Latent and Sensible Heat Flux Anomalies over the Northern Oceans:
Driving the Sea Surface Temperature

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(Manuscript received 27 December 1990, in final form 25 October 1991)

ABSTRACT

A part of the large-scale thermodynamic forcing of the upper ocean is examined by relating monthly anomalous latent and sensible heat flux to changes in sea surface temperature (SST) anomalies over the North Atlantic and North Pacific. The fluxes are estimated using bulk formulas from a set of about four decades of marine observations from the COADS dataset from 1946 to 1986. Monthly anomalies are constructed by removing the long-term monthly means. The latent and sensible flux anomalies are strongly correlated over most of the ocean, so they are considered together as a sum.

The heat flux estimates contain large spatial-scale anomalies consistent with both atmospheric circulation anomalies and with month-to-month changes (tendencies) in monthly SST anomalies. The monthly flux anomalies and the SST anomaly tendency are significantly correlated over much of the oceans, with anomalous positive/negative fluxes associated with anomalous cooling/warming. The connection between the flux and the SST tendency anomalies is strongest in the extratropics during the cool season when the latent and sensible fluxes and their variability are greatest, and the radiative fluxes are weakest.

While the heat flux forcing of the SST anomalies operates locally, the flux and SST tendency anomalies are organized over spatial scales that often span major portions of the North Atlantic and North Pacific. For each basin, canonical correlations expose large-scale, collocated anomaly patterns in the two fields. These patterns reflect the control exerted by the large-scale atmospheric circulation, inferred from sea level pressure (SLP) modes. Evidence for this result is the strong similarity in the configuration of anomalous flux and SST tendency patterns in their association with major SLP modes. Typical flux anomalies of 50 W m$^{-2}$ are associated with monthly SST anomaly changes of order 0.2°C. The surface-layer thickness inferred from a simplified model relating the flux anomalies to the temperature anomalies of a slab ocean is consistent in magnitude and seasonal cycle with the observed mixed-layer depth in middle latitudes.

1. Introduction

Latent and sensible heat transfers are primary mechanisms by which the ocean vents the heat absorbed from solar radiation. Globally, approximately 60% of the solar radiation absorbed at the earth’s surface is released by latent and sensible heating, primarily from the ocean (Sellers 1965). Gill and Niler (1973) invoked scaling arguments in the governing equations to infer that, over large scales ($\geq 1000$ km meridionally and $\geq 3000$ km zonally), anomalous open-ocean surface temperature changes are dominated by changes in heat fluxes through the surface, in comparison with advective influences. The present study is an effort to test a portion of the surface heat budget by relating monthly anomalies of the latent and sensible fluxes to anomalous tendencies of the sea surface temperature (SST) field.

There is ample evidence for significant anomalies of the latent and sensible fluxes on time scales over the weather to short-period climate range. In midlatitudes, largest latent and sensible flux values occur preferentially in the cold sectors of storms (Bond and Flesagle 1988). Bunker’s (1981) analysis of monthly latent and sensible flux in the North Atlantic reveals significant monthly to interannual variations. Zhao and McBean’s (1986) analysis revealed coherent flux anomalies over hundreds of kilometer scales in the North Pacific. Weare (1984) showed that in the tropical Pacific, anomalies of the latent flux are systematically related to the El Niño–Southern Oscillation phenomenon.

This study is the third portion of the author’s Ph.D. dissertation (Cayan 1990; hereafter designated C1), which analyzed the anomalous variability of bulk formulas latent and sensible fluxes on monthly scales over the global ocean. The first two parts of this work, which are summarized in separate papers, concern an analysis of variance of the anomalous fluxes (Cayan 1991a; hereafter designated C2), and their relationship to the atmospheric circulation (Cayan 1991b; hereafter des-
igned C3). Because C2 and C3 provide background useful to the present study, pertinent results from them are briefly summarized here.

Cayan (1991a) explored the dependence of the flux anomalies on the mean conditions and the anomalies of the fundamental surface variables. Mechanistically, to understand the flux anomalies, the joint behavior of mean values and anomalies of wind speed (\(w\)), the ocean surface saturation humidity–air humidity \(\Delta q\), and the sea–air temperature difference (\(\Delta T\)) must be included. Spatial and temporal changes in the mean and anomalous fields lead to strong geographical and seasonal variability of the latent and sensible fluxes (\(F_l\) and \(F_s\)). In middle latitudes during winter, largest contributions to the variance of monthly \(F_l\) anomalies involve \(\bar{w}\Delta q\); that is, the effect of \(\Delta q\) anomalies in the presence of high mean \(w\). The \(F_l\) anomalies become important poleward of about 35°, where their variance is dominated by the term involving \(\bar{w}\Delta T\); that is, the effect of \(\Delta T\) anomalies in the presence of high \(\bar{w}\). Both means and variances of the fundamental variables are greatest near the continental margins in the western extratropical ocean basins, so that the monthly variance of \(F_l\) and \(F_s\) have marked western intensified patterns. Factors driving tropical latent flux anomalies often differ from those in middle latitudes, with greater contributions involving \(\Delta q\sqrt{w}\); that is, the effect of \(\sqrt{w}\) anomalies in the presence of high \(\Delta q\).

The \(\Delta q\) and \(\Delta T\) monthly anomalies are correlated (cooler air is usually drier), especially in the extratropics, so that latent and sensible flux anomalies usually reinforce, yielding relatively large net heating anomalies in winter. Cayan (1991a) exhibited combined \(F_l\) and \(F_s\) anomalies that often exceed 100 W m\(^{-2}\) over several hundred kilometer regions. These estimates suggest that monthly anomalies of \(F_l\) and \(F_s\) are usually larger than those of the radiative components. Comparing the variances of \(F_l\) and \(F_s\), with those of bulk formulas estimated net solar radiative flux (\(F_{SW}\) and infrared flux (\(F_{IR}\)), C2 found that 1) during winter in the extratropics north of 30°N, \(F_l\) and \(F_s\) anomalies dominate the monthly anomalous surface heat budget; 2) from about 15°N to about 30°N, \(F_l\) anomalies dominate; 3) from 10°N to 30°S, \(F_l\) and \(F_{SW}\) anomalies are about equally large.

Finally, C2 showed that a major portion of the \(F_l\) and \(F_s\) anomalies are coherent over basin-scale regions in both the North Atlantic and North Pacific. This constitutes strong evidence that the monthly flux anomalies contain a short-period "climate signal." The first four empirical orthogonal functions (EOFs) of the sum of the \(F_l\) and \(F_s\) anomalies account for about half of the total variance of this field in each basin. The EOFs represent regionally coherent anomaly packets, with typical maxima ranging from 40 to 100 W m\(^{-2}\).

In C3 the influence of the atmospheric circulation on monthly anomalies of ocean surface latent and sensible heat fluxes was explored. In the extratropics, flux anomalies are linked to the local wind direction, but not in the tropics. In the North Atlantic and North Pacific (north of about 15°N), largest positive anomalies are associated with northerly to northwesterly winds. This appears to be a response to advection of lower atmospheric humidity and temperature from north to south, and also to strong wind speeds associated with westerly winds. In the tropics, there is little relationship between the wind direction and the \(F_l\) and \(F_s\) anomalies since horizontal gradients of humidity and temperature are weak and the wind direction is relatively steady.

One of the most convincing aspects of the climate variability exhibited by the estimated flux anomalies is their organization with respect to monthly atmospheric circulation patterns. Cayan (1991b) showed that the dominant, atmospheric circulation anomaly modes, represented as EOFs of the sea level pressure (SLP), have systematic patterns of pertinent surface marine variables in the extratropical North Atlantic and North Pacific. These variables include the wind speed (\(w\)), surface saturation humidity–air humidity difference (\(\Delta q\)), and sea surface temperature–air temperature difference (\(\Delta T\)). In concert, these yield large-scale patterns in \(F_l\) and \(F_s\). Negative SLP anomalies tend to have positive \(w\) anomalies to their south and negative \(w\) anomalies to their north, probably because of shifts in storm tracks and changes in the mean wind field. Positive \(\Delta q\) (and \(\Delta T\)) tend to occur to the west of negative SLP anomalies, and positive \(\Delta q\) to their east, apparently because of meridional advection of air temperature and humidity. The resultant flux anomalies are positioned meridionally and zonally relative to the SLP centers, with enhanced sea-to-air fluxes to the southwest and diminished fluxes to the east of negative SLP anomalies.

Like C1, C2, and C3, the present study employs bulk formula parameterized heat fluxes, calculated from standard marine surface meteorological observations. These are the only means to estimate a multyear time history of the heat exchange over broad regions of the oceans. Similarly, SST has been measured routinely by merchant ships for several decades, so SST is used to infer the variability of the upper-ocean heat content. Although the structure of upper-ocean heat content can be complex, during winter the upper ocean is quite well mixed and SST is a good indicator (White and Walker 1974).

The relationship between the climatological mean latent and sensible heat flux and that of SST appears to support two different points of view. On one hand, the fluxes may be driven by the SST: High ocean-to-atmosphere fluxes are found in the tropics where there is warm water and high-saturation vapor pressure, and low fluxes occur in high latitudes and along the eastern side of the ocean basins where surface temperatures are cool. For the anomaly case, there is evidence for the SST-forcing flux mechanism in the warm season
extratropics and in the tropics (C1; Liu and Gautier 1990).

On the other hand, in the extratropics the fluxes appear to force the temperature (Gill and Niler 1973); SST is coolest in late winter following the highest ocean-to-atmosphere fluxes. For the flux-forcing SST case to be valid, the SST anomaly must play a negligible role in determining the flux, so that the flux is driven by atmospheric conditions—that is, $w$, $T_a$, and $q$ variations dominate those of SST. Indirectly, this scenario has been investigated by relating SST anomalies to the large-scale atmospheric circulation (Bjerknes 1964; Namias 1972; Davis 1976; Wallace et al. 1990). While all of these studies established a definite connection between the monthly or seasonal mean circulation and the SST anomalies, they could only infer the role of anomalous heat flux in driving SST anomalies. Most investigations of the heat flux on effects on the upper-ocean thermal structure have considered the heat budget at particular locations, such as weather ships (Tabata 1965; Husby and Seckel 1975; Talley and Raymer 1982). In addition to local impacts documented in these studies, there is also evidence that the heat fluxes are associated with larger-scale patterns. Using ocean general circulation models, Haney (1985) and Lukesch et al. (1989) indicated that anomalous heating by surface fluxes is a major component in producing monthly thermal anomalies in the North Pacific Ocean. Correlations between bulk formula fluxes and anomalous temperature variations have been documented over the North Pacific by Clark (1967) and Frankignoul and Reynolds (1983), and in the western tropical Pacific by Meyers et al. (1990).

In this study we test the influence of the latent and sensible heat-flux anomalies upon monthly changes in SST anomalies over a large portion of the world oceans. When atmospheric fields are related to contemporaneous SST anomalies, it is difficult to determine whether SST is forcing the atmosphere, both fields are being forced by some other agent, or the atmosphere is forcing the SST. Here we remove this ambiguity by adopting a simple thermodynamic model to relate the flux anomalies to the tendency (time rate of change) of the SST anomalies. This is clearly a test of the atmosphere-forcing SST scenario. These analyses encompass a broad region of the World Ocean, using surface marine observations throughout approximately four decades since 1946. Results from the North Atlantic are compared with those for the North Pacific, and also to some extent with other ocean areas, including the tropics.

The study is organized as follows: data employed are described in section 2; aspects of the bulk formula latent and sensible flux estimates and their uncertainties are described in section 3; local relationships to the SST anomalies are discussed in section 4; the relation between patterns of the flux and SST anomalies is discussed in section 5; and the consistency between the magnitudes of the two variables is documented in section 6.

2. Data

The latent and sensible heat-flux and SST estimates are calculated from monthly mean marine surface data for the period 1946–86. These data are from the Monthly Summaries Trimmed (MST) subset of the COADS marine dataset (Slutz et al. 1985; Woodruff et al. 1987). To derive the fluxes, monthly means of several variables were employed, including the scalar wind ($w$), SLP, specific humidity ($q$), the surface saturation specific humidity ($q_s$), air temperature ($T_a$), and SST. The surface saturation humidity–air humidity difference ($\Delta q$) and the SST minus $T_a$ difference ($\Delta T$) are positive when the values at the surface ($q_s$ or SST) are greater than those in the air ($q$ or $T_a$). The bulk formulation of the fluxes employs the products ($w\Delta q$ and $w\Delta T$), which are supplied by the COADS MST as monthly averages of the products of individual pairs of observations.

For each of the North Atlantic and North Pacific basins from 1946 to 1979 there were, on average, 30 000–40 000 concurrent observations per month from which to calculate $w\Delta q$ or $w\Delta T$, yielding approximately 14 million flux total “observations” from this period for each basin. The sampling density generally increases with time; the period before about 1958 is quite sparse in the Pacific sector. A preliminary version of the second COADS release for 1980–86 (Woodruff et al. 1987) was later obtained and merged with the original observations to update the data to 1986.

Since our primary interest concerns air–sea interaction over several hundred kilometer scales, each of the COADS $2^\circ$ data fields was averaged onto a $5^\circ$ latitude–longitude grid, centered on $5^\circ$ latitude–longitude intersections. The data density and the spatial averaging scheme is described in C1 and C2. For comparison with the climatology of other COADS variables by Shea (1986), long-term means were calculated for 1950–79. Analyses here were based on monthly data, but in some cases, they were aggregated over a season, under the usual Northern Hemisphere convention where winter is the average of December, January, and February, and so on for the other seasons.

The quasi-global coverage of COADS allows comparison between behavior of different ocean basins. In addition to the global ocean grid, each variable employed in this study was subdivided into two different geographical arrays, one for the North Pacific and one for the North Atlantic, so that the flux, SST, and atmospheric circulation fields could be examined individually for each basin. Monthly SLP anomalies were employed to calculate EOFs, which were used to identify the prominent wintertime atmospheric circulation modes within each of these two basins (C3).
All of the marine data contain errors (Weare and Strub 1981; Taylor 1984). For example, SST is affected by both observation errors and biases due to sampling procedures. Tabata (1978) compared intake temperatures with bucket temperatures collected in the northeast Pacific on a Canadian research vessel and found that intake temperatures were on average higher than bucket temperature by 0.3°C. In addition, intake temperatures varied about the bucket temperature with a standard deviation of 1.2°C; Tabata (1978) indicates that much of this random error was caused by incorrect thermometer readings. There are also errors introduced into the surface wind speed because of the difficulty in calibrating changes over the record from sea-state wind force estimates to directly measured anemometer readings, inconsistencies from differences in anemometer height from one vessel to another, and trends from changes over time in the height of ships and their anemometers. Although individual corrections of these problems were beyond the scope of this study, the latent and sensible flux were adjusted for trends in the global average wind speed and ΔT from COADS (see C1 and C2 and the following discussion).

The random error in the SST and other marine variables is reduced by averaging the observations within a 5° latitude-longitude “square” over each month (many 5° squares in the North Atlantic and North Pacific have 100 or more observations per month). Also, the time-invariant bias is removed from the SST because the mean temperatures were subtracted to form anomalies. Furthermore, the analyses performed were insensitive to trends in the data because the SST anomalies were differentiated to form the time derivative. Errors in the heat flux estimates are discussed in section 3.

3. Bulk formula parameterization

a. Introduction

The only practical means of estimating the air–sea heat exchange on a routine basis over a several-decade time history is to use “bulk formula” parameterizations of routine weather observations measured near the sea surface. Here, bulk formula’s latent and sensible fluxes are derived over a four-decade dataset and monthly anomalies are constructed by subtracting the long-term monthly mean. A detailed description is provided in C1 and C2, but a brief summary of the flux derivation is also given here.

The bulk formulas for latent (F_l) and sensible (F_s) sea–air heat fluxes are

\[ F_l = \rho L C_w (q_s - q_a) \]
\[ F_s = \rho C_p C_H w (SST - T_a), \]

where \( w \), \( q_a \), and \( T_a \) are the scalar wind (wind speed), specific humidity, and temperature of the air in the boundary layer at a given height (observation level) above the sea surface; \( q_s \) and SST are the saturation specific humidity and sea surface temperature; \( \rho \) is the density of the air at observation level, calculated using the ideal gas equation with a correction for the virtual temperature; \( L \) is the latent heat of evaporation of water, which is calculated from the surface temperature, but is approximately \( 2.50 \times 10^6 \) J kg\(^{-1}\); \( C_w \) is the specific heat of air at constant pressure, taken to be constant at \( 1.0048 \times 10^3 \) J kg\(^{-1}\) °C; and \( C_E \) and \( C_H \) are the transfer coefficients for latent heat (water vapor) and sensible heat, respectively, taken from Isenmer and Hasse (1987, Table 4, p 14). These transfer coefficients generally increase with wind speed and with ΔT. The convention adopted here was that the flux is positive for heat given up by the ocean to the atmosphere.

As for those of the other marine variables employed in this study, the flux anomalies were constructed on a 5° latitude-longitude grid by averaging the pertinent monthly anomalies from 2° “squares” provided in the COADS MST data (C1; C2). Because the latent and sensible flux anomalies are well correlated (C1), the sum of their anomalies \( F_{l+s} \) was used in all of the analyses shown here.

b. Errors in the bulk formula latent and sensible fluxes

Ideally, the monthly heat-flux estimates would only contain short-term climate fluctuations having periods of two months and greater. However, because of uncertainties in the bulk formulas and errors in the marine observations, there is a significant error component that contaminates the monthly mean fluxes. Details of the errors in the fluxes are provided in Weare (1989), C1, and C2. In summary, there are three types of errors that affect the monthly bulk formula latent and sensible fluxes. Two of these, observation errors and weather sampling errors, are random in character and are reduced by averaging several observations together. The other error type is associated with uncertainties in the bulk formulas and by biases in the observations, which are not automatically reduced by averaging. Time-varying biases from trends in the fundamental observations caused by changes in instrumental practices are also a source of error; effects of global mean trends of wind speed and ΔT seem to be in this error category and were removed from the flux estimates (C1; C2).

The marine data sample density is relatively high in the extratropics of the North Atlantic and North Pacific, with many 5° squares having 100 or more observations per month. On the other hand, the density of observations is marginal to poor over the tropics and in the Southern Hemisphere, with many 5° squares having less than 20 observations per month, so that in these regions random errors in the monthly flux estimates are likely to be a serious problem.

Averaging of the individual observations removes a significant amount of the random error, however, so
the uncertainty of monthly mean for well-sampled 5° squares (50 or more independent samples per month) is reduced to approximately 30 W m⁻² for the latent flux and to less than W m⁻² for the sensible flux. Errors from time-varying biases in the fundamental variables are difficult to quantify, but the marine data have global trends in w and ΔT, which translate to changes of approximately 0.8 m s⁻¹ and -0.2°C from 1950 to 1986. Because these seemed likely to be at least partially artificial due to instrumental and calibration changes over the period, the fluxes were adjusted to remove the effects of the global average linear trends. This adjustment reduced the linear change in the latent and sensible fluxes from 1950 to 1986 by about 20 and 5 W m⁻², respectively (C1; C2). It is important to note that the analyses shown here are little affected by this adjustment. This is because the anomalous fluxes and SST tendencies have standard deviations substantially larger than these changes (C2) and because the phenomenon of interest has much of its energy at relatively high frequencies (e.g., greater than one cycle per year).

In many regions the aggregate of these errors considerably exceeds the 10 W m⁻² tolerance that has been targeted for climate studies (Taylor 1984). Nevertheless, it will be shown that there is still important monthly climate variability contained in the flux estimates. In reviewing individual maps of the fluxes, several months have anomalies with magnitudes of 50 W m⁻² or greater. By using primary anomaly modes of monthly SLP as an index of periods when the monthly atmospheric circulation has strong anomalies, it is evident that the flux anomalies are organized spatially (C3). In accord, it will be shown that the corresponding SST anomaly tendency patterns are consistent with these flux patterns.

4. Relationship between flux and SST tendency anomalies

a. Local correlations: Spatial and seasonal distribution

If the upper ocean is a constant-depth well-mixed slab that exchanges heat with the atmosphere, the SST is predicted by a highly simplified thermodynamic equation. Removing the time mean, and noting that the temperature anomaly of the slab is SST', the equation governing the temperature is:

\[
\frac{\delta \text{SST}'}{\delta t} = -\frac{F'}{\rho \beta C_p}.
\] (3)

In this one-dimensional description of the mixed-layer temperature, horizontal temperature advection and vertical and horizontal mixing have been omitted, so that only the effect of the anomalous heating (F') on the temperature anomaly is considered. Also, the radiative fluxes are ignored and (F') is taken to be the latent plus sensible flux anomalies. A full account of the mixed-layer thermodynamic equation is given by Frankignoul (1985). Note that this simplified temperature equation predicts the tendency of the SST anomaly (\(\frac{\delta \text{SST}'}{\delta t}\)), not the anomaly itself. Here \(\frac{\delta \text{SST}'}{\delta t}\) has a higher-frequency character than the anomaly itself, since the time derivative acts as a high-pass filter. Accordingly, there are nearly as many independent samples of \(\frac{\delta \text{SST}'}{\delta t}\) as there are months, so that four decades of record contain approximately 120 independent monthly samples within the December–February winter period. The anomaly tendency is approximated using a discrete rate of change of centered differences of the monthly anomalies; that is, \(\Delta \text{SST}'/\Delta t\), where \(\Delta t\) is two months. This approximation introduces an error from SST variability of the SST of the two neighboring months (before and after). A more accurate formulation requires shorter-period observations to form the difference of SST at the very end of the central month and that at the very beginning of the month.

The relationship between \(\text{F}_{\text{lat}}\) and \(\Delta \text{SST}'/\Delta t\) was first tested using basin-scale measures by correlating zonal-mean anomalies of the flux and the SST tendency over each of the Pacific, Atlantic, and Indian oceans. Regions where SST is governed by (3) will be indicated by negative correlations between \(\text{F}_{\text{lat}}\) and \(\Delta \text{SST}'/\Delta t\). To determine the relationships between these zonal-mean anomalies, correlations were calculated for each 5° latitude strip and for each month. Significant correlations have magnitudes greater than 0.30, based on the following test. There are 41 possible data for each of the 12 months in the 1946–86 dataset; neither \(\text{F}_{\text{lat}}\) or \(\Delta \text{SST}'/\Delta t\) is significantly autocorrelated at a time lag of 1 year, and for 41 samples a correlation of 0.30 is significantly different from zero at about the 95% level of confidence.

A large number of the latitude–month correlations between the zonal-mean quantities are statistically significant (correlations tabulated in Fig. 1) and they are consistent with the \(\text{F}_{\text{lat}}\), forcing SST scenario. Nearly all of the significant correlations are negative, indicating that increased sea-to-air flux leads to decreasing temperature, which is the result expected from the simplified SST model in (3). Most of the latitudes have strongest correlations in winter. These have magnitudes exceeding 0.30 between about 10°–15°N and 55°N in both basins, and many with magnitudes exceeding 0.5. In the Pacific significant correlations are most prevalent north of about 20°N, with highest correlations in the cool months between October and March. Significant correlations also appear in the tropics north and south of the equator and as far south as 30°S. In the North Atlantic from about 10°N to about 55°N, strongest correlations also occur during the cool season, but unlike the Pacific they are also significant in summer and weakest in fall. Significant negative correlations also occur in the South Atlantic at about 20°–25°S during most months. In the northern Indian Ocean, which only extends northward to 20°N, highest correlations
occur between November and March from 5° to 20°N. In the southern Indian Ocean, the strongest negative correlations occur between April and December from 10° to 30°S. Beyond the equatorial belt, this analysis suggests that the flux anomalies act to drive the SST anomalies, especially during cool-season months.

Near the equator between 5°N and 5°S in all three basins, \( F_{\text{eq}} \) and \( \Delta \text{SST} / \Delta t \) are not well correlated. Correlations between zonal average flux and SST anomalies indicate that in some regions of the tropics, the flux and SST anomalies (not the tendencies) tend to be positively correlated. This suggests that the flux is driven by SST (C1), or in other words, that equatorial SST anomalies are more strongly affected by internal ocean processes rather than by the heat flux.

Next, we examined the geographical distribution of the correlations between \( F_{\text{eq}} \) and \( \Delta \text{SST} / \Delta t \) for winters (Fig. 2). Again the 0.3 level is used as a threshold of statistical significance. Meaningful correlations on this map are almost everywhere negative, as they were for the zonal means (Fig. 1). Stronger correlations (\( \leq -0.5 \)) are found mostly between about 25° and 40°N within the anticyclonic subtropical gyres of the two oceans. In the North Atlantic, there are also relatively strong correlations along the eastern boundary between 25° and 60°N. For the North Pacific, strongest correlations appear east of 180° and extend to the California Current, but not in the western part of the basin where the flux variability is greatest. In the central North Atlantic, the flux and \( \Delta \text{SST} / \Delta t \) anomalies have strong correlations in the central subtropical gyre as well as in the high variance western North Atlantic region.

b. Spatial covariability

While the analyses in section 4a reveal the local connection between the flux and \( \Delta \text{SST} / \Delta t \) anomalies, it remains to determine the spatial patterns over which heat flux anomalies force the upper ocean. If the heat-flux forcing of SST anomalies is an important mechanism, there should be large-scale connections between the two fields, since the SST anomalies and their tendencies have large-scale patterns (Wallace et al. 1990). To extract the primary spatial patterns associated with the flux versus SST anomaly tendency connection, we used canonical correlation analysis, a technique that extracts patterns of two data fields having the highest correlations. Formulation and meteorological applications of this technique are described by Graham et al. (1987) and by Barnett and Preisendorfer (1987), but an outline of this method is provided in the Appendix.

Briefly, this analysis produces pairs of spatial patterns, denoted \( g_k \) and \( h_k \), representing the \( F_{\text{eq}} \) and \( \Delta \text{SST} / \Delta t \) members of a canonical correlation mode \( k \). These can be thought of as predictors and predictands in a linear prediction scheme. Associated with
these are the time series of each, denoted $u_k$ and $v_k$, respectively. The canonical correlation for mode $k$ is the square of the correlation between $u_k$ and $v_k$. The canonical modes are ordered by the magnitude of the canonical correlations. To limit the number of predictors entering the covariance of the two datasets, the two sets were first filtered, in this case by employing a truncated number of their leading winter-month EOFs. This truncation is 8 and 12 for $F_{t+k}$ and $\Delta SST^*/\Delta t$, respectively. Pertinent statistics describing these EOFs and the results of the canonical correlations are listed in Tables A1 and A2 in the Appendix.

Note that the canonical analysis only operates on the cross-correlation structure of the EOF time amplitudes, maximizing temporal correlations between the primary modes of the two fields; there is nothing built into the technique to provide for spatial correlation between the corresponding $g$ and $h$ maps. The result of this particular application, however, is that the canonical pattern pairs have remarkably coincident spatial features ($g_k$ vs $h_k$). For both oceans the analysis yields pairs of patterns that feature regions of decreasing/increasing SST anomalies mated with regions of positive/negative flux anomalies. This coincidence is evident in Figs. 3 and 4, where the canonical $F_{t+k}$, $g$ maps are superimposed on their $\Delta SST^*/\Delta t$, $h$ map counterparts for the North Atlantic and North Pacific, respectively. This spatial coincidence appears in each of the first five modes of both basins, although only the first three modes are illustrated here.

From the canonical correlation maps and from the cross correlations listed in Tables A1 and A2, three important features stand out: 1) the canonical (squared) correlations are relatively high, ranging from about 0.6 to 0.3 for the first five canonical pairs; 2) the canonical modes are quite strongly correlated with the leading EOF modes of the $F_{t+k}$ and $\Delta SST^*/\Delta t$ fields, confirming the important link between the flux and the SST tendency anomalies; and 3) the first two canonical patterns of $F_{t+k}$ strongly resemble the flux anomaly pattern associated with primary modes of the atmospheric circulation in the two basins. Major atmospheric circulation patterns in the North Atlantic and North Pacific can be inferred from the first two EOF modes of SLP anomalies in the two basins shown in Fig. 5. Concerning their link to the atmospheric circulation, in the North Atlantic, the first two canonical $F_{t+k}$ modes are best related to the North Atlantic oscillation (NAO) and the eastern Atlantic (EA) SLP EOFs, respectively, with correlations of 0.58 and -0.66. In the North Pacific, the first two canonical flux modes are best related to the Pacific/North American pattern (PNA) and the Bering Sea (BER) SLP EOFs, respectively (discussion of these modes is provided in the next section). Correlations between the first two canonical $F_{t+k}$ modes and the PNA and BER EOFs from 1950 to 1986 are -0.63 and 0.55, respectively. Relationships between the monthly SLP modes and the flux anomalies are discussed in C3, but are illustrated in the next section. It can also be seen from the following section that the associated $\Delta SST^*/\Delta t$ canonical patterns are likewise well related to the first two SLP EOFs, an indication that the large-scale flux–SST tendency connection is driven by the atmospheric circulation.

A note about the statistical reliability of the canonical correlations is in order, since this analysis may have extracted fortuitously correlated portions of the two fields. At the same time, we wish to estimate the true amount of $\Delta SST^*/\Delta t$ variability accounted for by the
FIG. 3. Spatial patterns for leading three North Atlantic canonical correlation modes: g maps (solid lines), representing $F_L$, patterns, vs $h$ maps (shaded), representing $\Delta SST'/\Delta t$ patterns. Contours and shading show relative amplitude of the $g$ and $h$ maps, based on winter months of 1950–86. Positive/negative values of $g$ maps are indicated by $+/-$ signs; positive/negative values of $h$ maps are indicated by stippling/hatching.
Fig. 4. Spatial patterns for the leading three North Pacific canonical correlation modes: $g$ maps (solid lines), representing $F_{ij}$, patterns, vs $h$ maps (shaded), representing $\Delta SST/\Delta t$ patterns. Contours and shading show relative amplitude of the $g$ and $h$ maps, based on winter months of 1950-86. Positive/negative values of $g$ maps are indicated by $+/-$ signs; positive/negative values of $h$ maps are indicated by stippling/hatching.
Fig. 5. First two EOFs of monthly SLP anomalies in the North Atlantic (NAO and EA) and in the North Pacific (PNNA and BER). Computed separately for each basin, from all months of 1950–86.
$F'_{tx}$ forcing. To this end, a cross-validation sampling scheme (Barnett and Preisendorfer 1987) was implemented to derive from subsets of the data a succession of canonical models. The variance of the $\Delta$SST'/$\Delta t$ data explained, denoted the skill, by the model-predicted values was calculated from the dataset withheld from the model development. Specifically, 37 different canonical models were derived to successively verify the three winter months of each year within the 1950–86 period. In deriving the model for a given year, data for that year and for its two adjacent years were withheld from the training set; that year’s data was reserved for the verification (skill calculation) of the model. The year before and the year after were removed to avoid contributions that might arise from persistence. However, there is little autocorrelation between adjacent months of the flux (C1) or the SST tendency anomalies, so this data is amenable to cross validation (Barnett and Preisendorfer 1987) and the number of independent cases entering the validation approaches the total number of months, 111.

Following the procedure outlined earlier, each model yielded five canonical modes from 8 $F'_{tx}$ EOFs and 12 $\Delta$SST'/$\Delta t$ EOFs. Using each model, three $\Delta$SST'/$\Delta t$ maps were predicted for the winter months omitted from the training set. These were calculated from the December, January, and February $F'_{tx}$'s for that year. The skill, or the variance of the observed $\Delta$SST'/$\Delta t$ explained by the $F'_{tx}$ predictors, is shown in Fig. 6 for the North Atlantic (above) and North Pacific (below). Skill is calculated as

$$1 - \frac{\sum_{i=1}^{111} (\text{predicted} - \text{observed})^2}{\text{predicted}^2}.$$  

The skill is nearly everywhere positive for both oceans with the only negative skills along the boundaries and at the lowest latitudes of the basins. When validated

---

**FIG. 6.** Skill (% of observed $\Delta$SST'/$\Delta t$ variance explained) of cross-validated canonical correlation model of $F'_{tx}$ vs $\Delta$SST'/$\Delta t$.  

---
against the original observed $\Delta SST'/\Delta t$ data (shown in Fig. 6), the average model skill was 0.10 over the North Atlantic and 0.13 over the North Pacific. Alternatively, when validated against the filtered (12 EOFs) $\Delta SST'/\Delta t$ data on which the model was based, there was an increase in the average skill to 0.15 and 0.21, respectively (maps not shown). For both calculations, skill was greatest in the subtropical gyre regions of the two oceans with maxima of about 40%, in agreement with local correlations between $F_{i,\pi}$ and $\Delta SST'/\Delta t$ (Fig. 2).

c. $\Delta SST'/\Delta t$ versus $F_{i,\pi}$, during strong atmospheric circulation modes

The strong connection between the anomalous atmospheric circulation and the flux anomalies (C3) provides a test of the consistency of patterns of the flux and SST tendency anomalies. If the SST tendency anomalies are caused by the fluxes, they should have corresponding patterns. This exercise compliments the canonical correlation analysis because it provides a physically based comparison of the two fields, in terms of the atmospheric circulation and the fundamental surface marine variables (C3). The mean monthly patterns of surface weather variables, including the wind speed, $\Delta q$, and $\Delta T$, are systematically arranged around the highs and lows of large-scale circulation features, so the latent and sensible fluxes tend to be generated in the form of spatially coherent anomalies. The North Atlantic and North Pacific exhibit similar configurations of flux anomalies with respect to the major high and low SLP anomalies (C3). In the present study, this organization was exploited by using the dominant SLP EOF modes as an index to compare flux anomaly and corresponding SST anomaly tendency patterns under cases of strong atmospheric circulation.

A brief digression on the major atmospheric circulation patterns is now presented. In winter, SLP anomalies are strongly linked to the upper-level pressure field, and hence to storm tracks, especially over the oceans. The inferred flow (nearly geostrophic) about these SLP patterns and attendant synoptic weather features are helpful in diagnosing anomaly patterns in pertinent surface variables. A substantial fraction of anomalous SLP in the North Atlantic and North Pacific is represented by a few EOFs (Davis 1976). The first two EOFs of North Atlantic SLP and their North Pacific counterparts are illustrated in Fig. 5. Together, these two EOFs account for 51% and 55% of the monthly SLP anomaly variance in the respective basins. Since there is ample separation between the fraction of variance explained by the first three modes, these EOFs were not rotated as in other studies, such as Barnston and Livezey (1987), who studied rotated EOFs of the 700-mb height field over the entire Northern Hemisphere. North Atlantic SLP EOF 1 is a north–south dipole structure (van Loon and Rogers 1978), which is well known as the North Atlantic oscillation (NAO), and is one of the most common modes of Northern Hemisphere atmospheric variability (Barnston and Livezey 1987). North Atlantic SLP EOF 2 is a large-scale feature centered in middle latitudes of the eastern North Atlantic and has been identified as the eastern Atlantic (EA) pattern (Barnston and Livezey 1987). The first two SLP EOFs in the North Pacific have spatial patterns similar to those in the North Atlantic. North Pacific EOF 1 resembles North Atlantic SLP EOF 2 (the EA pattern). North Pacific EOF 1 features a large pressure anomaly in the central North Pacific and is associated with the well-known Pacific–North American (PNA) pattern. On the other hand, North Pacific EOF 2 has a north–south anomaly dipole structure that is similar to North Atlantic EOF 1 (NAO). North Pacific EOF 2 resembles the Bering Sea (BER) pattern extracted from a longer monthly SLP dataset by Rogers (1990).

Returning to our central theme, we wish to determine the linkage between the flux anomalies and $\Delta SST'/\Delta t$ under cases of strong anomalous atmospheric circulation. To accomplish this, winter months with extreme positive and negative EOF amplitudes were identified for each SLP EOF (C1; C3). Using this criterion, between 10 and 30 months of each of the extreme EOF modes (positive and negative amplitudes) were chosen to examine the linkage to $\Delta SST'/\Delta t$. Composites of $F_{i,\pi}$ and $\Delta SST'/\Delta t$ were formed by averaging the fields during the respective extreme months for each of the two basins. For brevity, the composites were expressed as the difference between averages of positive (strong) versus negative (weak) phase months of the SLP EOFs.

Composite $F_{i,\pi}$ differences associated with positive versus negative extremes of the first two North Atlantic SLP EOFs are shown in Fig. 7, and the corresponding composite $\Delta SST'/\Delta t$ differences are shown in Fig. 8. Differences that are significantly different from zero in excess of the 95% level of confidence in Student's t-test are indicated by the shading in Figs. 7 and 8. For both SLP EOFs, the $\Delta SST'/\Delta t$ pattern is well matched to that for $F_{i,\pi}$. As expected, SST anomalies fall (negative anomaly tendencies) in regions of positive sea-to-air flow and rise (positive tendencies) in regions of negative flux. In accord with the NAO circulation pattern, the North Atlantic is partitioned into alternating anomalous cells: 1) positive $\Delta SST'/\Delta t$ differences in the northeastern North Atlantic between Great Britain and Scandinavia; 2) negative $\Delta SST'/\Delta t$ differences in the central North Atlantic southeast of Greenland between 50°N and 60°N; 3) a broad region of positive $\Delta SST'/\Delta t$ differences extends from the Gulf of Mexico and the eastern seaboard east to the central Atlantic at about 35°N, 30°W; 4) negative $\Delta SST'/\Delta t$ differences occupy the eastern tropical North Atlantic between about 50°W and North Africa. The major regions of significance of $\Delta SST'/\Delta t$ are closely matched to those
of $F_{\Delta SST}$, supporting the view that they are causally linked. For North Atlantic SLP EOF 2 (EA), the major $\Delta SST' / \Delta t$ features are 1) negative differences in the central North Atlantic east of Newfoundland; 2) positive differences in the Gulf of Mexico and the region just east of Florida; and 3) positive differences in the far northeastern Atlantic and the Mediterranean Sea. For both North Atlantic SLP EOFs, the anomaly difference centers exceed 0.2°C mo$^{-1}$ with maxima of approximately 0.5°C mo$^{-1}$. Corresponding $F_{\Delta SST}$ composite difference centers exceed 50 W m$^{-2}$, with maxima of about 200 W m$^{-2}$. The first two SLP EOF patterns strongly resemble the first two canonical correlation modes in Fig. 3.

In the North Pacific, patterns of $\Delta SST' / \Delta t$ are also well aligned with the flux anomaly patterns and consistent with the canonical correlation patterns. Composite $\Delta SST' / \Delta t$ differences and $F_{\Delta SST}$ differences corresponding to positive versus negative extremes of the PNA pattern are shown in Fig. 9 (left and right). The
PNA marks out a great spatial extent of atmospheric circulation anomalies more than halfway around the earth, and as was the case for the North Atlantic circulation modes, the PNA composites exhibit a close match between $F_{i+s}$ and $\Delta SST'/\Delta t$ features. Magnitudes of the North Pacific $\Delta SST'/\Delta t$ differences are about the same as those for the North Atlantic, with values of 0.2°C mo$^{-1}$ to 0.8°C mo$^{-1}$. Over the North Pacific, PNA incorporates a swath of positive $F_{i+s}$ differences across midlatitudes (centered at 40°N) from Asia to 140°W, and an arc of negative $F_{i+s}$ differences extending from the subtropics at 20°N, 160°E to the eastern North Pacific border, from California to the Gulf of Alaska. Downstream, the strong downstream arm of the PNA is indicated. Negative $F_{i+s}$ differences occupy the Gulf of Mexico and the eastern seaboard of the United States, and positive anomalies are found in the subtropical North Atlantic centered at (20°N, 50°W), and also in the region east of Newfoundland. The magnitude of the $F_{i+s}$ and $\Delta SST'/\Delta t$ differences...
in the western and central North Atlantic exceed 100 W m⁻² and 0.6°C mo⁻¹, respectively. These anomalies are actually larger than those upstream over the North Pacific. Throughout the North Pacific and North Atlantic, the PNA-associated ΔSST'/Δt differences mirror the flux differences, having collocated centers of opposite sign as the flux differences.

Results for North Pacific SLP EOF 2 (BER) are not shown, but reveal similar correspondence between the flux anomalies and ΔSST'/Δt (C1). For this mode, however, the major composite anomaly features are confined to the North Pacific.

d. Fluxes versus vertical mixing

Although the ΔSST'/Δt anomalies are well matched to the heat flux anomalies, it is possible that they are actually caused by another process, which is correlated with the heat flux. Mechanical mixing by the wind is a candidate for such a process, because it affects the ocean surface temperature and because wind speed enters the fluxes through the bulk formulas. Thus, the ΔSST'/Δt might simply be a signature of wind mixing—as w increases, the temperature drops. Vertical turbulent mixing is often parameterized as the cube of the wind speed (Haney et al. 1981). COADS does not contain monthly averages of w³, and here we use w as a surrogate. To justify this representation, an estimate of monthly mean w³, denoted 〈w³〉, was constructed from the monthly mean wind speed using an assumed Weibull distribution of wind speeds for each month. Mean Weibull parameters for ocean surface wind speeds from Paviá and O’Brien (1986) were used to calculate 〈w³〉. In relating wind speed parameters to the SST tendency anomalies, the analyses shown below were performed twice: using w and also using 〈w³〉. Since the results using w were very similar to those using 〈w³〉, we present those using w.

Composite w' field differences for strong versus weak extremes of the first SLP EOFs in the North Atlantic and North Pacific are shown in Fig. 10. These composites are based on the same extreme winter months as were the F'₁₀ and ΔSST'/Δt composites above. Although some of their major features are similar to those of the F'₁₀ (Figs. 7 and 9), there are significant differences. The strongest w'are centered north and south of the SLP anomaly centers, while the major F'₁₀ are oriented zonally as well as meridionally relative to the SLP anomaly centers. Significant w anomaly differences for the NAO spread across the entire North Atlantic basin (top of Fig. 10). The ΔSST'/Δt composite pattern in Fig. 8 appears to be more coincident with the F'₁₀ than the w' pattern. This spatial relationship is also evident in the North Pacific, where the wide basin permits strong expression of the anomalies up- and downstream from the SLP centers over midocean. For the PNA mode, there are three regions with strong F'₁₀ and ΔSST'/Δt evident in Fig. 9, but w' values in these regions are weak (Fig. 10, bottom). These three regions are off the coast of Asia, off the west coast of the United States, and in the Gulf of Mexico.

There is support for the above patterns' qualitative evidence that the flux dominates the wind speed in producing SST tendency anomalies. This is evidence provided by a multiple regression analysis predicting ΔSST'/Δt from w' and F'₁₀ over winter months from...
the datasets. To determine the relative importance of these two predictors, they were entered as normalized anomalies:

\[
\frac{\Delta\text{SST}'}{\Delta t} = a_F\hat{F}' + a_w\hat{w}'
\]  

(4)

where \(\Delta\text{SST}'/\Delta t\) is the estimated \(\Delta\text{SST}'/\Delta t\), and \(\hat{F}'\), and \(\hat{w}'\) are the \(F_{F+}\) and \(w\) monthly anomalies normalized by their monthly standard deviations, respectively. For both the North Atlantic and North Pacific, this analysis indicates that the flux anomaly is a more important predictor of the SST anomaly tendency than the wind speed. The regression coefficients \(a_F\) and \(a_w\) are mapped in Fig. 11 over the North Atlantic. Where coefficients are large, both predictors have negative signs: positive \(F_{F+}\) and \(w\) produce negative \(\Delta\text{SST}'/\Delta t\). However, in most locations, the magnitude of \(a_F\) is greater than that of \(a_w\), indicating that \(F_{F+}\) is a stronger predictor of \(\Delta\text{SST}'/\Delta t\) than is \(w\). This result prevails throughout the tropics and the extratropics. Results for the North Pacific (not shown) were similar, in that nearly all predictors were negative, and the \(F_{F+}\) predictors dominated the \(w\) predictors.

e. Inferred mixed-layer depth

If, as in (3), we assume that the anomalous surface heat flux is transmitted from an isothermal slab of depth \(h\) at the top of the ocean, then the tendency in SST anomalies measures the change of heat within this mixed layer. From our "observations" of monthly \(F_{F+}\) and \(\Delta\text{SST}'/\Delta t\), we can estimate \(h\) statistically from a regression equation. This was accomplished by mul-
tiplying (3) through by $F'$ and computing the covariance $\langle \langle \rangle \rangle$ over the available samples:

$$h = \frac{\langle F'^2 \rangle}{\rho c_p \langle F' \Delta SST' / \Delta t \rangle}.$$ (5)

Since we are testing the effects of latent and sensible flux, the heat flux anomaly $F'$ is taken to be $F'_{r,s}$. As an alternative to (5), $h$ can also be estimated by correlating each side of (3) with $\Delta SST' / \Delta t$ and solving for $h$; however, this results in $\langle \Delta SST'^2 / \Delta t \rangle$ in the denominator. Since $\Delta SST' / \Delta t$ is the result of several other processes (besides the heat flux), we use (5) because both its numerator and denominator are associated with the flux.

Since the mixed-layer depth varies seasonally, $h$ was estimated separately for each of the 12 months using the series of data for each month. Here $h$ was calculated for each month when the connection between the flux and the SST anomaly tendency was judged to be significant, namely, when the correlation between area averages of $F'$ and $\Delta SST / \Delta t$ was less than $-0.30$. This connection is weak in several of the warm-season months, precluding a reliable estimate of $h$. For subtropical (20°–35°N) and middle-latitude (40°–55°N) midocean regions of the North Atlantic and North Pacific, area-average monthly estimates of $h$ and their standard deviations are graphed in Fig. 12.

Observations (not shown) of the mixed-layer depth from hydrographic observations mapped over the North Atlantic by Lamb (1984) and over the North Pacific by M. Robinson (unpublished charts) exhibit maximum values of 100–250 m in late winter and minimum values of 10–40 m in summer. Although the estimated $h$ in Fig. 12 are somewhat high, they have the correct order of magnitude and exhibit a realistic annual variation, with maxima in late winter and minima in summer. This method appears to provide mixed-layer depth estimates that are as good as another method using the phase between the annual cycle of SST and that of net solar heating (van den Dool et al. 1988). It is emphasized that the present calculations of $h$ were not intended to accurately determine the mixed-layer depth, but to confirm the validity of the flux anomaly–upper-ocean connection. Since reasonable magnitudes were obtained for $h$, it is concluded that the mixed-layer heat content anomalies are consistent with forcing by the flux anomalies.

5. Summary and conclusions

The effect of monthly anomalies of latent and sensible heat fluxes on the SST anomaly field is detectable over broad regions of the extratropical ocean surface. The focus in this study is on behavior during winter when the latent and sensible fluxes and their variance are large compared to radiative components of the surface heat budget and when much of the seasonal heat accumulated in the upper ocean is released to the atmosphere. The flux anomalies, derived from bulk formulations, exhibit large-scale patterns of variability, which are related to patterns of sea level pressure (SLP) variability and also to patterns of SST anomaly tendencies ($\Delta SST' / \Delta t$). Importantly, the heat-flux estimates are virtually independent of the SLP and $\Delta SST' / \Delta t$ observations. The consistency of the relationships between flux anomalies and 1) large SLP patterns and 2) the $\Delta SST / \Delta t$ field over both the North Atlantic and North Pacific is evidence of a realistic short-period climate variability in the bulk formula sea–air heat exchange estimates.

We considered a simple thermodynamic system in which the upper-ocean heat balance is driven by the latent and sensible heat flux; for example, positive flux anomalies produce negative SST anomaly changes. The sum of the latent and sensible flux anomalies ($F'_{r,s}$) was used in these analyses since the two components are well correlated in the extratropics. The signature of this process, negative correlations between the $F'_{r,s}$ and $\Delta SST' / \Delta t$, emerges nearly everywhere in the extratropics. Negative correlations are strongest during winter but they are detectable during all seasons. Exceptions occur in the tropics and in part of the extratropics during the warm season, where the flux anomalies tend to be in phase with the SST anomalies, suggesting that the flux anomalies are driven by the ocean thermal field rather than vice versa. The amount of $\Delta SST' / \Delta t$ variance accounted for by the flux anomalies during winter months ranges from 10% to 40% over much of the ocean. However, because both the flux and the SST tendency estimates contain errors, it is likely that this skill underestimates the strength of the real heat-flux forcing process.

Previous studies have demonstrated the influence of heat flux on SST anomalies in local or regional settings, but the broad-scale organization of the flux and $\Delta SST' / \Delta t$ anomalies has not been appreciated. Two separate analyses, canonical correlation analysis and composites according to atmospheric circulation anomaly modes, demonstrate that $F'_{r,s}$ and $\Delta SST' / \Delta t$ co-vary with patterns that span the ocean basins. Although the variance of the SST anomaly tendencies accounted for by the flux anomalies is modest, there is a remarkable consistency in the patterns of the two fields. Positive sea-to-air $F'_{r,s}$ and negative $\Delta SST' / \Delta t$ features coincide, as do the opposite sign features of each. Canonical correlation patterns illustrate the spatial configuration of this connection, as organized by the atmospheric circulation in the respective basins. Typical of 50 W m$^{-2}$ are associated with monthly changes in SST anomalies of order 0.2°C mo$^{-1}$. The structure of the patterns of these fields, their interrelationships, and the magnitude of their anomaly centers are very similar in the North Atlantic and North Pacific. Moreover, the patterns of the flux-related anomaly tendencies resemble primary patterns of SST anomalies that have been exposed by several other investigators in previous studies.
An alternative explanation for the large-scale changes in SST anomalies associated with the SLP modes might be that they are forced by wind mixing. This mechanism was tested by relating $\Delta\text{SST}'/\Delta t$ anomalies to a parameterization of mixing, represented by anomalies of the monthly mean wind speed ($w$). Although mixing probably does play a role, composites indicate that patterns of $\Delta\text{SST}'/\Delta t$ anomalies are better matched to those of the flux anomalies than to those of the $w$ anomalies. The impression that $F'_{\text{mix}}$ have greater impact on the surface temperature than does $w$ is confirmed by a multiple linear regression analysis. This analysis indicates that the $\Delta\text{SST}'/\Delta t$ variance explained by the flux anomalies dominates that explained by the scalar wind. However, it is emphasized that $w$ may not be an adequate representative of the wind mixing.

A crucial check on the consistency between the flux anomalies and the tendencies of the SST anomalies is whether the inferred depth of the upper-ocean heat-exchanging layer is consistent with observations. Again assuming a simple thermodynamic response in an isothermal upper ocean, the layer depth can be estimated using the magnitude of the flux anomalies and the associated $\Delta\text{SST}'/\Delta t$. A regression between $F'_{\text{mix}}$ and $\Delta\text{SST}'/\Delta t$ yields depths between 50 and 300 m, depending on month and location. Although this idealized mixed layer has rather large uncertainty, it has the same magnitude as observations and exhibits a realistic annual cycle that shoals in summer and deepens in winter.

While the monthly flux anomalies account for part of the $\Delta\text{SST}'/\Delta t$ variance, it is clear that other processes...
must also have an important influence on the SST anomalies. A more complete test of the connection of the fluxes and other mechanisms to the upper-ocean heat content will be possible when historical synoptic bathythermograph measurements become available. Preliminary comparisons relating the flux anomalies to variations in the near surface heat content from several years of ship-of-opportunity XBT data in the central North Pacific show fairly strong quantitative agreement (X. Yan and P. Niiler, personal communication).

Acknowledgments. This work was part of my Ph.D. dissertation, which was guided by Russ Davis, who provided valuable advice and high standards. Jerome Namias was instrumental in providing support and encouragement. Thanks are also due to the COADS marine dataset consortium, especially Scott Woodruff who supplied auxiliary information. Thanks to Emelia Bainto for data processing, Peggy Whan for word processing, and Marguerette Schultz for drafting. Discussions with Tim Barnett, Art Miller, Nick Graham, Peter Niiler, and John Roads were very useful. Comments by Mike Wallace and an anonymous reviewer were particularly helpful in improving this manuscript.

APPENDIX

Canonical Correlations

This analysis employs filtered versions of \( F_{t+s} \) and \( \Delta \text{SST}'/\Delta t \) for the winter months of 1950–86. The input data consisted of their leading EOFs (designated here as \( F \) and \( S \), respectively):

\[
F(x, t) = \sum_{i=1}^{M} \alpha_i(t) a_i(x) \quad (A1)
\]

\[
S(x, t) = \sum_{j=1}^{N} \beta_j(t) b_j(x). \quad (A2)
\]

The EOF series are truncated at \( M = 8 \) and \( N = 12 \) modes. This truncation produces a reasonable representation of the large-scale content of the two fields, capturing 68% and 73% of the variance in the North Atlantic and 63% and 60% of the variance in the North Pacific for the \( F_{t+s} \) and \( \Delta \text{SST}'/\Delta t \) fields, respectively. The EOF filtering tends to remove shorter space-scale variability of the two fields, some of which is probably error, as discussed previously. The structure of the \( F_{t+s} \) and \( \Delta \text{SST}'/\Delta t \) EOFs is shown in C1.

The derived canonical coefficients, \( u_k(t) \) and \( v_k(t) \), are a weighted combination of the \( \alpha \) and \( \beta \), respectively:

\[
u_m = \sum_{i=1}^{M} \alpha_i(r_{im}) \quad v_n = \sum_{j=1}^{N} \beta_j(s_{jn}), \quad (A3)
\]

where the \( r \) and \( s \) are eigenvectors of the product of the cross-covariance matrix (of the \( \alpha \) and \( \beta \)) and its transpose:

\[
[\text{CC}^T] r_m = \mu^2_m r_m \quad (A4)
\]

\[
[\text{CT}C] s_n = \mu^2_n s_n. \quad (A5)
\]

The new paired variables are constructed so that their time correlations \( \langle u_m v_n \rangle \) are maximum when \( m = n \) and are 0 otherwise. The \( u \)'s are orthonormal, as are the \( v \)'s, that is, \( \langle u_i u_j \rangle = \delta_{ij} \). The eigenvalues of this system \( \mu^2_k \) are called canonical correlations and represent the squared correlations between the \( u_k \) and \( v_k \)

The new variables have a spatial expression in the form of pairs of canonical maps, which Graham et al. (1987) call “\( g \)” and “\( h \)” maps.

\[
g_k(x) = \langle F(x, t) u_k(t) \rangle \quad (A6)
\]

\[
h_k(x) = \langle T(x, t) v_k(t) \rangle \quad (A7)
\]

In terms of the simplified thermodynamic system in (3), the predictor maps \( g_k \) represent the forcing by the latent and sensible flux anomalies, and the predicand maps \( h_k \) represent the response of the \( \Delta \text{SST}'/\Delta t \). In general, the \( g_k \) and \( h_k \) are not orthogonal. For this analysis, five pairs of canonical modes were derived. Pertinent statistics from this analysis (cross correlations between the original EOF modes, the variance explained, and the canonical correlations) are reported in Tables A1 and A2 for the North Atlantic and North Pacific, respectively.
### Table A1. North Atlantic canonical correlation statistics $F'_{\text{tie}}$ versus $\Delta\text{SST}/\Delta t$ for DJF 1950–86.

#### $F'_{\text{tie}}$ EOF cumulative variance explained

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<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
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#### $\Delta\text{SST}/\Delta t$ EOF cumulative variance explained

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<td>2.1</td>
<td>1.9</td>
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### Cross correlations of $F'_{\text{tie}}$ EOFs modes 1–8 versus $\Delta\text{SST}/\Delta t$ EOF modes 1–12

#### $\Delta\text{SST}/\Delta t$

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#### Canonical correlations, modes 1–5

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### Cross correlations of $g$ coefficients modes 1–5 versus $F'_{\text{tie}}$ EOF modes 1–8

#### $F'_{\text{tie}}$ EOF

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### Cross correlation of $h$ coefficients modes 1–5 versus $\Delta\text{SST}/\Delta t$ EOF modes 1–12

#### $\Delta\text{SST}/\Delta t$ EOF

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<th>8</th>
<th>9</th>
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<td>-0.23</td>
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### Table A2. North Pacific canonical correlation statistics $F_{*1}$ versus $\Delta SST/\Delta t$ for DJF 1950–86.

#### $F_{*1}$, EOF cumulative variance explained

<table>
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<th>3</th>
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<th>6</th>
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<th>8</th>
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#### $\Delta SST/\Delta t$, EOF cumulative variance explained

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<th>4</th>
<th>5</th>
<th>6</th>
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<th>10</th>
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<th>12</th>
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#### Cross correlations of $F_{*1}$, EOF modes 1–8 versus $\Delta SST/\Delta t$, EOF modes 1–12

#### $\Delta SST/\Delta t$, EOF

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**Canonical correlations, modes 1–5**

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#### Cross correlations of $g$ coefficients modes 1–5 versus $F_{*1}$, EOF modes 1–8

<table>
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<th>7</th>
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**Cross correlations of $h$ coefficients modes 1–5 versus $\Delta SST/\Delta t$, EOF modes 1–12

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**References**


Davis, R., 1976: Predictability of sea surface temperature and sea
level pressure anomalies over the North Pacific Ocean. *J. Phys. Oceanogr.*, 6, 249–266.


