x} + v \frac{\partial T}{\partial y} + w \frac{\partial T}{\partial z} = - \frac{1}{\rho c_p} \frac{\partial I}{\partial Z} \quad (3)$$

From mass continuity

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0; \quad T_x(\nabla \cdot \bar{v}_3) = 0 \quad (4)$$

Combining (3) and (4)

$$\frac{\partial T}{\partial t} + \frac{\partial(uT)}{\partial x} + \frac{\partial(vT)}{\partial y} + \frac{\partial(wT)}{\partial z} = \frac{-1}{\rho c_p} \frac{\partial I}{\partial z}$$

One now defines $(\bar{\quad})$ as a horizontal average over an area which is large compared to an individual convective element and small compared to the scale of interest.

Idealized cross section of mixed layer showing individual convective elements, (represented by loops), and averaging area. Heavy dashed line denotes the vertical extent of mixing.

$$\frac{\partial \bar{T}}{\partial t} + \frac{\partial(\bar{uT})}{\partial x} + \frac{\partial(\bar{vT})}{\partial y} + \frac{\partial(\bar{wT})}{\partial z} = \frac{-1}{\rho c_p} \frac{\partial \bar{I}}{\partial z} \quad (5)$$

One may expand the following terms as:

$$\overline{(uT)} = \overline{(\bar{u} + u')(T + T')} = \bar{u}\bar{T} + \underbrace{\bar{u}T'}_{\text{zero by definition}} + \underbrace{\bar{u}'\bar{T}}_{\text{zero by definition}} + \overline{u'T'}$$

where

$' \equiv$ represents deviation from horizontal average

$\overline{u'T'} \equiv$ non-zero only if there is a correlation between u' and T'

$\overline{(vT)}$ and $\overline{(wT)}$ may be expanded in a similar manner.

Assuming horizontal homogeneity in these fluxes which is basic to one-dimensional models, one can write

$$\frac{\partial \overline{u'T'}}{\partial x} = \frac{\partial \overline{v'T'}}{\partial y} = 0$$

and since w' and T' are correlated, $\overline{w'T'} \neq 0$ and $\frac{\partial \overline{w'T'}}{\partial Z} \neq 0$. Eq. 5 becomes

$$\frac{\partial \bar{T}}{\partial t} + \frac{\partial \bar{u}\bar{T}}{\partial x} + \frac{\partial \bar{v}\bar{T}}{\partial y} + \frac{\partial \bar{w}\bar{T}}{\partial Z} = - \frac{\partial \overline{w'T'}}{\partial Z} - \frac{1}{\rho c_p} \frac{\partial \bar{I}}{\partial Z} \quad (6)$$

Since $T(\frac{\partial \bar{u}}{\partial x} + \frac{\partial \bar{v}}{\partial y} + \frac{\partial \bar{w}}{\partial Z}) = 0$, Eq. 6 becomes

$$\frac{\partial \bar{T}}{\partial t} + u \frac{\partial \bar{T}}{\partial x} + v \frac{\partial \bar{T}}{\partial y} + w \frac{\partial \bar{T}}{\partial Z} = - \frac{\partial \overline{w'T'}}{\partial Z} - \frac{1}{\rho c_p} \frac{\partial \bar{I}}{\partial Z} \quad (7)$$

where $\frac{\partial \bar{T}}{\partial x} = \frac{\partial \bar{T}}{\partial y} = 0$ in Eq. 7 again due to horizontal homogeneity. Since within the mixed layer $\frac{\partial \bar{T}}{\partial Z} = 0$, one may write an equation which applies at any depth within the mixed layer.

$$\frac{\partial \bar{T}}{\partial t} = - \frac{\partial \overline{w'T'}}{\partial Z} - \frac{1}{\rho c_p} \frac{\partial \bar{I}}{\partial Z} \quad (8)$$

Integrating Eq. 8 over the mixed layer from $Z = 0$ to $Z = -h$, where $\bar{I} = I_0 e^{vZ}$ and v is an attenuation coefficient which varies between $0.03m^{-1}$ in clear waters to $0.3m^{-1}$ in murky coastal waters yields

$$\int_{Z=-h}^{Z=0} \frac{\partial \bar{T}}{\partial t} dZ = - \int_{Z=-h}^{Z=0} \frac{\partial \overline{w'T'}}{\partial Z} dZ - \frac{1}{\rho c_p} \int_{Z=-h}^{Z=0} \frac{\partial \bar{I}}{\partial Z} dZ \quad (9)$$

Since $\frac{\partial \bar{T}}{\partial t}$ is not a function of Z Eq. 9 can be rewritten as:

$$\frac{\partial \bar{T}}{\partial t} \int_{Z=-h}^{Z=0} dZ = - \int_{\overline{w'T'}(Z=-h)}^{\overline{w'T'}_0} \overline{w'T'} - \frac{1}{\rho c_p} \int_{I(Z=-h)}^{I_0} \partial I$$

$$\frac{\partial \bar{T}}{\partial t} (0 - (-h)) = -(\overline{w'T'}_0 - \overline{w'T'}_{Z=-h}) - \frac{1}{\rho c_p} (I_0 - I_0 e^{-vh}) \quad (10)$$

The surface flux of heat can be specified in most cases with reasonably good approximation by the following bulk aerodynamic formula

$$\rho c_p \overline{w'T'}_0 = R_0 - I_0 + SH_0 + LH_0 \quad (11)$$

where

$R_0 \equiv$ surface flux of solar and terrestrial radiant energy

$I_0 \equiv$ penetrating component of solar radiation at surface

$SH_0 \equiv$ surface flux of sensible heat

$LH_0 \equiv$ surface flux of latent heat

All these fluxes are considered positive when directed upwards from water into the air. To determine the heat flux at the bottom of the mixed layer one integrates Eq. 8 from the top ($H + \frac{\Delta H}{2}$) to the bottom ($H - \frac{\Delta H}{2}$) of the quasi-discontinuity and then assumes that $\Delta H \rightarrow 0$. One can see more clearly how these integration limits are obtained by viewing the schematic diagram below.

One must also use the fact that $(vh) \geq 1$, where h is the depth of the mixed layer. Thus, e^{-vh} is negligibly small. It follows that $\frac{\partial \bar{T}}{\partial Z}$ can also be neglected at the bottom of the mixed layer.

$$\int_{H-\Delta H/2}^{H+\Delta H/2} \frac{\partial \bar{T}}{\partial t} dz + \int_{H-\Delta H/2}^{H+\Delta H/2} \frac{\partial}{\partial Z} \overline{w'T'} dz = 0 \quad (12)$$

The term $\bar{w} \frac{\partial \bar{T}}{\partial Z}$ is not included in this integration because we have assumed herein the absence of a mean vertical velocity ($\bar{w} = 0$). This assumption is used consistently throughout this model.

The first term in Eq. 12 can be expanded using Leibnitz's Theorem since $H=H(t)$

$$\frac{\partial}{\partial t} \int_{H - \frac{\Delta H}{2}}^{H + \frac{\Delta H}{2}} \bar{T} dZ - SST \frac{\partial H}{\partial t} + T_b \frac{\partial H}{\partial t} + \overline{w'T'}_{Z=H+\frac{\Delta H}{2}} - \overline{w'T'}_{Z=H-\frac{\Delta H}{2}} = 0 \quad (13)$$

where

$SST = T_o \equiv$ sea surface temperature

$T_b \equiv$ sea temperature at base of thermocline.

The last term in Eq. 13 is zero since the region below the mixed layer is non-turbulent. The first term in Eq. 13 becomes,

$$\frac{\partial}{\partial t} \int_{H-\Delta H/2}^{H+\Delta H/2} T dZ = \frac{\partial \bar{T}}{\partial t} (H+\frac{\Delta H}{2}) - \frac{\partial \bar{T}}{\partial t} (H-\frac{\Delta H}{2}) = \frac{\partial \bar{T}}{\partial t} (\Delta H) = 0 \quad \text{as } \Delta H \rightarrow 0 \quad (14)$$

From Eq. 13 and Eq. 14 one can write,

$$(T_b - SST) \frac{\partial H}{\partial t} + \overline{w'T'}_{Z=-h} = 0 \quad (15)$$

where

$$\Delta T = T_b - SST$$

$we = \partial H / \partial t \equiv$ rate at which water is entrained from below the mixed layer.

Eq. 13 becomes,

$$\overline{w'T'}_{Z=-h} + we \Delta T = 0 \quad (16)$$

Eq. 16 is referred to as the mixed layer bottom jump condition.

Inserting this last boundary condition along with Eq. 11 into Eq. 10, one obtains upon rearrangement:

$$\frac{\partial T}{\partial t}(h) = - \left[\frac{(R_o - I_o + SH_o + LH_o)}{\rho c_p} + w_e \Delta T \right] - \frac{(I_o - I_o e^{-vh})}{\rho c_p} \quad (17)$$

Since $T = SST$ and cancelling the I_o terms, Eq. 17 becomes

$$\frac{\partial(SST)}{\partial t} = \frac{-1}{\rho c_p h} (R_o + SH_o + LH_o - I_o e^{-vh}) - \frac{w_e \Delta T}{h} \quad (18)$$

Eq. 18 is the mixed layer temperature equation in differential form.

APPENDIX 2

Frankignoul and Hasselmann (1977) chose the simplest possible oceanic model, where the uppermost ocean layer is assumed to be completely mixed and horizontally homogeneous down to depth "h". The prognostic equation governing the change in the mixed layer temperature for their model is given as:

$$\frac{dT}{dt} = -\frac{1}{\rho c_{ph}} (R_o + SH + LH - I_o e^{-vh}) - \frac{W_e \Delta T}{h} \quad (1)$$

where:

$T \equiv$ mixed layer temperature

$\rho \equiv$ water density

$h \equiv$ depth of mixed layer

$R_o \equiv$ surface flux of solar and terrestrial radiant energy

$SH \equiv$ surface flux of sensible heat

$LH \equiv$ surface flux of latent heat

$I_o e^{-vh} \equiv$ penetrating solar radiation which passes into layers below "h"

$W_e = \frac{dh}{dt} \equiv$ rate at which water is entrained from below mixed layer

$\Delta T \equiv$ quasi-discontinuity in temperature across bottom of mixed layer

This model is further simplified by assuming the mixed layer depth remains constant. Results from Thompson (1976) demonstrate that a one-dimensional mixed layer model, which assumes a constant mixing depth, will generate SST's that are too warm in summer and too cold in winter, if the mixed layer is assumed to be too shallow (e.g. 15 meters). On the other hand, Thompson's results showed that with the

depth adjusted to 30 meters, the longer-term trends (e.g. seasonal) will be correct, but the short-term (shallow) fluctuations in SST's will not be visible. Furthermore, this model ignores the variations in radiation flux, which could lead to significant discrepancies between actual and model simulated SST anomalies, especially in summer when radiation flux into the mixed layer maximizes. Based on these assumptions, Eq. 1 can be written as:

$$\frac{dT}{dt} = \frac{1}{h} f(T) \quad (2)$$

where

$$f(T) = [f(T)] + f'(T) = -\frac{1}{\rho c_p} (SH + LH)$$

[] - long-term mean ' - departure from mean

$f(T)$ is now expanded about an equilibrium temperature ($T=0$) for which $[f]_{T=0} = 0$.

$$f(T) = f(T=0) + \left(\frac{\partial f}{\partial T}\right)_{T=0} T + \dots \text{higher order terms that can be neglected}$$

$$f(T=0) = [f]_{T=0} + f'_{T=0} = f'$$

Eq. 2 can be rewritten as:

$$\frac{dT}{dt} = \frac{f'}{h} - \lambda T \quad (3)$$

where

$$\lambda = -\left(\frac{\partial f}{\partial T}\right) h^{-1}$$

The " λ " in Eq. 3 represents a negative feedback process leading, in this approximation, to an exponential decay back to the climatological temperature of the system.

Equation 3 has the form of a first-order autoregressive process. The spectra generated from the time series solution of this equation can be obtained by taking the Fourier transform of both sides of Eq. 3 as shown below.

$$\frac{1}{2\pi} \int_{-\infty}^{\infty} e^{-i\omega t} dt \times \left[\frac{dT}{dt}(t) = \frac{f'}{h} - \lambda T \right] \quad (4)$$

Integration of Eq. 4 yields:

$$i\omega T(\omega) = F(\omega) - \lambda T(\omega),$$

or upon rearrangement

$$T(\omega) = \frac{-F(\omega)}{(i\omega + \lambda)} \quad (5)$$

Multiplying both sides of Eq. 5 by the appropriate complex conjugates (denoted with a *) results in the following expression for the power spectrum.

$$G(\omega) = \langle T^*(\omega) \cdot T(\omega) \rangle = \frac{\langle F^*(\omega) \cdot F(\omega) \rangle}{\omega^2 + \lambda^2} \quad (6)$$

$\langle \rangle$ in Eq. 6 represents the ensemble average over all realizations of the stochastic process representing the random force $f'(T)$ or $F(\omega)$.

The aliased form of the power spectrum $G(\omega)$ in Eq. 6 is given by Reynolds (1977) as:

$$G(\omega) = \frac{2\sigma}{1 + \alpha^2 + 2\alpha \cos(2\pi f)} \quad (7)$$

where

$\sigma \equiv$ variance of white noise input spectrum

$\alpha \equiv$ autocorrelation first lag

$f \equiv$ frequency

Frankignoul and Hasselmann (1977) have shown empirically that the spectrum of forcing, $\langle F^*(\omega) \cdot F(\omega) \rangle$ in Eq. 6, is independent of frequency over the range of SST anomaly response of interest here. It would follow that the spectral density of this white noise input

spectrum, $\langle F^*(\omega) \cdot F(\omega) \rangle$, is proportional to the variance of its spectrum, σ , in fact $\sigma = 2\langle F^*(\omega) \cdot F(\omega) \rangle$.

" α " in Eq. 7 can be shown to be the lag "one" autocorrelation of the SST anomaly time series by writing the discrete analogue of Eq. 3 as:

$$T_{t+1} = \frac{f'}{h} \Delta t - \lambda \Delta t T_t + T_t$$

$$\text{or } T_{t+1} = F_t + \alpha T_t \quad (8)$$

where

$$T_{t+1} \equiv \text{SST at time } (t+1)$$

$$T_t \equiv \text{SST at time } (t)$$

$$F_t = \frac{f'}{h} \Delta t \equiv \text{which essentially is white noise input spectrum}$$

$$\alpha = (1 - \lambda \Delta t), \text{ and is a dimensionless quantity.}$$

Multiplying Eq. 8 by T_t and averaging the various terms with respect to time, one obtains:

$$\langle T_t \cdot T_{t+1} \rangle = \alpha \langle T_t^2 \rangle + \langle F_t \cdot T_t \rangle \quad (9)$$

Since F_t and T_t are uncorrelated due to the white noise characteristics of F_t , the last term in Eq. 9 is zero. Therefore, upon rearrangement:

$$\alpha = \frac{\langle T_t \cdot T_{t+1} \rangle}{\langle T_t^2 \rangle}$$

which, by definition, is the autocorrelation at the first lag.