Variability of Eddy Heat Fluxes Over the Northwestern Gulf of Mexico.

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VARIABILITY OF EDDY HEAT FLUXES OVER THE NORTHWESTERN GULF OF MEXICO

A Dissertation

Submitted to the Graduate Faculty of the Louisiana State University and Agricultural and Mechanical College in partial fulfillment of the requirements for the degree of Doctor of Philosophy in

The Department of Oceanography and Coastal Sciences

by

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ABSTRACT

Eddy heat flux variability over the Louisiana-Texas shelf was investigated using satellite-derived surface velocity and temperature data between October 1993 and October 1994. Assuming the product of sea water density and specific heat is relatively constant, velocity-temperature covariance reflects eddy heat flux (the fluctuating part of the 1x1 degree, 90 day mean heat flux). Available velocity and temperature fields, however, are not synchronous. Temperature "observations" at velocity positions were optimally estimated using the Gauss Markoff Theorem. The error estimate is comparable to the error resulting from the application of the widely accepted SST correction algorithm. The trend that instantaneous temperature flux principal axes become more isotropic offshore is significant at $\alpha = 0.10$ in all seasons but October-December. Across the shelf, eddy heat flux is directed upcoast. In winter, the innershelf upcoast eddy heat flux is induced by cool downcoast transport associated with cold air outbreaks; while near the shelf break, it is induced by warm upcoast transport probably associated with an anticyclonic ring shed from the Loop Current. In the summer, the innershelf upcoast eddy heat flux is induced by warm upcoast transport. Eddy heat transport may be an important term in the winter heat budget. Heat is lost downcoast primarily because of the longshore mean-velocity gradient.
INTRODUCTION

1.0 How the dissertation is organized

This dissertation investigates eddy heat flux variability over the Louisiana-Texas shelf. It consists of five chapters: Introduction, Literature Review, Materials and Methods, Results and Discussion, and Conclusions. In the first chapter, the study's economic and academic value is addressed, followed by the study's objectives and research questions. In Chapter 2, the study area and its bathymetric characteristics are described. Freshwater sources, wind effects on the local currents and sea surface temperature (SST), and local current variability are discussed. A review of prior eddy heat flux determinations, along with methodologies and results of other large scale projects with similar objectives are presented. In Chapter 3, the data sources, software and computer facilities utilized are outlined. Data processing algorithms are then described. In Chapter 4, the observation errors are presented, and the variabilities of the mean and eddy temperature fluxes are compared. In this chapter's last two sections, the relative importance of eddy heat flux is presented and the dominant surface temperature divergence term identified. Chapter 5 provides a summary, list of conclusions, and suggestions for future efforts.

1. Eddy, as used in this dissertation, refers to a deviation from the mean as defined by Osborne Reynolds (e.g., Pond and Pickard 1991).
1.1 Importances of the study

A better understanding of spatial and temporal variability of eddy heat flux over the Louisiana-Texas shelf is economically and academically important. Eddies can transport, trap, and disperse passive matter in the sea [Robinson 1982], including salt, nutrients, planktonic larvae [Olson and Backus 1985], pollutants, and sediments. These transports can significantly alter the biological, chemical, and physical characteristics of the source and receiving environments. Additionally, the horizontal and vertical eddy length scales affect mixed layer properties, species concentration profiles, and daily migration of species [Robinson 1982].

It is hoped that the horizontal scales of eddy heat flux will reflect the scales of other transported material. Heat, however, besides being an important physical property of the sea affecting biological, chemical, and physical processes, is also dynamically important. Thus, some caution may be necessary in extrapolating the results of this study.

Eddy heat transport can be particularly important over the shelf where irregular topography, frontal instability, shearing of swift currents, current reversals, strong SST fluctuations, and sharp frontal boundaries are common. Over the Oregon shelf, for example, the onshore eddy heat transport balances the offshore mean heat transport. The eddy heat transport term is vital to the low
frequency heat budget equation for that shelf [Bryden et al. 1980]. This balance is not true everywhere. Off the northwest Africa shelf, cross-shelf eddy heat transport is an order of magnitude smaller than the mean heat transport [Richman and Badan-Dangon 1983]. The relative importance of eddy heat transport over the Louisiana-Texas shelf has not previously been determined. The magnitude of eddy heat transport may be expected to be relatively large along the Texas shelf where the local longshore currents reverse every 1-2 weeks [Smith 1980] and SST temporal and spatial gradients can be significant. In March, for example, the SST gradient can be O(0.25 °C km⁻¹) near the shelf break [Barron and Vastano 1994] (where O denotes "order of").

Lack of observations on the temporal and spatial variability of eddy heat flux has led to incorrect representations in numerical models. In the conservation of heat equation, eddy heat flux terms are often parameterized as proportional to the horizontal gradient of the mean temperature, analogous to molecular diffusion. Such Fickian diffusion representations, however, are invalid because eddy fluxes are sensitive to the properties of the flow as well as the fluid [Townsend 1956; Gosman et al. 1969]. In the initial analysis of over 300 drifting buoys in the northwestern Gulf of Mexico, the mean and eddy portions of buoy-velocities have been found to be uncorrelated [P. Niiler, LATEX meeting in Baton Rouge, February 1995].
This study also offers an approach to the analysis of a rare, large-scale study of eddy heat flux. The approach is described in Chapter 3. As the cost of satellite data continues to decrease, the proposed approach becomes economically more appealing. Additionally, the study develops a framework for future efforts. With slight modification, the methodological technique would be applicable to investigations of large scale variability of sediment or chlorophyll fluxes. Finally, the estimated fluxes can provide initial model conditions and/or data for verification of model results.

1.2 Objective

The objective of this study is to investigate the spatial and temporal variability of eddy heat flux over the Louisiana-Texas shelf between October 1993 and October 1994. This investigation is focused on characteristics of the shelf’s eddy heat fluxes, as well as the shelf’s mean horizontal heat fluxes. Questions addressed include:

- What are the seasonal and record mean distributions of the mean and eddy heat fluxes?
- What processes contribute to the variability of eddy heat fluxes over the Louisiana-Texas shelf?
- Is the approach taken reliable and practical?

1.3 How the objective was met

Assuming \( gC_p \) is constant over the Louisiana-Texas shelf, where \( g \) is sea water density and \( C_p \) is specific heat, the surface velocity-temperature covariances repre-
sent surface eddy heat fluxes. In order to compute the covariance terms, multiple synoptic and coincident fields of near surface temperature and velocity are required. Buoys equipped with ARGOS transmitters may be tracked by the ARGOS systems on NOAA satellites. The Advanced Very High Resolution Radiometer (AVHRR) sensors also on NOAA satellites can measure sea surface albedos and radiation temperatures. Processed ARGOS and AVHRR data are capable of providing synoptic yet accurate estimates of near surface velocities and sea surface temperatures [McClain et al. 1985], respectively. The two fields are, however, seldom synchronous. Furthermore, clouds over the sea surface prohibit a reliable SST estimate.

To make the two fields coincide, SST "observations" at velocity positions were optimally estimated. Optimal analysis is a linear interpolating technique evolved from the Gauss Markoff theorem (see Sections 3.6 and 3.7). An optimal estimate is one whose expected squared-error is minimal. The theorem requires that the expected mean of the variable of interest be zero. Thus, in this study the theorem was applied to estimate detrended SST "observations" given numerous detrended SST's fields. The gross SST "observation" was the sum of the detrended "observation" and the mean. Velocity and temperature means and subsequently covariance were computed from SST "observations" and velocities within that region. Estimation of the variability of velocity-temperature covariances is a major
step forward to satisfying the objectives mentioned above.
2.0 Chapter outline

This chapter describes the physical characteristics of the study area, the effects of winds on local currents and SST's, and current variability. Prior eddy heat flux determinations, parallel studies with similar objectives and methodologies, and characteristics of AVHRR are also discussed.

2.1 Study area and physical characteristics

The study area which includes the Louisiana-Texas shelf is shown in Figure 2.1. Its northern and western boundaries are the coastline; its southern and eastern boundaries are indicated by bold solid line in Figure 2.1.

The study area consists of two adjacent shelves of different bathymetric characteristics. East of 96 °W, the isobaths are oriented in the east-west direction. The shelf is broad and shallow and mainly covered with fine grained sediments [Rezak and McGrail 1983]. The shelf width varies along the coast, being a mere 50 km near the Mississippi River mouth and roughly 200 km near 93.5 °W. The average shelf slope is O(1/1000). In contrast, west of 96 °W, the isobaths are oriented north-south. The shelf is much narrower and deeper. Shelf width is relatively constant between 26 and 28 °N, O(100 km). The shelf slope is approximately twice as steep at O(2/1000).
Figure 2.1 The Louisiana-Texas shelf study area. The offshore contours are in meters. The bold solid line is the eastern and southern boundary of the study area.
The Louisiana-Texas shelf receives over ninety percent of its freshwater from the Atchafalaya and Mississippi Rivers [Cochrane and Kelly 1986]. The Mississippi-Atchafalaya river system releases an annual average of 20,000 m$^3$s$^{-1}$ of water to the shelf. The annual discharge peaks around April and reaches its low around October.

### 2.2 Effects of winds on currents and sea surface temperatures

Wind is the dominant agent driving currents over the Louisiana-Texas Shelf [Cochrane and Kelly 1986; Crout et al. 1984]. Easterly winds are dominant for most of the year [Murray 1976]. Wind speeds are weakest in the summer and strongest in the winter [Crout et al. 1984]. Longshore and cross-shelf winds affect near-surface currents differently. Over the Texas shelf, near-surface currents respond strongly to longshore wind. Over the Louisiana shelf, currents respond to the longshore wind, as well as to cross-shelf wind in the vicinity of broad nearshore shoals. Chuang and Wiseman (1983) attribute the difference in response to bottom friction characteristics of the two shelves.

Cochrane and Kelly (1986) gathered historic wind, current meter, and hydrographic data to form a coherent picture of low frequency shelf circulation over the northwestern Gulf of Mexico. They hypothesize that, except for June and July, a cyclonic circulation pattern exists over the Louisiana-Texas shelf. Its eastern boundary lies between Atchafalaya Bay and the Mississippi River delta while
its southern boundary lines along the shelf break. Its western boundary varies seasonally. The shoreward end of the western boundary is a point where the mean wind vector is perpendicular to the coast, and coastal currents converge [Rezak and McGrail 1983]. The western boundary lies quasi-parallel to the US-Mexican border in September and moves upcoast thereafter. By May, the western boundary is close to Cameron, LA, and west of it an anticyclonic cell has formed. In June, the western boundary of the cyclonic cell has disappeared and the anticyclonic cell moves upcoast. The center of this anticyclonic cell is around 29°N, 93°W. Abruptly in September the cyclonic cell with its western boundary near the US-Mexican border is formed and the cycle continues.

Higher frequency current variability can be generated by storms and hurricanes. These relatively short but very severe wind events drive strong currents, reverse flow direction, intensify inertial oscillations, generate upwelling and thus lower sea surface temperature, and force transport almost parallel with the wind direction. At 20 m deep, 50 km south of Galveston Bay, the 1973 tropical storm Delia produced 2 m s⁻¹ current speed compared to the local mean speed of 15 cm s⁻¹ [Forristall et al. 1977]. In addition to hurricanes, cold air outbreaks, which begin in October, also can cause current reversals and SST fluctuations. Before a cold air outbreak, southeasterly wind pushes shelf water against the coast. This is followed by
cold, dry, northerly winds which strip heat from the sea surface [Huh et al. 1978; Huh et al. 1984], and deepen the surface mixed layer. In a two week period centered around a cold front passage, nearshore SST off Galveston can drop as much as 4°C [Nowlin and Parker 1974].

Cold air outbreaks and lower frequency winds can also induce SST variability via upwelling/downwelling events [e.g. Dagg 1988]. In a low frequency heat budget study of the Louisiana-Texas shelf, Etter, et al. (1985) found horizontal heat flux divergence of 0(30 Wm⁻²) in June and July when upwelling favorable winds prevail; and heat flux convergence between 0(30 Wm⁻²) to 0(100 Wm⁻²) for the remainder of the year when downwelling favorable wind prevails.

Current velocity and SST not only vary in time but also in space as well. Barron and Vastano (1994) observed that currents over the outer shelf are slower than currents near shore. Velocity principal axes elongate parallel to the local topography nearshore and become more circular offshore [Johnson and Niiler 1994]. In the summer along the Texas shelf, Smith (1980) notes that although mid-depth currents near Port Aransas, Port-O'Connor, and Port Mansfield reverse at 0(1-2 weeks), there is a phase lag of 0(3 hr) at frequencies higher than 1/3 day⁻¹. Surface convergence or divergence can develop during these transient periods. Cold air outbreaks can generate sharp SST spatial gradient because shallow inner shelf water cools more rap-
idly than deeper offshore water [Huh et al. 1978]. After a cold air outbreak, maximum gradients nearshore and near the shelf break off Galveston can be 0.14 and 0.43 C km⁻¹, respectively [Nowlin and Parker 1974]. Furthermore, since cold air outbreaks occur every 3-10 days through the winter months, successive outbreaks are capable of advecting previously conditioned cool water downcoast [Nowlin and Parker 1974] and thus changes temperature in both the source and receiving environments.

The wind's ability to cause both current and temperature fluctuations on the continental shelf makes it, potentially, a prime driving agent of eddy heat flux. Over the California shelf, a large fraction of the average eddy heat flux is wind-driven [Send 1989].

2.3 Determination of eddy heat flux

The majority of eddy heat flux may be concentrated in the surface layer since most driving mechanisms responsible for both velocity and temperature fluctuations are input through the surface and random fluctuations are attenuated by friction near the bottom [Rezak and McGrail 1983]. Over the Oregon shelf, Bryden, et al. (1980) observed that 99% of the depth-integrated eddy heat flux occurs in the upper 1/3 of a 100 m deep water column. Over the California shelf, Send (1987) observed that temperature decays rapidly with depth. In his later work, he only used the sum of the upper 30 m eddy heat fluxes to estimate depth-integrated eddy heat flux [Send 1989]. Over the equa-

The proportional factor between temperature and heat is $q_{c_p}$. Although $q_{c_p}$ was not determined in this study, velocity-temperature covariance may be used to infer surface eddy heat flux because $q_{c_p}$ is relatively constant over the shelf. Specific heat is mainly dependent on salinity [e.g. Hill 1962] and the local density is more sensitive to salinity than temperature. Considering two extreme Louisiana-Texas shelf conditions, at surface salinity of 35 ppt and temperature of 10 °C, $q_{c_p}$ is 0.957 cal cm$^{-3}$ °C$^{-1}$; while at 10 ppt and 30 °C, $q_{c_p}$ is 0.973 cal cm$^{-3}$ °C$^{-1}$. The difference is merely 1.6%.

2.4 Methodologies of other parallel studies

Other spatially intensive shelf studies exist. These studies [Freeland et al. 1975; Kundu and Allen 1976; Davis 1985; Poulain and Niiler 1989; Brink et al. 1991] were designed to investigate the spatial variability of shelf dynamics and kinematics. Mean velocities were computed to characterize the shelf environment. The velocity principal axes were also estimated in order to reflect the degree and direction of anisotropy in the fluctuating flow field. Some studies [Freeland et al. 1975; Davis 1985; Poulain and Niiler 1989] also computed the correlation estimates to determine phase velocities and decorrelation scales of currents.
From the variability of these statistical properties, Poulain and Niiler (1989) conclude that the velocity field is very inhomogeneous in time and space over the California shelf. They attribute the spatial inhomogeneity to the synoptic presence of several oceanographic features of different horizontal scales: straight jets, eddies, and trapping regions. Furthermore, the flow is more anisotropic within 50 km from shore than further offshore. From the correlation estimates of buoy-velocities, they observed decorrelation scales of O(80 km) and O(12 days) and a phase speed of 2 cm s\(^{-1}\) from east to west.

Over the California inner shelf, Brink, et al. (1991) observed a decrease in eddy kinetic energy from north to south. North of 33 °N, the mean regional velocities are relatively strong. Primary orientations of velocity principal axes and mean regional velocities are quasi-parallel. In contrast, south of 33 °N, the mean regional velocities are weak and the orientations of velocity principal axes appear to be random. Brink, et al. (1991) propose that the spatial change could have been caused by (1) the broader shelf south of 33 °N or (2) seasonal variations, since the experiment endured for more than one season.

One difference between these studies and the present work is that a SST "observation" has been attached to each buoy-velocity in the present work. The SST "observations" were estimated from AVHRR-derived SST's.
2.5 Advanced Very High Resolution Radiometers (AVHRR)

AVHRR data are a component of High Resolution Picture Transmission (HRPT) telemetry data. The AVHRR data are remotely measured by scanning sensors on board the NOAA satellites. Each sensor responds to five wavelength channels, ranges 0.58-0.68, 0.7-1.1, 3.5-3.9, 10.5-11.5, and 11.5-12.5 µm. The difference between channel 4 and channel 5 radiation temperatures is proportional to the temperature offset induced by atmospheric water vapor [Deschamps and Phulpin 1980]. The AVHRR-derived SST using multiple channels are often referred to as multi-channel sea surface temperature (MCSST). The MCSST can approximate true sea surface temperature to within 1 °C [McClain et al. 1985].

In addition to AVHRR's ability to provide relatively accurate SST estimates, the AVHRR sensor has a swath width of 2800 km and thus provide broad spatial coverage. Its single path swath can cover an area as large as 2,800 x 5,400 km with a 0(1.1 km) spatial resolution at nadir. A single AVHRR image over the Gulf of Mexico, for example, can contain over a million individual MCSST values (1400 x 800). Furthermore, since the NOAA satellites orbit on a near-polar track which is sun-synchronous, the same area can be captured about the same time every day with nadir passes occurring around 0300 and 1500 and 0730 and 1930 local time. AVHRR data have been successfully used to monitor temporal and spatial variability of river plumes and coastal processes [Muller-Karger et al. 1991; Rucker et al.].
1990].
MATERIALS AND METHODS

3.0 Chapter outline

In this chapter, the data and data processing are described. Three data sets were used: preprocessed ARGOS buoy-velocity data, raw AVHRR data, and conventional oceanographic data. These data are described in Sections 3.1 through 3.3. The cloud screening procedure and the principles of the multi-channel temperature correction algorithm are presented in Sections 3.4 and 3.5, respectively. Gauss Markoff principles and optimal analysis are discussed in Sections 3.6 and 3.7. Section 3.8 defines instantaneous, mean, and eddy temperature flux. Section 3.9 describes how the study's resolution (bin size) was selected. The next three sections describe the statistical procedures for the analysis. In Sections 3.13 and 3.14, methods for estimating the relative importance of eddy heat flux in the heat budget equation and a method for identifying the dominant component of surface temperature divergence are discussed.

3.1 Buoy-velocity data

Scripps Institute of Oceanography (SIO) and the Minerals Management Service (MMS) deployed more than 300 surface ARGOS drifting buoys over the northwestern Gulf of Mexico between October 1993 and October 1994 [Johnson and Niiler 1994]. These data are used in this dissertation. From October through May, most buoys were located over the
western half of the Louisiana-Texas shelf [Johnson and Niiler 1994]. In June, buoys began to flow upcoast, some continued east past the Mississippi delta. These data represent the first large scale Lagrangian observations of the near surface current field over the northwestern Gulf of Mexico shelf. SIO provided optimally interpolated velocity "observations" at 1115 GMT for each buoy-day. There are 19,299 buoy-days of observations over the northwestern Gulf of Mexico (Figure 3.1).

3.2 Satellite data

The Earth Scan Laboratory of Coastal Studies Institute captures and archives raw AVHRR telemetry data 6-7 times a day. The percent clear area over the study area between October 1993 and July 1994 is shown in Figure 3.2. Terascan™, a satellite data processing software package, was used to extract each AVHRR image from HRPT telemetry data and convert 10 bit data from visible and infrared channels to albedos and radiation temperatures, respectively. An optimal analysis (OA) package [Mariano and Carter in preparation] was used to objectively estimate SST "observations" at positions coincident with buoy-velocity data. Additionally, a cloud-free movie of SST fields over the northwestern Gulf of Mexico from October 1993 through October 1994 was produced as a by-product.

Optimal analysis is described in Section 3.7. An optimal estimate is a linear combination of the surrounding MCSST data. In the experimental stage of the study,
Figure 3.1 Locations where buoy-velocity data were obtained, Oct 2 1993-Sep 27 1994
Figure 3.2 Percent clear area over the Louisiana-Texas shelf, Oct 93-Jul 94
weights of the surrounding data were determined by a Gaussian correlation function. When both day and night time MCSST data were used to produce night-time fields of SST "observations", and when night-time data are sparse, the night-time SST "observations" are warm like daytime temperatures. This is because SST has a strong diurnal signal. Thus, either only the day or the night time data should be used. Since the buoy-velocity data were optimally to night time, only the night-time MCSST data were used.

3.3 LATEX-A density cruise data

Texas A&M University has furnished 0.5 m, filtered, sea water density data from four 2-week cruises: late November 1993, early February 1993, late April 1994, and late July 1994. Station locations are shown in Figure 3.3. Only the early February cruise failed to sample the entire study area. The cruise data were used to estimate the Burger Number [e.g., Brink 1989] and mixed layer depth (MLD) fields discussed in Chapter 4.

3.4 Cloud screening

Albedos and radiation temperatures of cloud contaminated pixels produce unreliable SST estimates. Thus, cloud-screening is an essential pre-processing step prior to optimal analysis. A series of tests developed by McClain, et al.(1985) was applied to AVHRR data to identify cloud-contaminated pixels. A pixel was classified as cloud by either the uniformity test or channel intercomparison test.
Figure 3.3a LATEX-A CTD stations, Nov 7-21, 1993
Figure 3.3b LATEX-A CTD stations, Feb 6-13, 1993
Figure 3.3c LATEX-A CTD stations, Apr 24-May 7, 1994
Figure 3.3d LATEX-A CTD stations, Jul 27-Aug 5, 1994
Fields of cloud-contaminated pixels exhibit higher temperature variability than fields of unobscured sea surface temperature [McClain et al. 1985]. Clouds occupying only 10% of the field of view can lower radiation temperature by 1-2 °C [Maul and Sidran 1973]. The uniformity test classifies a pixel as "cloud" if the average of the 8 channel-4 gradients within a given 3x3 pixel box exceeds a prespecified threshold. The 8 gradients are the gradients between the center pixel and the 8 neighboring pixels. Thresholds of 0(1 °C/pixel) were specified in this study. The specific threshold value was set subjectively after visual inspection of each particular image.

The uniformity test may not be able to detect low stratus cloud because they have extremely uniform cloud top temperatures [McClain et al. 1985]. The channel intercomparison test labels a pixel as "cloud", if the difference between 3.75 μm and 10.8 μm radiation temperatures is less than a prespecified threshold. The test exploits the difference between cloud and sea surface emissivity characteristics. Sea surface emissivities at 3.75 μm and 10.8 μm are 0.975 and 0.993, respectively [Katsaros 1980]; while cloud emissivities at the same wavelengths are 0.75 and close to unity [Hunt 1973]. For any atmospheric condition, the inter-radiation temperature difference

\[ (T_{3.75\mu m} - T_{10.8\mu m}) \]

of clouds is lower (more negative) than that of the sea surface.
This inter-radiation difference varies depending on the amount of water vapor in the atmosphere because percent transmissivity is also wavelength dependent. Thermal infrared transmissivity is much more sensitive to the amount of atmospheric water vapor than is mid infrared transmissivity [Deschamps and Phulpin 1980]. Due to the presence of unresolved clouds, water vapor, aerosols, sea surface's quasi black body property, radiation temperature underestimates true sea surface temperature [Bernstein 1982]. As the amount of water vapor increases, the deviation between infrared radiation temperature and true temperature increases more than the deviation between mid-infrared radiant temperature and true temperature. For dry and moist conditions, the inter-radiation temperature differences of the sea surface are $0(0 \, ^\circ C)$ and $0(2 \, ^\circ C)$, respectively [Ter-ascan™ Reference Manual 1993]. The inter-radiation temperature differences of clouds is usually less than $-2 \, ^\circ C$. The prescribed separation threshold varied depending on atmospheric conditions when the image was captured.

"Cloud" pixels were eliminated at this point. Before saving the remaining pixels as good data, cloud streaks that may have gone undetected were visually identified and manually removed from further processing.

3.5 Multi-channel sea surface temperature

Among the 3.75, 10.8, and 12 $\mu$m spectral bands, the 3.75 and 12 $\mu$m transmissivities are least and most sensitive to the amount of water vapor in the atmosphere. Un-
fortunately, the 3.75 \( \mu m \) radiation temperature sensor is noisy. The difference between 10.8, and 12 \( \mu m \) radiation temperatures was used to estimate the temperature offset that results from atmospheric absorption. In practice, the multi-channel sea surface temperature (MCSST) can estimate true SST to less than 1 °C. The time and space coordinates of each sea surface pixel, MCSST were recorded.

Cloud-filtered MCSST's were employed to optimally estimate SST "observations" at buoy-velocity positions. Optimal analysis is a linear interpolation technique evolved from the Gauss Markoff theorem [Liebelt 1967]. The principles of Gauss Markoff theorem and optimal analysis are described in the next two sections.

3.6 Gauss Markoff principles

The Gauss-Markoff theorem is the basis for a linear interpolation technique for a homogeneous variable with zero mean [Liebelt 1967]. The definition of an optimal estimate is one for which the expected value of the squared difference between the estimate and the true value is minimized.

Thus, \( \hat{T} = AX \) where \( \hat{T} \) is a vector of estimates at the points of interest. The vector of corresponding true values is \( T \) and \( X \) is a data vector. The Optimal Analysis package estimates \( \hat{T} \) one point at a time. Thus, \( \hat{T} \) will be described as a 1x1 matrix, \( X \) is an nx1 matrix of \( X \)'s, and \( A \) is a 1xn linear operator.
Following Liebelt (1976), by minimizing $\mathbb{E}[(\hat{T} - T)(\hat{T} - T)^T]$ where $\mathbb{E}[]$ denotes the expected value and $(\cdot)^T$ denotes the transpose of $(\cdot)$, $A$ and $T$ become:

$$A = C_{TX}C^{-1}_{XX} \quad [3.1]$$
$$T = C_{TX}C^{-1}_{XX}X \quad [3.2]$$

$$\mathbb{E}[(\hat{T} - T)(\hat{T} - T)^T] = C_{TT} - C_{TX}C^{-1}_{XX}C_{TX} \quad [3.3]$$

where covariance matrix is denoted as $C$, e.g., $C_{TX} = \mathbb{E}[T^T]$., and it is assumed that $C_{XX}$ is positive definite.

3.7 Optimal analysis (OA)

Optimal analysis slightly modifies the technique previously described in 3.6. Assume for the moment, $\mathbb{E}[T]$ is zero. In practice, error-free measurements cannot be made. In this study, the data vector consists of MCSST's. They are subject to instrumental error and error induced by sub-pixel processes and features such as unresolved clouds. The sum of the error from these sources will be denoted by $e$, such that

$$X = T + e \quad [3.4]$$

where $T$ is the true value of the field at the measurement site. Substituting $T + e$ for $X$, $C_{TX}$ and $C_{XX}$ become:

$$C_{TX} = C_{TT} + C_{Te} \quad [3.5]$$
$$C_{XX} = C_{TT} + 2C_{Te} + C_{ee} \quad [3.6]$$

The Gauss Markoff theorem is unaffected by replacing the covariance matrices with the covariance matrices normalized by their variances at zero lag [Mariano and Carter in preparation]. The normalized matrices are known as correlation matrices. Furthermore, the $T$ field is as-
sumed to have uniform variance, so the number of variance parameters needed is reduced to 1 [Mariano and Brown 1992]. The correlation matrix $C_{Te}$ may be eliminated by assuming that $T$ and $e$ are uncorrelated. Thus, $C_{TX}$ and $C_{XX}$ are reduced to:

$$C_{TX} = C_{TT} \quad [3.7]$$

$$C_{XX} = C_{TT} + C_{ee} \quad [3.8]$$

The technique must also be modified if the observed field is not mean-zero. Natural SST fields are seldom mean zero. The technique, however, may be applied to optimally estimate a detrended temperature, $\hat{S}$, given detrended measurements $Z$. The estimate $\hat{T}$ will then be the sum of the mean or trend field (in the following, we will refer to the mean field but imply the possible existence of a trend) and $S$ as follows:

$$\hat{T} = \text{mean (trend) of } \hat{T} + \hat{S} \quad [3.9]$$

and similarly,

$$X = \text{mean (trend) of } \hat{T} + Z \quad [3.10]$$

Analogous to $\hat{T}$ in Eqn 3.2,

$$\hat{S} = BZ = C_{SZ}C_{ZZ}^{-1}Z \quad [3.11]$$

where $B$ is a linear operator, and $S$ is the true detrended SST at the estimation point.

In this study, monthly mean SST fields were estimated by fitting a smooth surface to all observations in a given month. It is doubtful that a monthly mean field can be portrayed adequately by a constant because eddies with various length and time scales reside over the Louisiana-
Texas shelf. When a second order polynomial expression was employed to depict the October and November 1993 monthly mean fields, the variances of the resulting $ fields were relatively high implying that the mean fields probably failed to capture some high frequency structures such as frontal boundaries.

When a bi-cubic spline function was used to represent the monthly mean fields, the variances of the $ fields were reduced significantly. Mariano and Brown (1992) found a bi-cubic spline to better represent the SST mean field over the tropical Pacific between 30 °S and 30 °N than a polynomial type expression.

The bi-cubic spline used in this study was developed by Inoue (1986). This spline function may be pictured as a flexible plate forced to pass near a set of data points. Inoue's algorithm allows the analyst to control the following characteristics of the plate: degree of roughness, degree of tension at the boundary of the plate, and grid resolution. The OA package used in this study also computes root-mean squared fitting errors of the mean field. In an attempt to determine the parameter values that best describe the Louisiana-Texas shelf mean SST fields, the root-mean squared fitting errors of 27 December-1993 mean fields were compared. The 27 mean fields were parameterized by the combinations of low, intermediate, and high parameter values shown in Table 3.1.
Table 3.1 Experimented parameter values.

<table>
<thead>
<tr>
<th></th>
<th>low</th>
<th>intermediate</th>
<th>high</th>
</tr>
</thead>
<tbody>
<tr>
<td>deg of roughness</td>
<td>10</td>
<td>500</td>
<td>1000</td>
</tr>
<tr>
<td>deg of tension</td>
<td>0.1</td>
<td>0.5</td>
<td>0.9</td>
</tr>
<tr>
<td>grid resolution in km</td>
<td>0(15)</td>
<td>0(25)</td>
<td>0(50)</td>
</tr>
</tbody>
</table>

The root-mean squared fitting errors of these mean fields were very similar. The intermediate parameter values were then chosen to describe all monthly mean fields in the study.

The monthly mean SST fields for October 1993 through September 1994 are shown in Figures 3.4 through 3.7. They reproduce relatively large and persistent oceanographic features known to occur over the Louisiana-Texas shelf. During the winter months, cold front passages pass through the Louisiana-Texas shelf every 3 - 10 days. In the October through March fields, the shallow innershelf water is particularly cool and innershelf sea surface temperature gradients are sharp. Such gradients may be generated because of the differences between shallow and deeper water cooling rates [Huh et al. 1978]. In the November through February fields, the coastal boundary current is well defined by strong SST gradients. The width of the coastal boundary current appears to be relatively constant from the Louisiana-Texas border down to the US-Mexican border. The Mississippi and Atchafalaya River plumes are di-
Figure 3.4 SST monthly mean fields: (a) Oct, (b) Nov, (c) Dec 1993
Figure 3.5 SST monthly mean fields: (a) Jan, (b) Feb, (c) Mar 1994
Figure 3.6 SST monthly mean fields: (a) Apr, (b) May, (c) Jun 1994
Figure 3.7 SST monthly mean fields: (a) Jul, (b) Aug, (c) Sep 1994
rected westward. A warm patch (Eddy Vasquez) is also evident just out off the Texas continental shelf. These rings can transport heat from the eastern to the western Gulf. Also apparent is the offshore advection of cool coastal water over the south Texas shelf in March. This observation supports conclusions reached by Cochrane and Kelly (1986) and Dinnel and Wiseman (1986) that freshwaters originating from the Mississippi and Atchafalaya Rivers are transported offshore across the south Texas shelf and turn upcoast along the shelfbreak. In April, the Atchafalaya plume is separated from the cool water downcoast. Beginning in May, the sea surface temperatures are relatively uniform. Müller-Karger, et al. (1991) also observed poorer spatial structure in their 2-week, mean AVHRR-derived, SST fields from May through October. The lack of pattern in the May field suggests that May may be a transition period. By June, the surface temperatures of river plumes and ambient shelf water are indistinguishable. The cool coastal current has also disappeared. If the supply of cool water has come to a halt, then the elongated cool patch along the south Texas shelf in the June, July, and August fields is not likely a remnant of Louisiana cool water because the nighttime SST decay scale is relatively short (<2 weeks). In the summer, the innershelf current along Texas reverses and flows upcoast [Crout et al. 1984]. The cool patch well-confined against the coast may be induced by upwelling along the Texas shelf. Cochrane, et al. [in preparation]

After the mean fields have been established, they may still fail to adequately represent the local mean near an estimation point. While estimating a particular $S$, the mean field could be improved by accounting for the local mean. The local mean $\bar{z}$ may be computed by satisfying the condition:

$$\sum B_i = 1 \text{ [Bretherton 1976]}$$

where $B_i$ are the elements of the linear operator for $B$.

Thus, $\hat{T}$ now becomes:

$$\hat{T} = \text{monthly\_mean} + \bar{z} + C_{zz}C^{-1}_{zz}[Z - \bar{z}] \quad [3.12]$$

and $E[(\hat{T} - T)(\hat{T} - T)^T]$ becomes:

$$C_{ss} - C_{sz}C^{-1}_{zz}C_{sz}^T + (1 - C_{sz}C_{zz}^{-1}U)^2 / (U^T C_{zz}^{-1}U) \quad [3.13]$$

where $U$ is an nx1 vector of ones. The new third term accounts for the uncertainty of the estimated mean.

Constructing $C_{zz}$

Since $C_{zz}$ is unknown, $\hat{S}$ cannot yet be optimally estimated. Analogous to the simplification of $C_{sx}$ and $C_{xx}$ in Eqns 3.7 and 3.8, $C_{sz}$ and $C_{zz}$ become:

$$C_{sz} = C_{sg} \quad [3.14]$$

$$C_{zz} = C_{sg} + C_{ee} \quad [3.15]$$

where $g$ is the true detrended value at the measurement site.

The term $C_{sg}$ reflects the second order statistics of $S$. Like most random variables' correlation functions,
$C_{ss}$ should decay rapidly away from the zero lag value. A common form of correlation function is the product of sinusoidal and Gaussian terms. The Fourier transform of such a function is positive at all frequencies, which assures that the correlation matrix is positive definite. The $C_{ss}$ used in this study is characterized by a user-adjustable analytical function developed by Mariano and Brown (1992) shown below.

$$C(dx,dy,dt) = C(1)[1.-(dx-C(2)*dt)/C(4)^2 - (dy-C(3)*dt)/C(5)^2]$$

$$*exp(-[(dx/C(6))^2 + (dy/C(7))^2 + (dt/C(8))^2])$$

where

dx is the east-west lag;
dy is the north-south lag;
C(1) is the correlation at zero lag (a value less than one accounts for variability below the resolution of the measuring device);
C(2) is the mean east-west phase speed;
C(3) is the mean north-south phase speed;
C(4) is the zero-crossing scale in the direction of the primary axis;
C(5) is the zero-crossing scale in the direction of the secondary axis;
C(6) is the e-folding scale in the direction of the primary axis;
C(7) is the e-folding scale in the direction of the secondary axis;
C(8) is the time decay scale; and
C(9) is the orientation of the the principal axis. Although not shown in the equation, C(9) rotates C(dx,dy,dt) in space.

The nine parameters for C_{ss} were estimated from the correlation estimates of the detrended measurements, Z, because the population of S is not available. For each 1x1 degree block, the time-independent parameters \{C(4), C(5), C(6), and C(9)\} were estimated from detrended-measurements of individual images between July 1993 and December 1993, one image at a time. Also for each 1x1 degree block, time-dependent parameters \{C(1), C(2), C(3), and C(8)\} were estimated from detrended-measurements obtained from a series of images between July 1993 and December 1993. Since none of the parameters show apparent seasonal variability, only first and second order statistics of the parameter values are shown in Table 3.2.

The principal axes of the estimated correlation functions usually elongate parallel to the local bathymetry. A non-zero phase velocity was not apparent in the data. Parameters C(2) and C(3) were assumed to be zero in the correlation function model. Although phase velocity may not actually be zero, the irregularly, but frequently (twice a day), sampled AVHRR data can partially account for the true phase velocity. The time decay scale, C(8), exhibited no consistent pattern of variability. It was
Table 3.2 First and second order statistics of correlation parameter values.

<table>
<thead>
<tr>
<th></th>
<th>median</th>
<th>mean</th>
<th>min</th>
<th>max</th>
<th>std. dev</th>
</tr>
</thead>
<tbody>
<tr>
<td>C(1)</td>
<td>0.4</td>
<td>0.5</td>
<td>0.1</td>
<td>0.9</td>
<td>0.2</td>
</tr>
<tr>
<td>C(2) [deg/dy]</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>C(3) [deg/dy]</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>C(4) [deg]</td>
<td>0.5</td>
<td>1.1</td>
<td>0.4</td>
<td>4.0</td>
<td>0.5</td>
</tr>
<tr>
<td>C(5) [deg]</td>
<td>0.4</td>
<td>0.4</td>
<td>0.1</td>
<td>2.0</td>
<td>0.2</td>
</tr>
<tr>
<td>C(6) [deg]</td>
<td>0.6</td>
<td>0.7</td>
<td>0.2</td>
<td>2.0</td>
<td>0.4</td>
</tr>
<tr>
<td>C(7) [deg]</td>
<td>0.3</td>
<td>0.3</td>
<td>0.1</td>
<td>0.9</td>
<td>0.1</td>
</tr>
<tr>
<td>C(8) [dy]</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>C(9) [deg]</td>
<td>0.0</td>
<td>22</td>
<td>-80</td>
<td>120</td>
<td>43.5</td>
</tr>
</tbody>
</table>

assumed that the detrended measurements are no longer correlated after 8 days, a reasonable time scale considering that the periodicity of cold front passages is 3–10 days. The 8 day period also allows data at the measurement sites to influence the optimal analysis should clouds mask the areas of interest at the interpolation time.

Like $C_{gg}$, $C_{ee}$ is difficult to estimate. The term $C_{ee}$ may first be simplified by assuming that the errors at different positions are uncorrelated, thus all but the diagonal elements of $C_{ee}$ will be zero. Recall that $e$ represents the sum of the environmental and instrument errors. Assume the covariances of the two parts are additive, so $C_{ee}$ may be written as $[C_{ee}]_{\text{environment}} + [C_{ee}]_{\text{sensor}}$. 

Due to the constant changes in the natural environment, C(1) was assumed to be 0.9 which is consistent with the value chosen by Mariano and Brown (1992). In the Optimal Analysis package used in this study, $[\text{C}_{\text{ee}}]_{\text{environment}}$ was set to be equivalent to $1 - C(1)$, and thus $[\text{C}_{\text{ee}}]_{\text{environment}}$ became 0.1. The $[\text{C}_{\text{ee}}]_{\text{sensor}}$ is not derived directly from the AVHRR sensor, since the AVHRR sensor does not actually measure SST. Instead $[\text{C}_{\text{ee}}]_{\text{sensor}}$ is induced by the MCSST algorithm [McClain et al. 1985] as well as true sensor errors. Realistic MCSST estimation errors are $O(1 ^\circ C)$. By assuming a conservative MCSST field variance of $O(4 ^\circ C^2)$ (between October 1993 and December 1993, the MCSST field variances range from $O(4 ^\circ C^2)$ to $O(6 ^\circ C^2)$), $[\text{C}_{\text{ee}}]_{\text{sensor}}$ was initially set at 0.25.

In addition to accounting for the environment and sensor errors, $\text{C}_{\text{ee}}$ is also used as an adjustment parameter in the Optimal Analysis package [Mariano and Carter in preparation]. Recall that the Gauss-Markoff theorem requires that $\text{C}_{\text{zz}}$ in Eqn. 3.12 be positive definite. In the present study when $\text{C}_{\text{zz}}$ failed to satisfy this criterion, the diagonal elements of $\text{C}_{\text{ee}}$ would increase in increments of 0.1, and consequently increase the diagonal elements of $\text{C}_{\text{zz}}$. The resulting steeper shape of $\text{C}_{\text{zz}}$ signifies that $\text{C}_{\text{zz}}$ has become more positive definite. The diagonal elements of $\text{C}_{\text{ee}}$ continued to increase until $\text{C}_{\text{zz}}$ was positive definite.
Preliminary SST fields that were constructed by using different $C_{zz}$ values for each 1x1 degree block exhibit unrealistic SST gradients along the block boundaries. This problem was eliminated by using only one $C_{zz}$ for the entire Louisiana-Texas shelf. The median parameter values were selected as the shelf's parameter values. It was found that the medians represent the overall dispersion characteristics better than the means. In other Optimal Analysis studies, Mariano and Brown (1992) and Robinson et al. (1987) also designed their $C_{zz}$'s using the median parameter values.

Practical adjustments of optimal analysis

In practice, not all available measurements are needed to estimate $\hat{S}$, otherwise the run-time would be extremely long and the technique would be impractical. In this study, $Z$ consists only of a finite number, $n$, of influential data points, $Z$, with the highest correlation estimates. The size of $n$ can affect estimation error and run time. As $n$ increases, and if the additional observations lie not very distant from the estimation point, the second term on the right hand side of Eqn. 3.3, which measures the information content of observations [Mariano and Brown 1992; Bretherton et al. 1976], would increase. The squared error-estimate on the left hand side of the equation would decrease. Increasing $n$, on the other hand, means that more $Z$'s are taken into account and consequently computational costs increase.
An appropriate \( n \) is one where the \( \hat{S} \) field can be produced in a reasonable time with an acceptable error. Carter and Robinson (1987) empirically found that Optimal Analysis estimates are sensitive to noise for \( n < 6 \). At the other extreme, when \( n > 14 \), run-time increases drastically while the error remains relatively constant. Several \( n \)'s were tested in this study. An \( n \) of 15 was selected because the resulting \( \hat{S} \) field could be computed in a reasonable time with an acceptable error. In addition to selecting an appropriate \( n \), run time was further reduced by confining the search for the 15 \( Z \)'s to the time and space windows of \( S \) defined by zero-crossing spatial and temporal scales.

3.8 Definition of mean and eddy temperature flux

The OA package estimates \( \hat{T} \) one point at a time. Hereafter, \( \hat{T} \) will be written simply as \( T \) for convenience. The product of \( T \) and optimally estimated surface velocity \( U \) produces an estimate of the instantaneous temperature flux.

Using \( <> \) to symbolize the Eulerian mean, \( U' \) and \( T' \) are defined as \( U' \equiv U - <U>_E \) and \( T' \equiv T - <T>_E \). \( UT \) may be expanded as follows:

\[
UT = (U' + <U>_E)(T' + <T>_E)
\]

\[
= U'T' + U'<T>_E + <U>_E<T'_E + <U>_E<T>_E \quad [3.16]
\]

Therefore, \( <UT>_E \) may be expressed as

\[
<UT>_E = <U'T'_E + <U'_E<T>_E + <U>_E<T'_E + <U>_E<T>_E
\]

The velocity-temperature covariance term, \( U'T' \), may be
referred to as the eddy temperature flux. The Eulerian mean of $U'T'$ vectors is $<U'T'>_E$. Since $<T'>_E$ and $<U'>_E$ are both zero, $<UT>_E$ reduces to

$$<UT>_E = <U'T'>_E + <U>_E<T>_E \tag{3.17}$$

### 3.9 Bin selection

In this study, a bin refers to the time-space box in which the mean of the observations in the box is denoted by $<>_E$. Decorrelation scales were used to guide bin size selection. Bins with sizes smaller than decorrelation scales provide redundant information while possessing few degrees of freedom. On the other hand, bins with sizes larger than the decorrelation scales ignore small scale variabilities within the data in favor of increased statistical reliability. An aliasing problem may develop if the sub-bin variabilities are significant. Estimates of decorrelation scales and degrees of freedom were computed as follows.

The period over which the variable is correlated with itself may be estimated from the Lagrangian integral scale, $T_L$ [Tennekes and Lumley 1974]. It is assumed that the correlation decreases rapidly enough at large lag, $\tau$, so that $T_L$ is finite. The Lagrangian time scale of $u'T'$ will be denoted by $T_L(u'T')$ where $u'T'$ represents the east-west oriented component of $U'T'$. From Brink et al. (1991), $T_L(u'T')$ may be defined as follows:
\[ T_L(u'T') = \Sigma_0^\infty R(\tau) \]

where \( R(\tau) = \)

\[ \Sigma_0^N \left[ \frac{\{u(t)T(t) - \Sigma u(t)T(t)\} \{u(t+\tau)T(t+\tau) - \Sigma u(t+\tau)T(t+\tau)\}}{\sigma(t)\sigma(t+\tau)} \right] \]

The autocorrelation function of \( u'(t)T'(t) \) is denoted by \( R(\tau) \) and \( N \) is the number of observations. For an expression of \( T_L(v'T') \), where \( v'T' \) is the north-south oriented component of \( U'T' \), replace \( u'T' \) by \( v'T' \) in the equation for \( R(\tau) \). The autocorrelation functions for \( u'T' \) and \( v'T' \) are shown in Figure 3.8(a,b). The resulting \( T_L(u'T') \) and \( T_L(v'T') \) are 3.3 and 4.3 days, respectively. An independent sample was then defined as one 4.5 day segment of each buoy track.

Like \( T_L(u'T') \) and \( T_L(v'T') \), \( T_L(u') \) and \( T_L(v') \) can also be computed. \( R(\tau) \) for \( u' \) and for \( v' \) are shown in Figure 3.8c and d. The resulting \( T_L(u') \) and \( T_L(v') \) are 3.5 and 4.7 days, respectively. The similarities between \( u'T' \) and \( u' \) and also between \( v'T' \) and \( v' \) autocorrelation functions suggest decorrelation scales are primarily determined by the velocity structure. The time scales \( T_L(u') \) and \( T_L(v') \) can be further utilized to approximate the Lagrangian length scales, \( L_L \) of \( u' \) and \( v' \) where \( L_L \approx \sigma T_L \) [Poulain and Niiler 1989]. The length scales \( L_L(u') \) and \( L_L(v') \) are 54 and 50 km, respectively, suggesting that the structure of the velocity field is fairly complex. The scales are
Figure 3.8 Autocorrelation functions for the entire study area
(a) $u'T'$, (b) $v'T'$, (c) $u'$, (d) $v'$
slightly higher than the scales computed from the LATEX-A hydrographic cruise data, $O(30 \text{ km})$ [Yongxiang et al. 1996], but of the same order of magnitude as $L_L(u')$ and $L_L(v')$ over the North Atlantic [Freeland et al. 1975] and slightly smaller than $L_L(u')$ and $L_L(v')$ over the California shelf [Davis 1985].

Since $L_L(u'), L_L(v'), C(4)$, and $C(5)$ are $O(50) \text{ km}$, the spatial dimension of the bin was chosen at $1\times1$ degree, roughly twice the characteristic length scale. The time scale for this study was determined by computing the degrees of freedom for estimates in each $1\times1$ degree box at 30, 35, 40, ... 95 days. At time intervals greater than 65 days, there were at least 5 degrees of freedom associated with variable estimates in most boxes on the shelf. The natural timescale of the same order is the seasonal scale, 90 days. Thus, bin dimensions of 1 degree by 1 degree and 90 days were selected. Beginning in October 1993, the four seasons to be studied are: October–December, January–March, April–June, and July–September. The degrees of freedom of bin estimates, placed at the mean latitudes and longitudes, for each bin are depicted in Figure 3.9.

Note that the bin time dimension was selected based solely on statistical reliability. The period of 90 days is much longer than $T_L(u'T')$ and $T_L(v'T')$. The estimated Eulerian means cannot resolve sub-seasonal variabilities.

---

2. Mean Eulerian characteristics can be biased towards those of a sub-region with numerous samples; e.g., a convergence zone in the bin. The mean positions of the data points may indicate possible biases in the estimates of each mean Eulerian characteristic.
Figure 3.9a Degrees of freedom, Oct-Dec 1993. The numbers are placed at the mean latitudes and longitudes computed for all the samples in the 1° by 1° bin.
Figure 3.9b Same as Figure 3.9a, except for Jan-Mar 1994
Figure 3.9c Same as Figure 3.9a, except for Apr–Jun 1994
Figure 3.9d Same as Figure 3.9a, except for Jul-Sep 1994
An aliasing problem may develop if sub-seasonal variabilities are significantly larger than the lower frequency variabilities. The magnitude of this problem may be determined by inspecting spectra and cross spectra of the local, near surface velocities and temperatures.

3.10 Confidence ellipses of the bin average vectors

Whether or not $\langle UT \rangle_E$ and $\langle U'T' \rangle_E$ vectors adequately represent the vectors in the bin depends on the variability of the local vectors. Confidence regions for $\langle \rangle_E$ vectors can indicate how well they represent the means of the local background distribution. Among other factors, confidence regions are dependent on the underlying distribution of the samples.

Initially, no assumption was made concerning the distribution shape. For each bin, the Chebyshev's test [Feller 1957], a non-parametric test, was applied to determine the 80% confidence intervals of $\langle uT \rangle_E$ (and $\langle u'T' \rangle_E$) and $\langle vT \rangle_E$ (and $\langle v'T' \rangle_E$) separately. The north-south and east-west confidence intervals were then combined to make a box within which the expected mean lies.

To evaluate the reliability of a $\langle \rangle_E$ vector, the box's center was placed at the vector's tip. For most $\langle \rangle_E$ vectors, their lengths were completely covered by these boxes. Consequently, one cannot define even the sign of these $\langle \rangle_E$ vectors with any acceptable level of significance.
It was then assumed that the UT and U'T' vectors in each bin were normally distributed. For each bin, this assumption was tested by applying a bivariate normality test to the local UT and U'T', separately. A normal distribution of a bivariate sample is expected to have at least 50% of the squared-generalized-distances lie within $\chi^2(50\%)$ of the origin [Johnson and Wichern 1988]. Respectively, the squared-generalized-distance of UT and U'T' in the bin are:

$$\begin{align*}
[UT- <UT>_E]^T G^{-1} [UT- <UT>_E] \\
[U'T'- <U'T'>_E]^T H^{-1} [U'T'- <U'T'>_E],
\end{align*}$$

where G is the covariance matrix of UT; and H is the covariance matrix of U'T'.

It was found that, for most bins, more than 50% of the squared general distances for both UT and U'T' lie within $\chi^2(50\%)$ of the origin. Therefore, the assumption of bivariate normality cannot be rejected for either UT or U'T', in general.

The confidence region of the mean of a bivariate normal distribution is dependent on the lengths of the principal axes of the covariance matrix and the number of degrees of freedom present. The shape of the confidence region is actually an ellipse instead of a box.

The reliability of a $<>_E$ vector may be gauged by the relative size of the confidence ellipse. When the confidence ellipse blankets a large fraction of the $<>_E$ vector, the $<>_E$ vector does not adequately represent the back-
ground vector field. Whereas an ellipse that is oriented along $\langle \rangle_E$ vector and covers a small portion of the $\langle \rangle_E$ vector indicates that the $\langle \rangle_E$ vector adequately represents the background vector field.

Additionally, the confidence ellipse also describes the spatial structure of internal (sub-bin) variability. More specifically, the confidence ellipse's shape and principal orientation symbolize the internal variability's degree of anisotropy and preferred orientation. The relative magnitude of the internal variability may also be important. The confidence ellipse does not represent this magnitude because the size of the confidence ellipse is modified by the number of degrees of freedom. The principal axes will have the same shape and orientation as the confidence ellipse but their size will be independent of number of degrees of freedom. Therefore, the principal axes were used to signify the internal variability's degree of anisotropy, preferred orientation, and its magnitude. As shall be seen in the next chapter, these characteristics may be linked to some physical properties of the flux.

3.11 Rayleigh test and confidence interval for preferred direction

The Rayleigh test [Mardia 1972] was applied to $\mathbf{UT}$ and $\mathbf{U'T'}$ vectors to test the null hypothesis: a preferred direction does not exist at $\alpha = 0.05$. Since the mean direction of the data vectors is not necessarily the same as the direction of the mean of these vectors, the vectors were normalized to unit length so that all directions pos-
sess equal weights. The Rayleigh test assumes the directions have a von Mises distribution. The von Mises distribution is unimodal and symmetric about the mean direction. The mode and antimodes for a von Mises distribution are at the mean direction and mean direction ± π, respectively. The mode/antimode ratio is e^{2k}, where k is the concentration parameter. As k increases, more directions are tightly clustered near the mean direction.

If k is significantly different from zero, the null hypothesis can be rejected and the mean direction may be referred to as a statistically preferred direction. The concentration parameter is significantly different from zero when the magnitude of the mean normalized vector exceeds a critical value which is dependent on the number of degrees of freedom. Only when a preferred direction exists, was the 95% confidence range about the preferred direction computed. Should the direction of the $<E_x$ vector lie within the confidence interval, it would adequately reflect the preferred direction.

3.12 Student's t and Kendall's Tau tests

Two other tests were applied to strengthen the description of observations. A student's t test was applied to determine whether or not any selected pairs of mean eddy heat fluxes (Section 4.4) were significantly different given a significance level, α. The test's normality assumption was examined using the quantile-quantile (q-q) test [Johnson and Wichern 1988]. The q-q test compares the
sorted observation $x_j$ with quantile $q_j$ of an expected normal distribution.

$$P[Z \leq q(j)] = \int_{-\infty}^{q(j)} \left( \frac{1}{\sqrt{2\pi}} \right) e^{-z^2/2} \, dz$$

where $j$ is the sample count. As $j$ increases, probability also increases. The correlation coefficient, $r_Q$, of the observed and expected quantiles is

$$r_Q = \frac{\left[ \sum (x(j) - \langle x \rangle)(q(j) - \langle q \rangle) \right]}{\sqrt{\sum (x(j) - \langle x \rangle)^2} \sqrt{\sum (q(j) - \langle q \rangle)^2}}$$

The normality assumption may be suspect if $r_Q$ fails to exceed the q-q plot critical values [Table 4.2, Johnson and Wichern 1988].

Additionally, Kendall's Tau ($\tau_K$) was used to test the null hypothesis: the two samples do not have a trend given a significance level [Siegel and Castellan, 1988; Press et al., 1988]. Kendall's Tau may be expressed as follows:

$$\tau_K = \frac{c-d}{\sqrt{c+d-t_1} \sqrt{c+d-t_2}}$$

where $c$ is the count when the rank difference of two variables ($v_1(y_1)$ and $v_1(y_2)$) and the rank difference of two corresponding variables ($v_2(y_2)$ and $v_2(y_2)$) have the same sign. Conversely, $d$ is the count when these two rank differences have opposite signs; $t_1$ is the count when the rank difference between the two $v_1$'s; and $t_2$ is the count when the rank difference between the two $v_2$'s is zero. When $n$ is
greater than 10, the null distribution of $\tau_k$ is approximately normal, the expected mean is zero, and $\sigma_k^2 = (4n + 10)/(9n(n-1))$. The number of standard deviations from zero, $\tau/\sigma_k$, is used to determine probability of independence. A statistical table of probabilities for $n$ is less than 10 may be found in Siegel and Castellan (1988). The alternate hypothesis is that the two variables may be dependent. Positive and negative $\tau_k$'s indicate increasing and decreasing trends in the data, respectively.

3.13 Relative importance of eddy heat flux

Lack of air-sea humidity measurements prevent the estimation of latent heat flux, usually an important term in the heat budget equation. Thus, an analysis of the shelf's heat budget could not be made. The relative importance of net eddy heat flux in the seasonal heat budget equation was gauged from the the ratio of net eddy heat flux and net radiative flux. Over the Gulf of Mexico, net radiative flux is an important heat gain term in the low frequency heat budget [Etter 1983]. Assuming the shelf's net radiative flux may be represented by the Gulf's net radiative flux, the monthly heat fluxes per unit area estimated by Etter (1983) were combined to produce estimates of the shelf's seasonal radiative heat flux rate of 97, 178, 161, and 60 Watts m$^{-2}$ for October-December, January-March, April-June, and July-September, respectively.

The net eddy heat flux over the Louisiana-Texas shelf can be estimated by integrating eddy heat fluxes in
the mixed layer around a closed boundary. Such a representation is valid under the following assumptions: (1) during the 90 day interval, the net change in mass and composition of the shelf waters may be neglected (i.e., heat transported by 1 ton of 30 °C water should not be compared with heat transported by 30 tons of 1 °C water, Montgomery 1974); (2) the vertical shear of eddy heat flux is relatively weak so that the near surface eddy heat flux can represent the sub-surface eddy heat fluxes [W. Wiseman, personal communication]; and (3) the eddy heat flux integrated over the mixed layer accounts for the majority of the total depth integrated eddy heat flux [Bryden et al. 1980]. Thus, the net eddy heat flux over the shelf may be written as:

\[ \int H \varphi_{cp} n' T' \, ds \]  

[3.18]

where \( n' T' \) is a component of \( U' T' \) that is normal to a defined boundary \( s \) and \( H \) is the mixed layer depth. The smallest increment of \( s \), is denoted by \( ds \). It was set at 1 km. Since \( \varphi_{cp} \) may vary at most by 1.6% (Sec. 2.3), it was taken out of the integral. The value of \( \varphi_{cp} \) over the Gulf of Mexico is 0.966 cal cm\(^{-3}\) °C\(^{-1}\) [Etter 1983].

The boundaries are made up of several longitudinal and latitudinal segments, so \( n' T' \) is either the north-south or east-west components of \( U' T' \). The \( n' T' \) values were optimally estimated from \( U' T' \) vectors estimated earlier. Error-estimates of \( n' T' \) were computed in the same manner as the error-estimates of SST "observations" (see Sec 4.1),
using Eqn 3.13. The total error-estimate was determined by replacing $n'T'$ in Eqn 3.18 with its error-estimate.

A desirable boundary is one that demarcates the largest area possible, so that the integral best represents the eddy heat flux of "the shelf" rather than some local region. At the same time, a statistically desirable boundary is one that continuously passes through clusters of observations so that the error-estimates will be small. With these two desirable properties in mind, 10 km gridded error fields of $u'T'$ were constructed for each season to assist in selecting appropriate boundaries. I drew the boundaries just inside the outer edges of the observations (Fig 3.10). Two additional boundaries were also made to investigate the seasonal variability of net eddy heat fluxes. The two boundaries were drawn to minimize the error of the first three seasons and all four seasons, respectively (Figs 3.11a, b). The second boundary was drawn only around a central region of the Gulf because it is the only common area with abundant "observations" in all seasons.

The mixed layer depths, $H$, chosen for this study were determined from the depths of maximum Brunt Vaisala frequencies during LATEX-A cruises. For each boundary segment (ends of segments are marked by dark dots in Figs 3.10, 3.11), a single $H$ was assigned for all ds increments in that segment. The mixed layer depths for October-December 1993, April-June 1994, and July-September 1994 were
Figure 3.10a Boundary chosen for the determination of net eddy heat flux off the shelf, for the first season, Oct-Dec 1993. The grey dots represent locations where $U'T'$ was calculated during this period.
Figure 3.10b Same as Figure 3.10a, except for Jan-Mar 1994.
Figure 3.10c Same as Figure 3.10a, except for Apr–Jun 1994.
Figure 3.10d Same as Figure 3.10a, except for Jul-Sep 1994.
Figure 3.11a Common boundary chosen for the determination of seasonal variability of net eddy heat flux off the shelf, for the first three seasons, Oct 93-Jun 94.
Figure 3.11b Same as Figure 3.11a, except for the four seasons, Oct 93-Sep 94
obtained from mixed layer depths of the late November 1993, late April 1994, and late July 1994 cruises, respectively. The mixed layer depths between January–March 1994 were obtained from a combination of the February 1993 and the late November 1994 cruises, because the February cruise covered only the eastern portion of the study area. Note that cruise data may sometimes not represent the seasonal mean values. They, however, are the only corresponding subsurface data available.

3.14 Surface temperature flux divergence

Spatial gradients of $\langle UT \rangle_E$ vectors can be used to estimate surface temperature flux divergence. Since $qC_p$ is relatively constant, the divergence term may indicate the relative amount of heat gain/loss from a region. Divergence of temperature flux at the center of 2 by 2 degree-bins are estimated by:

$$\Delta \langle UT \rangle_E / \Delta X + \Delta \langle VT \rangle_E / \Delta Y =$$

$$\frac{1}{2} \left\{ \frac{(\langle UT \rangle_E)_{i+1,j} - (\langle UT \rangle_E)_{i,j}}{X_{i+1,j} - X_{i,j}} + \frac{(\langle UT \rangle_E)_{i+1,j+1} - (\langle UT \rangle_E)_{i,j+1}}{X_{i+1,j+1} - X_{i,j+1}} \right\} +$$

$$\frac{1}{2} \left\{ \frac{(\langle VT \rangle_E)_{i,j+1} - (\langle VT \rangle_E)_{i,j}}{Y_{i,j+1} - Y_{i,j}} + \frac{(\langle VT \rangle_E)_{i+1,j+1} - (\langle VT \rangle_E)_{i+1,j}}{Y_{i+1,j+1} - Y_{i+1,j}} \right\}$$

where $X_{i,j}$ and $Y_{i,j}$ are the mean latitude and longitude of bin$_{i,j}$.

Along with determining whether the shelf gains or loses heat, insight into the statistics and mechanisms of heat gain/loss may be identified through a series of decom-
positions and order of magnitude comparisons of these terms.

The gradient term $\Delta \langle UT \rangle_E$ may be decomposed into $\Delta \langle U \rangle_E \langle T \rangle_E$ and $\Delta \langle U' T' \rangle_E$. By comparing the magnitudes of the mean and eddy parts, the dominant part may be identified. The dominant part may once again be decomposed into two terms (analogous to partial derivatives) that are induced by different mechanisms: a velocity-gradient term and a temperature-gradient term. For example, $\Delta \langle U \rangle_E \langle T \rangle_E / \Delta X$ is approximately equal to $\langle T \rangle_E \Delta \langle U \rangle_E / \Delta X + \langle U \rangle_E \Delta \langle T \rangle_E / \Delta X$, where $\langle \rangle_E = 0.25 [\langle \rangle_E i, j + \langle \rangle_E i, j+1 + \langle \rangle_E i+1, j + \langle \rangle_E i+1, j+1]$. The terms $\Delta \langle U \rangle_E / \Delta X + \Delta \langle T \rangle_E / \Delta X$ are finite approximations of $\partial u / \partial x$ and $\partial T / \partial x$, respectively. If $X_{i,j}$ and $X_{i+1,j}$ are located at the center of each 1x1 degree bin, then the finite approximation scheme is called centered difference and it is second-order accurate (e.g., Pond and Pickard 1991). Since $X_{i,j}$ and $X_{i+1,j}$ are position at the sample mean latitude and longitude, the finite approximation is close to second-order accurate.
RESULTS AND DISCUSSION

4.0 Chapter outline

In this chapter, the results of the study are described and discussed. Beginning with Section 4.1, SST estimation errors are presented. Section 4.2 describes the general pattern of the \( \langle UT \rangle_E \) vectors and their internal variability. Physical processes that may have induced the variability are then discussed. The format of Section 4.3 is similar to that of Section 4.2, except that the focus is shifted from \( \langle UT \rangle_E \) to \( \langle U'T' \rangle_E \). Section 4.4 addresses the relative importance of the eddy flux term in the low frequency heat budget equation and describes the seasonal variability of net eddy heat flux. Section 4.5 presents results of the surface temperature divergence calculations. Decomposition of surface temperature flux divergence leads to the suggestion of the dominant processes responsible for the shelf's heat gain/loss.

4.1 Estimation error

Error-estimates of the SST "observations" indicate the reliability of these "observations". Whether or not the error is acceptable depends on the study's objective. To investigate the variability at the bin scales selected, \( O(100) \) km and \( O(90) \) days, SST's of meso-scale features should be well estimated. The error-estimates are best obtained by comparing SST "observations" with independent synoptic SST data (not the AVHRR data because SST "observa-
tions" were derived from them). Unfortunately, for this study, independent synoptic data were unavailable.

However, assuming that our correlation function of the detrended SST field adequately represents the true correlation function, then the squared error-estimates normalized by their variance may be computed (Eqn 3.13). The normalized squared error-estimate from the SST movie are relatively constant at \( O(0.25) \) as shown in Figure 4.1. The dates of the error-estimates are 1993-94 Julian dates. The 1993-94 Julian day to calendar day table is shown in the appendix. The error-estimates are of the same order of magnitude as the normalized squared error-estimate found by Mariano and Brown (1992) for an optimally interpolated SST field that reproduced multiple water masses observed in the western Atlantic.

Given the variance of the error-estimate, the absolute error may be computed. From Eqn 3.3, the error-estimate variance is the field variance minus a term that describes the information content. Thus, a conservative error-estimate variance is the field variance [A.J. Mariano personal communication; R.E. Macchiavelli personal communication]. The conservative daily mean error-estimates for the movie are shown in Figure 4.2, while the monthly mean error-estimates for SST "observations" at buoy-velocity positions are shown in Table 4.1. The errors peak in the winter when the field variances are relatively high (Figure 4.3). High variance implies that SST structure is complex.
Figure 4.1 Night-time mean $E[(\hat{T}-T)(\hat{T}-T)^T]$ for the entire Northwestern Gulf of Mexico normalized by the variance of the $(\hat{T}-T)$ field. The 1993-94 Julian days to calendar day table is shown in the appendix.
Figure 4.2 Night-time mean conservative error-estimate for the entire Northwestern Gulf of Mexico. The 1993-94 Julian days to calendar day table is shown in the appendix.
Figure 4.3 Night-time variance of the Northwestern Gulf of Mexico T field. The 1993-94 Julian days to calendar day table is shown in the appendix.
The errors reach their minimum values in the summer when SST's are more uniform. The conservative error-estimates are comparable to the error, in practice, resulting from the application of the widely accepted MCSST algorithm, $O(1 ^\circ C)$. These two indirect comparisons suggest that the errors are acceptable.

Error-estimates of UT vectors were not computed because the error-estimates of the individual $U$ vectors were unavailable. The reliability of $\langle UT \rangle_E$ and $\langle U'T' \rangle_E$ vectors may be evaluated through their confidence ellipses which will be described in the following sections.

Table 4.1 Monthly mean estimation errors from SST "observations" at buoy-velocity positions.

<table>
<thead>
<tr>
<th>Month</th>
<th>Number of Observations</th>
<th>Error-estimate(°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>October</td>
<td>432</td>
<td>0.79</td>
</tr>
<tr>
<td>November</td>
<td>1923</td>
<td>1.39</td>
</tr>
<tr>
<td>December</td>
<td>3085</td>
<td>1.68</td>
</tr>
<tr>
<td>January</td>
<td>3266</td>
<td>1.76</td>
</tr>
<tr>
<td>February</td>
<td>2254</td>
<td>1.54</td>
</tr>
<tr>
<td>March</td>
<td>1868</td>
<td>1.22</td>
</tr>
<tr>
<td>April</td>
<td>1418</td>
<td>1.18</td>
</tr>
<tr>
<td>May</td>
<td>746</td>
<td>0.67</td>
</tr>
<tr>
<td>June</td>
<td>1046</td>
<td>0.56</td>
</tr>
<tr>
<td>July</td>
<td>1156</td>
<td>0.45</td>
</tr>
<tr>
<td>August</td>
<td>1123</td>
<td>0.36</td>
</tr>
<tr>
<td>September</td>
<td>982</td>
<td>0.36</td>
</tr>
</tbody>
</table>
4.2 Observation of $\langle UT \rangle_E$

**General pattern**

The $\langle UT \rangle_E$ vectors for each season are displayed in Figure 4.4. For each bin, the origin of the vector is placed at the sample mean latitude and longitude. In general, the $\langle UT \rangle_E$ vectors on the shelf are oriented along the local bathymetry. Using the Kendall’s Tau test, the trend that innershelf (<50 km from shore) magnitudes west of 92 $^\circ$W increase downcoast is significant at $\alpha = 0.10$ in all seasons but October-December (probability, $p$, for seasons 1, 2, 3, and 4 = 0.167, 0.007, 0.086, and 0.025, respectively). The trend that the magnitudes decrease offshore is significant at $\alpha = 0.10$ in all seasons but July-September ($p = 0.001, 0.001, 0.088, 0.317$). The overlaid $\langle UT \rangle_E$ vector field, shown in Figure 4.4e, suggests that $\langle UT \rangle_E$ vectors near the south Texas shelf are directed offshore, then turn upcoast along the shelf break. This finding supports Cochran and Kelly (1986) and Dinnel and Wiseman (1986)'s conclusions: freshwater originated from the Mississippi and Atchafalaya Rivers is advected offshore near the southern end of the Texas shelf and carried upcoast along the shelf break. Furthermore, heat may advect across the shelf break. Along the shelf break at 95.5 $^\circ$W and 93.5 $^\circ$W, the $\langle UT \rangle_E$ vectors are directed into and out of the shelf, respectively. Along the shelf break at 91.5 $^\circ$W and around 89.5 $^\circ$W, the $\langle UT \rangle_E$ vectors are directed into and out of the

3. October-December, January-March, April-June, and July-September denote the four seasons.
shelf, respectively. The magnitudes of the $\langle \text{UT} \rangle_E$ vector pair at 95.5 °W and 93.5 °W are smaller than the pair at 91.5 °W and around 89.5 °W. Elliott (1982) observed that anticyclonic rings which were shed from the Loop Current migrated westward near 26 °N. Since the field in Figure 4.4e consists of vectors from different seasons, the westward decreasing flux magnitudes may reflect both ring age (i.e., distance from the Loop Current) and shelf-slope exchange seasonality (i.e., warmer SST in the summer).

Along with the general pattern observed, the vectors' confidence ellipses (Fig 4.5), confidence interval for preferred direction (Fig 4.6), and principal axes (Fig 4.7) also provide valuable information. The trend that principal axes become more isotropic offshore is significant at $\alpha = 0.10$ in all seasons but October-December ($p = 0.169, 0.005, 0.014, 0.027$). West of the Atchafalaya Bay, the $\langle \text{UT} \rangle_E$ vectors are directed downcoast. The vectors are large compared to the size of the 95% confidence ellipses, and their directions lie within the 95% confidence interval for preferred directions. The principal axes are anisotropic and elongated parallel to the local bathymetry. The average major/minor length ratio within the shelf west of 92 °W is 2.44.

East of the bay, on the other hand, the $\langle \text{UT} \rangle_E$ vectors are directed upcoast but preferred directions are absent. The vectors are small compared to the size of the 95% confidence ellipses. The principal axes are more isotrop-
Figure 4.4a Field of instantaneous temperature flux Eulerian mean, \( <\text{UT}>_E \), for Oct-Dec 93. The unit is °C m/s.
Figure 4.4b Same as Figure 4.4a, except for Jan-Mar 94.
Figure 4.4c Same as Figure 4.4a, except for Apr-Jun 94.
Figure 4.4d Same as Figure 4.4a, except for Jul-Sep 94.
Figure 4.4e Same as Figure 4.4a, for all four seasons.
(Oct-Dec 93, Jan-Mar 94, Apr-Jun 94, and Jul-Sep 94)
Figure 4.5a Field of instantaneous temperature flux Eulerian mean, $\langle UT \rangle_E$, for Oct-Dec 93, with the axes of confidence ellipses in gray.
Figure 4.5b Same as Figure 4.5a except for Jan-Mar 94.
Figure 4.5c Same as Figure 4.5a except for Apr-Jun 94.
Figure 4.5d Same as Figure 4.5a except for Jul-Sep 94.
Figure 4.6a Field of instantaneous temperature flux Eulerian mean, $\langle UT \rangle_E$, for Oct-Dec 93, with the 95% confidence interval of the preferred direction in gray.
Figure 4.6b Same as Figure 4.6a, except for Jan-Mar 94.
95% confidence interval of preferred direction

Figure 4.6c Same as Figure 4.6a, except for Apr-Jun 94.
Figure 4.6d $\langle UT \rangle_E$ and 95% confidence interval, Jul-Sep 94
Figure 4.7a Principal axes of UT in each 1° by 1° bin, Oct-Dec 93. The major/minor axis ratio is in plain text; and the area of an ellipse with the calculated major and minor axes given in bold text. The units of the ellipse area are (°C m/s)².
Figure 4.7b Same as Figure 4.7a, except for Jan-Mar 94.
Figure 4.7c Same as Figure 4.7a, except for Apr-Jun 94.
Figure 4.7d Same as Figure 4.7a, except for Jul-Sep 94.
ic, especially in bins near the river mouths. East of 92 °W, the average major/minor length ratio is 1.75.

**Seasonal deviations from this pattern**

The forementioned pattern is not time independent. Deviations were encountered both west and east of the Atchafalaya Bay in April–June, and July–September. In both seasons, the vectors north of 29 °N and west of the bay are small compared to the 95% confidence ellipses (Figs 4.5c,d). Furthermore, in July–September, north of 29 °N, preferred directions no longer exist (Figs 4.6d). For the bin immediately south of Galveston Bay, in particular, the lack of a statistically significance <UT>_E may result from the convergence of upcoast and downcoast coastal currents previously reported by Smith (1980).

Although over the south Texas inner shelf, preferred directions exist, the widening of the associated confidence intervals beginning in April implies that the directions vary more in April–June and July–September than they do in October–December and January–March. The widening of the confidence interval may also reflect the data sparsity over the south Texas shelf. Areas of ellipses drawn around the principal axes (hereafter referred to as size of principal axes) in April–June and July–September (Figs 4.7d) are larger than their counterparts in October–December and January–March (Figs 4.7a,b). The mean sizes of shelf principal axes (area) for October–December, Janu-
ary—March, April—June, and July—September are 1149, 932, 2801, and 4408 C² m²s⁻², respectively.

Deviations from the norm are also found east of Atchafalaya Bay. In April—June, the \( <\text{UT}>_E \) vectors are statistically significant (Fig 4.5c) and their directions lie within the 95% confidence interval for preferred direction (Fig 4.6c). The principal axes are anisotropic and elongated parallel to the local bathymetry (Fig 4.7c). In July—September, except for the region adjacent to the mouth of the Atchafalaya Bay, the \( <\text{UT}>_E \) directions lie within the 95% confidence interval for preferred direction.

Discussion

In general, the principal axes are often isotropic near the freshwater sources. Their sizes are particularly large in the summer. These patterns led to the suggestion of two hypotheses: (1) the level of stratification partly determines the degree of anisotropy and the degree of bathymetric steering; and (2) mixed layer thickness (MLD) partly controls the size of the principal axes.

Since stratification is a barrier between the upper and lower layers, a weakly stratified water column allows the upper layer flow to 'feel' the bottom and subsequently its orientation be guided by bottom topography. A highly stratified column, on the other hand, can isolate the upper and lower layers so that the orientation of the upper layer flow becomes independent of bottom topography. One measure of stratification is the Burger Number, which may be ex-
pressed as the squared ratio of the Rossby deformation radius and a geometric length scale, $L$ [Pedlosky 1987; Brink 1989]:

$$S = \frac{(N^2 D^2 f^{-2})}{L^2}$$

where $S$ is the Burger number, $D$ is the water depth, $N$ is the Brunt Vaisala frequency, and $f$ is the Coriolis parameter. Depending on the "observation" position, $L$ was the smaller value for the cross-shelf distance [Münchow and Garvine 1993a] or the Lagrangian velocity scale, 50 km. In this study, the Brunt-Vaisala frequency, $N$, was defined as the maximum Brunt-Vaisala frequency, so that $S$ represents the most stratified condition. Caution must be taken when interpreting $S$, since it does not account for velocity scale or distance from a freshwater source. Burger Numbers of $O(10^0)$ and $O(10^{-1})$ suggest that the water column is moderately and weakly stratified, respectively [Hogg 1973; Münchow and Garvine 1993a].

The second hypothesis links the size of the principal axes to mixed layer thickness (MLD) because the flow's sensitivity to external forcing depends on its MLD. A thinner mixed layer is expected to respond more rapidly to a given wind impulse than a thicker mixed layer. Therefore, the size of the principal axes of a thin mixed layer is expected to be larger than that of a thicker mixed layer. Data from the LATEX-A cruises were used to compute $S$ and MLD shown in Figures 4.8 and 4.9, respectively.
Figure 4.8a Contours of the Burger Number for the LATEX-A cruise, Nov 7-21, 1993. The dots are locations of the CTD sampling stations.
Figure 4.8b Same as Figure 4.8a, except for Feb 6-13, 1993.
Figure 4.8c Same as Figure 4.8a, except for Apr 24-May 7, 1994
Figure 4.8d Same as Figure 4.8a, except for Jul 27-Aug 5, 1994
Figure 4.9a Mixed layer depth contours, in meters, for the LATEX-A cruise, Nov 7-21, 1993. The dots are locations of the CTD sampling stations.
Figure 4.9b Same as Figure 4.9a, except for Feb 6-13, 1993.
Figure 4.9c Same as Figure 4.9a, except for Apr 24-May 7, 1994.
Figure 4.9d Same as Figure 4.9a, except for Jul 27-Aug 5, 1994.
Burger Number and mixed layer thickness variability

In early November 1993, except for the region near the Atchafalaya River mouth, the water column was generally weakly stratified north of 28.3 °N in water depths shallower than 50 m and moderately stratified offshore (Fig 4.8a). However, the trend that stability strengthens offshore is not significant at α = 0.10 (p=0.131). The trend that MLD thickens offshore is significant at α = 0.10 (p=0.001).

During early February 1992, a mid shelf patch is weakly stratified water was observed (Fig 4.8b). South of the Atchafalaya Bay, the water column was moderately stratified and its mixed layer was thin (Fig 4.9b). The highly stratified and thin mixed layer south of Terrebonne Bay, LA, around 28.5 °N may be induced by the Mississippi River plume. The trend that stability weakens away from the Mississippi River mouth is significant at α = 0.10 (p=0.001). The trend that MLD thickens offshore is significant at α = 0.10 (p=0.001).

In late April 1994, the weakly stratified patch over the Louisiana shelf may represent an anomaly rather than a seasonal mean condition (Fig 4.8c). The trend that stability strengthens offshore is significant at α = 0.10 (p=0.030). The trend that innershelf (<50 km from shore) stability west of 92 °W strengthens downcoast is significant at α = 0.10 (p=0.001). The mixed layer is thinnest along the innershelf. The trend that MLD thickens offshore is significant at α = 0.10 (p=0.001).
During late July 1994, the northern inner Texas shelf is weakly stratified (Fig 4.8d). A highly stratified water column occurs southwest of the Mississippi River mouth and over the south Texas shelf. The trend that stability strengthens offshore is significant at $\alpha = 0.10$ ($p=0.001$). The trend that stability weakens away from the Mississippi River mouth is significant at $\alpha = 0.10$ ($p=0.001$). The trend that MLD thickens offshore is significant at $\alpha = 0.10$ ($p=0.001$). The mixed layer is thinnest west of the Atchafalaya Bay. The trend that MLD thickens downstream from the Mississippi River mouth is significant at $\alpha = 0.10$ ($p=0.054$).

Since the cruises and the $\langle \rangle_{\text{E}}$ fields do not coincide, $S$ and MLD were indirectly compared with principal axis characteristics. Caution is necessary in comparing synoptic cruise and seasonal mean fields. The link between stability level and degree of anisotropy is supported in January-March and July-September. The link between MLD and the size of the principal axes is not supported in any seasons. The comparisons are described below.

In October-December, all principal axes are anisotropic and elongated parallel to the local bathymetry (Fig 4.7a). The mean and $\sigma$ of the major/minor length ratios are 2.39 and 0.76, respectively. The minimum and maximum ratios of 1.35 and 4.45 occur immediately west of the Atchafalaya Bay ($29.15^\circ\text{N}, 92.69^\circ\text{W}$) and over the inner south Texas shelf ($26.70^\circ\text{N}, 97.07^\circ\text{W}$), respectively. The trend
that principal axes become more isotropic offshore is not significant at $\alpha = 0.10$ (p=0.169), thus the first hypothesis is not supported. The second hypothesis may be rejected because the size of the principal axes may actually increase offshore.

In January-March, consistent with the first hypothesis, the least anisotropic principal axes, length ratios of $0(1.5)$, reside in the 2 bins west of the Atchafalaya Bay and Mississippi River mouths (Fig 4.7b). Excluding these two 2-bins, the length ratio's mean and $\sigma$ are 2.37 and 0.41 suggesting that the principal axes are anisotropic. Furthermore, the principal axes are elongated parallel to the local bathymetry. The most anisotropic principal axes occur over the south Texas shelf. Supporting the first hypothesis, the trend that principal axes become more anisotropic away from the Mississippi River mouth is significant at $\alpha = 0.10$ (p=0.005). Not supporting the second hypothesis, the size of the principal axes may actually increase offshore.

In April-June, the principal axes are anisotropic but highly variable and elongated parallel to the local bathymetry (Fig 4.7c). The mean and $\sigma$ of the length ratios are 2.63 and 1.21, respectively. The minimum and maximum length ratios, 1.34 and 5.65, occur southwest of the Atchafalaya Bay (29.67 °N, 92.56 °W) and south of Matagorda Peninsula (28.29 °N, 96.32 °W), respectively. Not supporting the first hypothesis, the innershelf (<50 km from shore)
principal axes may actually become more anisotropic down-coast. Not supporting the second hypothesis, the trend that the size of the principal axes decreases offshore is not significant at $\alpha = 0.10$ ($p=0.159$).

Compared to other seasons, the least anisotropic principal axes occur in July-September. The length ratio's mean and $\sigma$ are 2.20 and 0.70, respectively. Supporting the first hypothesis, the least and most anisotropic principal axes coincide with the high and low stability regions, respectively. Along the shelf, the least anisotropic principal axes, length ratios of 0(1.5), occur east of 92 °W, where a significant part of the variability may be induced by the Atchafalaya and Mississippi River plumes. For example, the major axis in the bin immediately south of the Atchafalaya Bay is not elongated parallel to the local bathymetry, but instead along the Atchafalaya River plume often seen on satellite imagery. The more anisotropic principal axes are distributed along the innershelf. They are also elongated parallel to the local bathymetry. The length ratios' mean and $\sigma$ of the innershelf (<50 km from shore) principal axes west of 92 °W are 2.71 and 0.54, respectively. Supporting the first hypothesis, the trend that these principal axes north of 28 °N become more anisotropic away from the Mississippi River mouth is significant at $\alpha = 0.10$ ($p=0.002$). Not supporting the second hypothesis, the trend that size of the principal axes north of 28 °N de-
creases away from the Mississippi River mouth is not significant at $\alpha = 0.10$ ($p=0.196$).

The lack of innershelf $\langle UT \rangle_E$ vectors directed up-coast in July-September should be noted. During the study period, innershelf $UT$ vectors did reverse and flowed up-coast from late May through July. The upcoast characteristics were spread out in April-June and July-September and consequently do not reverse the seasonal means. In July-September, the relatively strong downcoast $\langle UT \rangle_E$ vectors over the Texas shelf consist mainly of vectors from mid September 1994. However, redefining the seasons, so that innershelf $\langle UT \rangle_E$ vectors are directed up-coast, would require making a weak assumption: the October 1993 velocity field adequately represents the October 1994 velocity field.

4.3 Observation of $\langle U'T' \rangle_E$

General pattern

The $\langle U'T' \rangle_E$ vectors generally flow upcoast (Fig 4.10 series), opposing the $\langle UT \rangle_E$ vectors. The trend that the innershelf (<50 km from shore) magnitudes west of 92 °W increase downcoast is significant at $\alpha = 0.10$ in October-December and July-September ($p=0.019, 0.167, 0.298, 0.061$). The trend that magnitudes decrease offshore is significant at $\alpha = 0.10$ in the same two seasons ($p=0.027, 0.301, 0.159, 0.066$). The mean magnitudes are highest in April-June at 0.187 °C ms$^{-1}$ and lowest in July-September at 0.030 °C ms$^{-1}$. 
The $<U'T'>$ vectors, their confidence ellipses, confidence intervals for preferred directions, and principal axes are shown in Figures 4.11 through 4.13. The trend that principal axes become more isotropic offshore is significant at $\alpha = 0.10$ only in January-March and April-June ($p = 0.239, 0.044, 0.025, 0.317$). West of Galveston Bay, the $<U'T'>$ vectors are larger than the associated 80% confidence ellipses and their direction lies within the 95% confidence interval for preferred direction. The principal axes are anisotropic and elongated parallel to the local bathymetry. The mean major/minor length ratio west of 95 °W (on the shelf) is 2.99. East of Galveston Bay, on the other hand, the $<U'T'>$ vectors are smaller than the 80% confidence ellipses. Preferred directions do not exist at the 95% confidence level. The mean major/minor length ratios east of 95 °W (on the shelf) is 2.10. Thus, in the eastern region, the principal axes are more isotropic and less dependent of bathymetric steering.

Seasonal deviations from this pattern

The general pattern described above is not constant in time. Only the major deviations encountered between seasons are described below. During October-December, in twelve of the thirteen bins (on the shelf), the $<U'T'>$ vectors are smaller than the 80% confidence ellipses (Fig 4.11a) and preferred directions are absent (Fig 4.12a). In January-March and April-June, no deviation from the general pattern is observed. In July-September, innershelf $<U'T'>$
Figure 4.10a Field of eddy temperature flux Eulerian mean, $\langle U'T' \rangle_E$, in $1^\circ \times 1^\circ$ bins, Oct-Dec 93. The unit is $^\circ$C m/s.
Figure 4.10b Same as Figure 4.10a, except for Jan-Mar 94.
Figure 4.10c Same as Figure 4.10a, except for Apr-Jun 94.
Figure 4.10d Same as Figure 4.10a, except for Jul-Sep 94.
Figure 4.10e Same as Figure 4.10a, except for Oct- Dec 93, Jan-Mar 94, Apr-Jun 94, Jul-Sep 94.
Figure 4.11a Field of eddy temperature flux Eulerian mean, $<U'T'>_E$, in 1°x1° bins, Oct-Dec 93, with the axes of confidence ellipses in gray.
Figure 4.11b Same as Figure 4.11a, except for Jan-Mar 94.
Figure 4.11c. Same as Figure 4.11a, except for Apr-Jun 94.
Figure 4.11d Same as Figure 4.11a, except for Jul-Sep 94.
Figure 4.12a Field of eddy temperature flux Eulerian mean, $\langle UT' \rangle_E$, in $1^\circ \times 1^\circ$ bins, Oct-Dec 93, with the 95% confidence interval of the preferred direction in gray, Oct-Dec 93.
Figure 4.12b Same as Figure 4.12a, except for Jan-Mar 94.
Figure 4.12c Same as Figure 4.12a, except for Apr-Jun 94.
Figure 4.12d Same as Figure 4.12a, except for Jul-Sep 94.
Figure 4.13a Principal axes of U'T' in 1°x1° bins, Oct-Dec 93.
The major/minor axis ratio is in plain text; and the area of an ellipse with the calculated major and minor axes given in bold text. The units of the ellipse area are (°C m/s)^2.
Figure 4.13b Same as Figure 4.13a, except for Jan-Mar 94.
Figure 4.13c Same as Figure 4.13a, except for Apr-Jun 94.
Figure 4.13d Same as Figure 4.13a, except for Jul-Sep 94.
vectors are directed downcoast. These vectors are also not statistically different from zero (Fig 4.11d) and preferred directions are absent (Fig 4.12d).

**Relationship between stratification, mixed layer depth, and principal axis characteristics**

Like the previous section, the hypothesized links (1) between stratification and degree of anisotropy and bathymetric steering and, (2) between mixed layer thickness and size of the principal axes are compared. The focus is shifted from $<UT>_E$ to $<U'T'>_E$. Distributions and trends of the Burger Number and mixed layer thickness have been described in the previous section. In general, both linkage hypotheses are not supported. The principal axis characteristics and their trends are described below.

In October-December, the principal axes are anisotropic and elongated parallel to the local bathymetry (Fig 4.13a). The length ratio's mean and $\sigma$ are 2.30 and 1.20, respectively. Not supporting the first hypothesis, the trend that principal axes become more isotropic offshore is not significant at $\alpha = 0.10$ ($p=0.239$). Not supporting the second hypothesis, size of the principal axes may actually increase offshore.

In January-March, the two smallest length ratios, 1.21 and 1.43, occur in the two bins adjacent to the Mississippi River mouth which supports the first hypothesis (Fig 4.13b). The remaining principal axes are elongated parallel to the local bathymetry. Their length ratio's mean and $\sigma$ are 2.62 and 0.85, respectively. The trend that
principal axes become more anisotropic away from the Mississippi River mouth is significant at $\alpha = 0.10$ ($p=0.002$). Not supporting the second hypothesis, the trend that size of the principal axes decreases offshore is not significant at $\alpha = 0.10$ ($p=0.101$).

In April-June, except for the bin immediately south of Sabine Pass, TX, principal axes are anisotropic and elongated parallel to the local bathymetry (Fig 4.13c). The length ratio in the bin south of Sabine Pass is 1.33. The strong cross shelf variability was induced by one strong southerly $U$ vector in the middle of mostly east-west oriented $U$ vectors. It is unlikely that the local velocity field would exhibit such great variability, thus the readings are likely erroneous. Without this bin, the length ratio's mean and $\sigma$ are 2.80 and 1.19, respectively. Not supporting the first hypothesis, the innershelf ($<50$ km from shore) principal axes may actually become more anisotropic downcoast. Not supporting the second hypothesis, the trend that size of the principal axes decreases offshore is not significant at $\alpha = 0.10$ ($p=0.132$).

The July-September mean size of the principal axes is merely 6% of the annual mean size (Fig 4.13d). Not supporting the second hypothesis, size of the principal axes north of 28°N may actually increase away from the Mississippi River mouth. Not supporting the first hypothesis, the trend that principal axes become more isotropic away from the Mississippi River mouth is not significant at $\alpha = 0.10$
(p=0.196). East of 92 °W, except for the bin immediately south of the Mississippi River mouth, the length ratios are O(1.5). The principal axes south of the Atchafalaya Bay are elongated parallel to the principal orientation of the Atchafalaya river plume often seen on satellite imagery. Thus, a large part of this internal variability may be induced by the variability of the plume’s current or temperature fields.

In addition to comparing the Burger Number and mixed layer thickness with the principal axis characteristics, the upcoast-directed \( <U'T'>_E \) vectors were also examined. The \( <U>_E<T>_E \) vectors consistently flow downcoast. Three conditions may cause a \( U'T' \) vector to flow upcoast against the local \( <U>_E<T>_E \) vector: (CONDITION 1) \( T \) is cool (i.e., \( T' < 0 \)) while the current flows downcoast faster than the mean current; (CONDITION 2) \( T \) is warmer than \( <T>_E \) but the current flows downcoast slower than the mean current; and (CONDITION 3) \( T \) is warmer than \( <T>_E \) but the current flows upcoast. In the first two conditions, \( <UT>_E \) and \( <U>_E<T>_E \) flow downcoast, while \( <U'T'>_E \) flows upcoast. In the third condition, \( <UT>_E \) and \( <U'T'>_E \) flow upcoast while \( <U>_E<T>_E \) continues to flow downcoast. The upcoast \( U'T' \) vectors were separated into three groups based on the conditions forementioned. In general, the upcoast-directed \( U'T' \) vectors may be attributed to previously observed oceanographic processes and features including rapid cooling of shallow innershelf water following a cold front pas-
sage [Huh et al. 1978], upcoast innershelf current in the summer months [Crout et al. 1984; Cochrane and Kelly 1986], and upcoast current along the shelf break [Cochrane and Kelly 1986; Dinnel and Wiseman 1986].

The resulting filtered $U'T'$ fields reveal that from October through April (Fig 4.14a) cool downcoast transport (CONDITION 1) induces the upcoast flow. Innershelf (<50km from shore) vectors possess severe negative $T'$ (Fig 4.15a, the 1993-94 Julian day to calendar day table is available in the appendix) and a strong southerly $U$ component (Fig 4.15b). During this period, cold front passages visit the Louisiana-Texas shelf frequently. The dates that frontal squall lines make contact with the Louisiana-Texas coast are shown in Figure 4.16. The squall line tracks were obtained from NOAA weather maps. Cold air outbreaks typically begin in October, last through the winter months, and end sometime in May. They can cool the innershelf waters in particular because cooling rate is more rapid in the thin mixed layer or in shallow water [Huh et al. 1978]. The trend that innershelf $|<U'T'>|$ in October-December increase downcoast, significant at $\alpha = 0.10$ ($p=0.019$), may be attributed to the effects of successive cold air outbreaks, which not only cool SST, but also transport previously cool-conditioned water downcoast as well [Nowlin and Parker 1974]. The relatively slow, warm downcoast transport (CONDITION 2) is responsible for the $U'T'$ vectors over the midshelf (Fig 4.17). Barron and Vastano (1994) also ob-
served a decrease in surface current speed from innershelf to midshelf. The warm upcoast transports (CONDITION 3) are responsible for the upcoast-directed $U'T'$ vectors along the shelf break between October and March (Fig 4.18a). Upcoast flows have been observed at Flower Garden [Rezak and McGrail 1983]. The warm upcoast flow can be part of the cyclonic circulation cell described by Cochrane and Kelly (1986). The warm upcoast flow near the shelf break can also be transported, through friction, by the anticyclonic rings off the shelf break. The anticyclonic rings are shed from the Loop Current and migrate westward [Elliott 1982]. Warm upcoast transports are also responsible for the innershelf upcoast-directed $U'T'$ vectors in the summer months (Fig 4.18b) when nearshore current reverses and flows upcoast [Crout et al. 1984; Cochrane and Kelly 1986]. Among the $U'T'$ vectors that are induced by warm upcoast transports, the current speeds of these vectors that are located within 20 km from the shelf break (Fig 4.19a) and 20 km from shore (Fig 4.19b) are both 0.14 ms$^{-1}$. There is, however, less variability near the shelf break ($\sigma = 0.094$) than near shore ($\sigma = 0.189$). Between July and September, $|U'T'|$ are small because both SST and velocities are relatively uniform.

In general, $U'$ follow the seasonal variability of the shelf circulating described by Cochrane and Kelly (1986). In the winter, the innershelf $U'$ vectors are directed downcoast but the associated $U'T'$ vectors flow up-
Figure 4.14a $U' T'$ vectors that are in the opposite direction of $<U>_E <T>_E$ for CONDITION 1, Oct 2 1993-Apr 30 1994. The units are °C m/s.
Figure 4.14b Same as Figure 4.14a, except for May 1 1994-Sep 27 1994.
Figure 4.15a $T'$ associated with $U'T'$ vectors that are located within 50 km from shore for CONDITION 1. The vertical lines mark the beginning and ending of each season.
Figure 4.15b The north/south component of $\mathbf{U}$, $v$, that are associated with $\mathbf{U}'\mathbf{T}'$ vectors for CONDITION 1. The vertical lines mark the beginning and ending of each season.
Figure 4.16 Dates when cold front passages made contact with the Louisiana-Texas coast.
Figure 4.17 $U'T'$ vectors that are in the opposite direction of $<U>_E <T>_E$ for CONDITION 2, Oct 2 1993-Sep 27 1994.
Figure 4.18a $\mathbf{u}'\mathbf{T}'$ vectors that are the opposite direction of $<\mathbf{u}>_E<T>_E$ for CONDITION 3, Oct 2 1993-Mar 30 1994.
Figure 4.18b Same as Figure 4.18a, except for Apr 1–Jun 29, 1994.
Figure 4.18c Same as Figure 4.18c, except for Jun 30-Sep 27, 1994.
Figure 4.19a Current speeds, $|U|$, in m/s, associated with $U'T'$ vectors that are in the opposite direction of $<U>_E<T>_E$, and also are located within 20 km from shore for CONDITION 3. The horizontal line marks the mean of these $|U|$. The vertical lines mark the beginning and ending for each season.
Figure 4.19b Same as Figure 4.19a, except for the domain of interest is now ±20 km from shelf break.
coast because $T' < 0$. Later between April and June, the innershelf $U'$ vectors are directed upcoast. The timing of the upcoast-directed $U'$ vectors is earlier than Cochrane and Kelly's hypothesized anticyclonic cell centered around 29°N 93°W. Their hypothesized cell does not develop until July, in a climatological sense.

4.4 Relative importance of eddy heat flux

Relative importance

The ratios of eddy heat flux to net radiative heat flux (estimates for different regions of the shelf) suggest that eddy heat transport may be relatively important to the shelf's heat budget in all seasons but July-September (Fig. 4.20 and Table 4.2). In October-December, January-March, and April-June, eddy heat flux can be as large as 20% of the net radiative heat flux. A large fraction of the winter net eddy heat flux may be attributed to cold air outbreaks, since they are capable of generating both large temperature and current fluctuations. The relative error in the estimated eddy heat flux is, however, greater than one, so the directions of the net eddy fluxes are not statistically significant. The error arises partly because the structure of the time-independent eddy field is extremely complex.
Table 4.2 Net eddy heat fluxes over Louisiana-Texas shelf.

<table>
<thead>
<tr>
<th>Season</th>
<th>Net eddy heat flux per unit area</th>
<th>Surface area of eddy heat flux integrated</th>
<th>Error/estimate of eddy flux</th>
<th>Net eddy flux</th>
<th>Net radiative flux</th>
<th>Net eddy/radiative flux</th>
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<tr>
<td>oct-dec</td>
<td>97 W m$^{-2}$</td>
<td>53800 km$^2$</td>
<td>-</td>
<td>10$^9$ W</td>
<td>5218 W</td>
<td>0.195 W</td>
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<tr>
<td>jan-mar</td>
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<td>68697 km$^2$</td>
<td>5.96$^{-10}$</td>
<td>3951 W</td>
<td>12228 W</td>
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<tr>
<td>apr–jun</td>
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<td>41743 km$^2$</td>
<td>4.84$^{-10}$</td>
<td>3420 W</td>
<td>6720 W</td>
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<tr>
<td>jul-sep</td>
<td>60 W m$^{-2}$</td>
<td>51261 km$^2$</td>
<td>-10.25$^{-10}$</td>
<td>-302 W</td>
<td>3075 W</td>
<td>-0.098 W</td>
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</table>

Seasonal variability

Seasonal variability among the first three seasons were evaluated by comparing net eddy heat fluxes integrated along a common path shown in Figure 3.11a. The resulting eddy heat fluxes are shown in Figure 4.21 and summarized in Table 4.3. Using Chebyshev's inequality test, the 80% confidence intervals of the three seasons' mean eddy heat fluxes overlap. Using a student's $t$ test, only the mean eddy heat fluxes in October-December and April-June seasons are significantly different at $\alpha = 0.15$ ($p=0.108$). The normality assumption, however, may be suspect as the correlation between the sorted eddy heat fluxes and the normal distribution quantiles in these two seasons are only $0(0.7)$. Assuming between difference in the two seasons is real, the upcoast eddy heat fluxes over the

4. Net radiative flux estimates were obtained from Etter (1983).
south Texas shelf is responsible for much of the heat gain in April-June. The upcoast eddy heat flux may be induced by cool downcoast currents \((v', T', v < 0)\) or warm upcoast currents \((v', T', v > 0)\). Along the innershelf (<50 km from shore) south of 27.5 N, both cool downcoast and warm upcoast currents produce similar \(|\mathbf{U}'T'|\) (Fig 4.22a,b). From October through February, cool downcoast currents are responsible for the upcoast eddy heat flux. Both the cool downcoast and warm upcoast currents produce upcoast eddy heat flux from March through early April. Afterwards, upcoast eddy heat flux is induced primarily by warm upcoast currents.

Table 4.3. Seasonal variability of net eddy heat fluxes in the first three seasons.

<table>
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<tr>
<th></th>
<th>net radiative flux per unit area</th>
<th>surface area of eddy heat flux integrated</th>
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<tr>
<td></td>
<td>W m(^{-2})</td>
<td>km(^2)</td>
<td>10(^9) W</td>
<td>10(^9) W</td>
<td>-</td>
<td>-</td>
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<td>0.326</td>
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Seasonal variability between all four seasons for a small mid and inner shelf region near the Louisiana-Texas border was evaluated in the same manner and summarized in Table 4.4. The area around the integrated path is reduced to assure sufficient number of observations in all seasons. The smallest seasonal q-q correlation between the sorted
eddy heat fluxes and normal distribution quantiles is 0.80. The net eddy heat flux in April–June is significantly greater than net eddy heat fluxes in October–December and January–March at α = 0.20 (p=0.196; Student’s t test). The net eddy heat flux in January–March is also significantly greater than the net eddy heat fluxes in October–December (p=0.081) and July–September (p=0.001) at α = 0.10. Much of the eddy heat gain in April–June comes across the boundary between 94 and 95 °W (Fig 4.23). In a region bounded by 28.5–29.5 °N and 94–95 °W, upcoast eddy heat flux from October through April is primarily induced by cool downcoast current (Fig 4.24a,b). Both cool downcoast and warm upcoast currents generate upcoast eddy heat flux from March through early April. Afterwards, upcoast eddy heat flux is induced primarily by warm upcoast currents.

Table 4.4. Variability of net eddy heat fluxes in all seasons.

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<th>net eddy/radiative flux</th>
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<td>10⁹ W</td>
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<td>82</td>
<td>1987</td>
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</table>
Figure 4.20a The product of the mixed layer depth, H, and the component of $\mathbf{U}\cdot\mathbf{T}'$ (for Oct-Dec 93) normal to the boundary defined in Figure 3.10a. The unit is °C m²/s.
Figure 4.20b The product of the mixed layer depth, H, and the component of U'T' (for Jan-Mar 94) normal to the boundary defined in Figure 3.10b. The unit is °C m²/s.
Figure 4.20c The product of the mixed layer depth, $H$, and the component of $U'T'$ (for Apr-Jun 94) normal to the boundary defined in Figure 3.10c. The unit is °C m$^2$/s.
Figure 4.20d The product of the mixed layer depth, H, and the component of \( U'T' \) (for Jul-Sep 94) normal to the boundary defined in Figure 3.10d. The unit is °C m²/s.
Figure 4.21a The product of the mixed layer depth, $H$, and the component of $U'T'$ (for Oct-Dec 93) normal to the boundary defined in Figure 3.11a. The unit is °C m²/s.
Figure 4.21b Same as Figure 4.21a, except for Jan-Mar 94.
Figure 4.21c Same as Figure 4.21a, except for Apr-Jun 94.
Figure 4.22a Magnitudes of $U'T'$ that are associated with the northward components of $U'$ and $U$ being negative, $T' < 0$, and lie within 50 km from shore and south of 27.5°N.
Figure 4.22b Magnitudes of $U'T'$ that are associated with the northward components of $U'$ and $U$ being positive, $T' > 0$, and lie within 50 km from shore and south of 27.5 °N.
Figure 4.23a The product of mixed layer depth, H, and the component of $u'T'$ (for Oct-Dec 93) normal to a boundary defined in Figure 3.11b. The unit is °C m$^2$/s.
Figure 4.23b Same as Figure 4.23a, except for Jan-Mar 94.
Figure 4.23c Same as Figure 4.23b, except for Apr-Jun 94.
Figure 4.23d Same as Figure 4.23c, except for Jul-Sep 94.
1993-94 Julian dates

Figure 4.24a Magnitudes of $U'T'$ that are associated with the northward components of $U'$ and $U$ being negative, $T' < 0$, and lie within the region 28.5-29.5 °N, 94-95 °W.
Figure 4.24b Magnitudes of $U'T'$ that are associated with the northward components of $U'$ and $U$ being positive, $T' > 0$, and lie within the region 28.5-29.5 °N, 94-95 °W.
4.5 Surface heat flux divergence

Fields of surface heat divergence reveal that the sum of the longshore and cross-shore gradients are generally positive (Figure 4.25). Thus, the shelf loses heat. Since the cross-shore gradients are generally negative, while longshore gradients are positive, heat is lost along the shelf. In all seasons, heat is lost along the shelf in the downcoast direction because \( \langle UT \rangle \) vectors are directed downcoast and their magnitudes increase downcoast. There is, however, one noteworthy deviation. In April–June (Fig 4.25c), the cross-shore gradient is responsible for the heat loss at 27 °N 97 °W. The associated \( \langle U \rangle_E \) vector is directed offshore. Cochrane and Kelly (1986) and Dinnel and Wiseman (1986) previously concluded that coastal water is transported offshore over the south Texas shelf then turns upcoast along the shelf break.

In addition to describing the pattern of surface heat divergence, the dominant component, mechanism, and orientation responsible for the heat loss may be identified through a series of decompositions and order of magnitude comparisons. First, the dominant mechanism responsible for heat loss may be identified by decomposing \( \Delta \langle UT \rangle_E / \Delta x \) into \( \Delta \langle U \rangle_E \langle T \rangle_E / \Delta x \) and \( \Delta \langle U' T' \rangle_E / \Delta x \). Since \( \Delta \langle U \rangle_E \langle T \rangle_E / \Delta x \) is generally an order of magnitude larger than \( \Delta \langle U' T' \rangle_E / \Delta x \), \( \Delta \langle U \rangle_E \langle T \rangle_E / \Delta x \) is the dominant term responsible for the heat loss. The term \( \Delta \langle U \rangle_E \langle T \rangle_E / \Delta x \) was further decomposed into \( \langle U \rangle_E \Delta \langle T \rangle / \Delta x \) and \( \langle T \rangle_E \Delta \langle U \rangle_E / \Delta x \). In general, \( \langle T \rangle_E \Delta \langle U \rangle_E / \Delta x \) is
Figure 4.25a $d<vT>_E/dy$ and $d<uT>_E/dx$ values for Oct-Dec 93. The black dots mark the mid points of the four surrounding $<uT>_E$ vectors used to compute $d<vT>_E/dy$, $d<uT>_E/dx$. The $d<vT>_E/dy$ and $d<uT>_E/dx$ are shown above (bold) and to the right (plain) of the black dots, respectively. The $<uT>_E$ vectors are shown in the background.
Figure 4.25b Same as Figure 4.25a, except for Jan-Mar 94.
Figure 4.25c Same as Figure 4.25a, except for Apr-Jun 94.
Figure 4.25d Same as Figure 4.25a, except for Jul-Sep 94.
O(10^{-3}) to O(10^{-2} \text{ C m}^{-1}), whereas, \langle U \rangle_E \Delta \langle T \rangle_E / \Delta x is O(10^{-4} \text{ C m}^{-1}). Thus, \langle T \rangle_E \Delta \langle U \rangle_E / \Delta x is the term responsible for much of the heat loss. The observed similarity between temperature flux and velocity autocorrelation functions also suggests that temperature flux is governed primarily by velocity variability (Figure 3.8).

The primary orientation too can be determined in a similar procedure. The term \langle T \rangle_E \Delta \langle U \rangle_E / \Delta x was decomposed into its longshore and cross-shore components. The magnitudes of the longshore term consistently exceed the magnitudes of the cross-shore term. Since the direction of \langle U \rangle_E \langle T \rangle_E vectors is downcoast, heat was lost primarily because of the mean longshore velocity gradient. Furthermore, in each season the trend that innershelf (\langle 50 \text{ km from shore} \rangle) \mid \langle U \rangle_E \mid west of 92 \^\circ W increases downcoast is significant at \alpha = 0.10 (p=0.092, 0.052, 0.085, 0.025). Since the width of the coastal boundary current from the Louisiana-Texas border to US-Mexican border is often observed to be relatively constant on satellite images, the increasing downcoast current speed suggests that there is a net volume transport from the ambient shelf water into the coastal boundary current.
5.0 Summary

Eddy heat flux variability over the Louisiana-Texas shelf was investigated. Assuming the product of sea water density and specific heat is constant, velocity-temperature covariance were used to infer eddy heat flux. Surface velocity and temperature were derived from ARGOS and AVHRR data. The two fields, however, are not synchronous. Temperature "observations" at velocity positions were optimally estimated. (Optimal analysis is evolved from the Gauss Markoff theorem.)

The study's spatial and temporal resolution (bin size), 1 degree and 90 days, was selected based on decorrelation scale estimates and statistical reliability. Confidence ellipse, confidence intervals for preferred direction, and principal axes were computed for each bin's mean and eddy temperature flux estimate. Trends of the principal axis characteristics were compared with trends of the Burger Number and mixed layer thickness to evaluate the dependence of principal axis characteristics on degree of stratification and mixed layer thickness.

Net shelf eddy heat flux was also estimated for sub-regions of the study area. Eddy heat transport may be important in all seasons but July-September. Among the seasonal net eddy heat fluxes, the winter net eddy heat flux is relatively large. The majority of the heat gain
comes from the south Texas shelf. Surface temperature divergence estimates were also computed. The shelf loses heat primarily because of the mean longshore velocity gradient.

5.1 List of conclusions

1. The conservative SST error-estimates due to the optimal analysis are comparable to the error resulting from the application of the widely accepted, multi-channel sea surface temperature algorithm, $O(1 \, ^\circ C)$. The normalized squared-error-estimate is relatively constant at $O(0.25)$ throughout the year. It is of the same order of magnitude as the normalized squared-error-estimate found by Mariano and Brown (1992) for an optimally interpolated SST field that reproduced multiple water masses observed in the western Atlantic. These two indirect comparisons suggest that the error-estimates are acceptable.

2. The trend that the principal axes of the instantaneous temperature flux become more isotropic offshore is significant ($\alpha = 0.10$) in all seasons but the October-December season. The hypothesis that water column stability controls the orientation and degree of anisotropy of the principal axes of instantaneous temperature flux is supported in January-March and July-September. The hypothesis that mixed layer thickness affects the size of the principal axes of instantaneous temperature flux is generally not supported.

3. In the winter months, the innershelf $U'T'$ vectors directed upcoast are induced by cool downcoast transports
which may be associated with cold air outbreaks. The up-coast-directed $U'T'$ vectors near the shelfbreak are induced by warm upcoast transports which may be driven by an anticyclonic ring beyond the shelf break. Between May and June, the innershelf upcoast-directed $U'T'$ vectors are driven by warm nearshore upcoast transports.

4. Eddy heat transport may be an important term in the winter heat budget equation. Among the seasonal net eddy heat fluxes, net eddy heat flux is highest in April-May. Much of the heat gain comes from the south Texas shelf where the heat gain is generated by both cool downcoast and warm upcoast currents.

5. The shelf loses heat downcoast. The dominant contribution to surface temperature divergence is from the product of the local temperature and the longshore mean-velocity gradient. Innershelf current speed also increases downcoast. The increasing current speed coupled with the often observed relatively constant width of the coastal boundary current suggests there is a net volume transport of ambient water into the coastal boundary current.

5.2 Future recommendations

The success of the study relies heavily on the ability of the proposed technique to adequately estimate temperature flux. Error-estimates are dependent on the data distribution. Although the data distribution may be improved greatly using microwave radiometers, their field of view is much coarser at 75 km and thus would not be ap-
propriate for a shelf scale study. Microwave data may be an ideal alternative for larger scale studies, particularly those near the equator where clouds are ubiquitous [Bernstein and Morris 1983].

An estimate of the seasonal net heat flux may also be improved. In many instances, the buoy-velocities were clustered together providing accurate local but poor synoptic fields. Between July and September, there were few observations over the south Texas shelf. A more synoptic coverage may be achieved by deploying groups of buoys into each 1x1 degree bin on the shelf every month. A simpler task may be to release two groups of buoys every two months, one group south of Terrebonne Bay and the other over the south Texas shelf.

Some improvements may also be made in the processing procedure. A systematic but a more conservative cloud screening procedure will reduce the number of cloud contaminated pixels in the data used for optimal analysis. Length scales vary over the shelf [Yongxiang et al. 1996] and in time as well. A pilot experiment conducted using different correlation functions for each 1x1 degree box produced unrealistic gradients along the boundaries. A smoothly varying space and time correlation function may reduce estimation error.

The methodology described in this study may be applicable to other flux studies. In addition to SST, AVHRR data can also estimate relative suspended sediment con-
concentration [Stumpf and Pennock 1989] and mixed layer depth [Yan et al. 1990]. The upcoming Sea-Viewing Wide Field-of-view sensor (SeaWifs) data can potentially estimate low (0.05 - 1 mg m\(^{-3}\)) and high (1-40 mg m\(^{-3}\)) chlorophyll a concentrations to within 20% and 30%, respectively [Tasssan 1994]. Velocity fields may be obtained from current meters and models, as well as the ARGOS buoy tracks.
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### APPENDIX

#### 1993-94 Julian dates to calendar day table

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VITA

Varis Ransibrahmanakul was born in Bangkok, Thailand. He came to study in the United States beginning 1980. He graduated from St. Mary High School, Natchitoches, Louisiana, May 1984. He earned his Bachelor degree in Civil Engineering, Louisiana State University, in May 1988, and Master degree in Agricultural Engineering, Louisiana State University, August 1990. He then worked at the United States Department of Agriculture, New Orleans, for one year. He re-entered Louisiana State University in 1991 and received his Ph.D. in Oceanography and Coastal Sciences in May 1996.
DOCTORAL EXAMINATION AND DISSERTATION REPORT

Candidate: Varis Ransibrahmanakul

Major Field: Oceanography and Coastal Sciences

Title of Dissertation: Variability of Eddy Heat Fluxes Over the Northwestern Gulf of Mexico

Approved:

[Signatures]

Major Professor and Chairman

Dean of the Graduate School

EXAMINING COMMITTEE:

[Signatures]

Co-Chairman

Date of Examination: 3/22/96