VARIABILITY OF NORTH PACIFIC OCEAN SURFACE SENSIBLE
AND LATENT HEAT FLUXES

By
Roy A. S. Hourston
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Atmospheric Science
The University of British Columbia
2075 Wesbrook Place
Vancouver, Canada
V6T 1W5

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Abstract

Interannual variability in North Pacific Ocean (20°N – 56°N) winter-time surface sensible and latent heat fluxes and their relation to Northern Hemisphere sea-level pressure and 500 mb height fields is explored. Observational data based on the Comprehensive Ocean-Atmosphere Data Set (COADS) and National Meteorological Center (NMC) gridded data set over the 1950–89 period are analysed and compared with 10 years of simulated data from the Canadian Climate Centre (CCC) second generation general circulation model (GCM). Two regions of the North Pacific Ocean are examined closely: the Kuroshio Current region (KCR), where the mean and anomalies of the heat fluxes are largest (implying a strong link between the ocean and atmosphere); and a region in the central North Pacific (NPR) near 31°N 165°W where the heat flux means and anomalies are moderately large and where several authors have found a strong correlation between sea surface temperatures and the Northern Hemisphere atmospheric circulation.

In the COADS/NMC data, positive KCR heat flux anomalies are associated with significant differences from the climatological mean atmospheric fields: there are stronger ridges over Eastern Asia and Western North America, and deeper troughs over the Northwest Pacific and Northwest Atlantic. These patterns are similar to previous modelling studies of the atmospheric response to positive heat flux and sea surface temperature anomalies, and to a dominant mode of variability observed in the 500 mb height field known as the Western Pacific (WP) pattern. The corresponding GCM relations are quite different. Positive KCR heat flux anomalies are associated with a deeper trough over the Northeast Pacific and Northwest Atlantic at sea-level pressure, and a deeper trough over the Northwest Pacific at 500 mb.
The atmospheric anomalies associated with NPR heat flux anomalies are generally weaker than those accompanying KCR heat flux anomalies. In the COADS/NMC data, positive NPR latent heat flux anomalies are associated with deeper troughs over the Northeast Pacific and Eastern North America, stronger jets, a stronger ridge over Western North America, and a more intense Aleutian Low. However, positive NPR sensible heat flux anomalies are associated with weaker atmospheric anomalies in the opposite sense: a shallower trough over Eastern North America, a weaker North American Jet, a weaker ridge over Western North America, and a less intense Aleutian Low. The corresponding GCM relations are somewhat similar, with positive NPR heat flux anomalies accompanied by a deeper trough over the Northeast Pacific and a stronger ridge and weaker trough over North America.

VARIMAX rotated principal component analysis shows the two dominant modes of variation of COADS and GCM sensible and latent heat fluxes are strongly associated with variability over KCR and NPR, respectively. The greater complexity of the GCM principal components may reflect the smaller signal-to-noise ratio in the model data and the limitations of the mixed-layer ocean model. It also suggests spatial autocorrelation within the COADS and GCM heat flux fields is very different. This difference may have important implications for a more complete assessment of the model’s simulation of both the observed atmospheric fields ("current" climate) and the anticipated changes in atmospheric circulation due to anthropogenic emissions of greenhouse gases.
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Chapter 1

Introduction

1.1 Introduction

It is recognised that short-term (one season to 2-3 years) climate variability represents the net response to a variety of internal and external forcing mechanisms in the global climate system, of which air-sea interaction is an important component. This is because our weather and climate are driven by solar energy absorbed at the earth’s surface, where four times as much is absorbed by the oceans compared to land (Woods 1984). Energy absorbed by the oceans is either stored, increasing the ocean temperature; advected by ocean currents; or released into the atmosphere via sensible and latent heat fluxes and longwave radiation. It is the interaction between the ocean and the atmosphere on a large scale, with spatial dimensions of several hundred kilometres or more and time scales of a month to decades, that dominates climate variability.

This is illustrated in Figure 1.1 which shows that over the spectrum of climate variability, most variance between one month and 10 years occurs on spatial scales on the order of 1000 km. Over this frequency range, climate variability depends on slow thermal adjustments in the atmosphere (atmospheric thermal relaxation), and even slower thermal adjustments in the oceanic mixed layer (oceanic mixed-layer thermal relaxation). Atmospheric thermal relaxation refers to the thermal state of the atmosphere which is constantly perturbed by shorter time-scale processes. After each perturbation the atmosphere relaxes, partly through comparatively slow internal radiative adjustments, and
partly through thermodynamic interactions with the more slowly responding oceanic mixed layer. Oceanic mixed-layer thermal relaxation is similar, but in this case the oceanic mixed layer is perturbed by shorter time-scale atmospheric processes and interacts with the more slowly responding deeper ocean layers. The atmospheric thermal state depends on thermodynamic interactions with the ocean surface which in turn tend to generate convection within the oceans. This motion results in a transfer of heat between the ocean surface and deeper layers which may later affect the amount of heat at the ocean surface available to the atmosphere on time scales characteristic of overturning of the oceanic mixed layer (on the order of one to 10 years). Analysis of large-scale air-sea heat transfer (the link between the thermal state of the atmosphere and oceanic mixed layer)
may provide insight into the dynamics of interannual and decadal climate fluctuations, an especially important concern regarding fisheries (Mysak 1986), agriculture (Liverman 1986), and society in general (Chen et al. 1983, Kellogg and Schware 1981).

Energy is transferred from ocean to atmosphere by sensible and latent heat fluxes and longwave radiation. Upward surface sensible heat flux warms the atmospheric boundary layer (ABL) locally by conduction through adjacent sea and air molecules and this heat is then carried into the rest of the ABL by turbulent transfer. Upward surface latent heat flux heats the air when water vapour condenses (and only then warms the atmosphere irreversibly when the water is precipitated back to the ocean). This warming occurs locally in the ABL or long distances away and higher in the atmosphere. The upward longwave radiative heat flux is absorbed by clouds or water vapour directly above the ocean surface in the ABL, or higher up in the free atmosphere, or escapes directly to space.

Although it is small-scale turbulent processes that govern details of air-sea heat transfer, it is large-scale phenomena that ultimately drive the temporally and spatially averaged exchanges of heat between ocean and atmosphere. Where the flow of heat between ocean and atmosphere is strong, we would expect a relationship between the surface heat flux and both the temperature of the ABL and the structure of the surface and upper air pressure fields. When the flow of heat is weak, we would expect a much weaker relation between the ocean and atmosphere, and the surface heat flux fields and atmospheric pressure and temperature fields to vary essentially independently of each other. The temporal and spatial patterns of variability of the individual fields can be assessed and then compared with each other in a search for similarity and possible linkages. It is through an analysis of the spatial and temporal variability of air-sea fluxes and upper air fields that we hope to better understand how the atmosphere and ocean are linked, how they influence one another, and how their interaction affects climate variability.
Chapter 1. Introduction

Variability in the surface heat flux at the air-sea interface has been examined previously by many investigators using both observations and models. These approaches have usually been based on analyses of sea surface temperatures which serve as an index of the air-sea heat exchange (see Frankignoul (1985) for a review). Sea surface temperatures in different regions can be strongly interrelated, and this has added to the difficulty of clearly establishing the oceanic locations that contribute most to climate predictability. Bjerknes (1969) was one of the first to emphasize the important influence of the tropical oceans on interannual climate variability on a global scale. He suggested that in the tropics, fluctuations in the inputs of latent and sensible heat at the air-sea interface modify convection in the equatorial regions, thus influencing the north-south Hadley circulation cell, and therefore mid-latitude climate.

The effect of mid-latitude oceans on the atmosphere and climate has proven more difficult to ascertain. Namias (1959) was first to observe correlations between sea surface temperatures in the North Pacific Ocean and the overlying atmospheric circulation in the troposphere. More advanced statistical analyses have shown that mid-latitude North Pacific Ocean sea surface temperatures in some regions have weak predictive skill associated with North American air temperature changes, comparable with the predictive skill associated with tropical Pacific sea surface temperatures (Barnett 1981, Harnack 1982). The basis of this predictive skill, i.e. the link between sea surface temperatures and the atmosphere, is not as clearly understood for mid-latitude regions as for the tropics. This is perhaps because oceanic forcing in Northern Hemisphere mid-latitude regions is not as strong: the oceanic heat source is not as great in total area or magnitude per unit area compared to the tropics. It is not surprising then, that the modelling results of Frankignoul and Molin (1988), and Webster (1981, 1982), and reviews by Barnett and Sommerville (1983) and Namias and Cayan (1981) indicate atmospheric anomalies associated with mid-latitude sea surface temperature anomalies are weak at best, making
them very difficult to detect.

The sea surface temperature, however, does not fully represent the thermal forcing of the ocean on the atmosphere, which is determined by the sum of the sensible and latent heat fluxes and the longwave radiative flux. Over the North Pacific Ocean the sensible and latent heat fluxes constitute 65 to 85% of the total annual ocean-to-atmosphere heat transfer (Zhao and McBean 1986). Namias (1976) has suggested that variability in the latent heat flux strongly influences the intensity and position of the Aleutian Low. The Aleutian Low covers an extensive region in the Gulf of Alaska where low pressures dominate over monthly or longer averaging periods during the cold season as a result of cyclones intensifying as they move eastward from the east coast of Japan. Since the weather and climate of Western North America is strongly influenced by these cyclones, it is worthwhile examining the variability of the North Pacific Ocean surface heat fluxes and associated Northern Hemispheric circulation patterns with a view to better understanding their relationship with the Aleutian Low and regional climate variability.

In the North Pacific Ocean, two areas have been identified as having particularly strong connections with the overlying atmosphere and will be examined in detail: the Kuroshio Current region; and the 10° latitude by 10° longitude region centered on 31°N 165°W. The Kuroshio Current and Kuroshio Extension consist of a warm ocean surface current flowing eastward from Japan. This is a region of high ocean-to-atmosphere heat transfer, especially in winter when air-sea temperature and humidity differences can be great as a result of cold dry air flowing eastward off the continent of Asia and over the Pacific Ocean. Zhao and McBean (1986) have suggested that heat transfer variability in this region may be linked to the strength and position of the Aleutian Low. In contrast, the region centered on 30°N 165°W is not an area of especially high heat transfer, but was identified by Namias and Born (1972), Rasmusson and Carpenter (1982), Wallace and Jiang (1987), and Lau and Nath (1990) as having strong links with the tropical Pacific
Chapter 1. Introduction

Ocean heat source and Northern Hemisphere circulation.

Estimates of the surface heat fluxes are important for modelling studies where they can be used both to drive oceanic and atmospheric circulation models which require such inputs, and as diagnostics for models which predict them. For example, using the climatology of Esbensen and Kushnir (1981), Gates et al. (1985) determined that the Oregon State University general circulation model surface sensible and latent heat fluxes over the mid- and high-latitude oceans during winter were systematically over-estimated. Estimates of the surface heat fluxes based on observations (such as those by Esbensen and Kushnir) provide the only method with which to gauge model output, but known errors in the observed oceanic surface sensible and latent heat fluxes alone are on the order of 10% (Esbensen and Reynolds 1981). This error corresponds to 30 W m⁻² for a typical winter-time latent heat flux over the Northwest Pacific Ocean. It is expected that this error will be reduced in the near future since a goal of the World Ocean Circulation Experiment (WOCE) is to produce global estimates of air-sea heat fluxes having a systematic error of less than 10 W m⁻² (WOCE 1986).

In this research, the relation between surface heat fluxes and the overlying atmosphere will be explored. Variability in observed data will be examined, with special attention paid to the Kuroshio Current region and the area centered on 31°N 165°W. The results will be compared with analyses of the same fields generated by the Canadian Climate Centre (CCC) second generation general circulation model (GCM).

Although complex in its representation of the atmosphere, the CCC GCM’s oceanic component is highly simplified, consisting of a 50 metre thick slab of sea water. There are no currents but heat is transferred horizontally within this ocean surface layer and vertically through the top and bottom of the layer. Observed sea surface temperature climatology is used to specify horizontal heat advection within the ocean surface layer and the vertical heat flux through the bottom of the layer, but the sea surface temperature
used to compute vertical heat flux at the air-sea interface is determined by the model. Given these constraints it is nonetheless worthwhile determining how well the model reproduces variability in observed surface and upper air fields (and their correlations).

The rest of this chapter deals with research objectives which are presented in Section 1.2, methodology in Section 1.3, and a thesis outline in Section 1.4.

1.2 Research Objectives

The goal of this research is to determine variability of the North Pacific Ocean winter-time surface sensible and latent heat fluxes and their relation to Northern Hemisphere atmospheric circulation as represented by sea-level pressures and 500 mb heights. Only the winter season will be considered since this is when the ocean-to-atmosphere heat fluxes are greatest (implying a strong link between the ocean and atmosphere), and the sensible and latent heat fluxes together constitute most of the total ocean-to-atmosphere heat flux. Close scrutiny of the sensible and latent heat fluxes over the Kuroshio Current region and the area centered on 31°N 165°W will allow verification of their supposed strong relations to the Northern Hemisphere atmospheric circulation. Comparison of observations to output from the Canadian Climate Centre second generation general circulation model will allow verification of the model’s ability to simulate observed fields.

1.3 Methodology

Observations of surface and near-surface variables will be used to estimate winter-time (average of December, January, and February) North Pacific Ocean surface sensible and latent heat fluxes using bulk formulae. The long-term means and anomalies of the sensible and latent heat fluxes will be examined in conjunction with the fundamental variables used to compute them. They will be compared with the corresponding fields
generated by the Canadian Climate Centre second generation general circulation model. The most energetic modes of the “observed” and modelled surface sensible and latent heat flux variability will be identified using principal component analysis. The degree of correlation between the North Pacific Ocean heat flux anomalies/principal components and Northern Hemisphere sea-level pressure and 500 mb height anomalies will allow quantitative assessment of the statistical and possible physical links between the “observed” and modelled North Pacific Ocean and Northern Hemisphere atmosphere.

1.4 Thesis Outline

In Chapter 2, theory behind air-sea heat transfer is presented. The advantages and disadvantages of the bulk method for estimating the turbulent transfers of sensible and latent heat on climatological scales are discussed in detail. The problem of identifying forcing and response between the atmosphere and ocean is also addressed, and previous observational and modelling studies on the North Pacific Ocean are reviewed.

The observational and model data are described in Chapter 3. The heat flux computation schemes are presented and errors and trends in the observed data are described.

Observational and model data are analysed in Chapter 4. The long-term mean and annual anomalies of the North Pacific Ocean sensible and latent heat fluxes and fundamental variables are presented. The spatial and temporal variability of the sensible and latent heat fluxes is assessed using principal component analysis. Correlations between Northern Hemisphere sea-level pressure and 500 mb height anomalies and both the anomalies and principal components of the sensible and latent heat fluxes are presented. Similarities and differences between observed and model fields are discussed.

A discussion and conclusions follow in Chapter 5. Appendix A deals with the difficult problem of distinguishing between real and artificial trends in the observational data.
Chapter 2

Air-Sea Heat Transfer

2.1 Introduction

Heat is continuously exchanged between the atmosphere and the ocean in both directions through the sea surface in the form of sensible heat, latent heat, and longwave radiation. In mid-latitude oceans the flow of heat is almost always upwards from the ocean to the atmosphere (Oberhuber 1988, Isemer and Hasse 1987). Sensible heat is absorbed in the very lowest layers of the atmosphere (as this heat transfer process is just heat conduction through adjacent air and sea molecules) but is then carried into the rest of the atmospheric boundary layer by turbulent transfer. The latent heat from evaporation is carried upwards in water vapour and may be released directly overhead or may be carried many thousands of kilometres away and released when the water vapour condenses. Longwave radiation is transferred directly upwards and is absorbed by clouds or water vapour if at all.

The direct effect of ocean-to-atmosphere heat transfer on the atmosphere is difficult to determine. It may result in a warming and (assuming no horizontal divergence) a vertical expansion of the air. This would modify the upper air pressure gradient and flow aloft, and thus the position and amplitude of the quasi-stationary long wave train downstream. This modification of flow aloft may alter storm tracks, influencing surface flow and climate downstream of ocean-to-atmosphere heat transfer. The modified surface flow may then directly and indirectly affect air-sea sensible and latent heat fluxes: directly
as it relates to scalar wind speed, and temperature and humidity of the air as will be
described in Section 2.2 and Equations 2.3 and 2.4; and indirectly through its effect on
the evolution of sea surface temperatures as will be described in Section 2.4 and Equation
2.11.

This atmosphere-ocean feedback is in reality much more complicated, which makes it
difficult to determine whether the atmosphere is driving the ocean or *vice versa*. This
problem may never be definitively resolved.

The structure of this chapter is as follows: the bulk method for estimating turbu-
lent surface sensible and latent heat fluxes is described in Section 2.2; the applicabil-
ity of the bulk method for estimating climatological surface sensible and latent heat
fluxes is discussed in Section 2.3; the problem of identifying forcing and response in the
atmosphere-ocean climate system is addressed in general terms in Section 2.4; and air-sea
heat transfer over the North Pacific Ocean is discussed in Section 2.5.

### 2.2 The Bulk Formulae For Estimating The Surface Heat Fluxes

The turbulent Reynolds' fluxes of sensible and latent heat are:

\[
Q_H = \rho_a C_p \overline{w'\theta'} 
\]

\[
Q_E = \rho_a L_E \overline{w'q'} 
\]

where \( \overline{w'\theta'} \) and \( \overline{w'q'} \) are kinematic fluxes of temperature and water vapour; \( \rho_a \) is air
density; \( C_p \) is the specific heat at constant pressure of dry air; and \( L_E \) is the latent heat
of evaporation of water.

The bulk aerodynamic formulae are applicable for estimating heat fluxes over short
periods of time (about 30 minutes to 1 hour), and parameterize the kinematic fluxes in
terms of more easily measured mean (or bulk) quantities:
where $U(z)$, $T_a(z)$, and $q_a(z)$ are wind speed, air temperature, and specific humidity at a reference height $z$ which is within the boundary layer and usually taken to be 10 m. The variables $T_s$ and $q_s$ are sea surface temperature and saturation specific humidity of air at temperature $T_s$ and at sea-level pressure. The values for all variables are means over a period less than one hour. The variable $\rho_a$ is the average air density in the layer from the surface to height $z$. The non-dimensional transfer coefficients $C_H$ and $C_E$ (Stanton Number and Dalton Number) are functions of wind speed, stability, and $z$. Large errors in $C_H$ and $C_E$ (and hence $Q_{HB}$ and $Q_{EB}$) arise out of uncertainties in their estimates at neutrality of up to 10 and 20% respectively (Large and Pond 1982).

By examining air-sea temperature and humidity differences and wind speed in conjunction with bulk estimates of heat fluxes the most important factors that result in large (and small) amounts of heat transfer can be identified. For example, the bulk aerodynamic approach defines latent heat transfer to be dependent on scalar wind speed and on specific humidity of air and saturation specific humidity at sea surface temperature (the humidity difference). If latent heat transfer is high and humidity difference is low, it is high wind speed which results in high heat transfer. It is this isolation of factors responsible for high and low heat transfer that is an important step towards identifying physical processes that determine the interannual variability of air-sea heat transfer.
2.3 Applicability of Bulk Formulae For Estimating Climatological Surface Heat Fluxes

The bulk approach is the only practical means of estimating air-sea heat exchange on a routine basis over large areas and a several decade time history. Climatological sensible and latent heat fluxes are typically defined in terms of monthly averages using two different methods. The sampling method involves computing sample means of flux estimates from simultaneous observations of $U(z)$, $T_s$, $T_a(z)$, $q_s$, and $q_a(z)$, and using a transfer coefficient based on $U(z)$ and stability (determined from $T_s$, $T_a(z)$, $q_s$, and $q_a(z)$), and $\rho_a$ determined from pressure and temperature. The values for each variable usually represent measurements averaged over a period of less than one hour. These estimates of the heat fluxes are then averaged over a month to give the monthly mean. Alternatively, the classical method involves estimating the monthly mean heat fluxes from monthly means of $U(z)$, $T_s$, $T_a(z)$, $q_s$, $q_a(z)$, the transfer coefficient, and $\rho_a$ determined from monthly mean pressure and temperature. The differences between these approaches are shown below:

$$\frac{Q_{HB}}{\rho_a C_p} = \overline{CU\Delta T}$$  \hspace{1cm} (2.5)

$$= \overline{C U \Delta T} + \overline{C U' \Delta T'} + \overline{U C' \Delta T'} + \overline{\Delta T C' U'} + \overline{C' U' \Delta T'}$$  \hspace{1cm} (2.6)

$$\frac{Q_{EB}}{\rho_a L_E} = \overline{CU\Delta q}$$  \hspace{1cm} (2.7)

$$= \overline{C U \Delta q} + \overline{C U' \Delta q'} + \overline{U C' \Delta q'} + \overline{\Delta q C' U'} + \overline{C' U' \Delta q'}$$  \hspace{1cm} (2.8)

where we have assumed $C_H = C_E$ and dropped the subscripts for clarity. Also $\Delta T = T_s - T_a(z)$ and $\Delta q = q_s - q_a(z)$. The overbar denotes the mean over one month and the prime denotes a deviation from the monthly mean. The variables $L_E$, $C_p$, and $\rho_a$ have been moved to the left hand sides of these equations since their variations play a
negligible role in the surface flux estimates (Esbensen and Reynolds 1981). The right
hand sides of Equations 2.5 and 2.7 are the estimates of the monthly mean of the triple
products obtained by the sampling method, while the classical method is represented by
the first term on the right hand sides of Equations 2.6 and 2.8. The difference between the
two methods is due to the three covariances and the triple covariance in both Equations
2.6 and 2.8. It is obviously easier to calculate fluxes with the classical method but the
procedure is only acceptable if the temporal covariances are small or almost cancel. This
will depend on the variance of the wind speed and temperature difference, the correlation
of the wind and the temperature, and the functional form of the transfer coefficients. For
example, if constant transfer coefficients are used, only the first two terms on the right
hand side of Equations 2.6 and 2.8 will contribute, the latter three will be zero.

When the classical approach is used, the monthly mean of the transfer coefficient is
seldom computed as this would require all original data to determine. Instead, a new
transfer coefficient, $C^*$, is used which is computed from the monthly mean wind and
monthly mean stability. The functional difference between the two transfer coefficients
is:

$$
\bar{C} = F(U, T_s, T_a, q_s, q_a) \quad (2.9)
$$

$$
C^* = F(\bar{U}, T_s, \bar{T}_a, \bar{q}_s, \bar{q}_a) \quad (2.10)
$$

Due to the nonlinearity of their dependence on wind speed and stability, $C^*$ will typically
be smaller than $\bar{C}$, resulting in an underestimation of monthly mean surface heat fluxes
(Hanawa and Toba 1985; Esbensen and Reynolds 1981; Fissel, Pond, and Miyake 1977).
This error can be large since it grows with averaging time (significantly past two days
for sensible heat flux over the Northeast Pacific Ocean according to Fissel, Pond, and
Miyake (1977)), and grows most rapidly when the instantaneous flux magnitudes are
large (Marsden and Pond 1983). The need for a transfer coefficient greater than $C^*$ in
using the classical approach arises because the bulk Equations (2.3 and 2.4) and typical transfer coefficient formulations apply to short-term (30 minutes to 1 hour) turbulent transfers of heat when the covariances in Equations 2.6 and 2.8 are insignificant. Over the longer term (one month) these covariances can be significant, on the order of 10% of the total surface sensible and latent heat flux in the North Pacific Ocean (Esbensen and Reynolds 1981).

So far we have only discussed problems in the bulk equations (2.3 and 2.4) due to averaging in time, but similar arguments apply for averaging in space. Typically climatological data are presented as representative of an area, rather than values at a point. If we look at Equations 2.6 and 2.8 and regard the overbar as a spatial averaging operator, by looking only at the area averaged fluxes one ignores the contributions from the spatial covariances. Mahrt (1987) has shown that during the day over a mid-latitude land surface with uneven topography, these spatial covariances in the fluxes can be significant compared to the spatial averages and do not cancel, but these errors are essentially eliminated by formulating the transfer coefficient in terms of area averaged variables (like Equation 2.10 again regarding the overbar as a spatial averaging operator). This reduction of the spatial covariances and virtual elimination of the error occurred with systematic sampling of points within an area (i.e., Mahrt used regularly spaced grid-ded data). However, marine surface observations have positive sampling biases towards shipping lanes and negative sampling biases with regard to ocean storms, so it should be investigated whether spatial covariances of these data are also significant and non-zero. This may be particularly important in regions of large gradients of the surface parameters.

Differences between fluxes computed using the sampling and classical approaches range from 2% (Budyko 1974), 4% (Reed 1985), 10% (Adamec and Elsberry 1984, Kraus and Morrison 1966), 23% (Robinson 1966), and in excess of 50% over the sparsely sampled
Southern Hemisphere oceans (Simmonds and Dix 1989). It should be emphasized that flux data used in these comparisons were not direct measurements but based on various assumed functional forms of the transfer coefficients and using either micro-meteorological or climatological data to determine them. These studies were also conducted over a range of time periods, seasons, and locations around the world. As a result of such diversity it is perhaps not surprising that such a range of differences was found, when one considers the conceivable variation with latitude, season, proximity to coasts, etc., of the covariances in Equations 2.6 and 2.8.

In the North Pacific Ocean, the temporal and spatial covariances of the surface heat fluxes appear to be ≤ 10% of the total in both observations (Hanawa and Toba 1985; Marsden and Pond 1983; Esbensen and Reynolds 1981; Fissel, Pond, and Miyake 1977) and models (Simmonds and Dix 1989). Thus the errors in the classical approach appear tolerable in the North Pacific. In fact, Esbensen and Reynolds (1981) suggest that the classical method may provide a more reliable estimate of heat flux when the number of observations is small. They reason that because the covariance and the correlations of $U$ with $\Delta T$ and $\Delta q$ are known to be relatively small, the non-linearities in the flux estimates using the sampling method may lead to more statistical uncertainty in estimating the long-term average heat fluxes.

It appears the use of the bulk formulae (Equations 2.3 and 2.4) and classical approach is adequate for estimating the air-sea heat fluxes. Crucial for determining the heat fluxes is the specification of the form of the transfer coefficients. Although these coefficients may be determined in a completely empirical manner using regression analysis, Monin-Obukhov similarity theory for the boundary layer allows $C_H$ and $C_E$ to be specified using a rational, physically motivated approach and minimizing their dependence on empiricism. Esbensen and Reynolds (1981) suggest that the use of different “physically reasonable” transfer coefficient schemes should result in no significant variation in the flux
estimates. However, Blanc (1985) found a large range in the functional dependence in a review of ten different empirical and theoretical transfer coefficient schemes using one year of data from Ocean Weather Ship C in the North Atlantic Ocean. He found a maximum scheme-to-scheme variation of 70% for an average sensible heat flux of 25 W m\(^{-2}\) and 45% for an average latent heat flux of 40 W m\(^{-2}\). He suggests that much of this variation would be reduced by reducing the averaging time of the measurements and using the more difficult to obtain but more reliable direct measurements of the fluxes provided by eddy correlation instruments.

At wind speeds less than 10 m s\(^{-1}\) both \(C_H\) and \(C_E\) are strongly dependent on both wind speed and temperature differences. However, at wind speeds greater than 10 m s\(^{-1}\) the boundary layer approaches neutrality and the transfer coefficients' dependence on \(U\) and \(\Delta T\) is weak and \(C_H\) and \(C_E\) approach their neutral values. An examination of Figure 2.1 (and Table 2.1) shows that at neutral or slightly unstable conditions, while most schemes show the transfer coefficients to be constant or increasing with wind speed, those of Smith (1980)/ Friehe and Schmitt (1976), Liu et al. (1979), and Kondo (1975) show a decrease. This may be due to these schemes emphasizing the decreased instability at higher wind speeds.

2.4 Identifying Forcing and Response in the Atmosphere-Ocean Climate System

In this research we are concerned only with the sensible and latent surface heat fluxes and their anomalies, which are defined as monthly or seasonal deviations from their long term means. As shown in Equations 2.3 and 2.4 these fluxes depend on the atmospheric variables \(T_a\), \(q_a\), and \(U\) which have a time scale of a few days. The heat fluxes also depend on the oceanic variables \(T_s\) and \(q_s\) (which is a function of \(T_s\)). Since the sea
Figure 2.1: Sensible and latent heat coefficients ($C_H$ and $C_E$) for ten selected schemes under neutral or slightly unstable conditions as a function of the wind speed ($U(z)$) at an altitude of 10 m. Scheme acronyms are given in Table 2.1. Source: Blanc (1985).
surface temperature extends throughout the oceanic mixed layer (which has a large heat capacity and thermal inertia), it is rather persistent and its anomalies ($T'_s$) develop on a time scale of a month to a season or longer. Thus the heat fluxes and their anomalies develop on a time scale between that of the atmospheric and the oceanic anomalies, *i.e.*, somewhere between a few days and a season.

Once $T'_s$'s develop, they influence the magnitude of the local heat transfer between the atmosphere and the ocean (Equations 2.3 and 2.4). This in turn induces changes in the local atmospheric circulation (as described in Section 2.1), which can influence the local surface heat fluxes and sea surface temperature. Furthermore, due to the large-scale circulation of the atmosphere, influences on the local atmosphere due to local sea surface temperature anomalies may be transmitted long distances, and then have an impact on the atmospheric circulation, surface heat fluxes, and sea surface temperature very far from the initial sea surface temperature anomaly. Due to the constant feedback between ocean and atmosphere, it is of considerable difficulty to assign specific cause and effect (or forcing and response) in air-sea heat transfer.
This feedback between atmosphere and ocean on seasonal time scales is illustrated by considering the equation governing the time rate of change of $T'_s$ (Frankignoul 1985):

$$\frac{dT'_s}{dt} = - \frac{Q'}{\rho_s C_s h} - \frac{(hv)' \cdot \nabla(T_s + T'_s)}{\frac{h}{\bar{h}}} - \frac{\bar{h} \partial \bar{T}_s}{\bar{h} \partial t}$$

$$- \frac{(T'_s - T'_{s+}) (\bar{w}_e + w'_e)}{\frac{h}{\bar{h}}} - \frac{(T_s - T_{s+}) w'_e}{\frac{h}{\bar{h}}} + K \nabla^2 T'$$

(2.11)

where the terms on the right hand side describe the effect of the following processes:

A anomalies in the net surface heat flux;
B temperature advection by anomalous current;
C anomalies in the mixed-layer depth;
D anomalies in the temperature jump in the entrainment zone;
E anomalies in the entrainment velocity;
F horizontal mixing.

In this equation the seasonally varying mean is denoted by an overbar and the seasonal anomaly is denoted by a prime, the subscript plus indicates values at the bottom of the entrainment zone; the time derivative following the mean motion (or lagrangian derivative) is $\frac{d}{dt} = \frac{\partial}{\partial t} + \nabla \cdot \nabla$ where the horizontal velocity is $\mathbf{v} = (u, v)$ and the horizontal gradient is $\nabla = \left(\frac{\partial}{\partial x}, \frac{\partial}{\partial y}\right)$; $\rho_s$ and $C_s$ are the density and specific heat of sea water; $h$ is the mixed layer depth; the entrainment velocity is $w_e = \Lambda \left(\frac{\partial h}{\partial t} + \nabla \cdot (hv)\right)$ where $\Lambda = 1$ if $\frac{\partial h}{\partial t} + \nabla \cdot (hv) > 0$ and 0 otherwise; and $K$ is a horizontal diffusion coefficient to parameterize horizontal mixing. $Q$ is the sum of the sensible heat flux $Q_H$, latent heat flux $Q_E$, shortwave radiation $Q_{SW}$, and longwave radiation $Q_{LW}$. The variability of each heat flux component depends on location and time of year. $Q'_H$ and $Q'_E$ vary.
Chapter 2. Air-Sea Heat Transfer

the most in fall and winter, and $Q'_{SW}$ varies most in spring and summer (Frankignoul 1985), however, it is the ratio $Q'/h$ that determines the effectiveness of $T'_s$ forcing. This equation is important because it shows atmospheric forcing through $Q$ on the time rate of change of the seasonal sea surface temperature anomalies (or damping) by the heat flux.

This is in contrast to Equations 2.3 and 2.4 which show atmospheric forcing (through $T_a$, $q_a$, and $U$) and oceanic forcing (through $T_1$) on the seasonal surface heat flux anomalies. In mid-latitudes in winter, $T_s$ is almost always greater than $T_a$, thus by Equations 2.3 and 2.4 turbulent heat transfer is always from the ocean to the atmosphere. This transfer decreases the heat content in the mixed layer and thus reduces $T_s$. Therefore the feedback due to the turbulent heat fluxes is always negative.

The strength of the feedback depends on the degree of adjustment of $T_a$ to a change in $T_s$, which in turn is determined by the thermodynamics of the atmospheric boundary layer and on the dynamical responses of the troposphere to vertical transports of heat. With no adjustment in $T_a$ the change in the sensible and latent heat flux forcing induced by $T'_s$ has a decay time of $\sim 1.8$ months (Frankignoul 1985). With air temperature adjustment the negative feedback should be weaker and result in a longer decay time.

The forcing of surface heat fluxes by $T'_s$ may also cause feedback in the radiation fluxes. Lau and Nath (1990) and Namias (1976) have argued that a positive $T'_s$ results in increased transfers of sensible and latent heat which would increase cloudiness and possibly cyclogenesis, and that a negative $T'_s$ would have the opposite effects. Since shortwave radiation is generally the dominant component of $Q$, $T'_s$ has a negative feedback effect on $Q$. Cause and effect in this case has proven difficult to separate clearly, and Frankignoul (1985) reports of no known robust observations of the relation between cloudiness and $T'_s$’s in the middle latitudes. Correlations between observed December-January-February averages of North Pacific ($20^\circ$N – $56^\circ$N) $T'_s$ and total cloud cover from
the COADS data set (Chapter 3) over the 1950–89 period are not significant at the 95% level.

Due to the differing response times and persistence of the atmosphere (represented by $U'$ and $T'_s$), the ocean ($T'_s$), and the surface heat flux, it may be possible to discern a degree of cause and effect in the atmosphere-ocean climate system. When dealing with monthly mean data, the atmospheric response to $T'_s$ should be simultaneous, and exhibit highest values for simultaneous correlations. Wallace and Jiang (1987) examined correlations between $T'_s$ near 32°N 165°W and the index of the PNA pattern (a linear combination of the normalized geopotential height anomalies at four centers: near Hawaii; over the central North Pacific Ocean; over central Alberta; and over the Gulf Coast region of the U.S.), and found that due to the persistence of $T'_s$, lagged correlations were strongest with the atmosphere apparently leading the $T'_s$. Because $T'_s$ are slow to change, and if they are forcing the atmosphere, the geopotential height anomalies will also be slow to change. Wallace and Jiang found $T'_s$ and the geopotential height anomalies to be strongly correlated over a long range of lags (depending on the time of year) of up to ten months, with the strongest correlation occurring when the atmosphere leads the ocean by one month.

In the case of atmospheric forcing of $T'_s$, the atmosphere is varying more or less independently of the ocean, so atmospheric circulation anomalies will be quick to change compared to $T'_s$. Due to the higher frequency fluctuations of the atmosphere (and neglecting feedback from the ocean and damping for the moment), we would expect that on average $T'_s$ would tend to be preceded by atmospheric circulation anomalies of one sign (which generated them) and followed by equally strong atmospheric circulation anomalies of the opposite sign (which would subsequently erase them). Wallace and Jiang examined correlations between $T'_s$ and sea-level pressure anomalies near 32°N 165°W and found that even with feedback and damping occurring (as is probably most often the
case), the two were strongly correlated over a short range of lags of only one month, with highest values for the atmosphere again leading the $T'_s$ by one month. Even stronger correlations were found between $T'_s$ and sea-level pressure anomaly tendencies.

Knowledge of the relation between these two distinct types of lag correlation signatures and forcing and response will be exploited in an investigation of cause and effect with respect to surface heat flux anomalies and atmospheric circulation anomalies. Since surface heat flux anomalies exhibit time scales of variation between those of the atmosphere ($U'$ and $T'_a$) and those of the ocean ($T'_s$), it is expected that the range of lags over which surface heat flux anomalies and atmospheric anomalies are strongly correlated will not be as great as those between $T'_s$ and atmospheric anomalies and that strongest correlations in three month averaged seasonal data will be simultaneous with no lag.

### 2.5 North Pacific Ocean Air-Sea Heat Transfer

Most analyses of North Pacific air-sea heat transfer have focussed on the relationship between $T'_s$ and the atmosphere instead of the heat fluxes and the atmosphere. Observational studies have mainly concentrated on establishing correlations between $T'_s$ and atmospheric fields such as pressure and temperature either locally or downstream over North America. Namias (1976) demonstrated via cross-correlations that $T'_s$ in the Aleutian area in summer are associated with specific climatic parameters the following fall: the position and intensity of the Aleutian Low and air temperature and precipitation over North America. Davis (1978) found that up to 20% of the variance of fall and winter sea-level pressure anomalies over the North Pacific could be predicted from $T'_s$ of the fall and previous summer, while there was no skill in other seasons. Barnett (1981) found tropical $T'_s$ to be most effective as predictors of North American temperature changes, although significant predictive skill was attained using mid-latitude $T'_s$. Douglas et al.
(1982) showed that decadal scale changes in the North Pacific $T_s$ and 700 mb height are related to variability at the surface and aloft over North America. Reviews by Namias and Cayan (1981) and Barnett and Somerville (1983) indicate that the $T_s'$s are effective predictors of short-term climate changes, and have significant forecast skill over certain regions, particularly during the cold season.

To understand the physical basis for the forecast skill associated with $T_s'$s, observational studies have examined the nature of $T_s'$s themselves, their temporal and spatial extent, and mechanisms resulting in their generation. Namias et al. (1988) found North Pacific $T_s'$s to persist for 3-5 months on average with sizable lag correlations out to 12 months for winter $T_s'$s when the mixed layer is deepest. Their results suggest that the occurrence of persistent and non-persistent $T_s'$s is not random, but is related to the initial monthly or seasonal $T_s$ pattern which in turn depends on the initial atmospheric flow patterns and their associated wind fields. They found that strongly persistent $T_s'$ patterns were associated with 700 mb height anomaly patterns that were also persistent whereas non-persistent $T_s'$ episodes were associated with radically changing 700 mb height monthly anomaly patterns. This suggests oceanic forcing of the atmosphere during persistent $T_s'$ cases as noted in Section 2.4.

In looking for the trigger of these persistent or non-persistent $T_s'$s it is valuable to examine the initial atmospheric flow patterns and their associated wind fields. Large et al. (1986) and Frankignoul (1985) have shown that wind driven entrainment and vertical mixing during discrete episodes of high wind speed that occur in association with storms are important forcing mechanisms of $T_s'$s at higher latitudes, and can result in rapid changes in $T_s$ on the order of 2 or 3 degrees in 2 to 3 days. Wallace et al. (1990) argue that the patterns of atmospheric circulation anomalies responsible for the persistent $T_s'$s should be the ones that most strongly modulate the surface wind speed, such as the Western Pacific (WP) pattern described by Wallace and Gutzler (1981). This is because
the WP pattern so closely resembles the climatological mean surface wind and sea-level pressure fields, and thus modulates the surface wind speed (and thus $T_s'$) over broad expanses of the westerlies and the tradewind belts.

Models using prescribed $T_s'$s have been employed to examine the atmospheric response to oceanic forcing. In a review of mid-latitude $T_s'$s and their effect on the atmosphere Frankignoul (1985) concluded that results from linear wave models and general circulation anomalies indicated that realistic mid-latitude $T_s'$s have a weak influence on climate fluctuations amounting to changes in the geopotential height on the order of 10 to 30 m at most. Using the Goddard Institute of Space Studies (GISS) GCM Frankignoul and Molin (1988) showed the atmospheric response to North Pacific $T_s'$s was weak but significant, and characterized by changes in the geopotential height at zonal wave numbers 3-5 (large scales comparable to the scales of the $T_s'$s) detectable in the middle and upper troposphere on the order of tens of metres, which is small compared to the observed variability of monthly means. Lau and Nath (1990) found that over the entire North Pacific, $T_s'$s near 31.5°N 161°W were associated with the most robust and well-defined atmospheric response. They show that strong $T_s'$s at this point are associated with the reorientation of winter storm tracks, resulting in considerable remote precipitation anomalies (in contrast to tropical regions where precipitation anomalies are co-located with $T_s'$s). They argue that the presence of a warm (cold) $T_s'$ in a baroclinic zone such as at the confluence of the warm Kuroshio and cold Oyashio currents east of Japan tends to shift the site of strongest zonal $T_s$ gradients poleward (equatorward). Since the synoptic disturbances tend to thrive in areas of intense baroclinicity, such shifts in the pattern of $T_s'$ gradients result in displacement of the storm tracks from their climatological positions. The shifting of the storm tracks is accompanied by changes in the spatial patterns of the sources and sinks of latent and sensible heat, which are in turn linked to the anomalous behavior of the seasonally averaged circulation. Furthermore, changes in the cloud cover
that occur in conjunction with storm track displacements could also lead to systematic alterations in the radiative forcing of the atmosphere.

While much effort has been devoted to determining the relation between $T'_s$ and the atmosphere, relatively little has been done examining the relation between surface heat fluxes and the atmosphere. This is primarily due to the unavailability of data and is unfortunate since surface heat fluxes are a direct measure of air-sea heat transfer, while $T'_s$ are an indirect indicator. Observations of the surface heat fluxes have indicated that much of the annual total heat transfer occurs in the winter when the air-sea temperature differences are greatest and wind speeds are generally high due to increased storm activity (Zhao and McBean 1986). It is expected that in terms of ocean-to-atmosphere heat transfer, air-sea coupling will be strongest in winter.

Modelling results of Weaver (1987) suggest that the atmospheric response to Kuroshio heating anomalies is strongest at low levels. The response is small and positive upstream of the Kuroshio (weak high pressure anomalies over Eastern Asia for positive heat flux anomalies over the Kuroshio), and large and negative (low pressure anomalies) downstream over the central North Pacific Ocean. The ridge-trough structure in turn forces Kuroshio heating anomalies and eastern Pacific cooling anomalies. The heating/cooling anomalies then feedback and intensify the low in the lower troposphere over the central North Pacific.

In this research two areas will be focussed on: the Kuroshio Current region; and the 10° latitude by 10° longitude region centered on 31°N 165°W. The Kuroshio Current and Kuroshio Extension together consist of a warm ocean surface current flowing eastward of Japan. It is a region of high ocean-to-atmosphere heat transfer, especially in winter when air-sea temperature and humidity differences can be great as a result of cold dry air flowing eastward off the continent of Asia and over the Pacific Ocean. Zhao and McBean (1986) have suggested that heat transfer variability in this region may be linked to the
strength and position of the Aleutian Low. While not an area of especially high heat transfer, the region centered on 31°N 165°W was identified by Namias and Born (1972), Rasmusson and Carpenter (1982), Wallace and Jiang (1987), and Lau and Nath (1990) as having strong links with the tropical Pacific Ocean heat source and Northern Hemisphere circulation.
Chapter 3

Data Description

3.1 Introduction

Observational data are from two sources. North Pacific Ocean surface sensible and latent heat fluxes are estimated from surface observations from the Comprehensive Ocean-Atmosphere Data Set, or COADS (Slutz et al. 1985), which has been culled from observations since 1854 made aboard merchant marine "ships of opportunity" plying the world oceans. "Observed" Northern Hemisphere sea-level pressures and 500 mb heights consist of observations that have been smoothed by model analyses and interpolated onto an octagonal grid of 1977 equally spaced points when viewed on a polar stereographic map, giving them a resolution of about 2.5° at 20°N and 10° at 60°N. These atmospheric fields were produced by the Department of Atmospheric Sciences, University of Washington, and the Data Support Section, National Center for Atmospheric Research, based on the grid point data of the US National Meteorological Center (NMC).

Model surface sensible and latent heat fluxes and sea-level pressures and 500 mb heights were generated by the Canadian Climate Centre (CCC) second generation general circulation model (GCM) described by McFarlane et al. (1991) which is based on an earlier version of the model (Boer et al. 1984). The new version of the model has 10 vertical levels and employs a triangular spectral truncation having 32 longitudinal waves (T32/L10). This gives horizontal resolution of 3.75° in the east-west direction and about 3.71° in the north-south direction. Important improvements in the new version of the
model include an interactive cloudiness parameterization, improved solar and terrestrial radiative heating calculations, a more sophisticated treatment of land surface processes, and a simple ocean mixed-layer model with a thermodynamic sea-ice component.

### 3.2 Errors And Trends In The COADS Data

The COADS data suffer from spatial and temporal errors, some of which are random and are greatly reduced by averaging, and some which are not random and represent biases in the data. Random errors include those that are associated with each stage of observation, recording, and data processing, many of which have been corrected or eliminated by "trimming" of the data (Woodruff et al. 1987). Random errors also include the bias imparted on the data due to different ship size and shape, such as the height at which $T_a$ and $U$ are measured. Many of these errors are reduced by averaging. The effect of random observation errors on the heat flux estimates was examined by Cayan (1992) who assumed the magnitude of observation error for each variable was the same everywhere. Using a Monte Carlo simulation of a stochastic normally distributed random error imposed on the long term monthly means he found errors in the North Atlantic Ocean $Q_H$ and $Q_E$ fluxes to be less than 5 and 10 W m$^{-2}$ respectively for grid boxes of 25 samples. Since over the North Pacific Ocean, the long-term average number of observations per grid box is greater than 25, and the long-term mean winter-time $Q_H$ and $Q_E$ differ by 10% or less from those in the North Atlantic (Oberhuber 1988), errors in the North Pacific heat fluxes due to observation error will be comparable.

Variability in the sampling density leads to a non-random error in time and space and its effect on the heat flux estimates depends on the temporal and spatial variability of the ocean and atmosphere which are functions of geographical location and season. Regions where variability in the ocean and atmosphere is high exhibit large errors. The spatial
variability of sampling between (and within) ocean grid boxes represents a positive bias towards shipping routes and a negative bias regarding ocean storms. (In regard to this investigation this latter bias is unfortunate since this is where and when large amounts of heat are transferred between the ocean and atmosphere.) These factors also influence the temporal variability of sampling as new shipping routes open and the newer, larger ships are less likely to avoid ocean storms.

On the basis of this problem, Taylor (1984a) indicates that the number of observations per 5° by 5° latitude-longitude grid square per month needed to provide “barely adequate” heat flux estimates is 20, and that 100 will give “reasonable” heat flux estimates in the absence of significant measurement error. This translates to about 13 and 64 observations respectively for smaller 4° by 4° latitude-longitude grid squares. Using these criteria, sampling coverage over the 1950–89 period in the North Pacific Ocean (20°N – 56°N) gives “reasonable” heat flux estimates except during the early part of the period over two small areas where sampling coverage is only “barely adequate”: off the west coast of the Baja Peninsula, extending to 160°W from 20°N to 30°N; and in the subtropical Western Pacific from 180°E to 150°E, 20°N to 25°N. The average number of observations of $U\Delta q$ per 4° by 4° latitude-longitude grid square per month over the winter season (December, January, and February) in the North Pacific Ocean over the period 1950–89 is shown in Figure 3.1. Observations of $U\Delta q$ are shown because they represent a conservative estimate of the flux variable observations since they require simultaneous observations of both $U$ and $\Delta q$ and they number less than $U\Delta T$ observations. The North Pacific Ocean study area covers most of the North Pacific Ocean from 20°N to 56°N; it is outlined in bold in the figure and contains 198 4° by 4° latitude-longitude grid squares. The regions of interest that will be examined in detail are also shown in this figure: the Kuroshio Current region (hereafter referred to as KCR) and the central North Pacific region (hereafter referred to as NPR). Regions of maximum variability in sampling occur
Figure 3.1: Average number of observations of $U\Delta q$ per 4° by 4° latitude-longitude grid square per month over the winter season (December, January, and February) in the North Pacific Ocean over the period 1950–89. Contour interval is 20.0. The North Pacific Ocean study area, the Kuroshio Current region (KCR), and the central North Pacific region (NPR) are all outlined in bold. Lines of latitude and longitude are every 20° along the basin boundaries where the standard deviation is 30 to 50 samples per month in the winter season, and as high as 80 samples per month over KCR.

Errors in the flux estimates increase where the number of samples is small and this error is amplified in areas where both the magnitude of the fluxes is large, and the fluxes (atmospheric and oceanic variables) exhibit a high degree of natural variability. By assuming the time scale of flux events in the extratropics to be about one day, Cayan (1992) showed that sampling errors in the North Pacific Ocean can contribute to maximum errors in the monthly mean $Q_H$ along the western boundary of $\sim$20 W m$^{-2}$, while the minimum errors occur over much of the northeast of the basin and are less than 10 W m$^{-2}$. Sampling errors in the monthly mean $Q_E$ are a maximum of $\sim$40 W m$^{-2}$ along the western boundary and at a minimum of less than 20 W m$^{-2}$ over most of the northeast of the basin.
Changes in measurement method (primarily) represent non-random errors in time. They are reported to have caused the observed increases in $U$, decreases in $T_s$, and increases in $T_a$ over the extratropical oceans. The observed increase in $U$ is apparently due to a conversion from Beaufort Scale wind estimates to wind speeds which result in an underestimation of $U$, and an increasing use of anemometers measurements which do not suffer from this problem (Cardone et al. 1991). A small fraction of the observed decrease in $T_s$ is due to less measuring of surface water, and more measurement of deeper, cooler water through ship engine intakes (Tabata 1978). Increases in $T_a$ may in part be due to increasing ship size and its proportional heat island effect (Taylor 1984b). The net effect of these temporal biases (increase in $U$ producing an increase in the heat fluxes, and a decrease in $\Delta T$ producing a decrease in $Q_H$) was shown by Cayan (1992) to be $\pm 2$ W m$^{-2}$ for the monthly mean $Q_H$ over the extratropics in winter ($U$ and $\Delta T$ effects approximately cancelling) and $+20$ W m$^{-2}$ for the monthly mean $Q_E$ over the period 1950-79. This is comparable to results of Isemer and Hasse (1991) who found the winter-time zonally averaged monthly mean $Q_E$ over the North Atlantic Ocean was underestimated by 35 W m$^{-2}$ at 20°N and 10 W m$^{-2}$ at 60°N as a result of underestimation of $U$ in the Beaufort Scale conversion.

The types of errors in the COADS data and the magnitude of the maximum resultant error in the computed winter-time monthly mean $Q_H$ and $Q_E$ over the North Pacific Ocean are shown in Table 3.1. For the monthly mean $Q_H$ and $Q_E$ the combined errors often exceeds the 10 W m$^{-2}$ error tolerance for climate studies (Taylor 1984a, WOCE 1986). But many monthly anomalies exceed 50 W m$^{-2}$. Furthermore, in this study we are examining the winter-time surface heat fluxes (average of the monthly means of December, January, and February) which reduces the sampling error by increasing the number of samples. If we assume this reduces the sampling errors by half, maximum errors in the winter-time mean $Q_H$ and $Q_E$ are around 20 W m$^{-2}$ and 50 W m$^{-2}$,
### Table 3.1: Sources of error in the COADS data and their maximum effect on the magnitude of the estimated monthly mean surface sensible and latent heat fluxes over the North Pacific Ocean in winter.

<table>
<thead>
<tr>
<th>Source of Error</th>
<th>Sensible Heat Flux</th>
<th>Latent Heat Flux</th>
</tr>
</thead>
<tbody>
<tr>
<td>Observation</td>
<td>±5 W m$^{-2}$</td>
<td>±10 W m$^{-2}$</td>
</tr>
<tr>
<td>Sampling</td>
<td>±20 W m$^{-2}$</td>
<td>±40 W m$^{-2}$</td>
</tr>
<tr>
<td>$U$ and $\Delta T$</td>
<td>±2 W m$^{-2}$</td>
<td>±20 W m$^{-2}$</td>
</tr>
</tbody>
</table>

Trends in the COADS data over the North Pacific Ocean, are examined in detail in Appendix A. Averaged over the North Pacific Ocean study area from 1950 to 1989, $U$ has increased 1.6 m s$^{-1}$, $T_a$ and $T_s$ have decreased by 0.4 °C and 0.7 °C, respectively, and sea level pressure has decreased about 3 mb. Some localized trends are of course stronger than these averages. Evidence is presented in Appendix A that shows a significant part (if not most) of these observed trends reflect real changes in the variables. For this reason (and due to the difficulty of filtering out only the artificial part of the trends), no adjustments were made to the data to try to reduce the effects of changing observation practices.

### 3.3 COADS Surface Sensible And Latent Heat Fluxes Computation

Monthly mean observations from the Monthly Summaries Trimmed (MST) subset of COADS marine surface data averaged over 4° by 4° latitude-longitude grid squares of the North Pacific Ocean (20°N – 60°N) over the period 1950–89 were used to estimate $Q_H$ and $Q_E$ using the bulk parameterizations (Equations 2.3 and 2.4). $Q_H$ and $Q_E$ were computed from observations of $U(z)(T_s - T_a(z))$ and $U(z)(q_a - q_a(z))$ where symbols are as in Section 2.2, and units are m s$^{-1}$ K and m s$^{-1}$ kg kg$^{-1}$, respectively. The variables were measured simultaneously (averaged over a period less than one hour) and
their product was averaged over one month as indicated by the overbar. Monthly mean values for December, January, and February were averaged to obtain a value for the winter season. Air density was computed from the ideal gas law \( \bar{\rho}_a = \bar{P}/(R\bar{T}_a) \); where \( P \) is pressure in Pa; \( T_a \) is the air temperature in degrees K; and \( R \) is the ideal gas constant 287.04 J kg\(^{-1}\) K\(^{-1}\). The constants used for the specific heat of dry air and the latent heat of evaporation both at 0 °C were \( C_p = 1.005 \times 10^3 \) J kg\(^{-1}\) K\(^{-1}\) and \( L_E = 2.5008 \times 10^6 \) J kg\(^{-1}\).

The transfer coefficients for sensible and latent heat (Stanton Number and Dalton Number) were computed from monthly mean stability variables \( (\Delta T, \Delta q, \text{ and } U) \) and based on Monin-Obukhov similarity theory after Oberhuber (1988) who used relationships prescribed by Large and Pond (1981, 1982):

\[
C_H = \frac{C_{HN}\sqrt{C_M/C_{MN}}}{1 - C_{HN}C_{MN}^{-1/2}\psi_H(Z/L)} \tag{3.12}
\]

\[
C_E = \frac{C_{EN}\sqrt{C_M/C_{MN}}}{1 - C_{EN}C_{MN}^{-1/2}\psi_E(Z/L)} \tag{3.13}
\]

\[
\sqrt{C_M/C_{MN}} = (1 - \sqrt{C_{MN}^{-1}\psi_M(Z/L)})^{-1} \tag{3.14}
\]

\[
C_{MN} = \frac{\kappa^2}{\ln^2(Z/Z_0)} \tag{3.15}
\]

\[
C_{HN} = \frac{0.0327\kappa}{\ln(Z/Z_0)} \tag{3.16}
\]

\[
C_{EN} = \frac{0.0346\kappa}{\ln(Z/Z_0)} \tag{3.17}
\]

\[
Z_o = c_{char}\frac{u_2}{g} \tag{3.18}
\]

\[
u_2^2 = C_MU^2 \tag{3.19}
\]

\[
T_o = T_a(1 + 1.7 \times 10^{-6}T_aq_a) \tag{3.20}
\]
Chapter 3. Data Description

The von Kármán constant ($\kappa$) is 0.4, and Charnock's constant ($c_{\text{char}}$) is 0.032. The transfer coefficient for momentum is $C_M$. The subscript $N$ indicates the case for neutral conditions ($T_s = T_a$). The functions $\psi$ and $(Z/L)$ are defined for stable conditions ($T_s < T_a$) as:

$$\psi_M = \psi_H = \psi_E = -7(Z/L)$$

$$\frac{Z}{L} = \frac{-70Z}{U^2T_o} (\Delta T + 2.5 \times 10^{-6}T_o^2 \Delta q)$$

and for unstable conditions ($T_s > T_a$):

$$\psi_M = 2\ln[(1 + X)/2] + \ln[(1 + X^2)/2] - 2\arctan X + \pi/2$$

$$\psi_H = \psi_E = 2\ln[(1 + X^2)/2]$$

$$X = 1 - 16(Z/L)^{1/4}$$

$$\frac{Z}{L} = \frac{-100Z}{U^2T_o} (\Delta T + 1.7 \times 10^{-6} T_o^2 \Delta q)$$

The computed heat fluxes thus represent the sum of the $UC \Delta T$, and $CU' \Delta T'$ terms (and corresponding $\Delta q$ terms) on the right-hand side of Equations 2.6 and 2.8, and omit the $UC' \Delta T'$, $\Delta T' CU'$, and $CU' \Delta T'$ terms (and corresponding $\Delta q$ terms). These latter covariances are small, however, and contribute less than 10% to the total heat flux in the North Pacific Ocean (Esbensen and Reynolds 1981).

The winter-time mean over 1950–89 of the surface air density is fairly constant along a line of latitude, and increases with latitude, from about 1.20 kg m$^{-3}$ at 20°N to about 1.27 kg m$^{-3}$ at 56°N. Excluding regions of low sampling (the north and northeast boundaries of the basin) the magnitude of interannual variability is small, with the standard deviation of the winter-time means less than 1% of the long-term mean.

The winter-time means of $C_H$ and $C_E$ over 1950–89 are shown in Figure 3.2. They both show similar patterns with a maximum off the east coast of Japan (over KCR)
Figure 3.2: Mean winter-time air-sea sensible and latent heat flux transfer coefficients over the period 1950–89. Values have been multiplied by $1.0 \times 10^3$ and contour interval is 0.05.
of about $1.5 \times 10^{-3}$ and $1.6 \times 10^{-3}$ for $C_H$ and $C_E$, respectively, and decreasing values to the south and east with minimum values of about $1.3 \times 10^{-3}$ and $1.4 \times 10^{-3}$ off the Baja Peninsula in the southeast. Assuming that the coefficient parameterizations are reasonable, these two figures show that while very similar in pattern, the coefficients are systematically different in magnitude, with $C_E$ everywhere $0.10 \times 10^{-3}$ greater than $C_H$. Interannual variability of the coefficients is small, with the standard deviation of each coefficient less than 3% of their long term mean values.

### 3.4 GCM Surface Sensible And Latent Heat Flux Computation

The GCM turbulent fluxes of surface sensible and latent heat over the oceans are based on Monin-Obukhov similarity theory and are almost identical to the bulk aerodynamic formulae (Equations 2.3 and 2.3):

$$Q_H = \rho_a C_p C_{DH} U_L (T_s - T_a)$$  \hspace{1cm} (3.27)

$$Q_E = \rho_a L_E C_{DH} U_L (q_s - q_a)$$ \hspace{1cm} (3.28)

except the subscript $L$ indicates variables at the lowest prognostic level. That the transfer coefficient is the same for $Q_H$ and $Q_E$ is in contrast to the formulations of Large and Pond (1981, 1982) used with the observational data in Section 3.3 where $C_E$ is ~8% larger than $C_H$.

The air density is $\rho_a = P/(RT_a)$ and the air temperature and specific humidity are given by:

$$T_a = \frac{T_{L+1/2}}{\sigma_{L+1/2}^k} \approx \frac{1}{C_p} S_{L+1/2}$$ \hspace{1cm} (3.29)

$$q_a = q_{L+1/2}$$ \hspace{1cm} (3.30)

where $\sigma$ is the vertical coordinate (1.0 is the bottom, 0.0 is the top) and $\sigma_{L+1/2} = 0.960$;
where the correction factor $c_f$ accounts for the fact that the mid-point of the lowest model layer is higher than the 10 m shipboard anemometer level on which observations are based. The height of the lowest prognostic level for momentum is $z_L$ and the von Kármán constant ($\kappa$) is 0.4. The drag coefficient ($C_D$) is specified and has a maximum of $2 \times 10^{-3}$ over the oceans.

The bulk Richardson number is defined as:

$$
R_i_B = 2 \frac{g z_L}{T_a} \frac{(T_a - T_s)}{U^2}
$$

(3.33)

where the factor 2 has been introduced to account for the staggering of the temperature and momentum levels in the model. The crude assumption used here is that the top of the surface layer is near the lowest prognostic level for temperature and moisture ($0.5z_L$) and that the wind speed at this point is approximately one half of that at the lowest prognostic level for momentum.

The functional dependence on $R_i_B$ is given by:

$$
F_H = \begin{cases} 
1 - 10 |R_i_B|/(1 + 10 \sqrt{|([R_i_B]/(87A_H^2)])|} & R_i_B < 0 \\
(1 - 5\varepsilon R_i_B)^2/(1 + 10(1 - \varepsilon)R_i_B) & 0 \leq R_i_B \leq 1/5\varepsilon \\
0 & R_i_B > 1/5\varepsilon
\end{cases}
$$

(3.34)

where:

$$
A_H = \frac{\kappa^2 \sqrt{z_o/z_L}}{C_{DH}}
$$

(3.35)
The factor $\varepsilon$ is 0.3 over the ocean and represents a crude attempt to account for unresolved turbulent transfer when the resolved part of the mean flow is sufficiently stable to repress it. The expression for $C(U_L)$ is based on the observational work of Smith (1980).

Since the winter-time air-sea turbulent transfers of sensible and latent heat over the North Pacific Ocean are almost always from the ocean to the atmosphere, it is also important to discuss how the model represents the storage and transport of heat in the ocean.

The GCM's oceanic component is highly simplified, consisting of a simple ocean mixed-layer model of 50 m depth. The heat budget is:

$$
\rho_s C_s H \left( \frac{\partial T_s}{\partial t} \right) = F_a - F_h - F_b
$$

(3.39)

where $H$ is the depth of the layer, $T_s$ is its temperature, and $\rho_s$ and $C_s$ are the density and specific heat of sea water. The terms on the right hand side of Equation 3.39 represent the net flux of heat entering the layer from the top ($F_a$) and leaving from the sides ($F_h$) and bottom ($F_b$). The net flux at the top of the ocean is:

$$
F_a = Q_{SW} + Q_{LW} + Q_H + Q_E
$$

(3.40)

where the first term on the right hand side of this equation is the net solar radiative flux at the surface, the second term is the net longwave radiative flux, and the last two terms are the net fluxes of sensible and latent heat due to turbulent vertical transfer in the atmospheric boundary layer. Positive values represent upward fluxes (a loss by the ocean) and negative values represent downward fluxes (a gain by the ocean).
The internal flux terms ($F_h$ and $F_b$ of Equation 3.39) represent the effects of horizontal heat transport and vertical exchange of heat with underlying water during periods of deepening or shallowing of the mixed layer. The simple slab ocean model does not account explicitly for these processes but specifies the sum of their effects ($F_h + F_b$, designated $R$) as a function of location and season. The value for $R$ is constrained such that Equation 3.39 gives a realistic simulation of the observed climatological $T_s$ field when forced with the $F_a$ field from a climate simulation using the observed monthly $T_s$ climatology. In other words, $R$ is specified on a monthly basis as a residual of the monthly mean net surface heat flux ($F_a$) and the implied change in the local heat storage of the slab as determined from the climatological $T_s$ field:

$$R = (F_a) - \rho_s C_s \left( \frac{\partial T_s}{\partial t} \right)$$

(3.41)

where overbars denote monthly mean values. Climatological $T_s$ fields were used to both estimate the $T_s$ tendency and as a lower boundary condition in the climate simulation from which $F_a$ was obtained.
Chapter 4

Data Analysis And Comparison

4.1 Introduction

Observational (COADS/NMC) and model (GCM) data are analysed and compared in this chapter. Long-term winter-time means (over the period 1950–89 for COADS/NMC data, and over 10 years of simulation for GCM data) are depicted by an overbar, and anomalies for each year are denoted by a prime. Data from a given data set is indicated by superscripts. The North Pacific study area and both KCR and NPR are essentially the same for the GCM data as for the COADS data shown in Figure 3.1. Due to the finer resolution of the model data, the North Pacific study area contains 243 grid squares compared to 198 grid squares for the COADS data.

4.2 Long-Term Means Of The Surface Sensible And Latent Heat Fluxes

\( \overline{Q_H}^{COADS} \) is shown in Figure 4.1a. Positive values correspond to upward fluxes and negative values correspond to downward fluxes. The maximum is about 120 W m\(^{-2}\) over KCR, and there is a strong gradient to the south and east, with values decreasing to a minimum of less than 20 W m\(^{-2}\) over much of the eastern half of the basin.

\( \overline{Q_H}^{GCM} \) is very similar and has a maximum of about 110 W m\(^{-2}\) over KCR with the strongest gradient to the south (Figure 4.1b). Lowest values of less than 30 W m\(^{-2}\) occur over much of the eastern half of the basin and a small region of negative values (\( \sim -5 \) W m\(^{-2}\)) lies over the Northeast Pacific.
Figure 4.1: Mean winter-time air-sea sensible heat flux. Contour interval is 10.0 W m$^{-2}$. KCR and NPR are outlined in the figures.
Figure 4.2: Mean winter-time air-sea latent heat flux. Contour interval is 25.0 W m$^{-2}$. KCR and NPR are outlined in the figures.
$Q_{E\text{COADS}}$ has a maximum of greater than 325 W m$^{-2}$ over KCR, and a strong gradient to the north and east where values decrease to a minimum of less than 75 W m$^{-2}$ over much of the eastern half of the basin north of 45°N (Figure 4.2a).

The pattern of $Q_{E\text{GCM}}$ is very similar to $Q_{H\text{GCM}}$, and has a maximum of 225 W m$^{-2}$ over KCR and the strongest gradient to the north and east (Figure 4.2b). Minimum values of less than 50 W m$^{-2}$ occur over the northeast part of the basin.

Over the North Pacific, $Q_{H\text{GCM}}$ is for the most part within the 20 W m$^{-2}$ maximum error of $Q_{H\text{COADS}}$ estimated in Section 3.2, and both $Q_{H\text{GCM}}$ and $Q_{H\text{COADS}}$ display marked zonal asymmetry. One notable difference is that $Q_{H\text{COADS}}$ is positive basin-wide, whereas $Q_{H\text{GCM}}$ has small negative values over a small region in the Northeast Pacific. Over the North Pacific, $Q_{E\text{GCM}}$ is also within the estimated 50 W m$^{-2}$ maximum error of $Q_{E\text{COADS}}$, except over KCR where $Q_{E\text{COADS}}$ is larger by 100 W m$^{-2}$. Also, $Q_{E\text{COADS}}$ is more zonal in nature than $Q_{E\text{GCM}}$ which is more zonally asymmetric (like $Q_{H\text{GCM}}$), and the influence of the Kuroshio Current extends further east for $Q_{E\text{GCM}}$. Despite the qualitative agreement between $Q_{H\text{GCM}}$ and $Q_{H\text{COADS}}$, and differences in $Q_{E\text{GCM}}$ and $Q_{E\text{COADS}}$, the significance of such a comparison is limited due to known errors in the COADS fluxes and limitations of the model ocean.

The overall pattern of $Q_{H\text{COADS}}$ and $Q_{E\text{COADS}}$ reflects their stability dependence. The region of maximum values over KCR is where $\Delta T\text{COADS}$ and $\Delta q\text{COADS}$ are greatest on average, about +5 °C and greater than 6 g kg$^{-1}$, respectively (Figures 4.3a and 4.4a, which are derived from the 40 year mean of the monthly mean differences). It is also where $U\text{COADS}$ is large, ranging from 9 to 11 m s$^{-1}$ (Figure 4.5a), which makes the turbulent transfer of heat at the air-sea interface more effective.

The maximum of $Q_{H\text{COADS}}$ occurs further north than the $Q_{E\text{COADS}}$ maximum, and the pattern of $Q_{H\text{COADS}}$ is identical to that of $\Delta T\text{COADS}$. While $\Delta T\text{COADS}$ is at a maximum over the Kuroshio and extends over the Northwest Pacific, the eastern half of
Figure 4.3: Mean winter-time air-sea temperature difference. Contour interval is 0.5 °C. KCR and NPR are outlined in the figures.
Figure 4.4: Mean winter-time air-sea humidity difference. Contour interval is 0.5 g kg$^{-1}$. KCR and NPR are outlined in the figures.
Figure 4.5: Mean winter-time wind speed. Contour interval is 1.0 m s⁻¹. KCR and NPR are outlined in the figures.
the basin is relatively constant with values between 0.5 and 1.5 °C. Why $\Delta T^{\text{COADS}}$ is constant and $\sim 1^\circ$C over half the North Pacific Ocean is not certain but may be the result of the heat flux reaching equilibrium between both an ocean with no strong horizontal temperature gradients and an atmosphere with no strong continental influences. The pattern of $\Delta q^{\text{COADS}}$ is markedly different from $\Delta T^{\text{COADS}}$ and shows a relatively uniform increase from north to south with a small disturbance in this pattern over KCR south of Japan where the maximum of 7 g kg$^{-1}$ is reached. In contrast to $\Delta T^{\text{COADS}}$ and $\overline{Q}_H^{\text{COADS}}$, this maximum in $\Delta q^{\text{COADS}}$ does not coincide with the $\overline{Q}_E^{\text{COADS}}$ maximum which is to the northwest and closer to the location of the $\overline{U}^{\text{COADS}}$ maximum. This suggests that $U^{\text{COADS}}$ may be more important in determining $Q_E^{\text{COADS}}$ than $Q_H^{\text{COADS}}$.

The patterns of $\overline{Q}_H^{\text{GCM}}$ and $\overline{Q}_E^{\text{GCM}}$ follow those of $\Delta T^{\text{GCM}}$ and $\Delta q^{\text{GCM}}$ (Figures 4.3b and 4.4b, which are derived from the 10 year mean of the differences of monthly means), while apparently bearing little relation to $\overline{U}^{\text{GCM}}$ (Figure 4.5b). The region of maximum heat flux values over KCR just east of Southern Japan is where $\Delta T^{\text{GCM}}$ and $\Delta q^{\text{GCM}}$ are greatest on average (+9 °C and 6 g kg$^{-1}$, respectively). Lowest values occur over the Northeast Pacific where $\Delta T^{\text{GCM}}$ and $\Delta q^{\text{GCM}}$ are at a minimum (less than +1.0 °C and 1.5 g kg$^{-1}$, respectively). However, $\overline{U}^{\text{GCM}}$ is highest over the Northeast Pacific (10 to 11 m s$^{-1}$) and lowest over KCR (3 to 6 m s$^{-1}$). This suggests that the mean fields of $\overline{Q}_H^{\text{GCM}}$ and $\overline{Q}_E^{\text{GCM}}$ are more strongly dependent on those of $\Delta T^{\text{GCM}}$ and $\Delta q^{\text{GCM}}$ than on $\overline{U}^{\text{GCM}}$.

A comparison of the GCM and COADS air-sea gradients shows that the pattern of $\Delta T^{\text{GCM}}$ is similar to $\Delta T^{\text{COADS}}$ with maximum values over KCR (+9 °C for $\Delta T^{\text{GCM}}$ and +5 °C for $\Delta T^{\text{COADS}}$) and minimum values over the Northeast Pacific ($\sim +0.7$ °C for both $\Delta T^{\text{GCM}}$ and $\Delta T^{\text{COADS}}$). The magnitude of the maximum values is different but it must be remembered that $T_a^{\text{COADS}}$ is an average of measurements made between approximately 5 and 25 m, depending on the ship, and $T_a^{\text{GCM}}$ is just a multiple of the
lowest model level air temperature. $\Delta q^{GCM}$ is very similar to $\Delta q^{COADS}$ with maxima over KCR (7 g kg$^{-1}$ and 6 g kg$^{-1}$, respectively) and minima over the Northeast Pacific (1.2 g kg$^{-1}$ and 0.9 g kg$^{-1}$, respectively). Another factor limiting direct comparison between the air-sea gradients is that $\Delta T^{GCM}$ and $\Delta q^{GCM}$ are biased low with respect to $\Delta T^{COADS}$ and $\Delta q^{COADS}$ because the GCM gradients are derived from the mean of the difference of monthly means, whereas the COADS gradients are derived from monthly mean differences. Despite this, the COADS and GCM gradients appear to be in qualitative agreement.

The mean wind speed fields $U^{COADS}$ and $U^{GCM}$ are very different. The $U^{COADS}$ maximum (11.9 m s$^{-1}$) is over the Kuroshio Current Extension, and minimum values (less than 7 m s$^{-1}$) occur across most of the basin at 20°N and off the west coast of North America’s Baja Peninsula. In contrast the $U^{GCM}$ maximum (11.2 m s$^{-1}$) is over the Northeast Pacific and the minimum (1.7 m s$^{-1}$) is southeast of Japan at about 27°N 165°E. Over the North Pacific, $U^{COADS}$ and $U^{GCM}$ differ by as much as 6 m s$^{-1}$. These differences are probably not significant since $U^{COADS}$ is a measurement of surface wind speed, and $U^{GCM}$ is the wind speed at the lowest model level. But the similarity of the $Q_{H}^{COADS}$ and $Q_{H}^{GCM}$ mean fields suggests a weak dependence on the very different $U$ fields and a much stronger dependence on the $\Delta T$ fields. The difference in the mean $Q_{E}$ fields appears to reflect the differences in the mean fields of $\Delta q$, and not $U$.

The time series of $Q_{H}^{COADS}$ and $Q_{E}^{COADS}$ at individual grid squares are strongly correlated, with an average correlation coefficient of $r \approx +0.8$ (Figure 4.6a). (The 95% significance level for a Student’s 2-tailed t-test with 38 degrees of freedom is $r = \pm 0.31$. Here significance assumes that the values for each season are independent of all others; this assumption is not strictly valid but we use significance level as an indication of the strength of the relation.) The strong correlation between the heat fluxes reflects the strong relation between $\Delta T^{COADS}$ and $\Delta q^{COADS}$ and the heat fluxes interdependence.
on $U_{COADS}$. The time series of $\Delta T_{COADS}$ and $\Delta q_{COADS}$ at individual grid squares are correlated with an average correlation coefficient of $r \approx +0.7$, with largest values over KCR ($r \approx +0.8$). Largest correlations between $Q^H_{COADS}$ and $Q^E_{COADS}$ ($r > +0.9$) also occur over the warm Kuroshio Current water where $\Delta T_{COADS}$ and $\Delta q_{COADS}$ can be large due to cold, dry air masses moving over the region from Asia. Over the eastern half of the basin the correlations between $\Delta T_{COADS}$ and $\Delta q_{COADS}$ are smaller since $T_{a,COADS}$ anomalies are more likely to be associated with moist maritime air masses that do not always contribute to $q_{a,COADS}$ anomalies.

The time series of $Q^G_{H,GCM}$ and $Q^G_{E,GCM}$ are also strongly correlated with an average correlation coefficient of $r \approx +0.8$, and most of the basin is greater than this average (Figure 4.6b). (The 95% significance level for a Student's 2-tailed t-test with 8 degrees of freedom is $r = \pm 0.63$.) This is due to the strong relation between $\Delta T_{GCM}$ and $\Delta q_{GCM}$ evident in their correlation pattern (not shown), which is very similar in pattern and magnitude to Figure 4.6b except along the west coast of USA where $\Delta T_{GCM}$ and $\Delta q_{GCM}$ are negatively correlated with maximum values of $r = -0.8$. This suggests that along the west coast of USA where $\Delta T_{GCM}$ and $\Delta q_{GCM}$ vary out of phase, variations in $Q^G_{H,GCM}$ and $Q^G_{E,GCM}$ are more closely related to $U_{GCM}$ than $\Delta T_{GCM}$ and $\Delta q_{GCM}$. This was verified by computing correlations between $Q^G_{H,GCM}$ and both $\Delta T_{GCM}$ and $U_{GCM}$, and between $Q^G_{E,GCM}$ and both $\Delta q_{GCM}$ and $U_{GCM}$. Correlations between $Q^G_{H,GCM}$ and $\Delta T_{GCM}$ are consistently much greater than between $Q^G_{H,GCM}$ and $U_{GCM}$. This was the case for $Q^G_{E,GCM}$ as well, except off the west coast of USA where $Q^G_{E,GCM}$ is more strongly correlated with $U_{GCM}$ than with $\Delta q_{GCM}$.

A common method of examining the relation between the sensible and latent heat fluxes is by computing the Bowen ratio ($\beta = Q_H/Q_E$) at individual grid points. $\beta_{COADS}$ is 0.1 basin-wide at 25°N and increases to the northwest to greater than 1.0 near the Kamchatka Peninsula (Figure 4.7a), showing the decreasing contribution by $Q^E_{COADS}$
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Figure 4.6: Correlation between winter-time air-sea sensible and latent heat fluxes. Contour interval is 0.1. (a) The 95% significance level is $r = \pm 0.31$ (Student’s 2-tailed t-test with 38 degrees of freedom). (b) The 95% significance level is $r = \pm 0.63$ (Student’s 2-tailed t-test with 8 degrees of freedom).
to the surface heat budget in the Northwest Pacific Ocean. This pattern reflects the
distribution of the ratio of $\Delta T_{\text{COADS}}$ to $\Delta q_{\text{COADS}}$ (Figures 4.3a and 4.4a).

Over the Northwestern Pacific from 20°N to 40°N, $\Delta T_{\text{COADS}}$ increases because $T_{a_{\text{COADS}}}$ decreases more rapidly than $T_{s_{\text{COADS}}}$. This is due to the influences of the warm Kuroshio Current on $T_{s_{\text{COADS}}}$ and of the cold Asian land mass on $T_{a_{\text{COADS}}}$. However, $\Delta q_{\text{COADS}}$ decreases over this region due to the nonlinear dependence of humidity on temperature (i.e., the small temperature differences at ~25 °C and 20°N result in larger values of $\Delta q_{\text{COADS}}$ than the larger temperature differences at ~10 °C and 40°N). From 40°N to 45°N over the Northwestern Pacific, $\Delta T_{\text{COADS}}$ decreases because $T_{s_{\text{COADS}}}$ decreases more rapidly than $T_{a_{\text{COADS}}}$ (due to the influences of the cool Oyashio Current on $T_{s_{\text{COADS}}}$ and of the cold Asian land mass on $T_{a_{\text{COADS}}}$). Northward of 45°N, $\Delta T_{\text{COADS}}$ becomes negative. Northward of 40°N, $\Delta q_{\text{COADS}}$ decreases rapidly as $T_{a_{\text{COADS}}}$ decreases and $q_{a_{\text{COADS}}}$ approaches zero. ($\Delta q_{\text{COADS}}$ does not increase north of 45°N like $\Delta T_{\text{COADS}}$ since $q_{a_{\text{COADS}}}$ cannot become negative like $T_{a_{\text{COADS}}}$.)

Over the Northeastern Pacific $\beta_{\text{COADS}}$ is much easier to explain. $\Delta T_{\text{COADS}}$ is relatively constant due to the lack of a continental influence on $T_{a_{\text{COADS}}}$, and the air is in relative thermal equilibrium with the ocean which exhibits no strong horizontal surface temperature gradients. While $\Delta T_{\text{COADS}}$ is constant, $\Delta q_{\text{COADS}}$ decreases approximately linearly with respect to latitude from 20°N to 45°N as $T_{a_{\text{COADS}}}$ and $T_{s_{\text{COADS}}}$ vary. As a result, $\beta_{\text{COADS}}$ increases from south to north over the Northeast Pacific. The interannual variability of $\beta_{\text{COADS}}$ is high, with the standard deviation approximately 10 to 30% of $\beta_{\text{COADS}}$ over most of the basin, and as high as 50 to 100% over the Northeast Pacific due to the higher variability in $\Delta T_{\text{COADS}}$ ($T_{a_{\text{COADS}}}$) than in $\Delta q_{\text{COADS}}$ ($q_{a_{\text{COADS}}}$) over this region.

The general pattern of $\beta_{\text{GCM}}$ (Figure 4.7b) is an increasing trend from 0.2 in the southeast to 1.0 in the northwest. There are also smaller scale departures from this
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(a) $\overline{\beta}^{\text{COADS}}$

(b) $\overline{\beta}^{\text{GCM}}$

Figure 4.7: Mean winter-time ratio of sensible to latent heat fluxes, or Bowen ratio ($\overline{\beta}$). Contour interval is 0.1.
pattern in the northeast consisting of negative values ranging from −0.1 to −0.4 due to small negative values of $Q_H^{GCM}$. Over the Northwest Pacific the increase in $\bar{Q}^{GCM}$ from south to north reflects the larger negative meridional gradient in $\Delta q^{GCM}$ ($\bar{Q}_E^{GCM}$) than $\Delta T^{GCM}$ ($\bar{Q}_H^{GCM}$) north of 35°N. South of 35°N, $\Delta q^{GCM}$ is approximately constant, while $\Delta T^{GCM}$ decreases southward. Interannual variability of $\beta^{GCM}$ is low over most of the North Pacific with the standard deviation ranging from 5 to 10% of $\bar{T}^{GCM}$. Over the Northeast Pacific, however, the standard deviation is 50 to 100% of $\bar{T}^{GCM}$ as a result of higher relative variability in $\Delta T^{GCM}$ ($T_a^{GCM}$) than $\Delta q^{GCM}$ ($q_a^{GCM}$).

In both GCM and COADS data sets the covariations of $Q_H$ and $Q_E$ are consistent with respect to their inter-correlations and ratios. $Q_H$ and $Q_E$ in both cases are highly correlated, with a basin-wide average correlation coefficient of $r \approx +0.8$ and largest values (greater than +0.9) over KCR (Figure 4.6). The Bowen ratios are also very similar, increasing from less than 0.2 in southeast of the domain to greater than 1.0 in the northwest, with the exception of a small region of negative values in the Northeast Pacific of the GCM Bowen ratio due to small negative values of $Q_H^{GCM}$.

4.3 Mean And Variability Of The Atmospheric Fields

Before we consider heat flux and atmospheric co-variability, we examine the mean and variability of the atmospheric fields. The observed long-term (1950–89) mean winter-time sea-level pressure ($\bar{SLP}^{NMC}$) pattern over the Northern Hemisphere is shown in Figure 4.8a. The most prominent features are the Asian High with a maximum of 1040 mb, and the Aleutian and North Atlantic Lows with minima of 999 mb. The pressure gradient between Asia and the North Pacific Ocean drives the near-surface winds which at 850 mb (not shown) are highly sub-geostrophic and strongest over the east coast of Asia (30°N – 50°N), and become almost geostrophic over the North Pacific Ocean.
as the Westerlies. This air moving off Asia and over the Pacific is very cold and dry (characteristic of its source region) which explains the large values of $\overline{dT}^{COADS}$ and $\overline{dp}^{COADS}$ over the Northwest Pacific Ocean. Maximum temporal variability occurs over the North Pacific Ocean to the east of the Aleutian Low (at about 165°W, 50°N where the standard deviation is almost 7 mb in Figure 4.8b) due to variation in the winter storm tracks. Variability is also high over the North Pole, Greenland, and Northern Europe.

The most prominent features of the GCM 10-year mean winter-time sea-level pressure (SLP$^{GCM}$) pattern are the Asian High with a maximum of 1039 mb, and the Aleutian and North Atlantic Lows with maxima of 992 mb and 994 mb, respectively (Figure 4.8c). The strongest pressure gradients are to the southwest and southeast of these centers. Maximum temporal variability occurs over the North Pacific Ocean where the standard deviation is almost 6 mb (Figure 4.8d). Regions of lesser variability occur over the north coast of Asia and over Greenland.

The observed long-term mean winter-time 500 mb height ($Z_{500}^{NMC}$) pattern (Figure 4.9a) shows lowest values over the North Pole and pronounced troughs extending southward over both Northeastern Asia and Eastern Canada. The strong meridional gradients in height associated with these semi-permanent troughs infer the strong zonal flow of the Asian and North American jet streams. Downstream of these jets (500 mb height troughs) are regions of frequent cyclogenesis that are associated with upward vertical motion (low pressure at the surface, high pressure aloft) and large temporal variability at the surface and at 500 mb (where the standard deviation is 6 mb and 80 m, respectively) due to the variation in cyclone tracks from year to year.

The GCM 10-year mean winter-time 500 mb height ($Z_{500}^{GCM}$) is at a minimum over the pole and prominent troughs extend southward over both Japan and Eastern Canada (Figure 4.9c). Maximum temporal variability occurs to the east of these troughs over the west coasts of Canada and Europe where the standard deviations are 56 m and $\sim$43 m
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Figure 4.8: Northern hemisphere (20°N to 90°N) winter-time sea-level pressure. Contour intervals: (a) 4.0 mb (b) 1.0 mb (c) 4.0 mb (d) 1.0 mb

(a) $\overline{\text{SLP}}^{\text{NMC}}$.

(b) $\text{SLP}^{\text{NMC}}$ standard deviation.

(c) $\overline{\text{SLP}}^{\text{GCM}}$.

(d) $\text{SLP}^{\text{GCM}}$ standard deviation.
Figure 4.9: Northern hemisphere (20°N to 90°N) winter-time 500 mb height. Contour intervals: (a) 100.0 m (b) 10.0 m (c) 100.0 m (d) 10.0 m.
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respectively (Figure 4.9d).

The 10-year mean GCM atmospheric fields realistically simulate those of the NMC analyses with all of the locations and magnitudes of the climatological features faithfully captured. These include the Asian High, Aleutian Low, and the North Atlantic Low in the SLP field, and the troughs over Eastern Asia and Eastern North America in the 500 mb height field. In general, however, the GCM fields are somewhat lower than the NMC fields: in the SLP field the Aleutian Low and the North Atlantic Low are too intense, and the ridge over Western Canada is weaker; and the 500 mb height field is uniformly too low over the Northern Hemisphere. Temporal variability in the NMC and GCM fields as represented by their standard deviations show good agreement. Maximum variability in both SLP^{NMC} and Z500^{NMC} fields occurs over the North Pacific, Northern Russia, and the North Atlantic, and minimum variability occurs around the Northern Hemisphere at 20°N. This pattern also occurs in the SLP^{GCM} field, but the region of maximum variability in Z500^{GCM} is shifted eastward over the west coast of North America, and variability over the North Atlantic is relatively weak.

4.4 Anomalies Of The Surface Sensible And Latent Heat Fluxes

Anomalies of the heat fluxes and fundamental variables at each grid point were computed by subtracting the long-term mean winter-time value from the value for each year. The mean magnitudes (absolute values) of the COADS sensible and latent heat flux anomalies (|\(Q'_H\)|^{COADS} and |\(Q'_E\)|^{COADS}) are shown in Figures 4.10a and 4.11a. Largest values of |\(Q'_H\)|^{COADS} and |\(Q'_E\)|^{COADS} occur over KCR just east of Japan, and are greater than 15 and 30 W m\(^{-2}\) respectively. Figures 4.3a, 4.4a, 4.5a, 4.12a, 4.13a, and 4.14a show that this is where \(\bar{U}^{COADS}\) is large and |\(\bar{U}'\)|^{COADS} is small (10 to 11 m s\(^{-1}\) and less than 0.5 m s\(^{-1}\), respectively); \(\Delta T^{COADS}\) and |\(\Delta T'\)|^{COADS} are large (5 to 6 °C and 0.4 to
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0.6 °C, respectively); and $\overline{\Delta q^{\text{COADS}}}$ is large and $|\overline{\Delta q'}^{\text{COADS}}|$ is largest (5 to 6 g kg$^{-1}$ and 0.3 to 0.6 g kg$^{-1}$, respectively).

The mean magnitudes of the GCM sensible and latent heat flux anomalies ($|\overline{Q_H'}^{\text{GCM}}|$ and $|\overline{Q_E'}^{\text{GCM}}|$) are shown in Figure 4.10b and 4.11b. Largest values of $|\overline{Q_H'}^{\text{GCM}}|$ (8 W m$^{-2}$) occur southwest of Japan within KCR where Figures 4.3b, 4.4b, 4.5b, 4.12b, 4.13b, and 4.14b show $\overline{\Delta T}^{\text{GCM}}$ and $|\overline{\Delta T'}^{\text{GCM}}|$ are large (6 to 8 °C and almost 0.5 °C respectively), and $\overline{U}^{\text{GCM}}$ and $|\overline{U'}^{\text{GCM}}|$ are both large (5 m s$^{-1}$ and 1.2 m s$^{-1}$ respectively). $|\overline{Q_H'}^{\text{GCM}}|$ is relatively constant over the eastern half of the basin (2 to 4 W m$^{-2}$) despite the great range between 20°N and 56°N of $\overline{\Delta T}^{\text{GCM}}$ (4 °C), $|\overline{\Delta T'}^{\text{GCM}}|$ (0.4 °C), $\overline{U}^{\text{GCM}}$ (7 m s$^{-1}$), and $|\overline{U'}^{\text{GCM}}|$ (1.1 m s$^{-1}$).

The largest values of $|\overline{Q_E'}^{\text{GCM}}|$ are almost 20 W m$^{-2}$, and occur south of Japan between 20°N and 25°N (Figure 4.11b) where $\overline{\Delta q}^{\text{GCM}}$ is large and $|\overline{\Delta q'}^{\text{GCM}}|$ is largest (5.5 g kg$^{-1}$ and 0.3 g kg$^{-1}$ respectively), and $\overline{U}^{\text{GCM}}$ is large and $|\overline{U'}^{\text{GCM}}|$ is largest (~9 m s$^{-1}$ and 0.8 m s$^{-1}$ respectively). Large values of $\overline{Q_E'}^{\text{GCM}}$ also occur over NPR (almost 15 W m$^{-2}$) and over the Northwest Pacific (almost 10 W m$^{-2}$). These regions are characterized by moderate values of $\overline{\Delta q}^{\text{GCM}}$ (3.0 to 3.5 g kg$^{-1}$), high and low values of $|\overline{\Delta q'}^{\text{GCM}}|$ (0.3 g kg$^{-1}$ and 0.1 g kg$^{-1}$, respectively), and moderate values of $\overline{U}^{\text{GCM}}$ (6 and 8 m s$^{-1}$) and $|\overline{U'}^{\text{GCM}}|$ (0.8 and 0.6 m s$^{-1}$).

The mean absolute values of the GCM sensible and latent heat flux anomalies are roughly half as large as their COADS counterparts. Largest values of $|\overline{Q_H'}^{\text{GCM}}|$ (6 to 8 W m$^{-2}$) and $|\overline{Q_H'}^{\text{COADS}}|$ (15 to 20 W m$^{-2}$) occur over KCR, although equally high values of $|\overline{Q_H'}^{\text{GCM}}|$ also occur in Northwest Pacific north of 60°N. Largest values of $|\overline{Q_E'}^{\text{GCM}}|$ occur over the Kuroshio Current Extension and south of KCR (18 to 20 W m$^{-2}$) where $|\overline{Q_E'}^{\text{COADS}}|$ is also highest (30 to 40 W m$^{-2}$). Maximum variability in both GCM and COADS heat fluxes appears closely linked to the Kuroshio Current, although relatively
Figure 4.10: Mean winter-time air-sea sensible heat flux anomaly magnitudes. Contour intervals: (a) 5.0 W m$^{-2}$ (b) 2.0 W m$^{-2}$. KCR and NPR are outlined in the figures.
Figure 4.11: Mean winter-time air-sea latent heat flux anomaly magnitudes. Contour intervals: (a) 5.0 W m$^{-2}$ (b) 2.0 W m$^{-2}$. KCR and NPR are outlined in the figures.
Figure 4.12: Mean winter-time air-sea temperature difference anomaly magnitudes. Contour interval is 0.1 °C. KCR and NPR are outlined in the figures.
Figure 4.13: Mean winter-time air-sea humidity difference anomaly magnitudes. Contour interval is 0.1 g kg$^{-1}$. KCR and NPR are outlined in the figures.
Figure 4.14: Mean winter-time wind speed anomaly magnitudes. Contour intervals: (a) 0.1 m s\(^{-1}\) (b) 0.2 m s\(^{-1}\). KCR and NPR are outlined in the figures.
large anomalies in the GCM heat fluxes also occur over NPR. Maximum values of $|Q'_H|$ and $|Q'_E|$ near KCR are linked to $\Delta T'$ and $\Delta q'$, $|\Delta T'|$ and $|\Delta q'|$ which are also at a maximum in this area in both GCM and COADS data sets. Over KCR $|\overline{U'}|^\text{GCM}$ is relatively high while $|\overline{U'}|^\text{COADS}$ is lowest in this region, but maximum values for both $|\overline{U'}|^\text{GCM}$ and $|\overline{U'}|^\text{COADS}$ occur over the Eastern Pacific. This implies that variability in both GCM and COADS $Q'_H$ and $Q'_E$ is determined by $\Delta T'$ and $\Delta q'$ basin-wide, and that the GCM heat flux anomalies are somewhat more dependent on $U'$ than the COADS heat flux anomalies.

The components of $Q'_H$ and $Q'_E$ are examined by first considering $Q'_H$. As discussed previously, the fluctuating components of $\rho_a$, $C_p$, and $C_H$ contribute little to the monthly mean sensible heat flux (Sections 2.3 and 4.2), thus:

$$\left(\frac{Q_H}{\rho_a C_p C_H} \right)' = \overline{U} \Delta T + \overline{U} \Delta T' + U' \Delta T + U' \Delta T'$$

(4.42)

where the overbar denotes the long-term mean and primes represent annual deviations from that mean. Subtracting the mean from both sides gives the anomaly and anomaly components:

$$\left(\frac{Q_H}{\rho_a C_p C_H} \right)' = \overline{U} \Delta T' + U' \Delta T + U' \Delta T' - U' \Delta T'$$

(4.43)

Large flux anomalies may occur when $\overline{U}$ and $\Delta T'$ and/or $U'$ and $\Delta T'$ are large.

An examination of each of the terms on the right hand side of Equation 4.43 in the COADS data shows that at times the anomaly components reinforce one another, but usually are of opposite sign and cancel one another. Averaged over 1950–89 and over the North Pacific study area, the $\overline{U} \Delta T'$ term is the largest and contributes most to $Q_H^{\text{COADS}}$, $U' \Delta T$ is the second largest term, $U' \Delta T'$ the third, and $\overline{U} \Delta T'$ is fourth. The relative sizes and spatial distributions of the $Q_H^{\text{COADS}}$ components are summarized in Table 4.1a. The size of the $Q_H^{\text{COADS}}$ component terms relative to $Q_H^{\text{COADS}}$ and $\overline{U} \Delta T'$ (the largest $Q_H^{\text{COADS}}$}
(a) $Q'_H^{\text{COADS}}$ components.

<table>
<thead>
<tr>
<th>$Q'_H$ Term</th>
<th>Size As % Of $Q'_H$</th>
<th>Size As % Of $\bar{U}\Delta T'$</th>
<th>Spatial Distribution</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\bar{U}\Delta T'$</td>
<td>100 to 200</td>
<td>100</td>
<td>Maximum to Northwest of study area, minimum to Southeast.</td>
</tr>
<tr>
<td>$U'\Delta T$</td>
<td>50 to 150</td>
<td>50 to 150</td>
<td>Maximum over KCR, minimum along Eastern boundary and Northeast Pacific.</td>
</tr>
<tr>
<td>$U'\Delta T'$</td>
<td>5 to 20</td>
<td>5 to 10</td>
<td>Approximately constant basin-wide.</td>
</tr>
<tr>
<td>$U'\Delta T'$</td>
<td>5 to 20</td>
<td>5 to 10</td>
<td>Maximum over KCR and Kuroshio Extension, near zero elsewhere.</td>
</tr>
</tbody>
</table>

(b) $Q'_E^{\text{COADS}}$ components.

<table>
<thead>
<tr>
<th>$Q'_E$ Term</th>
<th>Size As % Of $Q'_E$</th>
<th>Size As % Of $\bar{U}\Delta q'$</th>
<th>Spatial Distribution</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\bar{U}\Delta q'$</td>
<td>150 to 250</td>
<td>100</td>
<td>Maximum south of Japan and over KCR and Kuroshio Extension, minimum between 45°N and 55°N.</td>
</tr>
<tr>
<td>$U'\Delta q'$</td>
<td>100 to 200</td>
<td>50 to 150</td>
<td>Maximum south of KCR and NPR, minimum to northeast between 45°N and 55°N.</td>
</tr>
<tr>
<td>$U'\Delta q'$</td>
<td>5 to 20</td>
<td>5 to 10</td>
<td>Approximately constant basin-wide.</td>
</tr>
<tr>
<td>$U'\Delta q'$</td>
<td>5 to 20</td>
<td>3 to 10</td>
<td>Maximum south of Japan, near zero elsewhere.</td>
</tr>
</tbody>
</table>

Table 4.1: Relative sizes and spatial distributions of the COADS heat flux anomaly components.

Component term) given in the table represents about 90% of the maximum range in size over the North Pacific study area. Over a few isolated grid points the largest two $Q'_H^{\text{COADS}}$ component terms are many times the size of $Q'_H^{\text{COADS}}$ and/or the $\bar{U}\Delta T'$ term. This reflects the large magnitude/small scale variability of the largest $Q'_H^{\text{COADS}}$ component terms. The maximum of the $\bar{U}\Delta T'$ component in the Northwest Pacific is where $\bar{U}^{\text{COADS}}$ is large (the SLP gradient is strong) and $\Delta T^{\text{COADS}}$ is greatest due to anomalous cold air outbreaks over the Pacific. The maximum of the $U'\Delta T$ term over KCR is where $\Delta T^{\text{COADS}}$ is large and $U'^{\text{COADS}}$ is moderately large. These two components together account for most of
the $Q^H_{\text{COADS}}$ signal. The last two terms are much smaller, and $U^\text{COADS}$ and $\Delta T^\text{COADS}$ are weakly correlated (smaller than $\pm0.2$) so the $U'\Delta T'$ term is non-zero and is 5 to 20% of the magnitude of $Q^H_{\text{COADS}}$.

An equation analogous to 4.43 can also be written for the $Q^E_{\text{COADS}}$. Averaged over 1950–89 and over the North Pacific study area, $\overline{U}\Delta q'$ is the largest term and contributes most to $Q^E_{\text{COADS}}$, $U'^\Delta q$ is the second largest, $U'^\Delta q'$ the third, and $\overline{U}'\Delta q'$ is fourth (Table 4.1b). The maximum of the $\overline{U}\Delta q'$ component south of Japan and over KCR is where $\overline{U}^\text{COADS}$ is large (the SLP gradient is moderately strong) and $\Delta q^\text{COADS}$ is largest (due to cold/dry air outbreaks over the North Pacific), and the maximum of the $U'^\Delta q$ component south of KCR and NPR is where $U^\text{COADS}$ and $\Delta q^\text{COADS}$ are both large. These two components together account for most of the $Q^E_{\text{COADS}}$ signal. The last two terms are much smaller, and $U^\text{COADS}$ and $\Delta q^\text{COADS}$ are weakly correlated (smaller than $\pm0.2$) so $\overline{U}'\Delta q'$ is non-zero and is 5 to 20% of the magnitude of $Q^E_{\text{COADS}}$.

The GCM heat flux anomaly components are examined in a similar manner. An important difference from the COADS data computations is that $\Delta T^\text{GCM}$ and $\Delta q^\text{GCM}$ are differences of monthly means rather than monthly means of differences. As a result, one would expect the GCM heat flux anomaly components to be smaller than those of the COADS data. This problem is mitigated to a large extent since we are only concerned with the relative size of the anomaly components.

As for the COADS data, the GCM sensible and latent heat flux anomaly components at times reinforce one another, but usually are of opposite sign and cancel one another. For the GCM sensible heat flux anomalies the $U'^\Delta T$ term is largest and contributes most to $Q^G_{\text{GCM}}$, $\overline{U}\Delta T'$ is the second largest term, $U'^\Delta T'$ the third, and $\overline{U}'\Delta T'$ is fourth. The relative sizes and spatial distributions of the $Q^G_{\text{GCM}}$ components are summarized in Table 4.2a. The anomaly component size listed relative to $Q^G_{\text{GCM}}$ and $U'^\Delta T'$ represents about 90% of the maximum range in size over the North Pacific study area; for many of the
components their relative size is much greater over the northeast of the study area and is designated by "(NE higher)". This emphasizes that the heat flux anomaly components exhibit consistently different relative magnitudes over the Northwest and Northeast Pacific, and that the processes that determine heat flux anomalies differ between each side of the North Pacific Ocean. The maximum of the $U'\Delta T$ component over KCR and the Kuroshio Extension is where $\Delta T^{\text{GCM}}$ is largest and $U'^{\text{GCM}}$ is large, and the maximum of the $U\Delta q'$ component is dominated by and closely follows the distribution of $U^{\text{GCM}}$. These two components together account for most of the $Q_H^{\text{GCM}}$ signal. The last two terms are much smaller, and the weak correlation between $U'^{\text{GCM}}$ and $\Delta T^{\text{GCM}}$ (smaller than ±0.2) means the $U'\Delta T'$ term is non-zero and is almost everywhere 2 to 20% of the magnitude of $Q_H^{\text{GCM}}$.

Of the latent heat flux anomaly components, the $U'\Delta q$ term is the largest and contributes most to $Q_E^{\text{GCM}}$, $U\Delta q'$ is the second largest, $U'\Delta q'$ the third, and $U\Delta q'$ is fourth (Table 4.2b). The maximum of the $U'\Delta q$ component over KCR and the Kuroshio Extension is where $U'^{\text{GCM}}$ and $\Delta q^{\text{GCM}}$ are both large, and the maximum of the $U\Delta q'$ component is dominated by and closely follows the distribution of $U^{\text{GCM}}$. These two components together account for most of the $Q_H^{\text{GCM}}$ signal. The last two terms are much smaller, and $U'^{\text{GCM}}$ and $\Delta q^{\text{GCM}}$ are weakly correlated so the $U'\Delta q'$ term is non-zero and is almost everywhere less than 10% the magnitude of $Q_E^{\text{GCM}}$, except over the Northeast Pacific where it is 10 to 50% of the magnitude of $Q_E^{\text{GCM}}$.

The ordering of the GCM heat flux anomaly components differs from their COADS counterparts by the reversal of the relative size of the two largest terms. Of the $Q_H^{\text{GCM}}$ components, the $U'\Delta T$ term is at a maximum over KCR and the Kuroshio Extension, and is at a minimum over the Northeast Pacific, while the $U\Delta T'$ term is at a maximum over the Northeast Pacific and south of Japan, and at a minimum over most of the rest of the North Pacific. Of the two largest $Q_H^{\text{COADS}}$ components, the $U\Delta T'$ and $U'\Delta T$ terms have
(a) $Q_H^{GCM}$ components.

<table>
<thead>
<tr>
<th>$Q_H'$ Term</th>
<th>Size As % Of $Q_H'$</th>
<th>Size As % Of $U'\Delta T$</th>
<th>Spatial Distribution</th>
</tr>
</thead>
<tbody>
<tr>
<td>$U'\Delta T'$</td>
<td>100 to 200</td>
<td>100</td>
<td>Maximum over KCR and Kuroshio Extension, minimum over the Northeast Pacific.</td>
</tr>
<tr>
<td>$U'\Delta T'$</td>
<td>50 to 150</td>
<td>30 to 200 (NE higher)</td>
<td>Maximum over Northeast Pacific and south of Japan, minimum east of 160°E between 20°N – 30°N</td>
</tr>
<tr>
<td>$U'\Delta T'$</td>
<td>2 to 20</td>
<td>3 to 10 (NE higher)</td>
<td>Maximum over Northeast Pacific and KCR, uniformly low elsewhere.</td>
</tr>
<tr>
<td>$U'\Delta T'$</td>
<td>2 to 20</td>
<td>2 to 10 (NE higher)</td>
<td>Maximum over Northeast Pacific and south of Japan, low east of 140°E between 20°N – 40°N</td>
</tr>
</tbody>
</table>

(b) $Q_E^{GCM}$ components.

<table>
<thead>
<tr>
<th>$Q_E'$ Term</th>
<th>Size As % Of $Q_E'$</th>
<th>Size As % Of $U'\Delta q'$</th>
<th>Spatial Distribution</th>
</tr>
</thead>
<tbody>
<tr>
<td>$U'\Delta q'$</td>
<td>100 to 200 (NE higher)</td>
<td>100</td>
<td>Maximum over KCR and Kuroshio Extension, minimum over the Northeast Pacific.</td>
</tr>
<tr>
<td>$U'\Delta q'$</td>
<td>30 to 100 (NE higher)</td>
<td>50 to 200 (NE higher)</td>
<td>Maximum over Northeast Pacific and south of Japan, minimum between 20°N – 30°N, 170°E – 180°E</td>
</tr>
<tr>
<td>$U'\Delta q'$</td>
<td>2 to 10 (NE higher)</td>
<td>3 to 10 (NE higher)</td>
<td>Maximum over KCR, large over NPR and Northeast Pacific, minimum to northwest and southeast of study area.</td>
</tr>
<tr>
<td>$U'\Delta q'$</td>
<td>2 to 10 (NE higher)</td>
<td>3 to 10 (NE higher)</td>
<td>Maximum over Northeast Pacific, large south of Japan and over KCR, minimum to northwest and southeast of study area.</td>
</tr>
</tbody>
</table>

Table 4.2: Relative sizes and spatial distributions of the GCM heat flux anomaly components.
maxima over KCR and minima over the eastern boundary of the basin. This difference in the distribution of the $\overline{U} \Delta T'$ components of $Q_H^{GCM}$ and $Q_H^{COADS}$ (and perhaps the relative size of the components) is due to the very different $\overline{U}$ fields (Figure 4.5). The distribution of the $Q_E^{GCM}$ and $Q_E^{COADS}$ components is essentially the same as for $Q'_H$.

4.5 Correlations Between The Heat Flux And Atmospheric Anomalies

Time series of $Q_H^{COADS}$ and $Q_E^{COADS}$ averaged over the North Pacific Ocean study area both exhibit a generally increasing trend over the 40 year period (Figure 4.15a and 4.15b) and are correlated with $r = +0.62$. Averaged over KCR, $Q_H^{COADS}$ and $Q_E^{COADS}$ are increasing with time (Figure 4.15c and 4.15d) and are correlated with $r = +0.97$. Averaged over NPR, $Q_H^{COADS}$ shows a slight decrease with time while $Q_E^{COADS}$ is increasing (Figure 4.15e and 4.15f), and the two are correlated with $r = +0.75$. These figures show the basin-averaged and localized trends are small compared to the interannual variability of the fluxes.

Time series of $Q_H^{GCM}$ and $Q_E^{GCM}$ are shown in Figure 4.16. Averaged over the North Pacific Ocean study area, interannual variability is small ($Q_H^{GCM}$ and $Q_E^{GCM}$ are less than 0.5 W m$^{-2}$) and $Q_H^{GCM}$ and $Q_E^{GCM}$ are not significantly correlated with each other. Over KCR and NPR the flux anomalies show a decreasing trend, and their interannual variability is much greater ($|Q'_H|^{GCM} < 8$ W m$^{-2}$ and $|Q'_E|^{GCM} < 16$ W m$^{-2}$), and $Q_H^{GCM}$ and $Q_E^{GCM}$ are strongly correlated ($r = +0.93$ and $r = +0.99$ over KCR and NPR, respectively).

Figures 4.17a and 4.17c show time series of NMC winter-time sea-level pressure and 500 mb height anomalies (SLP$^{NMC}$ and Z500$^{NMC}$) averaged over the Northern Hemisphere ($20^\circ$N to $90^\circ$N). The time series of SLP$^{NMC}$ is relatively stationary, indicating no general trend. Z500$^{NMC}$ is decreasing an average of 0.2 m yr$^{-1}$, but appears to be
increasing in the latter part of the 1980’s. There appears to be no relation between these hemispheric-averaged atmospheric anomalies and the area-averaged $Q_{H}^{COADS}$ and $Q_{E}^{COADS}$ over the North Pacific Ocean basin as a whole, KCR, and NPR, as their correlations are not significant.

Time series of SLP$^{GCM}$ and Z500$^{GCM}$ for the Northern Hemisphere (20°N to 90°N) appear relatively invariant and stationary as they show very small anomalies and trends with respect to typical 10 year mean values (Figures 4.17b and 4.17d). Correlations between these atmospheric anomaly time series and those of $Q_{H}^{GCM}$ and $Q_{E}^{GCM}$ averaged over the North Pacific Ocean basin as a whole, KCR, and NPR are not significant.

Spatial variability of the significant correlations between the surface heat flux anomalies and atmospheric anomalies are considered next. Figure 4.18 shows the significant correlations between time series of $Q_{H}^{COADS}$ and $Q_{E}^{COADS}$ averaged over KCR (Figures 4.15c and 4.15d) and time series of SLP$^{NMC}$ and Z500$^{NMC}$ at individual grid points over the Northern Hemisphere (20°N – 90°N). The correlation patterns between $Q_{H}^{COADS}$ and $Q_{E}^{COADS}$ and SLP$^{NMC}$ are very similar, with maximum correlations of greater than -0.8 located about 5° north and 25° east of the KCR center. Significant positive correlations of greater than +0.4 are found over a broad band stretching from the coast of China northwest over Southern Russia to about 40°N 80°E (just north of the Tibetan Plateau), as well as over Northwest North America. Apart from their relation to KCR heat flux, these regions of positive correlations also appear to be related to topographic influences on the flow by the Tibetan Plateau in Southeastern Asia and the Rocky Mountains in Western North America. The southwestern edge of the positive correlation region over Southeastern Asia closely follows the northern boundary of the Tibetan Plateau which may act as a barrier to a KCR heat flux anomaly signal propagating southwestward. The positive correlation region over Northwest North America lies along the Rocky Mountains which may also act as a barrier to any eastward propagating signal and serve to amplify
Figure 4.15: Time series of COADS winter-time air-sea heat flux anomalies averaged over the given regions. Dotted line indicates zero anomalies. Solid straight line indicates a linear least-squares regression fit to the data.
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(a) $Q_{H}^{GCM}$ North Pacific Ocean study area.  
(b) $Q_{E}^{GCM}$ North Pacific Ocean study area.

(c) $Q_{H}^{GCM}$ Kuroshio Current Region.  
(d) $Q_{E}^{GCM}$ Kuroshio Current Region.

(e) $Q_{H}^{GCM}$ North Central Pacific Region.  
(f) $Q_{E}^{GCM}$ North Central Pacific Region.

Figure 4.16: Time series of GCM winter-time air-sea heat flux anomalies averaged over the given regions. Dotted line indicates zero anomalies. Solid straight line indicates a linear least-squares regression fit to the data.
Figure 4.17: Time series of winter-time atmospheric anomaly fields averaged over the Northern Hemisphere (20°N – 90°N). Dotted line indicates zero anomalies. Solid straight line indicates a linear least-squares regression fit to the data.
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The west-to-east positive-negative-positive correlation pattern corresponds to ridge strengthening over Eastern Asia and North America (which are heat sinks due to high radiation loss in winter) and trough deepening over the Northwest Pacific (a heat source), in association with warm heat flux anomalies over KCR. This pattern may be caused by standing oscillations with geographically fixed ridges and troughs, or may also be caused by propagating wave-like disturbances with preferred zonal and/or meridional scales.

The SLpNMc patterns agree with the modelling results of Weaver (1987) who found the response of the lower troposphere to anomalous Kuroshio Current region heating anomalies was weak positive upstream over Asia and strong negative downstream over the central North Pacific. Also, Pitcher et al. (1988) obtained a similar response to a Northwest Pacific (130°E to 150°W) warm sea surface temperature anomaly in January (corresponding to a positive heating anomaly) in the NCAR CCM (National Center for Atmospheric Research Community Climate Model). Significant SLP responses were negative over most of the North Pacific and positive over Western North America, and although no significant positive response was found over Eastern Asia, a significant negative response was also found over Western Asia.

Like the SLpNMc patterns, the correlation patterns between KCR \( Q_H^{COADS} \) and \( Q_E^{COADS} \) and Z500NMc are also very similar. The maximum correlation is greater than \(-0.8\) about 5° north and 10° east of the center of KCR for both \( Q_H^{COADS} \) and \( Q_E^{COADS} \). This negative correlation is significant over a broad spatial extent covering most of the North Pacific Ocean and stretching from Northeastern China eastward almost to the west coast of North America. There are significant positive correlations greater than \(+0.5\) over Northwest North America and Northeastern Asia. In the case of warm KCR heat flux anomalies, this pattern corresponds to ridge strengthening over Eastern Asia and Western North America, trough deepening over the Northwest Pacific, and a strengthening of
Figure 4.18: Correlations between time series of $Q_H^{\text{COADS}}$ and $Q_E^{\text{COADS}}$ averaged over KCR (outlined in figure) and $\text{SLP}^{\text{NMC}}$ and $\text{Z500}^{\text{NMC}}$ at individual grid points over the Northern Hemisphere ($20^\circ\text{N} - 90^\circ\text{N}$) over the 1950–89 period. The 95% significance level is $r = \pm 0.31$ (Student's 2-tailed t-test with 38 degrees of freedom). Only correlations where $|r| \geq 0.3$ are shown. Contour interval is 0.1.
Figure 4.19: Correlations between GCM time series of $Q^G_{H}$ and $Q^G_{E}$ averaged over KCR (outlined in figure) and SLP$^G_{GCM}$ and Z500$^G_{GCM}$ at individual grid points over the Northern Hemisphere (20°N — 90°N) over 10 years of simulation. The 95% significance level is $r = \pm 0.63$ (Student’s 2-tailed t-test with 8 degrees of freedom). Only correlations where $|r| \geq 0.6$ are shown. Contour interval is 0.1.
the Asian Jet (in relation to the mean \(Z_{500}^{\text{NMC}}\) field in Figure 4.9a), and *vice versa* for cool KCR heat flux anomalies.

A similar pattern was found in the GCM experiment of Pitcher *et al.* (1988) where significant 700 mb responses to a Northwest Pacific warm sea surface temperature anomaly were positive over Northwest North America and Northeast Asia, and negative (but not significant) over the North Pacific. (Significant negative responses also occurred over Europe and Eastern North America). The \(Z_{500}^{\text{NMC}}\) correlation patterns also appear related to observed major modes of variability in the monthly 500 mb height anomalies alone: (1) the Western Pacific (WP) pattern in the point by point intercorrelation analysis of Wallace and Gutzler (1981); and (2) the West Pacific Oscillation (WPO) and East Pacific (EP) patterns in the rotated principal component analysis of Barnston and Livezey (1987). The WP pattern index is defined as (Wallace and Gutzler 1981):

\[
WP = 1/2[Z^*(60^\circ \text{N}, 155^\circ \text{E}) - Z^*(30^\circ \text{N}, 155^\circ \text{E})]
\]  

(4.44)

where \(Z^*\) is the 500 mb height anomaly divided by the standard deviation at that grid point. The WP pattern differs from Figures 4.18c and 4.18d in that the positive centers over Northeastern Asia and Northwest North America are joined, leaving only a north-south dipole with broad longitudinal extent centered on the grid points in Equation 4.44. The WP index was computed using NMC winter-time 500 mb heights and correlations between the WP index and KCR \(Q_H^{\text{COADS}}\) and \(Q_E^{\text{COADS}}\) are significant but weak \((r = +0.41 \text{ and } r = +0.36)\). The KCR \(Z_{500}^{\text{NMC}}\) patterns are perhaps more similar to a combination of the WPO and EP patterns which consist of a north-south dipole over the West and East Pacific, respectively. However, the WPO is strongest from November to February, and the EP pattern is stronger in March, and the two collapse into an eastward displaced WP pattern when three month averages are examined.

The correlation patterns between KCR \(Q_H^{\text{GCM}}\) and \(Q_E^{\text{GCM}}\) and SLP\(^{\text{GCM}}\) are shown
in Figures 4.19a and 4.19b and appear very similar. Maximum values of \( r = -0.9 \) for \( Q^*_{GCM} \) and \( Q^*_{E GCM} \) occur 70° downstream from KCR, at 30°N 130°E. These maxima lie within a broad band of negative correlations greater than -0.6 stretching from 175°E to the west coast of North America. There are also regions of strong negative values over Eastern North America and the Western Atlantic (maximum values of -0.88 for \( Q^*_{H GCM} \), and -0.74 for \( Q^*_{E GCM} \)) and Northwestern Africa and the Eastern Atlantic (greater than -0.7 for both \( Q^*_{H GCM} \) and \( Q^*_{E GCM} \)). Large positive values greater than +0.6 occur over an area which extends from the Arabian Sea to Northern India and the Tibetan Plateau.

The most significant difference between the GCM and COADS/NMC patterns is the eastward shift of 45° in the maximum (negative) GCM values with respect to the COADS/NMC values. The maximum correlations between KCR \( Q^*_{H GCM} \) and \( Q^*_{E GCM} \) and SLP\(^*_{GCM} \) is negative 70° east of KCR, and is negative 25° east of KCR for COADS/NMC correlations (Figures 4.18a, 4.18b, 4.19a, and 4.19b). (There is also a small local negative correlation between \( Q^*_{H GCM} \) and SLP\(^*_{GCM} \).) The GCM patterns lack significant positive values over Eastern Asia and Western North America that appear in the COADS/NMC patterns, but this probably only reflects the smaller number of years of GCM data since positive but insignificant values are present. Both GCM and COADS/NMC patterns have negative values over Northeastern Canada but in the former case these negative values extend over the Western Atlantic. The GCM patterns also exhibit negative values over Northwestern Africa and positive values over the Arabian Sea which do not appear in the COADS/NMC patterns.

Correlations between KCR \( Q^*_{H GCM} \) and \( Q^*_{E GCM} \) and Z500\(^*_{GCM} \) are also similar (Figures 4.19c and 4.19d), but yield weaker significant values over much smaller areas. Maximum values of greater than -0.8 for \( Q^*_{H GCM} \) occur over KCR and smaller significant values occur over NPR (greater than -0.65), over Northern Africa (greater than +0.7), and over the Arabian Sea (greater than -0.7).
This differs slightly from the COADS/NMC data where maximum correlations between KCR $Q_H^{\text{COADS}}$ and $Q_E^{\text{COADS}}$ and $Z_{500}^{\text{NMC}}$ are negative but centered $10^\circ$ east of KCR (Figures 4.18c and 4.18d). Negative correlations also occur in both data sets over the Arabian Sea/Northern India. The prominent regions with positive values over Northeast Asia and Northwest North America in the COADS/NMC data are insignificant in the GCM data. Regarding the KCR correlation patterns in general, the GCM patterns are very different from COADS/NMC patterns at SLP. The (negative) maximum occurs $70^\circ$ east of KCR in the GCM data, and is $25^\circ$ east of KCR in the COADS/NMC data. They do, however, share a similar local (negative) maximum at 500 mb.

Figure 4.19 has little in common with the GCM experiment of Pitcher et al. (1988) described previously where the SLP and 500 mb response was largest, significant, and positive over Western North America, and negative but not significant over the North Pacific primarily west of $180^\circ$W and over Eastern Canada. These differences can not be treated as significant, however, as they may be attributable to the different forcing (sea surface temperature in the NCAR GCM as opposed to heat flux forcing) which is over slightly different regions, the short length of the model runs, and/or the many differences in the models themselves.

Next we consider the significant correlations between NPR $Q_H^{\text{COADS}}$ and $Q_E^{\text{COADS}}$ and SLP$^{\text{NMC}}$ and $Z_{500}^{\text{NMC}}$ (Figure 4.20). They are much weaker than those for KCR. The correlation patterns between $Q_H^{\text{COADS}}$ and SLP$^{\text{NMC}}$ and $Z_{500}^{\text{NMC}}$ are very similar. Positive maxima of greater than $+0.4$ lie over Northeastern North America and negative maxima of greater than $-0.3$ occur off the west coast of North America and over the Northwest Atlantic Ocean around $25^\circ$N. The correlation patterns between $Q_E^{\text{COADS}}$ and SLP$^{\text{NMC}}$ and $Z_{500}^{\text{NMC}}$ are stronger and also very similar, but differ from the $Q_H^{\text{COADS}}$ patterns. Negative maxima of greater than $-0.6$ occur just east of NPR and greater than $-0.4$ off the southeast coast of USA, while positive maxima of $+0.4$ and $+0.3$ lie over
Northern North America and east of Greenland, respectively.

In the case of warm NPR \( Q^\text{COADS}_H \) anomalies, the west-to-east negative-positive-negative correlation pattern corresponds to a weaker ridge over Western North America and a weaker North Atlantic Low and North American Jet due to reduced meridional gradients in SLP and 500 mb height over Northeast North America. For warm NPR \( Q^\text{COADS}_E \) anomalies, the associated atmospheric circulation is similar to that for \( Q^\text{COADS}_H \) at SLP, but at 500 mb the climatological mean features are not weakened, but enhanced: the troughs over the Northeast Pacific and Eastern North America are deeper, and the ridge over Western North America is stronger. Also, correlations between NPR \( Q^\text{COADS}_E \) and atmospheric anomalies show values of moderate magnitude and spatial extent, whereas those for NPR \( Q^\text{COADS}_H \) are much weaker. This suggests that the strong relation observed between sea surface temperature anomalies in this region and atmospheric anomalies (Namias and Born 1972, Rasmusson and Carpenter 1982, and Wallace and Jiang 1987) is based much more on \( Q^\text{COADS}_E \) than \( Q^\text{COADS}_H \) (\( Q^\text{COADS}_E \) is about 3 times the size of \( Q^\text{COADS}_H \) on average in Figures 4.10a and 4.11a). This may be because the sensible heat transferred into the atmosphere is effectively trapped within the marine boundary layer, but the latent heat is carried higher into the troposphere via convective cloud development during the passage of cyclones through this region (typical of the high amplitude positive mode of the second rotated empirical orthogonal function (EOF) of October to March cyclone track density described by Anderson and Gyakum (1989)).

The correlation pattern between NPR \( Q^\text{COADS}_E \) and \( Z_500^{\text{NMC}} \) (Figure 4.20d) resembles an observed major mode of variability in the 500 mb height field (Wallace and Gutzler 1981, Barnston and Livezey 1987) identified as the Pacific North America (PNA) pattern. The PNA pattern is characterized by amplification of the ridge over Western North America and troughs over the Northeast Pacific and Eastern North America in the mean \( Z_500^{\text{NMC}} \) field (Figure 4.9a). The PNA pattern index is defined as (Wallace and Gutzler
Figure 4.20: Correlations between time series of $Q^H_{\text{COADS}}$ and $Q^E_{\text{COADS}}$ averaged over NPR (outlined in figure) and SLP$^{\text{NMC}}$ and Z500$^{\text{NMC}}$ at individual grid points over the Northern Hemisphere ($20^\circ \text{N} - 90^\circ \text{N}$) over the 1950–89 period. The 95% significance level is $r = \pm 0.31$ (Student's 2-tailed t-test with 38 degrees of freedom). Only correlations where $|r| \geq 0.3$ are shown. Contour interval is 0.1.
Figure 4.21: Correlations between time series of $Q_H^{GCM}$ and $Q_E^{GCM}$ averaged over NPR (outlined in figure) and $SLP^{GCM}$ and $Z500^{GCM}$ at individual grid points over the Northern Hemisphere ($20^\circ N - 90^\circ N$) over 10 years of simulation. The 95% significance level is $r = \pm 0.63$ (Student's 2-tailed t-test with 8 degrees of freedom). Only correlations where $|r| \geq 0.6$ are shown. Contour interval is 0.1.
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PNA = \frac{1}{4}[Z^*(20^\circ N, 160^\circ W) - Z^*(45^\circ N, 165^\circ W) + Z^*(55^\circ N, 115^\circ W) - Z(30^\circ N, 85^\circ W)] \tag{4.45} (4.45)

and its correlation with NPR $Q^\text{COADS}_E$ is significant but weak ($r = +0.46$). This pattern is also evident to some extent in the responses of GCMs to warm sea surface temperature anomalies over this region (Kushnir and Lau 1992, Pitcher et al. 1988).

Significant correlations between NPR $Q^\text{GCM}_H$ and NPR $Q^\text{GCM}_E$ and atmospheric anomalies differ from the model KCR patterns chiefly by the presence of stronger correlations north of 60°N over Canadian and Russian Arctic regions. The $Q^\text{GCM}_H$ and $Q^\text{GCM}_E$ SLP$^\text{GCM}$ patterns are very similar, with maximum values of $r = -0.87$ over the Arctic Ocean off the Northeast Russian coast. Strong negative values ($r \sim -0.80$) also occur over the Northeastern Pacific centered on 35°N 140°W. Positive correlations greater than +0.6 occur over the North Atlantic and Northeastern Canada (with maxima of $r \sim +0.76$), as well as strong positive values of greater than +0.8 over the Persian Gulf/Arabian Peninsula.

The correlation patterns between NPR $Q^\text{GCM}_H$ and $Q^\text{GCM}_E$ and Z500$^\text{GCM}$ are also very similar. Maximum values of $r \sim -0.86$ occur over the Eastern Pacific Ocean just east of NPR at 35°N 140°W, and strong negative values occur over Northeast Russia ($r \sim -0.73$), and south of Japan (greater than –0.6). Strongest positive values of greater than +0.8 occur over North America and the Mediterranean Sea/North Africa.

The corresponding significant COADS/NMC NPR correlation patterns are much weaker, although many of the features in the GCM patterns are present. In the $Q^\text{GCM}_H$ correlations common features include negative values over the Northeast Pacific and Northwest Atlantic, and positive values over Northeastern North America. Similar features in the $Q^\text{GCM}_E$ patterns include negative values east of NPR and over the Northwest Atlantic southwest of USA, and positive values over North America. Strong correlations
(greater than +0.7) in the $Q^G_{H_{\text{CM}}}$ and $Q^G_{E_{\text{CM}}}$ patterns over the Persian Gulf/Arabian Peninsula at SLP and over the Mediterranean Sea/North Africa at 500 mb do not appear in the COADS/NMC fields.

The negative and positive maxima over the Northeast Pacific and North America give the GCM patterns a weak resemblance to the PNA pattern evident in the cross-correlations between the observed sea surface temperature at 32°N 165°W and 500 mb height (Wallace and Jiang 1987), and to the 500 mb height response to the sea surface temperature anomaly in the GCM experiment of Pitcher et al. (1988). Both NPR $Q^G_{H_{\text{CM}}}$ and NPR $Q^G_{E_{\text{CM}}}$ are weakly correlated to the PNA index ($r = +0.41$ and $r = +0.38$, respectively), which are insignificant but comparable in magnitude to the COADS/NMC values. Stronger similarity is found with the GCM sea surface temperature anomaly experiment of Kushnir and Lau (1992), where the strongest and most significant response in SLP and 500 mb height was negative, centered at 45°N and extending from about 160°W to the west coast of North America.

4.6 Principal Component Analysis Of The Heat Fluxes

In order to further understand the dominant spatial patterns of variability in the North Pacific Ocean sensible and latent heat fluxes, P-mode principal component analysis (PCA) was performed on the data. The term P-mode refers to the way the initial 2-dimensional matrix of raw data is constructed, in this case the stations (i.e., $4^\circ \times 4^\circ$ latitude-longitude grid squares) are designated as the “variables” (in regard to statistical nomenclature), and the observations (i.e., years) are designated as the “subjects” (e.g. see Willmott 1978). Each data value was normalised by subtracting the station mean and dividing by the station standard deviation, and then correlated with all other stations to form the correlation matrix. There exists some confusion in the literature over the
naming of PCA modes as Richman (1986) refers to this type of analysis as S-mode.

PCA of this correlation matrix yields a series of linear combinations of the normalized data in space:

\[
  z_1 = \alpha_{11}x_1 + \alpha_{12}x_2 + \ldots + \alpha_{1p}x_p \\
  \vdots \quad \vdots \quad \vdots \\
  z_p = \alpha_{p1}x_1 + \alpha_{p2}x_2 + \ldots + \alpha_{pp}x_p
\]

where \( p \) is the number of stations; \( x_j \) is the normalized data value (heat flux) at station \( j \); \( z_i \) is called the PC, PC score, or PC amplitude; and \( \alpha_{ij} \) is the coefficient or loading for PC \( i \) and station \( j \) (the vector of \( \alpha \)s for a given PC is called the eigenvector). The first PC is that linear function (combination of \( \alpha \)s) that explains the maximum possible variance, the second PC is the linear function with the maximum possible variance subject to being uncorrelated with the first PC, the third PC is the linear function which maximizes variances subject to being uncorrelated with the first and second PCs, and so on. The variance of each PC is called it’s eigenvalue, or eigenroot. Details of PC extraction can be found in texts such as Preisendorfer (1988) and Joliffe (1986).

At each stage of PC extraction, the sum of the variances (eigenvalues) of the PCs is as large as possible such that the PCs account for the maximum possible proportion of the variation in the original data. Altogether a maximum of \( p \) PCs can be constructed where \( p \) is the number of stations, however, only those PCs which account for an amount of variance greater than noise (random PCs) are selected for interpretation. Unfortunately, only heuristic rules exist for determining which PCs contain meaningful signal. We have used the "scree" test which "...considers that the last pronounced discontinuity in the eigenvalue series prior to its leveling off constitutes the division between the unrotated PCs that possess non-random signals (those with larger eigenvalues) and the ones that do not." (Richman and Lamb 1985).
Once the significant PCs were determined, they were rotated to: (1) reduce the effects of domain shape dependence; (2) increase subdomain stability; (3) reduce sampling errors; and (4) simplify the PC structure by isolating subdomain covariability which facilitates more meaningful interpretation of the PCs (Richman 1986). An infinite variety of rotation algorithms are possible, with a few coded and commonly available. We have used the VARIMAX rotation which rotates the PCs retaining their initial orthonormality, but unlike the initial PC extraction which maximizes the variances of the PCs in succession, the variances of the VARIMAX rotated PCs are maximized simultaneously. The result is that variances of the rotated PCs are more uniformly distributed than for the unrotated solution (which loads more heavily on the first or first and second PC), even though their mutual sums are equal. The computer code used for this PCA (including VARIMAX rotation) is the program FACTOR (Veldman 1967).

The advantages PCs have over the popular EOFs (empirical orthogonal functions) include: (1) if we begin with a correlation matrix, the coefficients $\alpha_{ij}$ are the correlation between the data time series for the station ($x_j$) and the time series of PC amplitudes ($z_i$); and (2) since the sums of squares of the PC $\alpha$s are not normalized by their PC amplitudes, rotated PCS better emphasize regional variability than rotated EOFs (Richman and Lamb 1985).

Of the 7920 grid squares (40 years $\times$ 198 grid squares) in the COADS data covering the North Pacific Ocean, 23 had missing values for both $Q_{COADS}^H$ and $Q_{COADS}^E$, all of these were in the 1950s and most were along the northern boundary of the study area near 54°N 180°W. For PCA, these missing values were replaced by an average of the surrounding 8 grid squares.

Application of the scree test to their respective eigenvalue series suggested the leading 2 PCs of both $Q_{COADS}^H$ and $Q_{COADS}^E$ contain non-random signals. The rotated first PC of $Q_{COADS}^H$ ($Q_{COADS}^H$ PC1) accounts for 18.5% of the total variance in $Q_{COADS}^H$ and is
shown with its associated time series of amplitude in Figure 4.22a. $Q_H^{COADS}$ PC1 shows the dominant mode of interannual variation in winter-time $Q_H^{COADS}$ is between years where $Q_H^{COADS}$ is high over KCR (where $\alpha > +0.6$) and $Q_H^{COADS}$ is low over much of the Northeast Pacific (where $\alpha < -0.6$), and years where the opposite situation occurs. (The sign of the $\alpha$ is arbitrary, and merely indicates opposite tendencies in the original data between regions of positive and negative sign.) This pattern suggests that interannual variability in the Kuroshio Current and Extension $Q_H^{COADS}$ dominates $Q_H^{COADS}$ variability basin-wide.

$Q_E^{COADS}$ PC1 accounts for 19.2% of the total variance in $Q_E^{COADS}$ and is shown with its associated time series of amplitude in Figure 4.22b. $Q_E^{COADS}$ PC1 is very similar to (and can be interpreted the same as) $Q_H^{COADS}$ PC1 except the region of high (or low) $Q_E^{COADS}$ (where $\alpha > +0.6$) extends further east past 180°W, and the region of low (or high) $Q_E^{COADS}$ (where $\alpha < -0.6$) is much smaller, and is located further east along the west coast of USA. This suggests that interannual variability in the Kuroshio Current and Extension $Q_E^{COADS}$ dominates $Q_E^{COADS}$ variability basin-wide.

Application of the scree test to the $Q_H^{GCM}$ and $Q_E^{GCM}$ eigenvalue series suggested the leading 2 PCs of both $Q_H^{GCM}$ and $Q_E^{GCM}$ contain non-random signals. $Q_H^{GCM}$ PC1 explains 17.6% of the total variance in $Q_H^{GCM}$, and is shown with its associated time series of amplitude in Figure 4.23a. $Q_H^{GCM}$ PC1 shows the dominant mode of interannual variation in winter-time North Pacific Ocean $Q_H^{GCM}$ is between years where $Q_H^{GCM}$ is high over the Northeast (20°N to 40°N) and far Northwest (40°N to 55°N) and low over the Northwest (Kuroshio Current and Extension, 20°N to 40°N) , and years for which the opposite pattern occurs.

$Q_E^{GCM}$ PC1 accounts for 17.3% of the total variance in $Q_E^{GCM}$, and is shown with its associated time series of amplitude in Figure 4.23b. $Q_E^{GCM}$ PC1 is very similar to (and can be interpreted the same as) $Q_H^{GCM}$ PC1 except: (1) the region of low (or high) $Q_E^{GCM}$ (the
Figure 4.22: First VARIMAX rotated principal components (PC1s) of $Q_H^{\text{COADS}}$ and $Q_E^{\text{COADS}}$, their percentage variance explained, and their associated time series of amplitude over the period 1950–89. Contour interval of the PCs is 0.2.
Figure 4.23: First VARIMAX rotated principal components (PC1s) of $Q_H^{\text{GCM}}$ and $Q_E^{\text{GCM}}$, their percentage variance explained, and their associated time series of amplitude over 10 years of simulation. Contour interval of the PCs is 0.2.
Kuroshio Current and Extension where $\alpha < -0.6$) extends further east than for $Q_E^{GCM}$ PC1 although the maximum value is smaller; and (2) the region of high (or low) $Q_E^{GCM}$ to the far northwest (where $\alpha > +0.6$) covers a greater area, extends further east, and has a higher maximum value. This suggests that interannual variability in the Kuroshio $Q_E^{GCM}$ may dominate $Q_E^{GCM}$ basin-wide, but is also strongly linked to variability in $Q_E^{GCM}$ north of 40°N.

$Q_H^{GCM}$ PC1 and $Q_E^{GCM}$ PC1 look very different than their COADS counterparts but they do share similarities. As noted earlier, $Q_H^{COADS}$ PC1 has a “tongue” of positive coefficients extending east from south of Japan past 180°W representing the Kuroshio Current effect, in $Q_H^{GCM}$ PC1 this tongue also appears but is split along 40°N with positive coefficients to the north and negative coefficients to the south. This bifurcation extends across to the eastern side of the basin. The Kuroshio tongue that extends further east past 160°W in $Q_E^{COADS}$ PC1 occurs in $Q_E^{GCM}$ PC1 as well, but again the GCM pattern is split along 40°N. The reason for this bifurcation in the GCM PCs along 40°N is not clear but this greater complexity in the GCM PCs may be a result of a smaller signal-to-noise ratio in $Q_H^{GCM}$ and $Q_E^{GCM}$ compared to $Q_H^{COADS}$ and $Q_E^{COADS}$ due to the fewer number of years analysed.

$Q_H^{COADS}$ PC2 accounts for 14.4% of the total variance and shows large positive $\alpha$s (greater than +0.6) over the central Northern Pacific (35°N – 50°N, 175°E – 160°W) and largest negative $\alpha$s (less than –0.6) off the west coast of USA (Figure 4.24a). This means that after removing the most dominant source of interannual variation in $Q_H^{COADS}$, the next major source of variation is a contrast between those years with a high $Q_H^{COADS}$ over the central Northern Pacific and a low $Q_H^{COADS}$ over the eastern boundary of the North Pacific, and years with the opposite tendency.

$Q_E^{COADS}$ PC2 (Figure 4.24b) accounts for 13.6% of the total variance and is very different from $Q_H^{COADS}$ PC2. The $\alpha$s are almost entirely positive with largest values
(\(\alpha > +0.6\)) in the southeast of the domain (in the region approximately bounded by
removing the dominant source of interannual variation in \(Q_E^{COADS}\), the next major source
of variability is a contrast between those years with high \(Q_E^{COADS}\) over the southeast part
of the domain, and years when \(Q_E^{COADS}\) over this area is low.

\(Q_H^{GCM}\) PC2 accounts for 16.2% of the total variance and shows large positive coeffi-
cients in the central North Pacific and over the Northwest Pacific south of Japan (Figure
4.25a). This means that after removing the first dominant source of variation in \(Q_H^{GCM}\,
the next major source of variation is a contrast between those years in which \(Q_H^{GCM}\) was
high over the central North Pacific and Northwest Pacific south of Japan and low over
the rest of the basin, and years with the opposite tendency.

\(Q_E^{GCM}\) PC2 accounts for 16.2% of the total variance and has large positive coefficients
in the central North Pacific and over the Northwest Pacific southwest of Japan, with
moderate negative values in between (Figure 4.25b). This means that after removing
the first dominant source of variation in \(Q_E^{GCM}\), the next major source of variation is a contrast between years in which \(Q_E^{GCM}\) was
high over the central North Pacific and southwest of Japan and low in between, and those years when the opposite case was
true.

The GCM and COADS \(Q_H\) PC2 and \(Q_E\) PC2 are quite different but share some
similarities. The area of highest loading of \(Q_H^{COADS}\) PC2 is centered at about 42°N
170°W, for \(Q_H^{GCM}\) PC2 it is about 20° south of the \(Q_H^{COADS}\) PC2 maximum, as well as
over KCR. Maximum loading of both \(Q_E^{COADS}\) PC2 and \(Q_E^{GCM}\) PC2 occurs 10° east of
NPR, thus all of the PC2s primarily reflect variability near NPR, while all PC1s are
dominated by the Kuroshio Current signal.

Correlation coefficients between the PC time series of amplitudes were computed and
shown in Table 4.3. There is a strong relation between the primary modes of variation of
Chapter 4. Data Analysis And Comparison

(a) $Q_{HC}^{COADS}$ PC2 14.4%

(b) $Q_{EC}^{COADS}$ PC2 13.6%

Figure 4.24: Second VARIMAX rotated principal components (PC2s) of $Q_{HC}^{COADS}$ and $Q_{EC}^{COADS}$, their percentage variance explained, and their associated time series of amplitude over the period 1950–89. Contour interval of the PCs is 0.2.
Chapter 4. Data Analysis And Comparison

(a) $Q_{HC}^{GCM}$ PC2 16.2%

(b) $Q_{EC}^{GCM}$ PC2 16.3%

Figure 4.25: Second VARIMAX rotated principal components (PC2s) of $Q_{HC}^{GCM}$ and $Q_{EC}^{GCM}$, their percentage variance explained, and their associated time series of amplitude over 10 years of simulation. Contour interval of the PCs is 0.2.
(a) COADS PCs

<table>
<thead>
<tr>
<th>(Q_{E}^{\text{COADS}}) PC1</th>
<th>(Q_{H}^{\text{COADS}}) PC1</th>
<th>(Q_{E}^{\text{COADS}}) PC2</th>
<th>(Q_{H}^{\text{COADS}}) PC2</th>
</tr>
</thead>
<tbody>
<tr>
<td>(+0.78^*)</td>
<td>NS</td>
<td>(-0.42^*)</td>
<td></td>
</tr>
<tr>
<td>(+0.55^*)</td>
<td>NS</td>
<td>(+0.29)</td>
<td></td>
</tr>
</tbody>
</table>

(b) GCM PCs

<table>
<thead>
<tr>
<th>(Q_{E}^{\text{GCM}}) PC1</th>
<th>(Q_{H}^{\text{GCM}}) PC1</th>
<th>(Q_{E}^{\text{GCM}}) PC2</th>
<th>(Q_{H}^{\text{GCM}}) PC2</th>
</tr>
</thead>
<tbody>
<tr>
<td>(+0.78^*)</td>
<td>NS</td>
<td>(+0.57)</td>
<td></td>
</tr>
<tr>
<td>(-0.59)</td>
<td>NS</td>
<td>(+0.72^*)</td>
<td></td>
</tr>
</tbody>
</table>

Table 4.3: Correlation coefficients between the PC time series of amplitudes. Values significant at the 95% level are denoted by a *. Values not significant at the 90% level are denoted by NS. (a) COADS PCs. The 90% and 95% significance levels are \(r = \pm 0.26\) and \(r = \pm 0.31\), respectively (Student's 2-tailed t-test with 38 degrees of freedom). (b) GCM PCs. The 90% and 95% significance levels are \(r = \pm 0.55\) and \(r = \pm 0.63\), respectively (Student's 2-tailed t-test with 8 degrees of freedom).

\(Q_{H}^{\text{COADS}}\) and \(Q_{E}^{\text{COADS}}\) (\(Q_{H}^{\text{COADS}}\) PC1 and \(Q_{E}^{\text{COADS}}\) PC1 are correlated with \(r = +0.78\)) as well as weaker covariability between the primary mode of variation of one COADS flux and the secondary mode of variation of the other COADS flux (Table 4.3a). Correlation coefficients between the GCM PC time series of amplitudes (Table 4.3b) suggest there is a strong relation between the primary modes of variation of \(Q_{H}^{\text{GCM}}\) and \(Q_{E}^{\text{GCM}}\) (\(Q_{H}^{\text{GCM}}\) PC1 and \(Q_{E}^{\text{GCM}}\) PC1 are correlated with \(r = +0.78\)) and the secondary modes as well (\(Q_{H}^{\text{GCM}}\) PC2 and \(Q_{E}^{\text{GCM}}\) PC2 are correlated with \(r = +0.72\)). There is weaker covariability between the primary mode of variation of one GCM flux and the secondary mode of variation of the other GCM flux.

The relation between the dominant modes of heat flux variability and atmospheric circulation anomalies are considered by examining the significant correlations between the time series of PC amplitudes and atmospheric anomaly fields. Figure 4.26 shows
correlations between the time amplitudes of $Q_H^\text{COADS}$ PC1 and $Q_E^\text{COADS}$ PC1 and SLP$^\text{NMC}$ and Z500$^\text{NMC}$ fields. The correlation patterns between $Q_H^\text{COADS}$ PC1 and SLP$^\text{NMC}$ and between $Q_E^\text{COADS}$ PC1 and SLP$^\text{NMC}$ are very similar (Figures 4.26a and 4.26b), with larger values for $Q_E^\text{COADS}$ PC1 than $Q_H^\text{COADS}$ PC1 (except over Western North America). Maximum values of greater than $-0.7$ occur in the Northwest Pacific just northeast of KCR, but because the $Q_E^\text{COADS}$ PC1 “tongue” (Figure 4.22b) extends further east than that of $Q_H^\text{COADS}$ PC1 (Figure 4.22a), the region of maximum correlation for $Q_E^\text{COADS}$ PC1 is located further east of the maximum for $Q_H^\text{COADS}$ PC1. There are also regions with negative values greater than $-0.5$ over Northeast North America, and positive values over China (greater than $+0.3$ for $Q_H^\text{COADS}$ PC1 and greater than $+0.5$ for $Q_E^\text{COADS}$ PC1), over Northeastern Russia (greater than $+0.5$ for $Q_H^\text{COADS}$ PC1 and $Q_E^\text{COADS}$ PC1), and over North America (greater than $+0.5$ for $Q_H^\text{COADS}$ PC1 and greater than $+0.4$ for $Q_E^\text{COADS}$ PC1), and weaker values over the UK (greater than $+0.3$) and Atlantic Ocean (greater than $+0.3$).

The correlation patterns between $Q_H^\text{COADS}$ PC1 and Z500$^\text{NMC}$ and between $Q_E^\text{COADS}$ PC1 and Z500$^\text{NMC}$ are also very similar. Maximum values of greater than $-0.8$ are centered just northeast of KCR but extend westward over Northern India and eastward almost to the west coast North America. There are strong positive correlations of greater than $-0.6$ over Northeast Asia and the west coast of North America.

The correlation patterns of $Q_H^\text{COADS}$ and $Q_E^\text{COADS}$ PC1 in Figure 4.26 are almost identical (and can be interpreted similarly) to those of KCR (Figure 4.18). This means that at the very least $Q_H^\text{COADS}$ PC1 and $Q_E^\text{COADS}$ PC1 are in phase with $Q_H^\text{COADS}$ and $Q_E^\text{COADS}$ over KCR, and it is most likely that basin-wide variability in $Q_H^\text{COADS}$ and $Q_E^\text{COADS}$ is dominated by $Q_H^\text{COADS}$ and $Q_E^\text{COADS}$ over KCR. There are two differences between Figures 4.26 and 4.18 worth noting: (1) the negative centers over the North Pacific are further east for $Q_E^\text{COADS}$ PC1 than KCR $Q_E^\text{COADS}$ because the Kuroshio “tongue” extends further
Figure 4.26: Correlations between COADS heat flux PC1 amplitude time series and NMC atmospheric anomalies at individual grid points over the Northern Hemisphere (20°N – 90°N) over the 1950–89 period. The 95% significance level is \( r = \pm 0.31 \) (Student’s 2-tailed t-test with 38 degrees of freedom). Only correlations where \(|r| \geq 0.3\) are shown. Contour interval is 0.1.
Figure 4.27: Correlations between COADS heat flux PC2 amplitude time series and NMC atmospheric anomalies at individual grid points over the Northern Hemisphere (20°N – 90°N) over the 1950–89 period. The 95% significance level is $r = \pm 0.31$ (Student’s 2-tailed t-test with 38 degrees of freedom). Only correlations where $|r| \geq 0.3$ are shown. Contour interval is 0.1.
east in $Q_E^{COADS}$ PC1 (Figure 4.22b) than as defined in Figure 3.1; and (2) the correlations are generally stronger for $Q_H^{COADS}$ PC1 and $Q_E^{COADS}$ PC1 than KCR $Q_H^{COADS}$ and KCR $Q_E^{COADS}$ for both the positive and negative nodes over North America, and significant negative correlations do not exist over Eastern North America for KCR $Q_H^{COADS}$ and KCR $Q_E^{COADS}$ correlations at 500 mb. This suggests KCR $Q_H^{COADS}$ and KCR $Q_E^{COADS}$ exert a stronger local atmospheric response but the leading modes of basin-wide variability in North Pacific $Q_H^{COADS}$ and $Q_E^{COADS}$ show a more robust signature over the Northern Hemisphere, which is what one would expect.

Figure 4.26 is very similar to the NCAR CCM response to a warm North Pacific sea surface temperature anomaly (Pitcher et al. 1988). The Z500$^{NMC}$ patterns (Figure 4.26c and 4.26d) are very similar to the WPO and WP pattern of 500 mb height variability (Barnston and Livezey 1987, Wallace and Gutzler 1981), although the time series of $Q_H^{COADS}$ PC1 and $Q_E^{COADS}$ PC1 are not significantly correlated with the WP pattern index.

Figure 4.27 shows correlations between the time amplitudes of PC2 of $Q_H^{COADS}$ and $Q_E^{COADS}$ and the NMC atmospheric anomaly fields. The two correlation patterns involving $Q_H^{COADS}$ PC2 are similar as are the two patterns involving $Q_E^{COADS}$ PC2. (This is in contrast to the PC1 patterns which showed similarities between both SLP$^{NMC}$ patterns and both Z500$^{NMC}$ patterns.) Similar features in the correlation patterns between $Q_H^{COADS}$ PC2 and SLP$^{NMC}$ and between $Q_H^{COADS}$ PC2 and Z500$^{NMC}$ consist of a large region of negative values over the North Pacific with a maximum of greater than −0.85, and a region of weak negative correlations of greater than −0.4 off the southeast coast of USA. The correlation pattern between $Q_H^{COADS}$ PC2 and Z500$^{NMC}$ also has regions of moderate positive correlations (greater than +0.6) over North America and the subtropical North Pacific. This wave-like feature over the Eastern Pacific and North America in Figure 4.27c resembles the PNA variability pattern identified in analyses of the observed
and modelled 500 mb height field (Wallace and Gutzler 1981, Lau and Nath 1990), and in cross-correlations between observed sea surface temperature anomalies at 32°N 165°W and 500 mb height anomalies (Wallace and Jiang 1987).

The correlation patterns between \( Q_{E}^{COADS} \) PC2 and \( SLP^{NMC} \) and between \( Q_{E}^{COADS} \) PC2 and \( Z500^{NMC} \) show weak negative values (-0.3 to -0.5) along a broad band over the western, southern, and eastern coastal waters of North America. For \( Q_{E}^{COADS} \) PC2 and \( Z500^{NMC} \), this weak negative correlation extends over the north coast of Asia, and is accompanied by weak positive values (+0.3 to +0.4) over the Northwest Pacific, North America, and Northwest Africa. The correlation pattern between \( Q_{E}^{COADS} \) PC2 and \( SLP^{NMC} \) also includes a broad band of weak positive correlations between 20°N and 30°N stretching from 180°W to 30°E (which seem to represent noise in the \( Q_{E}^{COADS} \) PC2 signal).

The \( Q_{H}^{COADS} \) PC2 and \( Z500^{NMC} \) pattern (Figure 4.27c) exhibits the strongest and most coherent PNA pattern although all patterns emphasize at least one PNA node. \( Q_{H}^{COADS} \) PC2 is most strongly correlated with the PNA index (\( r = +0.75 \)) although \( Q_{E}^{COADS} \) PC1 and \( Q_{E}^{COADS} \) PC2 are also significantly correlated the PNA index (\( r = +0.57 \) and \( r = +0.40 \)). The \( Q_{E}^{COADS} \) PC2 and \( Z500^{NMC} \) pattern (Figure 4.27d), however, is much more similar to the pattern associated with sea surface temperatures averaged over the El Niño “core region” from 180°W to the South American coast between 6°N and 6°S (Wallace and Jiang 1987). \( Q_{E}^{COADS} \) PC2 is moderately correlated to \( T_{s}^{COADS} \) over this region (\( r = +0.60 \)), but is even more strongly correlated to \( Q_{E}^{COADS} \) over this region (\( r = +0.69 \)). This emphasizes the strong link between the tropical Pacific and the secondary mode of heat flux variability in the extratropical Pacific.

The signature of this secondary mode of heat flux variability on the atmospheric circulation shares weak similarities to that of NPR (Figure 4.20) in that all patterns in both figures have regions of significant negative correlations over the Northeast Pacific.
Chapter 4. Data Analysis And Comparison

and Western Atlantic, and patterns at 500 mb all have significant positive correlations over North America. The $Q_{COADS}^H$ PC2 correlation values are much stronger than those of NPR $Q_{COADS}^H$. The similarities (and differences) in the patterns can be partially explained by realizing $Q_{COADS}^H$ PC2 and $Q_{COADS}^E$ PC2 deal with variability centered on regions close to but different than NPR. (Compare the areas of highest loading in $Q_{COADS}^H$ PC2 and $Q_{COADS}^E$ PC2 in Figure 4.24 with the location of NPR as defined in Figure 3.1.)

The GCM patterns are quite different. Correlations between $Q_{GCM}^H$ PC1 and $Q_{GCM}^E$ PC1 and SLP$^{GCM}$ and Z500$^{GCM}$ are shown in Figure 4.28. All correlation patterns in this figure are similar with the dominant feature being a north-south dipole pattern of a region of strong negative correlations (greater than -0.7) over the Bering Sea at 60°N and a region of strong positive correlations (greater than +0.6) directly south centered at 30°N. The similarity between the correlation patterns of $Q_{GCM}^H$ PC1 and $Q_{GCM}^E$ PC1 are a result of similar primary modes of variability of the two fields (Figure 4.23) which also have strong positive-negative north-south dipole patterns over the Northwest Pacific where the surface heat flux is greatest. If this correlation pattern primarily reflects variability in $Q_{GCM}^H$ and $Q_{GCM}^E$ over the Northwest Pacific, it appears the atmosphere is responding in the sense we would expect: trough deepening at 30°N downstream of the Kuroshio Current heat source (reflected in positive correlations with the negative PC1 values over the Kuroshio Current and Extension), and trough weakening at 60°N downstream of the Asian continent heat sink (negative correlations). This dipole pattern has been identified as a primary mode of variability in the observed 500 mb height and corresponding SLP fields and referred to as the Western Pacific (WP) pattern (Wallace and Gutzler 1981) and the West Pacific Oscillation (WPO) (Barnston and Livezey 1987).

It is not evident, however, in the responses of GCMs to warm sea surface temperature anomalies over the North Pacific (Kushnir and Lau 1992, Pitcher et al. 1988). It also bears no relation to the correlation patterns between KCR $Q_{GCM}^H$ and KCR $Q_{GCM}^E$ and
Figure 4.28: Correlations between GCM heat flux PC1 amplitude time series and atmospheric anomalies over 10 years of simulation. The 95% significance level is $r = \pm 0.63$ (Student's 2-tailed t-test with 8 degrees of freedom). Only correlations where $|r| \geq 0.6$ are shown. Contour interval is 0.1.
Figure 4.29: Correlations between GCM heat flux PC2 amplitude time series and atmospheric anomalies over 10 years of simulation. The 95% significance level is $r = \pm 0.63$ (Student’s 2-tailed t-test with 8 degrees of freedom). Only correlations where $|r| \geq 0.6$ are shown. Contour interval is 0.1.
SLP$^{GCM}$ and Z500$^{GCM}$ (Figure 4.19).

The correlation patterns between $Q_H^{GCM}$ PC2 and $Q_E^{GCM}$ PC2 and SLP$^{GCM}$ and Z500$^{GCM}$ in Figure 4.29 all show regions of negative values over both the Northeast and Northwest Pacific and appear similar to their corresponding correlation patterns of both KCR and NPR (Figures 4.19 and 4.21). This is reasonable given $Q_H^{GCM}$ PC2 and $Q_E^{GCM}$ PC2 are strongly positive over both KCR and NPR (i.e., heat flux variations over these regions are in phase). Upon close examination of these figures it appears the correlation pattern of $Q_H^{GCM}$ PC2 more closely resembles that of KCR $Q_H^{GCM}$, while $Q_E^{GCM}$ PC2 is most similar to that of NPR $Q_E^{GCM}$. This is because $Q_H^{GCM}$ PC2 loads heaviest over KCR while $Q_E^{GCM}$ PC2 loads heaviest over NPR. Figure 4.29 can be interpreted similarly to Figures 4.19 and 4.21: the most prominent features coincident with simultaneous positive heat flux anomalies over both regions are negative anomalies in both SLP$^{GCM}$ and Z500$^{GCM}$ locally and remotely over the Northwest Atlantic.

The GCM and COADS/NMC PC time series and SLP' and Z500' correlations are not strictly comparable since the PCs themselves are so different. But the PC/atmosphere correlations all reflect the atmospheric circulation associated with the dominant modes of heat flux variability. In the case of all PC1 correlation patterns (which reflect Kuroshio Current heat flux variability), a common feature is the large value centered around 30°N to 40°N and 160°E to 180°E which is negative in the COADS patterns and positive in the GCM patterns. They both represent negative atmospheric anomalies (trough deepening with respect to the mean SLP and 500 mb fields in Figures 4.8 and 4.9) in response to positive Kuroshio Current heating anomalies.

The GCM and COADS/NMC PC2 correlation patterns (Figures 4.27 and 4.29) which represent the atmospheric relation to heat flux variability near NPR show little similarity except all patterns show weak to strong negative values (trough deepening) just east of the PC2 centers of highest loading (corresponding to positive heating anomalies).
Most patterns also show weak to moderate negative values (trough deepening) over the Northwest Atlantic Ocean.
Chapter 5

Summary and Discussion

Interannual variability in North Pacific Ocean (20°N – 56°N) winter-time surface sensible and latent heat fluxes ($Q_H$ and $Q_E$) and their relation to Northern Hemisphere sea-level pressure (SLP) and 500 mb height ($Z_{500}$) fields has been explored. Data based on observations (COADS/NMC) over the 1950–89 period and 10 years of model (CCC GCM) simulations were analysed and compared. Both data sets possess characteristics that limit their direct comparison, but comparison does offer insight into both representations of the Northern Hemisphere climate system. The $Q_H^{COADS}$ and $Q_E^{COADS}$ fluxes contain maximum errors of 20 and 50 W m$^{-2}$, respectively, and display linear trends which averaged over the 1950–89 period and basin-wide amount to increases of +1.4 and +17.1 W m$^{-2}$, respectively. The $Q_H^{GCM}$ and $Q_E^{GCM}$ fluxes are constrained by a highly simplified oceanic model which consists of a 50 metre thick slab of quiescent sea water. This oceanic mixed-layer model contains a correction factor which ensures that the model simulates the average annual climatological cycle of sea surface temperatures.

The long-term mean COADS and GCM $Q_H$ and $Q_E$ fields agree within the maximum errors in the COADS fluxes except over the Kuroshio Current region (KCR) where $Q_E^{GCM}$ is 100 W m$^{-2}$ less than $Q_E^{COADS}$. Maximum values in both COADS and GCM $Q_H$ and $Q_E$ occur over KCR, and minimum values occur over the Northeast Pacific. This largely reflects the $\Delta T$ and $\Delta q$ fields, and to a much lesser degree, $\vec{U}$. The $Q_H$ and $Q_E$ fields are strongly correlated in time over the North Pacific Ocean ($r \approx +0.8$) with values of $r > +0.9$ over KCR as a result of the strong relation between $\Delta T$ and
Chapter 5. Summary and Discussion

$\Delta q$. Bowen ratios are lowest (0.1) over the Northeastern Pacific near 20°N, and increase fairly uniformly to 1.0 in the northwest.

The long-term mean SLP$^{{GCM}}$ and Z$^{500}_{0}^{{GCM}}$ fields reproduce the main climatological features of the NMC analyses: the Asian High, Aleutian Low, and the North Atlantic Low in the SLP field; and the troughs over Eastern Asia and Eastern North America in the Z$^{500}_{0}$ field. The GCM pressures and heights are generally lower than their NMC counterparts: the Aleutian Low and North Atlantic Low are too intense; and the ridge over Western Canada is weaker. Variability in the GCM and NMC fields as represented by their standard deviations is in good agreement with maximum values in the vicinity of the Aleutian and North Atlantic Lows, and over the north coast of Russia.

COADS heat flux anomalies are roughly twice as large as their GCM counterparts, with largest values of $|Q'_H|^{COADS}$ and $|Q'_H|^{GCM}$ over KCR (15 to 20 W m$^{-2}$ and 6 to 8 W m$^{-2}$, respectively), and minima (less than 4 W m$^{-2}$) over the southeast of the North Pacific study area. Maximum values of $|Q'_E|^{COADS}$ occur over the Kuroshio Extension and south of KCR (30 to 40 W m$^{-2}$) where $|Q'_E|^{GCM}$ is also largest (18 to 20 W m$^{-2}$), and minimum values occur basin-wide north of 45°N and off the west coast of USA (less than 15 W m$^{-2}$ and 6 W m$^{-2}$, respectively). These patterns roughly correspond to those of $|\Delta T'|$ and $|\Delta q'|$ which implies $Q'_H$ and $Q'_E$ are strongly related to the gradient anomalies. The heat flux anomalies are much more strongly correlated to the gradient anomalies than to $U'$, except over KCR and off the west coast of USA where correlations are stronger between $Q'_E^{COADS}$ and $U'^{COADS}$, and between $Q'_E^{GCM}$ and $U'^{GCM}$, respectively.

The covariations between $Q'_H$ and $Q'_E$ over KCR and the central North Pacific region (NPR) and SLP$'$ and Z$^{500}_{0}'$ in the COADS/NMC and GCM fields are markedly different, but show some similarities. Strongest correlations between SLP$^{NMC}$ and KCR $Q'_H^{COADS}$ and KCR $Q'_E^{COADS}$ are negative and 25° east of KCR, with moderate positive
correlations over Eastern Asia and Western North America, and weak negative correlations over Northeastern North America. This pattern corresponds to strengthening of the ridges over Eastern Asia and Western North America, and deepening of the trough over the Northwest Pacific (and Northwest Atlantic) in conjunction with positive heat flux anomalies. The pattern is similar to the modelling results of Weaver (1987) and Pitcher et al. (1988). The corresponding GCM patterns are quite different: the strongest correlations between SLPGCM and KCR $Q_H^{GCM}$ and KCR $Q_E^{GCM}$ are negative and 70° east of KCR, with weaker negative correlations over Northeastern North America/Northwestern Atlantic Ocean and Northwest Africa, and moderate positive correlations occur over the Arabian Sea/Northern India. The main atmospheric features associated with this pattern are deeper troughs over the Northeast Pacific and Northwest Atlantic in conjunction with positive heat flux anomalies.

In the COADS/NMC fields, the strongest correlations between Z500NMC and KCR $Q_H^{COADS}$ and KCR $Q_E^{COADS}$ are negative and centered 10° east of KCR, while there are moderate positive values over Northeastern Russia and Western North America. This corresponds to strengthening of the ridge over Eastern Asia and Western North America, a deeper trough over the Northwest Pacific, and a stronger Asian Jet in association with positive KCR heat flux anomalies. This pattern is similar to a dominant mode of variability observed in 500 mb height anomalies referred to as the western Pacific (WP) or West Pacific Oscillation (WPO) patterns (Wallace and Gutzler 1981, Barnston and Livezey 1987), and the GCM response to a warm North Pacific sea surface temperature anomaly (Pitcher et al. 1988). The strongest correlations in the GCM fields are local and negative, corresponding to trough deepening over the Northwest Pacific in association with warm KCR heat flux anomalies.

In general, the correlations between both COADS/NMC and GCM heat fluxes over NPR and atmospheric anomalies are weaker than those for KCR. NPR $Q_H^{COADS}$ and
Chapter 5. Summary and Discussion

$Q_E^{\text{COADS}}$ correlations with $\text{SLP}^{\text{NMC}}$ and $Z500^{\text{NMC}}$ are weakly negative over the Northeast Pacific and Northwest Atlantic Oceans, and weakly positive over Northern North America. The larger $Q_E^{\text{COADS}}$ correlations implies that the strong relation between observed sea surface temperature anomalies in this region and atmospheric anomalies (Namias and Born 1972, Rasmusson and Carpenter 1982, Wallace and Jiang 1987) is based much more on $Q'_E$ than $Q'_H$. This may be a result of the $Q_H$ anomalies being trapped within the marine boundary layer while the influence of the larger $Q_E$ anomalies is transported higher into the troposphere via convective cloud development during the passage of cyclones.

Regardless of the mechanism, the effect is for positive NPR $Q_E^{\text{COADS}}$ anomalies to be associated with stronger climatological features at 500 mb: the troughs over the Northeast Pacific and Eastern North America are deeper, the jets are strengthened, and the ridge over Western North America is stronger. However, positive NPR $Q_E^{\text{COADS}}$ anomalies occur in conjunction with a weaker ridge over North America, a weaker North Atlantic Low, and a weaker North American Jet.

The strongest correlation pattern (between NPR $Q_E^{\text{COADS}}$ and $Z500^{\text{NMC}}$) resembles a major mode of variability observed in the 500 mb height field identified as the Pacific North America (PNA) pattern (Wallace and Gutzler 1981, Barnston and Livezey 1987), but NPR $Q_E^{\text{COADS}}$ is only weakly correlated with the PNA index. The PNA pattern is also evident to some extent in the responses of GCMs to warm sea surface temperature anomalies over this region (Kushnir and Lau 1992, Pitcher et al. 1988). Correlations between NPR $Q_E^{\text{GCM}}$ and $Q_E^{\text{GCM}}$ and SLP$^{\text{GCM}}$ and $Z500^{\text{GCM}}$ only weakly resemble the COADS/NMC patterns and GCM studies quoted above. They show strong negative values east of NPR over the Eastern Pacific and positive values over Northern North America, corresponding to trough deepening over the Northeast Pacific and ridge strengthening over North America.

Principal component analysis (PCA) revealed that the 2 leading modes of variation
in North Pacific Ocean COADS and GCM $Q_H$ and $Q_E$ fields are strongly associated with variability over KCR and NPR, respectively. The first rotated principal components (PCs) of the COADS and GCM $Q_H$ and $Q_E$ reflect the dominant mode of variability of the heat fluxes and appear to have captured the signal of the Kuroshio Current and Kuroshio Extension. Both $Q_H^{COADS}$ PC1 and $Q_E^{COADS}$ PC1 explain 19% of the total variance and are characterized by a "tongue" of large positive coefficients extending eastward from south of Japan (over KCR) past 180°W (past 160°W for $Q_E^{COADS}$ PC1), and equally large negative coefficients over the eastern part of the basin. This means that the dominant mode of interannual variation in winter-time $Q_H^{COADS}$ and $Q_E^{COADS}$ is between years when the heat fluxes are large over the Kuroshio Current and Kuroshio Extension and small over the Northeast Pacific, and years when the opposite situation occurs. This is essentially the same interpretation for $Q_H^{GCM}$ PC1 and $Q_E^{GCM}$ PC1 (which explain 18% and 17% of the total variance), except the GCM pattern is split along 40°N. This more complicated pattern may be due to the smaller signal-to-noise ratio in the shorter GCM time series and the simplified representation of the ocean.

The second rotated PCs of the COADS and GCM $Q_H$ and $Q_E$ reflect the second largest source of interannual variation in the heat fluxes and are associated with variation at (or near) NPR. $Q_H^{COADS}$ PC2 and $Q_E^{COADS}$ PC2 explain 14% of the total variance each, and are characterized by large positive coefficients just to the north and east of NPR. $Q_H^{COADS}$ PC2 also has large negative coefficients off the west coast of USA. This means that the second largest source of interannual variation in $Q_H^{COADS}$ is between years when $Q_H^{COADS}$ is high north of NPR and low off the west coast of USA, and years with the opposite tendency. For $Q_E^{COADS}$ it is between years when $Q_E^{COADS}$ is high east of NPR, and years when $Q_E^{COADS}$ over this area is low. The interpretation of $Q_H^{GCM}$ PC2 and $Q_E^{GCM}$ PC2 which explain 16% of the total variance is somewhat similar. The PC2's load heaviest over NPR as well as south of Japan over KCR. This shows the GCM heat fluxes
over NPR vary in phase with those over part of KCR. Again, the difference between the GCM and COADS PC2s may reflect the different lengths of the time series and the simplified oceanic component of the GCM.

Correlations between the time series of amplitudes of $Q^\text{COADS}_H$ PC1 and $Q^\text{COADS}_E$ PC1 and $\text{SLP}^\text{NMC}$ and $Z500^\text{NMC}$ are almost identical to those of KCR $Q^\text{COADS}_H$ and KCR $Q^\text{COADS}_E$, and the associated atmospheric perturbations can be similarly interpreted. Two differences are worth noting: (1) the atmospheric response is weaker over KCR, but stronger over the Northern Hemisphere, because the PC pattern represents forcing over a larger scale; and (2) the strong negative correlations over the North Pacific are further east for $Q^\text{COADS}_E$ PC1 than KCR $Q^\text{COADS}_E$ because the Kuroshio tongue extends further east in $Q^\text{COADS}_E$ PC1. The similarity between KCR and PC1 patterns suggest that at the very least, KCR heat fluxes are in phase with the dominant mode of basin-wide heat flux variability, or probably that basin-wide heat flux variability is dominated by KCR $Q^\text{COADS}_H$ and $Q^\text{COADS}_E$.

There is some similarity between the COADS PC1 correlation patterns and those of the corresponding GCM fields which consist of a north-south dipole of negative values over KCR and/or the Kuroshio Extension, and positive values over the Bering Sea in conjunction with positive heating anomalies over the Kuroshio Current. This pattern also occurs over the Western Pacific in both the COADS PC1 and COADS KCR correlation patterns. The GCM PC1 correlation patterns resemble the WP/WPO modes of the Z500 and associated SLP fields, although they are not significantly correlated with the WP index. A WP-like pattern, however, is not evident in the responses of GCMs to warm sea surface temperatures over the North Pacific (Kushnir and Lau 1992, Pitcher et al. 1988).

The correlation patterns between the time series of amplitudes of $Q^\text{COADS}_H$ PC2 and $Q^\text{COADS}_E$ PC2 and $\text{SLP}^\text{NMC}$ and $Z500^\text{NMC}$ are all quite different, although they all have
regions of negative values over the Northeast Pacific and off the southeast coast of USA (similar to but generally stronger than the NPR correlations). Most notable are the Z500$^{\text{NMC}}$ patterns. For $Q^\text{COADS}_H$PC2 the Z500$^{\text{NMC}}$ correlation pattern also includes moderate positive values over Western North America and over the tropical Pacific, and it is strongly correlated with the PNA pattern identified in observed and modelled 500 mb height fields (Wallace and Gutzler 1981, and Lau and Nath 1990), and in cross-correlations between observed sea surface temperature anomalies at 32°N 165°W and 500 mb height anomalies (Wallace and Jiang 1987). For the $Q^\text{COADS}_E$PC2 correlations with Z500$^{\text{NMC}}$, the negative values extend almost around North America with positive values in the center. This pattern is strongly correlated with sea surface temperature anomalies and $Q^\text{NMC}_E$ over the eastern tropical Pacific, or El Niño signal area (Wallace and Jiang 1987). This emphasizes the strong link between variations over the tropical Pacific and the secondary mode of heat flux variability over the extratropical Pacific.

All the correlation patterns between $Q^\text{GCM}_H$PC2 and $Q^\text{GCM}_E$PC2 and SLP$^\text{GCM}$ and Z500$^\text{GCM}$ are similar with negative values over the Northwest and Northeast Pacific near KCR and NPR (and resemble the corresponding correlation patterns between the atmospheric circulation and both KCR and NPR). This is because this mode of variability consists of variations of $Q^\text{GCM}_H$ and $Q^\text{GCM}_E$ in phase near KCR and NPR as shown by $Q^\text{GCM}_H$ PC2 and $Q^\text{GCM}_E$ PC2. The PC2 correlation patterns resemble those of the regions over which they load heaviest: $Q^\text{GCM}_H$ PC2 is most similar to the correlation patterns of KCR $Q^\text{GCM}_H$; and $Q^\text{GCM}_E$ PC2 is most similar to $Q^\text{GCM}_E$.

In summary, we have shown that variability in the North Pacific Ocean sensible and latent heat fluxes as derived from COADS data are dominated by the heat fluxes over the Kuroshio Current region, and the second largest source of variation is associated with variability near the central North Pacific region. The former is associated with the WP/WPO mode of variability in the SLP and Z500 fields, and the latter with the
PNA variability mode in the Z500 field. Variability in the GCM sensible and latent heat fluxes also appears to be dominated by the heat fluxes over the Kuroshio Current region, but these are in phase with heat fluxes over the central North Pacific region, and the dominant modes of heat flux variability are more complicated. As a result, the dipolar atmospheric pattern associated with the dominant mode of heat flux variability is somewhat different than the atmospheric relation with variations over the Kuroshio Current region. However, the PC2/atmosphere correlation patterns are very similar to the NPR/atmosphere correlation patterns.

The GCM dipolar atmospheric pattern associated with the dominant source of heat flux variability is similar to the response of the GISS GCM to cool North Pacific Ocean sea surface temperature anomalies. We have considered the correlations between surface heat fluxes and the atmosphere, but concentrated discussion on the atmospheric anomalies associated with positive heat flux anomalies. Correlations of course are valid in both directions and by so considering them we have assumed the atmospheric relation to both warm and cold heat flux anomalies is equivalent but opposite. Although intuitive, this runs counter to the experiments of Kushnir and Lau (1992) and Pitcher et al. (1988) who found similar GCM responses to warm and cold North Pacific Ocean sea surface temperature anomalies. This suggests that both warm and cool surface forcing of the atmosphere have a similar effect, and that they should be examined separately with respect to the zero anomaly scenario. In the case of observational data analysis (and model data if a sufficient number of runs were available), composites of warm and cool heat flux anomaly cases compared with the mean may more clearly illustrate the distinct atmospheric relations to anomalous heat flux forcing. This is an objective of future work, and preliminary results from the COADS/NMC data show that the atmospheric anomalies associated with both strong positive and strong negative modes of KCR $Q_E$, $Q_E$ PC1, $Q_H$ PC1, and $Q_H$ PC2 are different, but not entirely opposite either. Some significant
nodes in the correlation patterns are due to either positive or negative heating anomalies, but not both. Also, in general the positive heating anomaly modes are associated with strong atmospheric anomalies upstream and weaker atmospheric anomalies downstream, and *vice versa* for negative heating anomalies. For example, the strong positive correlations between KCR $Q'_E$ and SLP' over Southeast Asia appear to be due to positive heat flux anomalies only, there are no significant SLP anomalies associated with negative KCR heat flux anomalies. This serves to illustrate the complexities of large scale air-sea interactions, and while simple correlation analyses as presented here are useful for identifying general patterns, more sophisticated methods may be necessary for obtaining more detail of these interactions.
Appendix A

Trends In Surface Marine Data

The identification of real trends in climatological data is very important for assessing climate change, but often the observed trends in climatological data may represent nothing more than changes in observing practices. More difficult to assess are the cases where data reflect changes in both observing practices and the variables themselves. This appendix deals with the nature of observed trends in North Pacific Ocean monthly mean surface wind speed, sea surface temperature, and surface air temperature over the 1950–89 period in the COADS data and attempts to determine whether the trends are real or artificial.

The mean and standard deviation of the observed wind speeds in winter (monthly average of December, January, and February) are shown in Figures A.1 and A.2. Linear changes in the wind speed were estimated by computing a linear least squares regression and multiplying the slope of the line of best fit by the length of the time series. Over the 1950–89 period the winds exhibit a linear increase everywhere in the North Pacific Ocean (Figure A.3) and averaged over the North Pacific Ocean study area (outlined in Figure A.3) have increased about 1.7 m s\(^{-1}\) (Figure A.4).

The significance of this trend is assessed in two ways. First, it can be compared to the temporal standard deviations at individual grid points. In other words, is the basin-averaged change over 40 years greater or less than the majority of annual anomalies
Figure A.1: Mean winter-time surface wind speed over the period 1950–89. Contour interval is 1.0 m s\(^{-1}\).

Figure A.2: Standard deviation of the mean winter-time surface wind speed over the period 1950–89. Contour interval is 0.2 m s\(^{-1}\).
at each grid point? Second, we can estimate the error in the linear trend by assuming: (1) a good linear fit to the data; and (2) equal errors in the individual observations/measurements equivalent to one standard deviation. The resultant combination of best linear fits to different combinations of the data points (within their assumed error) gives different slopes, the standard deviation of which is designated as the slope error. The slope error multiplied by the length of the time series is designated as the error in the linear change or error in the trend.

The linear increase in wind speed averaged over the North Pacific from 1950–89 of 1.7 m s\(^{-1}\) is greater than the standard deviation at most grid points (Figure A.2). Even when the 1950 data point (a large negative anomaly) is left out of the time series, the increase is 1.5 m s\(^{-1}\) which is still greater than the standard deviation at most grid points. The error in the linear change in wind speed ranges from 0.3 to 0.5 m s\(^{-1}\) over the North Pacific Ocean with an average of \(\sim 0.4\) m s\(^{-1}\). This suggests that the increasing trend in wind speed over the North Pacific Ocean is significantly different than zero. This trend may be artificial and merely reflect: (1) the change from estimates based on sea state or Beaufort Scale (100\% of observations in 1950) to anemometer measurements (50–80\% of observations in 1984) and their intercalibration (Isemer and Hasse 1991, Cardone et al. 1990); and (2) the increase in ship size which results in measurements made higher above the sea surface and more frequently in ocean storms since larger ships tend to avoid storms less.

The analyses of Cardone et al. (1990) and Posmentier et al. (1989) suggest that the increasing trend in wind speed over the Pacific Ocean is artificial. In a reexamination of individual extratropical Pacific wind reports Cardone et al. made corrections for measurement height greater than 10 m and the bias in the Beaufort Scale to wind speed conversion scheme (WMO code 1100) and found no increasing trend in wind speed. The WMO code 110 is biased low for wind speeds less than 15 m s\(^{-1}\) and biased high for wind
Figure A.3: Linear change in winter-time scalar wind speed over the period 1950–89. Contour interval is 1.0 m s\(^{-1}\). The North Pacific Ocean study area is outlined in bold.

Figure A.4: Time series of the winter-time scalar wind speed anomalies averaged over the North Pacific Ocean study area outlined in Figure A.3. Dotted line indicates zero anomalies. Solid straight line indicates a linear least-squares regression fit to the data.
speeds greater than 15 m s\(^{-1}\), and appears to be most dramatic for lower wind speeds (Cardone et al. 1990). For example, North Atlantic Ocean winter-time zonally averaged wind speeds at 20°N and 60°N of 8 and 12 m s\(^{-1}\) appear to be too low by 1.4 and 0.4 m s\(^{-1}\) respectively (Isemer and Hasse 1991). Posmentier et al. (1989) analysed wind data over the tropical Pacific Ocean in conjunction with tropical Pacific tidal gauge records and rejected the apparent increase in the strength of the easterlies because eastern tropical Pacific sea levels and sea surface temperatures both increased.

There is, however, evidence that at least part of the increasing trend in wind speed in the North Pacific is real. Sea level pressure (SLP) over the North Pacific Ocean over 1950–89 exhibits a linear trend with a maximum decrease of 10 mb in the northeast Pacific while along the western boundary and along 25°N there is no trend (Figure A.5). The error in the linear trend ranges from 0.5 mb in the southwest and southeast of the study area domain, to a maximum of 3.5 mb in the Northeast Pacific coinciding with the maximum linear change of 10 mb. This makes the linear changes of between −2.0 mb and −10 mb significant.

As a result of the SLP trend pattern, the magnitude of the meridional SLP gradient shows a linear increase over much of the central North Pacific between 25°N and 40°N (Figure A.6). The \(U\) (east-west) wind component also shows a linear increase over most of the North Pacific with a maximum value of 3.9 m s\(^{-1}\) at 38°N 168°W (Figure A.7) and corresponding error ranging from 0.5 to 1.5 m s\(^{-1}\). Correlations between the time series of the meridional SLP gradient and \(U\) wind component (Figure A.8) show a maximum value of 0.9, and most of the basin between 30°N and 50°N is greater than 0.6. (Almost all correlations greater than 0.1 are significant based on a Student’s t-test.) Although the trend in the \(U\) wind component may be partially due to larger ships avoiding oceans storms less (which also may affect SLP), and changes in measurement method, at least part of the increase in wind speed is probably due to the strengthening of the SLP
Figure A.5: Linear change in winter-time sea level pressure over the period 1950–89. Contour interval is 2.0 mb. Solid contours are positive, dashed contours are negative.

Figure A.6: Linear change in winter-time meridional sea level pressure gradient over the period 1950–89. Contour interval is 1.0 mb per 4° latitude.
Figure A.7: Linear change in winter-time $U$ (east-west) wind component over the period 1950–89. Contour interval is 1.0 m s$^{-1}$.

Figure A.8: Correlation between time series of meridional sea level pressure gradient and $U$ (east-west) wind component over the period 1950–89. Contour interval is 0.2.
Figure A.9: Linear change in winter-time $U_g$ (geostrophic) wind component due to the increase in the meridional sea level pressure gradient over the period 1950–89. Contour interval is 1.0 m s$^{-1}$.

gradient which drives the surface winds.

The effect of the increased meridional SLP gradient on the $U$ wind component can be estimated by looking at its effect on the $U$ component of the geostrophic wind:

$$\Delta U_g = \frac{1}{f \rho_a} \frac{\delta P}{\delta x}$$

(A.46)

where $f$ is the coriolis parameter ($1.46 \times 10^{-4}$ s$^{-1}$), $\rho_a$ is the air density in kg m$^{-3}$ (the long term season average for each grid point), and $\delta P/\delta x$ is the change in meridional pressure gradient in Pa m$^{-1}$. The contour interval of Figure A.6 of 1.0 mb ($4^\circ$ latitude$)^{-1}$ corresponds to 0.1/444 Pa m$^{-1}$. This is equivalent to about a 1.3 m s$^{-1}$ change in the $U_g$ wind component (Figure A.9). The actual effect on the $U$ wind component will be less than shown in Figure A.9, however, due to friction. But this calculation shows the effect on the $U_g$ wind component by the strengthening of the meridional SLP gradient to be on the same order of magnitude as the observed increase in the $U$ wind component and within 1 m s$^{-1}$ for most of the North Pacific Ocean.
More evidence for increasing wind speed over the North Pacific Ocean can be found in the National Meteorological Center (NMC) upper air analyses. Over the period 1964–89 the \( U \) wind component at 850 mb shows a maximum linear increase over the North Pacific of about 6 m s\(^{-1}\) at 35°N 155°W with most of the North Pacific basin showing an increasing trend. Like the SLP data, the 500 mb height field shows a linear decrease over most of the North Pacific basin with a maximum value of 120 m centered at 45°N 170°W. Along 25°N there is no trend which means the meridional pressure gradient and \( U \) wind component are increasing. Although the NMC analyses are based on limited data over the oceans and supplemented with model output, they support the possibility of an increasing trend in wind speed over the North Pacific Ocean. It should be noted that changes in analysis procedures in the NMC data have resulted in several artificial discontinuities in the northern hemispheric data, but data over the North Pacific Ocean appear to be relatively free from such influences (Lambert 1990).

Trends in the sea surface temperature (\( T_s \)) and surface air temperature (\( T_a \)) are best examined together, since these variables are closely coupled over the oceans. Decreases in the observed \( T_s \) have been attributed to the change from measurements of buckets of surface water to the engine intake which measures deeper, cooler water. In the North Pacific this change resulted in a decrease in observed \( T_s \) of about 0.3 °C (Tabata 1978). It has been suggested that observed increases in \( T_a \) can be attributed to an increase in ship size and its proportional heat island effect (Taylor 1984b).

The linear trend in \( T_s \) is decreasing over much of the North Pacific while temperatures are increasing along the eastern boundary (Figure A.10). Averaged over the North Pacific Ocean study area \( T_s \) shows a linear decrease of 0.7 °C over 1950–89 with an error ranging from 0.2 to 0.4 °C over the North Pacific Ocean. The spatial pattern of the trend in \( T_a \) (Figure A.11) is very similar to that of \( T_s \) but is mostly greater than the trend in \( T_s \) by \( \sim0.3 \) °C on average over the North Pacific (Figure A.12). In other words, averaged over
Figure A.10: Linear change in winter-time sea surface temperature over the period 1950–89. Contour interval is 0.5 °C.

Figure A.11: Linear change in winter-time surface air temperature over the period 1950–89. Contour interval is 0.5 °C.
Figure A.12: Linear change in winter-time surface air temperature minus the linear change in winter-time sea surface temperature over the period 1950–89. Contour interval is 0.5 °C.

the North Pacific basin the decrease in $T_s$ is greater than the decrease in $T_a$. (The error in the linear change in $T_a$ ranges from 0.3 to 0.5 °C over the North Pacific study area.) Because the difference in the trends is mostly non-zero suggests that one or the other or both of $T_s$ and $T_a$ are experiencing an artificial influence, otherwise their difference in trends should be near zero. But because the difference in trends varies over well-sampled regions (Figure 3.1) along 30°N from more than −0.5 °C south of Japan to greater than +0.5 °C off of California we cannot assume a basin-wide artificial trend in both $T_s$ and $T_a$. Furthermore, the fact that the magnitude of observed trends in $T_s$ and $T_a$ are generally greater than their difference indicates the real trends are stronger than the artificial trends. Further, the individual trends occur together and exhibit the same broad spatial pattern, an unlikely effect of separate, artificial processes. Thus the trends in $T_s$ and $T_a$ appear to be real. Since the trend patterns are similar to that of the leading principal component of $T_s$ (Namias et al. 1988; Iwasaka et al. 1987; Lanzante 1984; Hsiung and Newell 1983; Weare et al. 1976) it appears the dominant mode of $T_s$ is strongly tied to
low frequency natural variability.

It is also worthwhile examining changes in $T_a$ in relation to SLP since the two are related through the ideal gas law ($P = \rho RT_a$). Figures A.5 and A.11 support this relationship because both show decreases over most of the central North Pacific (30°N to 50°N, 150°E to 150°W). The signs are opposite, however, between SLP and $T_a$ along the western and eastern basin boundaries. Where the sign of the changes agree, the magnitude of the changes are not consistent. The maximum decrease in SLP is 8 mb which should correspond to a decrease in $T_a$ of $\sim 0.03 \, ^\circ C$, but the actual decrease is 0.3 \, ^\circ C. Decreases of 2 mb at 30°N 160°W and 40°N 165°E should correspond to decreases of 0.007 \, ^\circ C, but are actually realized in decreases of 0.4 \, ^\circ C and 1.2 \, ^\circ C. However, these inconsistencies are within measurement error and in general the trends in $T_a$ and SLP agree.

The trend in the specific humidity of the surface air ($q_a$) illustrates a similar pattern to that of $T_a$. This is because over the ocean, the moisture content of the air is only limited by its temperature (and not the availability of surface moisture) and over the short range of temperature change ($\sim 2\, ^\circ C$, Figure A.11) vapour pressure varies approximately linearly. As a result the difference between trends in $q_a$ and $T_a$ is close to zero over the North Pacific Ocean study area. The patterns of the trends in $(T_s - T_a)$ and $(q_s - q_a)$ are also very similar, and their difference is also approximately zero. Since trends in the humidity data over the North Pacific closely follow the trends in temperature (which we assume are mostly real in nature) and spatially coherent trends in humidity are absent in global data (Cayan 1992), it would be tenuous to assume a significant artificial trend in the humidity data.

These figures illustrate the problem in determining the nature of the significant trends observed in the marine surface data. The effects of changes in measurement technique alone have been clearly shown to induce artificial trends in the wind speed (Cardone et
al. 1990) and sea surface temperature (Tabata 1978). The efforts to isolate these effects, however, are confounded by the demonstrated changes in the sea level pressure gradient (above) and probable real changes in the air and sea surface temperatures. In the end one has to admit that the marine surface data record contains both artificial and real long-term trends. The increasing trend in wind speed appears to have both a significant real and artificial basis whereas the trends in sea surface temperature and surface air temperature are likely more real than artificial. Due to the evidence of real trends in the data and the difficulty of filtering out only the artificial trends, no adjustments were made to the data to try to reduce the effects of the latter.
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