Feedbacks Based on Hyperspectral Measurements of the Air-Sea Temperature Difference

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FEEDBACKS BASED ON HYPERSPECTRAL MEASUREMENTS OF THE AIR-SEA TEMPERATURE DIFFERENCE

By

Adam Chambers

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FEEDBACKS BASED ON HYPERSPECTRAL MEASUREMENTS OF THE AIR-SEA TEMPERATURE DIFFERENCE

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Ship-based measurements of the air-sea temperature difference (air-sea $\Delta T$) are analyzed and presented with a focus on the temperature difference probability distribution and causes of its stability and variability. By taking the measurements with a hyperspectral infrared radiometer, temperatures of both the lower atmosphere and the skin layer of the ocean are acquired by a single, well-calibrated instrument. Consider that for each radiometric cruise deployment, when the air-sea $\Delta T$ measurements are plotted in a histogram, a mean air-sea $\Delta T$ peak at about -1 K becomes apparent regardless of measurement location or time of year. The large amount of air-sea $\Delta T$ data centered around -1 K suggests that there is a ‘feedback’ loop involving oceanic and atmospheric variables that stabilizes the air-sea $\Delta T$ around -1 K.

Within a system, “feedback” is defined as the process in which a part of the output of the system is returned to its input to influence or regulate its further output. A feedback loop may be characterized as positive or negative. Following an initial disturbance, a positive feedback loop will lead to a growth in the initial perturbation, and could push a previously stable system past a critical condition and into another state. A negative feedback loop will act to lessen the initial disturbance, keeping the system within a stable
state. When describing climate systems, there are a number of variables that allow for the formation of both positive and negative feedback loops.

Data from the multiple cruises point to the negative ‘tail’ of the air-skin $\Delta T$ being caused by a combination of low wind speeds (less than 3 m/s) and intense solar radiation. Wind speeds from 3 m/s to 9 m/s cause an increase in sensible and latent heat fluxes that serve as negative feedbacks that stabilize the air-sea $\Delta T$ around its -1 K mean. At wind speeds above 10 m/s and low humidity levels, the latent heat flux acts as a positive feedback, and depending on the strength of other oceanic and atmospheric heat fluxes, may cause the air-skin $\Delta T$ to become more negative. A flow diagram of initial disturbances and the resulting feedbacks is presented, demonstrating the necessary conditions and consequences of the negative and positive feedback loops with regard to the air-skin $\Delta T$. 
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Chapter 1

Introduction

The air-sea temperature difference (air-sea ΔT), defined as the temperature difference between the skin layer of the ocean and the overlying atmosphere (traditionally referenced at 10m), is an important parameter of the air-sea fluxes of momentum, moisture, heat, and gases. Through these fluxes, the air-sea ΔT affects the stability of the lower atmosphere, and the stability of the atmosphere affects the air-sea fluxes. The fluxes are key factors in understanding the climate, cloud formation, and responses to changes in climate forcing. In addition, uncertainties in the air-sea ΔT lead to significant uncertainties in describing the heat balance of the global ocean, including the Tropical Warm Pool, which is known to be important in the climate system (Fairall, Bradley et al.
1996). The air-sea ΔT and the heat flux have a complex relationship in which the temperature difference between the skin layer of the ocean and the atmosphere both influences and is influenced by the heat flux (Fairall, Bradley et al. 1996).

At the interface between the ocean and the immediately overlying atmosphere, the temperature of the surface of the ocean is called the skin sea surface temperature (skin SST) (Minnett, Smith et al. 2010). The skin layer of the ocean may be described as two coexisting layers: the electromagnetic (EM) skin layer and the thermal skin layer. The EM skin layer occurs due to water’s EM properties, and the absorption and emission of infrared (IR) radiation occur in the EM skin layer. Infrared radiometers on ships, airplanes, and satellites may measure the emission from the EM skin layer with varying contributions from the atmosphere. The ocean emission comes predominantly from depths of less than 1 mm. The thermal skin layer, sometimes called the molecular sublayer, features a very thin temperature gradient that is controlled by molecular conduction. This thermal skin layer supplies the heat for the turbulent transfers to the atmosphere and the net infrared radiative heat loss from the ocean (Minnett, Smith et al. 2010). The process of molecular conduction is a less efficient process of heat transfer compared to the turbulence in the ocean beneath the skin layer, or the convection in the air immediately above the skin layer. This explains the strong temperature gradient seen in the thermal skin layer.

In the past, measurements of earth’s surface air temperature have been made by thermometers in specialized enclosures, called Stevenson screens, and are thus protected from direct solar radiation and wind (Goerss and Duchon 1980). Over the ocean, these
enclosures are not always very effective when deployed on ships or the equivalent protection for buoys. The thermal contamination of air by the structure of the buoy or ship occurs both due to the perturbed air flow and ‘heat island’ effect that often heats the thermometer enclosure to an unknown extent (Ramage 1984; Kent, Tiddy et al. 1993). The air and ocean have different thermal capacities, the temperature of which may not be accurately measured by traditional thermometers. In addition it is unreasonable to expect a large, robust thermometer to make an accurate measurement in the skin layer of the ocean that has a thickness of less than a millimeter. As will be shown later, the distributions of the air-sea ΔT vary greatly when measured by radiometers and conventional thermometers. While subsurface, traditional SST measurements are easier to make, the measurement itself may not be the best suited option for the determination of quantities that show a dependence on the sea surface temperature (Minnett, Smith et al. 2010). The distributions of the air-sea ΔT when measured by radiometers are much less variable, and it is the goal of this research to explain why this is so.

The calculations of air-sea fluxes of heat, momentum, and gases also rely on the accuracy of the air-sea ΔT measurement. The required accuracy of a temperature measurement to determine heat flux to useful accuracies is better than 0.2 K, and the uncertainties in the measurements of traditional thermometers are often too large to meet this requirement (Fairall, Bradley et al. 1996). These complications may be avoided by using a ship-based infrared hyperspectral radiometer with appropriate ship placement and infrared channel selection. However, the use of such radiometers demands extreme calibration accuracy and stability of the device. The cost of such accurate and reliable
instruments prevents them from being very widely used (Fairall, Bradley et al. 1996). Traditionally, the required accuracy of a marine lower atmosphere temperature measurement is ±0.1 K (HMSO 1994). The 0.1 K accuracy goal is an acceptable standard for weather forecasting; however, for climate research identification of long-term trends of ~0.1 K decade\(^{-1}\), more accurate measurements are required to detect the trends sooner and with more confidence (Minnett, Maillet et al. 2005). The Interim Sea Surface Temperature Science Team (ISSTST, 2010) specified that skin SST’s with an absolute accuracy of less than 0.1 K and a stability of less than 0.04 K/decade for satellite measurements are necessary for accurate projections of climate change (Ohring, Wielicki et al. 2005).

1.1 **Climate Feedbacks**

Within a system, “feedback” is defined as the process in which a part of the output of the system is returned to its input to influence or regulate its further output. A feedback loop may be characterized as positive or negative. Following an initial disturbance, a positive feedback loop will lead to a growth in the initial perturbation, and could push a previously stable system past a critical condition and into another state. A negative feedback loop will act to lessen the initial disturbance, keeping the system within a stable state (Bony, Colman et al. 2006). When describing climate systems, there are a number of variables that allow for the formation of both positive and negative feedback loops.
Within this research further details will be given with consideration to water vapor and cloud radiative feedbacks.

Within this research, cruise and model data will be analyzed in regard to the heat fluxes and air-skin $\Delta T$. A flow diagram of initial disturbances and the resulting feedbacks will be presented, demonstrating the necessary conditions and consequences of the negative and positive feedback loops with regard to the air-skin $\Delta T$. Given the small variance within the air-skin $\Delta T$ data, emphasis will be given to feedbacks explaining the causes of variance, and the physical processes that restore the shape of the air-skin $\Delta T$ distribution.
Chapter 2

Data

The data used in this study includes field data from numerous cruises and moorings, as well as data from models. The first section of this chapter describes the interferometer used to measure the air and sea temperature, including a discussion on the spectral selection of the interferometer, its traceability, and the validation of satellites. The final section offers a brief summary of the field deployments of the interferometer.
2.1 The Marine-Atmospheric Emitted Radiance Interferometer (M-AERI)

2.1.1 The Instrument

Ship based measurements used here were recorded using the Marine-Atmospheric Emitted Radiance Interferometer (Minnett, Knuteson et al. 2001). The M-AERI is a...

![Fig. 2.1: Spectra measured by the M-AERI at various angles while directed at the sky (top) and sea surface (bottom). Temperature is representative of the spectra, and the atmosphere is most transparent at the lowest sky temperatures. The blue bar shows which region is used to measure air temperature and the red bar shows which region is used to measure the skin sea-surface temperature. Note the change in temperature scales of the two panels. These data were taken in the tropical western Pacific during the Combined Sensor Program Cruise in 1996 (Post, Fairall et al. 1997; Minnett, Knuteson et al. 2001).](image-url)
Fourier transform spectroradiometer capable of measuring spectra in the infrared ($\lambda \sim 3 \ \mu m$ to $\sim 18 \ \mu m$) with a resolution of $\sim 0.5 \ cm^{-1}$. The noise equivalent temperature difference (NE\Delta T) of the instrument is below 0.1K, due to two infrared detectors being cooled to $\sim 78 \ K$ by a Stirling-cycle mechanical cooler (Minnett, Knuteson et al. 2001). The M-AERI is a single instrument that measures both the skin SST and the air temperature with real-time calibration accomplished using two internal blackbody cavities. Taking both air and sea temperature measurements with the same instrument avoids the artifacts inherent in using two separately calibrated thermometers. The M-AERI derives the skin SST and air temperatures from the same spectra, but using different spectral intervals. Spectral wavenumber ranges are carefully selected for radiometric measurement of skin SST and air temperature. The surface emission of the oceanic skin layer may be measured at any wavelength, as long as the atmosphere is transparent enough to allow the signal to propagate through its path between the M-AERI and the surface. However, due to the spectral properties of seawater there are wavelength intervals of the spectrum where the emission comes from a thicker section of the skin layer. While this may be useful in measuring the gradient of the skin layer, measurements from where the emission depth is very shallow must be chosen to retrieve the skin temperature (Friedman 1969).
2.1.2 Spectral Selection

Because of the spectral properties of seawater, infrared wavelengths longer than ~5 \( \mu \text{m} \) are emitted from shallow skin depths. Measurements taken in a narrow spectral interval of ~7.7 \( \mu \text{m} \) at a nadir angle of 55˚ are used to derive skin SST values from the M-AERI. Since the emission measured by the M-AERI while directed at the sea surface is a combination of the radiation emitted at the surface and the radiation from the sky reflected at the surface, the 7.7 \( \mu \text{m} \) window is chosen to improve the accuracy of the skin SST by reducing its dependence on the correction for reflected sky radiance because the relatively short atmospheric path length reduces sensitivity to changes in the sky emission caused by variable clouds (Smith, Knuteson et al. 1996). The 55˚ nadir viewing angle has been shown to have a small wind speed dependence on effective emissivity (Wu and Smith 1997; Hanafin and Minnett 2001). The viewing angle is reasonable to achieve at sea given the structural geometry of most ships, and allows the M-AERI to have a clear view of the sea surface ahead of the bow wave (Hanafin and Minnett 2001).

The radiometer obtains SST from measurements taken at a nadir angle of 55˚, with measurements from a 55˚ sky view needed to correct for reflections. The air temperature measurements are taken from data acquired at both the nadir and zenith angles of 55˚. The spectral intervals for the measurement of the air temperature are based on four key considerations. First, the transmissivity must be low enough to provide upward looking measurements with a signal from emission close to the instrument, with only a small contamination from higher in the atmosphere. For downward looking measurements, the
transmissivity should be low enough that the signal is not contaminated by the emission from the underlying seawater. Second, the spectral intervals must be transmissive enough to allow the M-AERI detectors to receive radiation emitted by the reference blackbody targets during the on-board calibration sequence. Third, to avoid correlation errors between the air temperature and atmospheric water vapor, the channels should be insensitive to humidity fluctuations. Finally, it is desirable to use multiple channels with similar transmittance to reduce instrument noise by averaging independent measurements. Standard M-AERI deployments average over 46 spectra for each sea and sky view (Minnett, Knuteson et al. 2001).

With these four considerations, a spectral interval of \(14 \mu m - 15 \mu m\) was chosen for the measurement of air temperature (Fig. 2.1) (Minnett, Maillet et al. 2005). Within this interval, the atmospheric emission is dominated by \(CO_2\), which is well-mixed in the lower atmosphere and has a small, mostly seasonal, variability that does not substantially contribute to uncertainties in the radiometric measurement of the air temperature. With proper consideration to the atmospheric conditions and ship geometry, the radiometer may measure the air and skin temperatures away from the perturbing effects of the ship to accuracy better than 0.1 K (Minnett, Maillet et al. 2005).

2.1.3 Traceability and Validation of Satellites

The M-AERI was designed for the validation (Minnett 2010) of the MODIS (Moderate Resolution Imaging Spectroradiometer); (Esaias, Abbott et al. 1998) satellite skin SST
retrievals. Confidence is given to the accuracy of the M-AERI measurements through the use of on-board blackbody calibration, traceability to the NIST standards (Rice and Johnson 1998), and through comparisons to other radiometers (Barton, Minnett et al. 2004). Aboard the R/V Mirai in 1999, a direct comparison between the M-AERI and the SISTeR (Scanning Infrared Sea Surface Temperature Radiometer), another radiometer described by Barton et al (2004), showed the mean and standard deviation of the differences of skin SST measurements (n=2414) from the two radiometers to be 0.003 K±0.057 K. Despite being built in difference facilities, having separate calibration processes, and different operation schemes, both the M-AERI and SISTeR are traceable to NIST standards and show notable agreement in their measurements of the skin SST (Minnett, Maillet et al. 2005).

2.2 Field Campaigns

2.2.1 M-AERI Cruises

Since 1996, M-AERI’s have gathered data from over 40 cruises in a variety of climatological regimes to improve the derivation and validation of satellite skin SST retrievals. Air-sea ΔT measurements taken by the M-AERI have a much narrower distribution than conventional bulk measurements, as seen in Fig. 2.2. This implies that the broader distribution from the conventional measurements must include additional sources of inaccuracies. The sources of these inaccuracies include heat island effects,
perturbed flow of the air being measured, and using a large thermometer to measure the subsurface sea temperature. The radiometric data have much smaller variability, and in this case do not show any data points to have a positive air-sea ΔT. The data gathered by the M-AERI from 2000-2006 on the Explorer of the Seas show 95% of measurements to have a negative air-sea ΔT. The Nauru’99 (Western Pacific, 1999), SAGE (Western Pacific, 2004), EGEE (Eastern Atlantic, 2006), and Cirene (Tropical Indian, 2008) cruise deployments of the M-AERI will be investigated in detail.

![Histogram of air-sea ΔT measured radiometrically (red) and by conventional, bulk, thermometers (black) from the R/V Mirai in the vicinity of the Equator in the Pacific Ocean (left). Histogram of all air-sea ΔT data gathered from 2000-2006 along western subtropical north Atlantic by the vessel Explorer of the Seas (right) with n=158 748, mean=-1.285 K±1.257 K.](image)
The stability of the atmospheric boundary layer is dependent on the air-sea temperature difference. Air-sea ΔT measurements are also used in the calculations of air-sea fluxes of gases, momentum, and heat. An unstable atmospheric boundary layer features convection due to heat flowing from the warm surface to a cooler overlying atmosphere. This convection significantly enhances the vertical transport of heat, momentum, and gases. In a stable air-sea boundary layer, turbulence in the lower atmosphere is suppressed, leading to much smaller fluxes. Consider that when all of the air-sea ΔT measurements are plotted in the same histogram, a mean air-sea ΔT peak at about -1 K becomes apparent regardless of measurement location or time of year (Figs. 2.2 and 2.3). Since the air-sea temperature difference is about ~-1 K, an uncertainty of a few tenths of a degree in either the air or sea temperature measurement can cause substantial error in the calculation of both the sensible and latent heat fluxes. These calculation errors may affect the magnitude, and of even more significance, the sign of the air-sea fluxes (Minnett 2004). Errors in the sign of the air-sea fluxes will undermine the purpose of taking such measurements. The large amount of air-sea ΔT data centered around -1 K suggests that there is a ‘feedback’ loop involving oceanic and atmospheric variables that stabilizes the air-sea ΔT around -1 K. This concept will be explored more thoroughly below. Increased accuracy in the air and skin temperature measurements serves to improve our understanding of the coupled nature of the ocean and atmosphere.
Fig. 2.3: The air-sea ΔT for all M-AERI research cruise tracks since 1996. Despite the latitude and seasonal temperature differences in the SST, world-wide measurements of the air-sea ΔT show very little variation.
CHAPTER 3

Methods and Parameterizations

This chapter gives an outline of the methods and principles used in the analysis of the data. An overview of the techniques to derive the heat fluxes is given, followed by a description of the cool skin. The final two sections describe the models used to explore the possibility of a feedback loop of heat fluxes that may govern the air-skin $\Delta T$. 
3.1 **Turbulent Heat Fluxes**

The planetary boundary layer, the lower portion of the atmosphere which is in contact with surface below, reacts to surface-atmosphere exchanges. This layer, extending up to 3 km above the surface of the earth, is characterized by turbulent flow. Surface friction interacts with the mean winds to cause convective instabilities, buoyancy fluxes, and vertical velocity shears. These vertical velocity shears and convective instabilities produce turbulent eddies varying in size from millimeters to kilometers. These eddies transport heat and pollutants. The sensible and latent heat fluxes are driven by this kind of turbulent transfer.

3.1.1 **Sensible Heat Flux**

The sensible heat flux $Q_s$ is given by:

$$Q_s = \rho c_p \overline{w'T'}$$

(1)

where $\rho$ is air density, $c_p$ is the heat capacity of air, $w$ is the vertical wind velocity, and $T$ is temperature of air. The time averaged quantity, $\overline{w'T'}$, represents the mean product of temperature and vertical wind fluctuations away from the average that occur due to passing turbulent eddies. The sensible heat flux may be non-zero; the sign and direction of the transport is a consequence of the temperature profile of the surface boundary layer (Friehe and Schmitt 1976).
The typical methods of determining the sensible heat flux are the eddy covariance method, bulk aerodynamic formula, and the gradient method. The details of these methods will be described in Latent Heat Flux section.

### 3.1.2 Latent Heat Flux

Latent heat flux describes the heat carried out of the ocean by evaporation of water. The mean vertical turbulent transport of moisture near the surface results in latent heat exchange at the air-sea interface. Latent heat flux, $Q_L$, is defined as:

$$Q_L = \rho L_e \overline{w'q'}$$  \hspace{1cm} (2)

Where $\rho$ is air density, $L_e$ is latent heat of vaporization, $w$ is the vertical component of wind, and $q$ is specific humidity. The latent heat turbulent transport is characterized by $\overline{w'q'}$, the mean product of the fluctuations of the vertical wind component and specific humidity. Typically, high winds and low humidity levels result in a higher evaporation rate from the sea surface and hence a greater heat loss from the ocean due to latent heat exchange. The sign of the latent heat exchange nearly always represents heat flowing from the ocean to the atmosphere, and is in general larger in magnitude than the sensible heat exchange. The heat flux rarely indicates heat flowing from the atmosphere to the ocean, which would cause condensation on the ocean surface i.e. fog.
3.1.3 Measurement of Turbulent Fluxes

Eddy Covariance Method

A direct measurement method for the sensible heat flux, the eddy covariance method measures the turbulent fluctuations of vertical wind and air temperature. Currently, there is no ‘standard’ process to implement a system to measure the eddy covariance, although networks like Fluxnet (Key and Schweiger 1998) are working towards a general standardization of set-up and data processing procedures. The Eddy Covariance method (Eddy Correlation, EC) was used in several oceanographic cruises, including those with the M-AERI, previously described. However, the EC method involves expensive instrumentation and requires assumptions to be made. The assumptions include that the flux is completely turbulent; that all of the net vertical transfer is due to eddies. Further assumptions are that density fluctuations are negligible, flow convergence and divergence are negligible, and that the surface is horizontal and uniform. In addition, measurements at the point of interest are assumed to represent an upwind region. While the assumptions are many, if they are satisfied the EC method provides direct sensible flux measurements. Even with the assumptions satisfied, the general data acquisition process may introduce frequency response errors due to time response, sensor separation, high/low pass filtering, and digital sampling. Since the instrument must be level, an additional challenge for oceanographers is to remove the ship’s motion from the data. Acquiring high quality data with the eddy covariance method is challenging because of the complexity of the system
design, implementation and the handling of the large amount of data (Burba and Anderson 2012).

**Bulk Aerodynamic Formulas**

Both the sensible and latent heat fluxes may be calculated with the bulk aerodynamic formulas. This technique is based on the Monin-Obukhov Similarity Theory (MOST). Under MOST, there is the assumption that in the lower ~10% of the planetary boundary layer (surface layer), turbulent fluxes are constant with height. MOST also assumes that the main mechanism for vertical heat exchange in the surface layer is turbulent transfer. The bulk formulas for the sensible and latent heat flux components are:

\[
Q_s = \rho c_p C_H (U - U_s)(T_s - T) \tag{3}
\]

\[
Q_L = \rho L C_L (U - U_s)(Q_s - Q) \tag{4}
\]

where \(\rho\) is air density, \(c_p\) is specific heat of air, \(C_H\) and \(C_L\) are the transfer coefficients of sensible heat and water vapor; \(L\) is the latent heat of evaporation; \(U, T,\) and \(Q\) are the wind speed, temperature, and specific humidity at a reference height; \(U_s, T_s,\) and \(Q_s\) are the wind speed, temperature, and specific humidity at the surface (Liu, Katsaros et al. 1979).

In terms of the bulk sensible heat formula, the eddy covariance of vertical wind and temperature is estimated as the difference between temperature at the given height and the surface multiplied by the difference in the mean horizontal wind speed between the surface and the given height. The latent heat bulk estimation uses a similar approach,
using specific humidity instead of temperature (Ramage 1984). The TOGA-COARE model, which will be described in further detail in section 3.5, uses the bulk formulae to calculate sensible and latent heat exchange.

The transfer coefficient for the latent heat flux has been shown to depend on the air-skin ΔT, as shown in Fig. 3.1. For wind speeds above 6 m/s, the transfer coefficient is approximately $1.2 \times 10^{-3}$. At wind speeds lower than 6 m/s, the transfer coefficient increases as the air-skin ΔT becomes more negative, and decreases as the air-skin ΔT becomes more positive.

![Figure 3.1: The latent heat flux transfer coefficient, $C_L$, plotted against the air-skin ΔT at various wind speeds (Kara, Hurlburt et al. 2004)](image)
**Gradient Method**

The gradient method allows for the calculation of both the sensible and latent heat fluxes. This method infers the turbulent fluxes above the surface roughness elements, and thus calculating the gradient only requires the wind, temperature, and humidity to be taken between two different heights. Both heights must be higher than the influence of surface roughness elements. The heights for field data may vary depending on the structure of the ship; Arya (1991) recommends that to minimize profile error, the ratio of the two measurement heights should not exceed 2. While the bulk formula requires standard meteorological data and parameterization of the wind-driven surface roughness, the gradient method relies on much fewer variables. The wind, temperature, and humidity profiles are assumed to be either linear or logarithmic, which lead to the following forms of the gradient method:

1. **Linear Approximation**
   
   \[
   \left( \frac{\partial M}{\partial z} \right)_{z_a} \approx \frac{\Delta M}{\Delta z} = \frac{M_2-M_1}{z_2-z_1} \quad (5)
   \]

2. **Logarithmic Approximation**
   
   \[
   \left( \frac{\partial M}{\partial z} \right)_{z_a} \approx \frac{M_2-M_1}{z_m \ln \left( \frac{z_2}{z_1} \right)} \quad (6)
   \]

Where \( Z \) is height and \( M \) represents scalar wind speed, temperature, or humidity values. Using the gradient approximations, one may solve for the gradient Richardson number and the Monin-Obukhov scaling parameters of \( u_* \), \( T_* \), and \( q_* \), and the Monin-Obukhov
stability parameters. These parameters in turn permit the calculation of the latent and sensible heat fluxes. While this is useful in estimating the turbulent fluxes without sea surface temperature or surface roughness, the approximations tend to be less precise than the bulk formula and EC methods because of the assumptions for the wind, temperature, and humidity profiles (Arya 1991).

For near-neutral to unstable surface layers, the logarithmic gradient profile approximation is favored, while the linear approximation gives a more accurate approximation of the gradient profile in a stable surface layer. The logarithmic approximation, with reasonable ratio and measurement of heights, has an error to be less than 2% for the gradient approximation, and about 4% error in the corresponding flux calculations. For general application, the logarithmic application is suggested. The error in the gradient profile using the linear approach may be as large as 8%, with the corresponding flux calculations having an error up to 16% (Arya 1991).

3.1 Radiative Heat Fluxes

Incoming radiative heat fluxes are conventionally split into two categories according to wavelength: the longwave (3 μm -100 μm) and shortwave (0.15 μm - 4 μm) radiative flux. The division is also related to the source, the sun or atmosphere.
3.2.1 Shortwave Radiation

Incident shortwave radiation from the sun contributes a heat gain to the ocean during the daytime. The intensity of incoming shortwave radiation is largely determined by latitude, time of day, season, and attenuation due to clouds, gas molecules, aerosols, and dust. The latitude, longitude, time of day, and season components may be combined and represented as the solar zenith angle (Stewart 2006). Radiation with shorter wavelengths may penetrate deeper into the ocean. The depth to which incident shortwave radiation may penetrate a body of water depends on the turbidity and amount of particulate matter present (Jerlov 1976). A portion of the incoming solar radiation is reflected at the surface and re-emitted into space. The amount of shortwave radiation reflected is dependent upon the albedo of water, which ranges from 0.05 to near 1. The higher the albedo, the more radiation is reflected. An albedo of 1 describes complete reflection. The albedo changes with the sea state and as the angle of the sun changes above the horizon (Stull 1984). Payne (1972) shows that the surface albedo may be determined by the atmospheric transmittance and the solar altitude (Payne 1972). In regions of shallow and clear water, it is possible that the incident radiation may penetrate to the bottom and be reflected upwards, perhaps to emerge and leave the ocean. Shortwave radiation that re-emerges at the surface does not heat the ocean.
3.2.2 Longwave Radiation

At the top of the electromagnetic skin layer, the ocean emits longwave radiation similar to a blackbody having the same temperature as water. Outgoing longwave infrared (IR) radiation from the ocean’s surface is dependent on the emissivity of the ocean’s surface ($\varepsilon$), the Stefan-Boltzmann constant ($\sigma$), and the temperature at the top of the skin layer ($T_{skin}$). The emissivity of the ocean has an angle and wavelength dependence as well. The emissivity of a body of water is typically 0.95-0.98, although it is often assumed to be 0.97, such as in TOGA-COARE model (Fairall 2003). The emitted IR radiation may be calculated using the Stefan-Boltzmann law:

$$Q_{LW1} = \varepsilon \sigma T_{skin}^4$$  \hspace{1cm} (7)

The spectral radiance distribution of emitted IR radiation from the sea surface is described by Planck’s equation:

$$B_\lambda(T_{skin}) = \frac{2hc^2}{\lambda^5} \cdot \frac{1}{e^{hc/(\lambda kT_{skin})} - 1}$$ \hspace{1cm} (8)

Where $B_\lambda(T)$ is the spectral radiance at for each wavelength $\lambda$ at a given temperature $T$, $h$ is Planck’s constant, $c$ is the speed of light, and $k$ is Boltzmann’s constant. The transmittance of the infrared spectrum through standard atmospheres at various latitude bands is shown in Fig. 3.2. The level of transmittance affects the amount of infrared radiation from the ocean that is absorbed by the atmosphere. Likewise, the transmittance regulates how much radiation from clouds may pass through the underlying atmosphere to reach the ocean. The atmospheric transmisivity is strongly affected by water vapor and
cloud cover, and the transmittance properties at a given wavelength play a role in selecting the appropriate spectral windows on measurement systems, such as the M-AERI. In the 8 µm to 13 µm window, the transmittance on a cloudless day depends mainly on the amount of water vapor present in the atmosphere. For other spectral intervals, such as from 3.5-4.0µm, the transmission depends on the amount of CO₂ and other gases in the atmosphere. For windows with a high CO₂ dependence, as the CO₂ concentration in the atmosphere increases, the spectral window closes, *ie* has less transmissivity, and more radiation is trapped in the atmosphere (Stewart 2006).

Figure 3.2: Transmittance of the infrared spectrum through the atmosphere at various latitudes (Stewart 2006).
Increased concentrations of CO₂, water vapor, and other greenhouse gasses lead to an greater greenhouse effect. This heightened greenhouse effect will include warming of the atmosphere, leading to an increase in evaporation, which will lead to additional water vapor in the atmosphere. This situation leads to a positive feedback loop of surface warming though greenhouse gasses. However, as the concentration of water vapor increases there is an increased likelihood of cloud formation (NESTA 2012). The effect of clouds on the weather and climate is twofold. The clouds increase earth’s albedo, reflecting incoming shortwave radiation and thus cooling the earth’s surface. However, clouds also serve to confine heat close to the surface, which leads to warmer temperatures. The dominant effect of clouds is to block incoming shortwave radiation.

The net IR radiation depends on a number of factors. The thicker the cloud, the more heat is trapped near the surface. The height of the cloud determines the radiation rate of the cloud, proportional to the cloud base temperature $T_b$, where lower clouds will be warmer than higher clouds. The water vapor content and surface temperature of the ocean affects the net IR radiation, as water vapor traps heat near the surface and the temperature of the ocean determines the outgoing IR radiation. Although it is not applicable to most of the ocean, ice and snow covered regions of the ocean are insulated sea from the atmosphere (Stewart 2006).
3.3 The Cool Skin (Thermal Skin Layer)

The interaction between the ocean surface and the lower atmosphere results in the exchange of heat. This heat exchange alters the temperatures in the immediate vicinity of the surface, causing the uppermost layer of the sea surface to be a few tenths of a degree cooler than the subsurface water. This millimeter-scale thermal layer that contains the concentrated temperature change is called the cool skin of the ocean (Minnett and Barton 2010). The study of the cool skin is key to understanding heat exchange between the ocean and overlying atmosphere. The sea surface temperature is a key climate variable and predictor for future weather patterns (Polyakova, Journel et al. 2006). The temperature change at the surface is defined by molecular thermal conductive properties. The thermal gradient is confined to a thin layer, \( \delta \), as beneath this the vertical transport of heat is done by turbulence. The cool skin effect exists between the subskin and atmosphere even when the subskin temperature is warmer than the water at depth, as illustrated in Fig. 3.3.

The exchanges at the air-sea interface, through the electromagnetic skin layer, are usually directed from the ocean to the atmosphere and act to cool the surface of the ocean. The cool skin layer absorbs longwave radiation, but absorbs only a small fraction of incoming solar radiation. The shortwave radiation is absorbed by subskin waters, creating a diurnal heating effect that influences the skin, subskin, and bulk water temperatures.
3.4 Description of Models

Due to the lack of spatial coverage of the radiometric temperature measurements, we explore the use of models to extend the datasets to be analyzed. The TOGA-COARE model is based on measurements from the Tropical Ocean Global Atmosphere Coupled Ocean Atmosphere Response Experiment (TOGA-COARE), an international field experiment conducted from 1992-1993 to study the oceanographic and atmospheric processes of the ‘warm pool’ region over the western Pacific (Fairall et al. 1996). The
TOGA-COARE model code is a result of the ultimate goal of the TOGA-COARE project: “to improve air-sea interaction and boundary layer parameterizations in models of the ocean and atmosphere, and to validate coupled models” (Webster and Lukas 1992). The models inputs and outputs are shown in Table 3.1.

<table>
<thead>
<tr>
<th>Inputs</th>
<th>Outputs</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tsea: sea surface temp (at about 0.05m depth)</td>
<td>Hf: sensible heat flux</td>
</tr>
<tr>
<td>Tair: air temperature (about 15 m)</td>
<td>Ef: latent heat flux</td>
</tr>
<tr>
<td>qair: air specific humidity (about 15 m)</td>
<td>SST: sea skin temperature</td>
</tr>
<tr>
<td>Rs: solar irradiance</td>
<td>Tau: surface stress</td>
</tr>
<tr>
<td>Ri: downwelling longwave irradiance</td>
<td>Wbar: mean Webb vertical velocity</td>
</tr>
<tr>
<td>Rain: precipitation</td>
<td>RF: Sensible heat flux due to precipitation</td>
</tr>
<tr>
<td>Lat: latitude (N=+)</td>
<td>Dter: cool skin effect deg.C</td>
</tr>
<tr>
<td>Lon: longitude (E=+)</td>
<td>Dt_wrm: warming across entire warm layer</td>
</tr>
<tr>
<td>Tdepth: temperature at 6m depth</td>
<td>Tk_pwp: warm layer thickness m</td>
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<tr>
<td></td>
<td>Tkt*1000:tkt=cool skin thickness</td>
</tr>
<tr>
<td></td>
<td>Wg: gustiness velocity m/s</td>
</tr>
</tbody>
</table>

Table 3.1: COARE model inputs and outputs

To further comprehend the coupled ocean-atmosphere system, TOGA-COARE focused on describing and understanding:

“1) The principal processes responsible for the coupling of the ocean and the atmosphere in the western Pacific warm-pool system;
2) The principal atmospheric processes that organize convection in the warm-pool region;
3) The oceanic response to combined buoyancy and wind-stress forcing in the western Pacific warm-pool region;
4) The multiple-scale interactions that extend the oceanic and atmospheric influence of the western Pacific warm-pool system to other regions, and vice versa.”

With these goals in mind, an intensive observation period occurred from November 1992 through February 1993 in the western Pacific region. The main component of the observational endeavors was an intensive flux array, which produced a high quality dataset necessary to calculate interfacial fluxes of heat, momentum, and moisture. This dataset also offers a baseline dataset for a number of remotely sensed variables (Webster and Lukas 1992).

The TOGA-COARE project allowed a thorough study of the western Pacific warm-pool region; however, the project also highlighted a number of gaps in our understanding of the area that prohibit the necessary detail needed to parameterize the fluxes at the air-sea interface. Many of these knowledge gaps relate to the air-sea $\Delta T$. Within tropical atmospheres, small variations in the sea surface temperature (SST) may have large impacts on the overlying atmosphere. These impacts are amplified in areas of highest SST, such as the western Pacific warm-pool. Within the western Pacific warm-pool, small changes in the SST have a greater effect on the overlying atmosphere, compared to other tropical atmospheric regions. It follows that uncertainties in SST measurements will profoundly restrict the ability to predict the state of the ocean-atmosphere system on several time scales (Webster and Lukas 1992). Numerous atmospheric circulation models support this behavior. The sensitivities tested by Palmer and Mansfield (1984) demonstrate that increased atmospheric perturbations occur when regions of the tropical
ocean are warmer than normal. Further comparisons with global climate models support findings of increased atmospheric sensitivity in areas of the warmest SST (Geisler 1985). Small changes in SST in the warmest ocean areas appear to trigger larger atmospheric responses than bigger SST changes in cooler water. It is still unclear how this sensitivity is related to the time and spatial scales of SST variability (Webster and Lukas 1992). Within the warm pool, the zonal advection of mean SST anomalies is thought to play a role in decadal oscillations of the SST (Zeng-Zhen Hu 2011). The same zonal advects also contribute to events on the seasonal time scale, due to interactions with wind patterns and the latent heat flux (Ueki 2011).

Further uncertainties arise because the SST of the tropical warm-pool areas reacts greatly to changes in the overlying atmosphere, in particular the latent heat flux and wind speeds. Gent (1991) used a primitive equation model to show that when the minimum wind speed for calculating latent fluxes was lowered from 4 m/s to 3 m/s, the net heat flux into the warm pool rose from 13.5 W/m² to 18 W/m² and the SST increased from 29.5°C to 32°C (Gent 1991). Using a more basic model, Seager et. al. (1988) demonstrated that the sensitivity of SST to changes in net heat flux is approximately 1°C per 12 W/m² (Seager, Zebiak et al. 1988).

With these complications in mind, it becomes apparent that many of the older air-sea ΔT datasets feature sea surface temperature data that are simply not as accurate as needed to adequately describe the air-sea interactions. Considering the previously mentioned 0.1 K SST accuracy goal, the need for a deeper understanding of the physics of the SST is apparent, especially in regards improved accuracy for satellite correction algorithms.
There is a demand for more accurate air-sea ΔT measurements, which may be met by hyperspectral radiometers that feature the necessary accuracy and traceability (Minnett, Maillet et al. 2005).

The TOGA-COARE bulk air-sea exchange algorithm was developed by the COARE community following the intensive observation period. The model considered the light wind, highly convective characteristics of the warm-pool region, and was based on the model of Liu, Katsaros, and Businger (LKB) (Liu, Katsaros et al. 1979). The original code (November 1993) includes changes to the LKB model to account for wind roughness length, Monin-Obukhov profile functions for strong convection, low-wind ‘gustiness’, and cool skin physics. Version 2.0 (August 1994) includes numerous additions, such as modeling of the ocean cool skin physics, ocean mixing, calculation of fluxes of momentum and sensible heat due to rainfall, and the Webb correction to latent heat. The Webb correction accounts for the behavior of the eddy covariance fluxes of water vapor and carbon dioxide due to the mean upward velocity caused by the evaporative flux of water and the effects of water vapor on the mixture density (Webb 1980). The code in version 2.0 allows for a conversion from bulk to skin temperature for calculating surface fluxes. A detailed description of the code and relevant updates may be found within the COARE_readme file on C.W. Fairall’s webpage (Fairall 2003).

STREAMER, a radiative transfer model, is another model that is useful in simulating radiometric measurements. Released in 2001, Streamer computes radiances for a broad range of atmospheric and surface conditions. The STREAMER inputs are the ocean and atmospheric states, and the outputs include surface and top of atmosphere radiative
fluxes, surface albedo, and cloud radiative effect. Output includes surface and top of the atmosphere radiative fluxes, surface albedo, and cloud radiative effect for flux calculations (Key 2001).

Utilizing the TOGA-COARE and STREAMER models permit an analysis of the heat exchange feedback loops that affect and are affected by the air-sea ΔT. Understanding the ship based and in-situ data may lead to increased accuracy in satellite retrievals of air-sea temperature differences.

3.5 **Feedback Loops and Air-Skin ΔT**

As previously stated, the large amount of air-skin ΔT data centered around -1 K suggests that there is potential ‘feedback’ loop of oceanic and atmospheric variables that stabilizes the air-sea ΔT around -1 K. The histogram distribution shows a sharp drop-off at air-skin ΔT equal to zero, but displays a long tail as the air-skin ΔT becomes more negative. The feedback loop will feature the sensible, latent, longwave, and shortwave fluxes, and may be dependent on any number of atmospheric and oceanic variables. It is the aim of this research to determine the strongest variable associations and to numerically characterize the contributions of each flux to such a repeatable distribution of the air-sea temperature difference. A conceptual diagram of the air-sea feedbacks is
Figure 3.4: Conceptual diagram for the fluxes involved in the air-skin $\Delta T$ feedback loops. Negative feedbacks serve to stabilize the air-skin $\Delta T$, while positive feedbacks will allow value of the air-skin $\Delta T$ to deviate from its $\sim 1$ K median value. Details are discussed in the text.

shown in Fig. 3.4. The diagram is centered around the air-skin $\Delta T$, and explores the consequences of an initial disturbance making the air-skin $\Delta T$ more negative by warming the skin layer (top half) and making the air-skin $\Delta T$ more positive though cooling the skin layer (bottom half), under night time conditions with constant cloud coverage. Red boxes represent processes that further encourage the initial disturbance (positive feedbacks) in the air-skin $\Delta T$, while blue boxes symbolize processes that dampen the initial air-skin $\Delta T$ disturbance, pushing the air-skin $\Delta T$ towards its mean state (negative feedbacks). The
white boxes describe a process that may increase or decrease the air-skin $\Delta T$ depending on the oceanic and atmospheric conditions. The goal of this research is to, using cruise and model data, quantify these feedback loops. For a given change in the air-skin $\Delta T$, each flux reacts in a specific manner. The sensible heat flux always acts as a negative feedback, dampening any sort of disturbance in the air-skin $\Delta T$. If the initial disturbance involves a change in the ocean temperature, the upwelling longwave radiation will act as a negative feedback as well. The quantification of these reactions, these feedback loops, will enhance understanding of the ocean-atmosphere coupling, especially in regards to the world-wide -1 K distribution of the air-skin temperature difference.
Chapter 4

Results

This chapter contains the simulation results and field data gathered over various cruises. The simulated data comes from the STREAMER and COARE models. The model data are then used as a starting point to guide investigation of the cruise datasets. Within these datasets are radiometric air skin ΔT measurements, as well as flux estimates. The air-skin ΔT data will be analyzed and presented in an effort to characterize the feedbacks between the air-skin ΔT, the sensible, latent, longwave, and shortwave heat
fluxes, and identify key oceanic and atmospheric conditions and variables that affect the air-skin $\Delta T$.

4.1 **COARE and STREAMER Preliminary Results**

The COARE and STREAMER models allowed for a large variable input range, simulating environmental conditions extending beyond those of the M-AERI cruise deployments. The possibility of a large range of input conditions allowed for the identification of trends of interest, which were then investigated more thoroughly in the field and ECMWF datasets. The models also function as a ‘sanity check’ to the field data, and vice-versa.

4.1.1 **STREAMER Results**

The STREAMER model was used to investigate a range of simplified conditions at the ocean-atmosphere interface. The inputs were cloud free night-time skies with negligible water vapor and aerosol content. The geographic and temporal characteristics were those of the Nauru’99 M-AERI deployment, as STREAMER considers both geographic location and solar zenith angle in its calculations. The STREAMER model calculates the longwave fluxes at given specified heights, considered ‘layers’ within the model. Figure 4.1 shows net longwave fluxes, calculated at various humidity values, for
constant temperatures of the ocean and atmosphere. In Fig. 4.1, the net longwave flux is calculated from the longwave radiation emitted from the skin layer and downward longwave radiation from a height of 1km.

### 4.1.2 COARE Results

The COARE model was used to simulate the non-solar heat fluxes over the western warm pool. The COARE model was run to analyze non-solar heat fluxes. For a given wind speed, the non-solar heat flux and skin-subskin $\Delta T$ COARE outputs are shown in Fig. 4.2. For the skin-subskin $\Delta T$, the non-solar heat fluxes are dominated by the wind speed, with a larger range of flux values for higher wind speeds. As the skin-subkin $\Delta T$ becomes more positive, or actually becomes positive, the range of values for the non-
solar heat fluxes increases. The model’s output of the non-solar heat fluxes is consistent with expectations and the bulk formulas.

The nighttime conditions shown in Fig. 4.3 resemble the COARE output data (Fig. 4.2). However, the daytime conditions in Fig. 4.3 are subject to intense solar radiation and as such are not as reliable in relating the wind speed, heat flux, and skin-subskin $\Delta T$ because there is not a fair comparison. The relationship between wind speed, non-solar heat flux, and air-sea $\Delta T$ is shown in Fig. 4.4. These studies lend confidence to the COARE model, and emphasize the importance of obtaining data free of the artifacts of diurnal heating.

![Figure 4.2: COARE model output showing the relationship of skin-subskin SST on wind speed and heat flux (Sensible and Latent Heat fluxes, contours are the result of discrete intervals used as model inputs)](image-url)
Figure 4.3: Contour plot of skin-subsurface ΔT as a function of wind speed and total (non-solar) surface heat flux for nighttime (upper) and daytime data (lower). Heat flux is positive from ocean to atmosphere (Murray, Allen et al. 2000)

Figure 4.4: COARE model output showing the relationship of air-skin SST difference with wind speed and heat flux (non-solar heat fluxes, contours are the result of discrete intervals used as model inputs)
The output from the COARE model was used to quantify the change in the latent and sensible heat fluxes per degree change in the air-sea $\Delta T$, for wind speeds ranging from $u=0$ - 10 m/s. These wind speeds are chosen as they cover the range that predominates over the ocean, especially in the tropics (Chandler 2004), and span the range experienced during M-AERI measurements. The relationship of the sensible and latent heat fluxes and air-skin $\Delta T$ at wind speeds from 0 m/s to 10 m/s are shown in Fig.4.5 and Fig. 4.6.

Figure 4.5: The change in latent heat flux per degree change in the air-skin $\Delta T$, at wind speeds 0-10 m/s in 1 m/s intervals. Data are from COARE model output.
Figure 4.6: The change in sensible heat flux per degree change in the air-skin $\Delta T$, at wind speeds 0-10 m/s, in 1 m/s intervals. Data are from COARE model output.

Figure 4.7: COARE skin temperature (calculated using ECMWF data as inputs) minus ECMWF skin temperature, for the Nauru’99 Region; n=525, mean difference $=-0.32 \text{ K} \pm 0.04 \text{ K}$
An instance of the COARE model was run using ECMWF data from the Nauru’99 region. All of the COARE input variables (see Table 3.1) were taken from the ECMWF data. The difference in oceanic skin temperatures from the ECMWF-based COARE and ECMWF is shown Figure 4.7. The ECMWF oceanic skin temperature is on average 0.32 K greater than that simulated by the COARE model using the ECMWF forcing as input. The difference in latent heat flux from the ECMWF-based COARE and ECMWF outputs is shown in Figure 4.8. The ECMWF latent heat flux is always greater than the COARE output for the Mirai region by a mean difference of 63.3 W/m$^2$. The possible explanation behind this difference will be explored in more detail in the heat flux section of this chapter.

Figure 4.8: COARE latent heat flux (calculated using ECMWF data as inputs) minus ECMWF latent heat flux, for the Nauru’99 Region; n=525, mean difference=-63.3 W/m$^2 \pm 17.4$ W/m$^2$
4.2 Cruise Data: The M-AERI Measurements

The data gathered by the Nauru’99 (1999), Explorer of the Seas (2000-2006), SAGE (2004), EGEE (2006), and Cirene (2008) cruise deployments of the M-AERI will be investigated with regards to the air-sea ΔT. The Explorer of the Seas offers the longest M-AERI dataset at six years, and the histogram of air-sea ΔT measurements from the Explorer of the Seas was shown in Figure 2.2. A lack of flux and wind speed measurements (or measurements of a complete set of variables that could be used in COARE to calculate fluxes) prevents an analysis of the relationship between air-sea ΔT and the heat fluxes and ocean-atmosphere boundary layer for the Explorer of the Seas data; similarly for the Cirene dataset.

The Nauru’99 cruise of the R/V Mirai was centered around the island of Nauru in the western Pacific during June of 1999. The dataset contains sensible, latent, longwave, and shortwave fluxes, in addition to the M-AERI oceanic and atmospheric temperature measurements. The air-sea ΔT histogram distribution from the Nauru’99 M-AERI data is shown in Figure 4.9, and the air-skin ΔT distribution from ECMWF data in the area of the Nauru’99 cruise is shown in Figure 4.10. The mean air-skin ΔT values are similar, with the M-AERI and ECMWF mean air-skin ΔT values of -0.92 K and -1.12 K respectively. Neither dataset contains positive air-skin ΔT values, although the ECMWF data have nearly double the variability, and contain negative values that are below the minimum air-skin ΔT measurements of the M-AERI. The ‘tail’ of the air-sea ΔT
Figure 4.9: Histogram of the air-skin ΔT for M-AERI data from the Nauru’99 cruise; n=924, mean=−0.92 K ± 0.41 K

Figure 4.10: Histogram of the air-skin ΔT for the Nauru’99 region of the ECMWF data with n=525, mean=−1.12 K ± 0.72 K. Note the change in x-axis compared to figure 4.8
Figure 4.11: Histograms of the air-sea ΔT from daytime Nauru’99 M-AERI data, binned by wind speeds of less than 3 m/s (a), 3 to 5 m/s (b), and 5 to 9 m/s (c)

Figure 4.12: Histograms of the air-sea ΔT from nighttime Nauru’99 M-AERI data, binned by wind speeds of less than 3 m/s (a), 3 to 5 m/s (b), and 5 to 9 m/s (c)
distribution occurs during times of low wind speeds (less than 3 m/s) and high solar insolation. The distribution of the air-sea ΔT binned into wind speeds of less than 3 m/s, 3 m/s to 5 m/s, and 5 m/s to 9 m/s for the daytime and night time M-AERI measurements and ECMWF data is shown in Figure 4.11, Figure 4.12 and Figure 4.13. For wind speeds less than 3 m/s, the daytime M-AERI and ECMWF data have similar mean air-skin ΔT values at -1.42 K and -1.43 K respectively. The large negative ‘tail’ of the air-sea ΔT histogram for the ECMWF data is present even at high wind speeds, and this is not so within the M-AERI data. As the wind speed increases, the mean air-skin ΔT value shifts towards in a positive direction and the variability of the M-AERI air-skin ΔT data decreases. The nighttime M-AERI data from the Nauru’99 campaign (Fig 4.11) do not
feature the negative tail seen in the daytime data, and 99% of the air-skin ΔT values are more positive than -1.5 K. As the wind speed increases, the air-skin ΔT distribution shifts towards zero.

Binning the M-AERI and ECMWF air-skin ΔT data by wind speed, it becomes apparent that at lower wind speeds the air-skin ΔT is more negative, and the distribution of the air-skin ΔT shifts in a positive direction as the wind speed increases. Although the mean air-skin ΔT distribution shifts in a positive direction with increasing wind speed, in both the M-AERI and ECMWF data there are still no positive values recorded for the air-sea ΔT. The positive shift in the air-sea ΔT suggests that the sensible and latent heat fluxes, which increase with wind speed, provide a negative feedback that stabilizes the air-sea ΔT at approximately -1K. Table 4.1 and Table 4.2 show this trend in the Nauru’99 M-AERI and ECMWF air-skin ΔT data, binned in increments of 1 m/s. This trend is seen within the field data for the Explorer of the Seas, and Cirene cruises as well; however, the ECMWF data from the Explorer of the Seas region fails to capture this trend, with the mean air-skin ΔT becoming more negative as the wind speed increases. In Table 4.2 the distribution differs from the traditional distribution seen in the Nauru’99, Explorer of the Seas, and previous cruise data discussed in chapter 2.2.1. The distribution is more Gaussian, and is due to the existence of the cold tongue. By separating the air-skin ΔT data into inside and outside the cold tongue region, shown in Figure 4.16, it is apparent that the cooler ocean temperatures within the cold tongue result in much smaller negative values of the air-skin ΔT, and even some positive values. The mean air-skin ΔT value for
<table>
<thead>
<tr>
<th>Wind Speed (m/s)</th>
<th>Data Points</th>
<th>Mean ΔT (K)</th>
<th>Std Dev ΔT (K)</th>
<th>Percentage ΔT &lt; -1</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>52</td>
<td>-1.28</td>
<td>0.44</td>
<td>0.88</td>
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<td>7</td>
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</table>

Table 4.1: Nauru’99; All M-AERI air-skin ΔT data binned by wind speed

<table>
<thead>
<tr>
<th>Wind Speed (m/s)</th>
<th>Data Points</th>
<th>Mean ΔT (K)</th>
<th>Std Dev ΔT (K)</th>
<th>Percentage ΔT &lt; -1</th>
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<td>5</td>
<td>144</td>
<td>-0.94</td>
<td>0.54</td>
<td>0.38</td>
</tr>
<tr>
<td>6</td>
<td>63</td>
<td>-0.95</td>
<td>0.53</td>
<td>0.30</td>
</tr>
<tr>
<td>7</td>
<td>49</td>
<td>-0.83</td>
<td>0.46</td>
<td>0.24</td>
</tr>
<tr>
<td>8</td>
<td>8</td>
<td>-1.62</td>
<td>0.54</td>
<td>0.75</td>
</tr>
</tbody>
</table>

Table 4.2: Nauru’99 region ECMWF air-skin ΔT binned by wind speed

8 m/s is abnormally high, but may be due to the small sample size of n=8 for this wind speed bin. The EGEE air-skin ΔT data differs in its distribution, and these differences will be explored further in their respective sections of the results.

The effects of diurnal heating on the air-skin ΔT are shown in Figure 4.14. As the solar insolation increases throughout the day, the subskin ocean waters absorb the radiation. At low enough wind speeds (<3 m/s) this heat is not thoroughly mixed by
wind-driven turbulence, and remains near the surface causing the skin temperature of the ocean to increase (Price, Weller et al. 1986). It should be noted that the cool skin effect still exists, even as the skin temperature rises.

The EGEE cruise took place aboard the R/V L’Atalante in the cold tongue region of the eastern Atlantic in June 2006. The M-AERI distribution of the air-skin ΔT (Fig. 4.15)

Figure 4.14: Cirene air-skin ΔT measurements taken by the M-AERI by local time of day, with windspeed colorbar
Figure 4.15: Histogram of the M-AERI air-skin ΔT data from the EGEE cruise with n=4170, mean=-0.70 K ± 0.42 K.

Figure 4.16: Histogram of the M-AERI air-skin ΔT data from the EGEE cruise outside (black) and inside (red) of the Cold Tongue. Region outside cold tongue (n=2869) has mean air-skin ΔT of -0.82 K ± 0.41 K. Region inside the cold tongue (n=1301) has mean air-skin ΔT of -0.45 K ± 0.33 K.
The most prominent difference in the air-skin $\Delta T$ in the cold tongue, compared to elsewhere, is the existence of positive air-skin $\Delta T$ values. In the Cirene and *Explorer of the Seas* air-skin $\Delta T$ studies, the M-AERI did not show any positive air-skin $\Delta T$ values; this is one of the major differences seen in M-AERI air-skin $\Delta T$ data compared to conventional thermometer air-skin $\Delta T$. The positive air-skin $\Delta T$ measurements occur while the *L’Atalante* is travelling through the cold tongue on June 4th (case one) and June 13th (case two).

![Figure 4.17 SST (K) measured by AMSR-E on Aqua with overlaid EGEE ship tracks through the cold tongue (black) and wind vectors (blue). The two cases of positive air-skin $\Delta T$ M-AERI data from the EGEE cruise may be explained by the ship’s position in the Cold Tongue given the wind direction and wind speed, as described in the text.](image-url)
of 2006. In the first case, the ship is heading due south, with constant winds from the south at ~5 m/s to 7 m/s. The first measurement of positive air-sea ΔT occurs as the ship approaches the southern boundary of the cold tongue. It is reasonable to conclude that warmer air from outside the cold tongue was blown into the M-AERI measurement path within cold tongue, causing a positive air-skin ΔT. When the ship first records a positive air-skin ΔT, at approximately 200 km from the southern boundary of the cold tongue, with a conservative wind speed assumption of 5 m/s, it would take the warmer air ~11 hours to blow to the position of the M-AERI. As the ship continues due south, the time for warmer air to reach the ship decreases, resulting in more positive air-skin ΔT measurements. The air-skin ΔT remains positive once the *L’Atalante* reaches the southern end of the cold tongue. Once the ship reaches the end of the cold tongue, the air-skin ΔT measurements become more negative, characteristic of the black histogram in Figure 4.16.

The second case of positive air-skin ΔT measurements occurred on June 13th, as the ship travels northeast through the cold tongue. The air-skin ΔT values become positive as the ship enters the cold tongue, with steady winds blowing from the southeast at 8m/s to 9m/s. As the ship crosses the cold tongue, the winds blow warmer air from outside the cold tongue into the M-AERI measurement path. At the northern boundary of the cold tongue, the winds would take ~6.5 hours to travel through the cold tongue from the south. As the ship enters the cold tongue and begins to record positive air-skin ΔT values, the warm air from outside the cold tongue reaches the ship’s position in less than one hour.
Table 4.3: EGEE cruise M-AERI air-skin ΔT data binned by wind speed

<table>
<thead>
<tr>
<th>Wind Speed (m/s)</th>
<th>Data Points</th>
<th>Mean ΔT (K)</th>
<th>Std Dev ΔT (K)</th>
<th>Percentage ΔT &lt;0</th>
<th>Percentage ΔT &lt;1</th>
</tr>
</thead>
<tbody>
<tr>
<td>2</td>
<td>4</td>
<td>-0.14</td>
<td>0.17</td>
<td>0.75</td>
<td>0.00</td>
</tr>
<tr>
<td>3</td>
<td>30</td>
<td>-0.42</td>
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<td>1.00</td>
<td>0.03</td>
</tr>
<tr>
<td>4</td>
<td>121</td>
<td>-0.41</td>
<td>0.37</td>
<td>0.90</td>
<td>0.05</td>
</tr>
<tr>
<td>5</td>
<td>221</td>
<td>-0.44</td>
<td>0.41</td>
<td>0.87</td>
<td>0.07</td>
</tr>
<tr>
<td>6</td>
<td>302</td>
<td>-0.48</td>
<td>0.42</td>
<td>0.89</td>
<td>0.11</td>
</tr>
<tr>
<td>7</td>
<td>345</td>
<td>-0.55</td>
<td>0.40</td>
<td>0.94</td>
<td>0.12</td>
</tr>
<tr>
<td>8</td>
<td>164</td>
<td>-0.62</td>
<td>0.45</td>
<td>0.92</td>
<td>0.16</td>
</tr>
</tbody>
</table>

Figure 4.17 illustrates the ship path and wind vectors for both cases of positive air-skin ΔT measurements.

Within the EGEE data, the mean air-skin ΔT values do not decrease with increasing wind speed, as the Explorer and Nauru’99 M-AERI measurements do. The air-skin ΔT values for the EGEE cruise, shown in Table 4.3, increase with increasing wind speed. The reasons behind this behavior will be explored in the discussion section.

4.3 The Air-Skin ΔT and Heat Fluxes

The Nauru’99, SAGE, and EGEE cruise datasets feature at-sea measurements of both the air-skin ΔT and surface heat fluxes. The Nauru’99 and SAGE cruises contain measurements of the solar insolation, latent, sensible, and longwave heat fluxes. The EGEE campaign contains at-sea measurements of longwave and shortwave radiation; the COARE model was used to calculate the sensible and latent heat fluxes based on the at-sea data. The wind speeds for the SAGE cruise were relatively high (over 10m/s) for the
majority of the campaign, which in addition to the proximity of the cruise to New Zealand and the surrounding islands makes the measurements difficult to compare to models.

The Nauru’99 cruise data for the sensible and latent heat fluxes are compared to the ECMWF data in Figure 4.18 and Figure 4.19. For the sensible heat flux, the ECMWF data matches the output of the bulk formula as expected. In comparison, the sensible heat flux field data exhibits a smaller magnitude. The field data reproduces the wind speed and air-skin $\Delta T$ dependence for the sensible heat flux, but at higher wind speeds the field data does not exhibit the exaggerated values seen in the ECMWF data. For the latent heat flux data, the magnitudes of the field and ECMWF data agree closely. The distinguishing feature between the field and ECMWF latent heat flux data is that the ECMWF data shows no dependence on the air-skin $\Delta T$, whereas the field data shows a definite correlation with the air-skin $\Delta T$. For each wind speed range, the field, COARE and sensible heat flux data from the Nauru’99 cruise are shown in Figure 4.20. The sensitivities of the heat flux to the air-skin $\Delta T$ represent the slope of the fitted lines. These represent how the sensible heat flux will, theoretically, change given a perturbation in the air-skin $\Delta T$. As the wind speed increases, the sensible heat flux sensitivity predictably increases for each dataset. The field sensible heat flux sensitivity dataset increases only slightly in magnitude at winds speeds higher than 5 m/s, whereas both models show a larger increase in magnitude at higher wind speeds. For each wind speed range, the field, COARE and ECMWF latent heat flux data and sensitivities from the
Figure 4.18: Sensible heat flux plotted against the air-skin ΔT with wind speed colorbar for ECMWF Nauru’99 region data (left) and Nauru’99 Field data (right)

Figure 4.19: Latent heat flux plotted against the air-skin ΔT with wind speed colorbar for ECMWF Nauru’99 region data (left) and Nauru’99 Field data (right)
Figure 4.20: Sensible heat flux plotted against the air-skin $\Delta T$ for Nauru’99 field data (black) and COARE (blue) and ECMWF (red) model data, binned by wind speeds of less than 3 m/s (a), 3 to 5 m/s (b) and greater and 5 m/s (c). The table shows the sensitivities of the sensible heat flux (ie the slopes of the fitted lines) with respect to the air-skin $\Delta T$ binned by wind speeds in m/s.

<table>
<thead>
<tr>
<th>Wind Speed (m/s)</th>
<th>U&lt;3</th>
<th>3&lt;U&lt;5</th>
<th>U&gt;5</th>
</tr>
</thead>
<tbody>
<tr>
<td>Field Data</td>
<td>2.1</td>
<td>5.2</td>
<td>5.3</td>
</tr>
<tr>
<td>COARE</td>
<td>4.2</td>
<td>6.7</td>
<td>10</td>
</tr>
<tr>
<td>ECMWF</td>
<td>4.6</td>
<td>6</td>
<td>8.9</td>
</tr>
</tbody>
</table>

Nauru’99 cruise are shown in Figure 4.21. The ECMWF data shows zero dependence on the air-skin $\Delta T$, whereas the field and COARE data show the latent heat flux sensitivity to increase with increasing wind speed. The field data, similar to the sensible heat flux, shows only a slight increase in the sensitivity when moving from wind speeds of 3 m/s -5
m/s to wind speeds above 5 m/s. The actual release of the latent heat flux back into the atmosphere, and its effect on the air-skin ΔT feedback loops, will be explored in Chapter 5.
Chapter 5

Discussion

The results described in Chapter 4 will be examined with respect to the literature concerning the models, air-skin $\Delta T$, heat fluxes, and feedback loops. The accuracy of the M-AERI has provided detailed measurements of the air-skin $\Delta T$, which point to consistencies in the air-skin $\Delta T$ distribution, regardless of location or time of year. Both negative and positive anomalies exist within these distributions, and they are accounted for through analysis of the data and comparison to literature. A system of atmospheric and oceanic radiative feedback loops is suggested to explain the restoration of the air-skin $\Delta T$ from anomalous values back to the mean value of -1 K. The nature of these loops, and
their combined effects, will be assessed in detail to determine their cumulative effects and the environmental processes largely responsible for the -1 K mean value of the air-skin ΔT distribution.

5.1 **Worldwide trends in Radiometrically Measured Air-Skin ΔT Distributions**

The air-skin ΔT distributions from the M-AERI deployments and the ECMWF data feature a mean value of approximately –1K. While the distributions are focused around -1K, there are numerous anomalous measurements found within the data. The air-skin ΔT must return to its mean value following the occurrence of anomalous values, and this chapter aims to explain the causes of the anomalies and the feedback processes by which the air-skin ΔT is returned to its mean value.

5.2 **Anomalous Air-Skin ΔT Measurements**

The negative ‘tail’ of the distribution of the air-skin ΔT for the M-AERI data (Fig. 4.11) may be caused by solar insolation combined with low wind speeds; resulting in diurnal heating. Another possibility for the large negative ΔT was proposed by Sikora (2012), showing that the ratio of buoyancy to shear in the boundary layer may be approximated by the ratio of air-sea ΔT to wind speed.

Figure 5.1 shows a case study of a negative deviation of the air-sea ΔT during the Nauru’99 cruise. In this case, the ocean is heated by SW radiation. As this value
increases, atmospheric cellular convection develops that carries the heat provided by the ocean upwards. In this case, the ocean is heated by SW radiation. Meanwhile, cool air sinks, is warmed, then lifted by the convection. This causes the air-sea ΔT to deviate from its peak distribution value of -1K, resulting in more negative air-skin ΔT values (Sikora 2012).

A difference between the Nauru’99 air-skin ΔT histograms when binned by wind speed is the large number of negative anomalies found within the ECMWF data, regardless of wind speed (Fig. 4.13). As shown in Fig. 4.11 and Fig. 4.12 the negative air-skin ΔT anomalies within the Nauru’99 field data exist almost exclusively at times of low wind speed and high solar insolation. The ECMWF Interim dataset has been known to, at times, contain excessive solar radiation that causes a net heat flux into the ocean, especially in convective tropical areas (Dee, Uppala et al. 2011). The ERA-Interim dataset has been shown to be substantially improved in regards to the seasonal and interannual variability in solar radiation, and also in the spatial structure of incoming solar radiation, compared to ERA-40 data (Balmaseda and Mogensen 2010). In one comparison of the ECMWF Interim and ECMWF-40 datasets, the quality of ERA surface fluxes were analyzed by using them to initialize the ocean elements of the ECMWF seasonal forecasting system. The skill of the forecast for the ECMWF Interim dataset of the tropical SST was found to improve consistently compared to the ERA-40 data, with the greatest improvements in the tropical Atlantic regions (Balmaseda, Vidard et al. 2008).
Despite this improvement, the achievable skill of the ECMWF seasonal forecasts depends largely on the quality of estimated fluxes of heat, momentum, and water at the atmosphere-ocean interface (Dee, Uppala et al. 2011). The excessive solar radiation within the model may explain the large number of negative anomalies seen in the ECMWF air-skin ΔT distributions, regardless of wind speed. The local time of day for the ECMWF Interim data used in the Nauru’99 comparison is 11 am.

Figure 5.1: Case study of a negative ΔT deviation from the Nauru’99 cruise, taken from days 173-174 (June 23rd -24th)
Figure 5.2: Case study of a positive $\Delta T$ deviation from the SAGE cruise, taken from day 92-94 (April 2$^{nd}$ -4$^{th}$). The red boxes emphasize the processes described in the text, missing periods of data signify rainfall, when the M-AERI was not operational.

As mentioned in Chapter 2, positive $\Delta T$ anomalies from the -1 K mean may be attributed to wind blowing over the ship, warm winds blowing offshore, and atmospheric fronts moving through the measurement path. An example of a positive $\Delta T$ anomaly is present in the data from the 2004 SAGE cruise, shown in Figure 5.2. A period of rain during days 91-92.5 prevents M-AERI measurements; however, once the M-AERI is operational on day 92.5 there is a pronounced positive air-skin $\Delta T$. As the air-sea $\Delta T$ becomes less positive (is pushed back towards its mean state) there is an increase in wind speed, increase in downwelling longwave radiation, and a westward shift in the wind direction. This implies the positive air-skin $\Delta T$ was due to the passage of an atmospheric
warm front around day 92.9. As the front passes, the warm air mass rises into the region of lower pressure, expanding and cooling. Water vapor condenses as the air cools, and extensive cloud coverage develops as revealed in the increase in downwelling longwave radiation. There is also a decrease in downwelling shortwave radiation.

On day 94.4, the air temperature decreases and the air-sea $\Delta T$ returns to its characteristic value of -1K. This occurs after a marked increase in the relative humidity and a resulting decrease in the latent heat flux. This decrease in latent heat flux coupled with increasing winds (deepening of the boundary layer) cools the atmosphere and serves as a negative feedback to restore the air-sea $\Delta T$.

The anomalous positive air-skin $\Delta T$ measurements observed during the EGEE cruise largely relate to the passage of the ship through the cold tongue. The Gulf of Guinea contains the largest amplitudes of sea surface temperature variability in the tropical Atlantic, with a maximum variability of 7 K depending on the time of year (Picaut 1983). Equatorial upwelling is primarily responsible for the seasonal variability, with the formation of a cold tongue in boreal summer (Weingartner and Weisberg 1991). The skin surface temperature cooling is related to the presence of the cold tongue is correlated with the northward migration of the intertropical convergence zone (Gu and Adler 2004).

The EGEE M-AERI dataset features an air-skin $\Delta T$ distribution unlike the majority of M-AERI deployments. The distribution of air-skin $\Delta T$ measurements for the EGEE cruise was more bell-shaped, contained positive air-skin $\Delta T$ values, and was lacking the narrow peak seen in most M-AERI deployments. The cruise path through the cold tongue yielded anomalously positive air-skin $\Delta T$ measurements, despite the relatively weak cold
tongue that developed in 2006. Conditions for the 2006 EGEE cruise featured a reduced intensity of the cold tongue compared to 2005, with the cold tongue not extending past 5° S. Since the cold tongue did not extend past 5° S, the Gulf of Guinea was disconnected from the cold upwelling waters further south in the basin, which resulted in a reduced zonal SST gradient along the southern boundary of the cold tongue compared to 2005 (Marin, Caniaux et al. 2008). The vast majority of the observed positive air-skin ΔT measurements occurred due to winds blowing warmer air from outside the cold tongue region which was measured from the ship while inside the cold tongue, see Fig. 4.17. It should be noted that for 2006 the period of the most intense cold tongue occurred from June to August (Marin, Caniaux et al. 2008). The recorded positive air-skin ΔT measurements occurred on June 4\textsuperscript{th} and June 13\textsuperscript{th}, at the beginning of this intense period.

5.3 Radiative Feedbacks and the Air-Skin ΔT

M-AERI cruise data and COARE/ECMWF model output data suggest that the repeatability of the -1K peak within the air-skin ΔT distribution is largely stabilized by radiative feedback loops. The small but nearly unvarying temperature difference is sustained across the air-skin interface by the impedance of the surface layer to the heat flux from the ocean to the atmosphere (Grassl 1976). It becomes apparent that as the wind speed decreases the molecular (Saunders 1967) and consequently convective heat transfer (Katsaros, Liu et al. 1977) become dominant. It has been shown that these processes may maintain a substantial temperature gradient within the skin layer. As seen
in the Cirene and Nauru’99 datasets, periods of high solar insolation and low wind speeds obscures the interpretation of the air-skin ΔT measurement due to the effects of diurnal heating (Donlon, Nightingale et al. 1999). These effects were represented by the Cirene and Nauru99 datasets; however, as previously discussed, the ECMWF data contains a considerable amount of anomalous negative air-skin temperature differences at higher wind speeds.

In areas of low wind speeds, solar insolation may yield strong warming within the upper few meters of the ocean. With diminishing wind speeds (for Nauru’99 and Cirene, less than 3 m/s), the trapping depth of the oceanic thermal response is approximately 1 m, with heating anomalies up to 3 K (Price, Weller et al. 1986). With higher wind speeds, vertical wind mixing causes the effects of solar heating to reach significantly greater depth than is reached directly by radiation. The increased wind speeds extend the trapping depth of the solar insolation to approximately 10m, and the amplitude of the surface heating is decreased to roughly 0.2 K (Price, Weller et al. 1986). It should be noted, once again, that even in environments with low wind speeds and large amounts of solar insolation, the skin layer remains cooler than the layer (~ 5cm) beneath it (Minnett, Smith et al. 2010). When both the solar insolation and wind speed are known, the time scale and strength of the diurnal cycle will depend on how the stabilizing surface heat flux and destabilizing surface stress affect the trapping depth, as described in detail by Price et al (1986).

The negative air-skin ΔT anomalies in the Nauru’99 and Cirene datasets, which peaked around 2pm local time (see Fig 4.14), may be further separated into those with
wind speeds of 2 m/s or less, and those with wind speeds of 2 m/s to 4 m/s. The characteristic feature of diurnal heating is the time of day when the maximum amount of heating is reached. The most negative air-skin $\Delta T$ anomalies during periods of low winds and high solar insolation match the time of most intense heating of the ocean, as measured at 5m depth from the Nauru’99 cruise. The data with wind speeds less than 2 m/s had the largest negative air-skin $\Delta T$ anomalies and the air-skin $\Delta T$ measurements returned to their characteristic value of $\sim -1K$ within 5 hours as the 5m sea temperature relaxed from its maximum value at the peak of solar insolation. The negative anomalies associated with winds of 2 m/s to 4 m/s were smaller. The time scale for the restoration of negative anomalies involving intense solar radiation and wind speeds of 2 m/s – 4 m/s to the mean air-skin $\Delta T$ value was on the order of $\sim 3-4$ hours. It should be noted that for both cases, winds less than 2 m/s and winds 2 m/s – 4 m/s, the timing of the maximum air-skin $\Delta T$ anomalies match that of the maximum of the 5 m sea temperature. The time scale for the Nuaru’99 and Cirene air-skin $\Delta T$ negative anomalies agrees closely with the diurnal heating and cooling rates for equatorial ocean waters, which are shown for the equinox in Figure 5.3 (Woods, Barkmann et al. 1984). This provides further evidence that the negative air-skin $\Delta T$’s during periods of low winds and high solar insolation are related to diurnal heating processes. This diurnal temperature rise of the skin will increase the sensible heat flux, and may significantly affect the lower mixed layer (Kawai and Wada 2007), perhaps even triggering shallow cumulus convection (Slingo, Inness et al. 2003). The changes in the atmospheric mixed layer and the surface wind field in
The anomalous negative air-skin $\Delta T$’s caused by the skin temperature increase return to their mean value of -1K within a few hours of the maximum heating due to solar insolation. One may imagine the ocean-atmosphere system to be at the -1K air-skin $\Delta T$ value, and the warming of the skin layer (in this case due to enhanced solar radiation) to be viewed as an imposed disturbance to the system. Because the ocean and atmosphere are constantly interacting, this initial disturbance in the temperature of the skin layer will then affect subsequent ocean-atmosphere interactions, creating a feedback loop in which
the initial disturbance may be exaggerated or dampened depending on the cumulative effects of physical processes on the ocean-atmosphere state. We seek now to examine the cumulative effects of air-skin $\Delta T$ radiative feedbacks, and the ocean-environment conditions that lead to net positive and net negative feedback loops. A flow diagram representing the various feedback loops is shown in Figure 5.4.

![Flow Diagram](Image)

**Figure 5.4:** Summary of the air-skin $\Delta T$ radiative feedbacks at night time for the sensible, longwave, and latent heat fluxes for initial disturbances that warm the skin of the ocean (top half) and initial disturbances that cool the skin of the ocean (bottom half)
The sensible heat flux acts as a negative feedback loop to stabilize the air-skin $\Delta T$ and dampen any perturbations to the ocean-atmosphere system. When the air-skin $\Delta T$ becomes more negative (positive) due to an increase (decrease) in the skin temperature, the sensible heat emitted from the ocean acts as a negative feedback that dampens the initial disturbance and serves to ‘push’ the air-skin $\Delta T$ towards its mean value of -1 K. The sensible heat flux is dependent on the air-skin $\Delta T$ (Eq. 9, Section 3.1.3), and increasing the air-skin $\Delta T$ will cause the ocean to lose more heat, thus acting to cool the ocean and return the air-skin $\Delta T$ to its mean state. Likewise, when the air-skin $\Delta T$ becomes more positive due to a decrease in the air-skin $\Delta T$, the ocean will lose less energy through the sensible heat flux, acting to stabilize the air-skin $\Delta T$ value at -1K. As expected, the sensible heat flux loss is dependent on the wind speed, with larger feedback sensitivities with increasing wind speed. The magnitude of the sensible heat flux feedbacks ranges from about 5 W/m$^2$ to 15 W/m$^2$ at wind speeds of 1 m/s and 9 m/s respectively.

The amount of sensible and latent heat fluxes due to the various heat flux-air skin $\Delta T$ feedbacks is summarized from the Nauru’99, COARE data, ECMWF data, and bulk formula calculations in Table 5.1, Table 5.2. These tables are feedback sensitivities to the heat fluxes, presented as change in flux per degree change in the air-skin $\Delta T$. The sensitivity of the sensible heat flux to changes in the air-skin $\Delta T$ increases with wind speed in every dataset, with the smallest range of sensitivity found within the Nauru’99 field data, and the largest range of sensitivity found within the COARE model simulations.
When the air-skin $\Delta T$ changes due to an increase (decrease) in the skin temperature, the longwave radiation emitted from the skin layer will act as a negative feedback, increasing (decreasing) the amount of radiation emitted and acting to cool (warm) the skin layer. The magnitude of this feedback may be calculated easily using Stefan-Boltzman law, and the feedback ranges from approximately 4 W/m$^2$ to 7 W/m$^2$ per degree change in the skin temperature.

The air-skin $\Delta T$ feedback loops involving the latent heat flux are more complicated than those with the sensible heat flux. Following an initial disturbance in the air-skin $\Delta T$,

<table>
<thead>
<tr>
<th>Wind Speed (m/s)</th>
<th>Sensible Heat Flux Sensitivity</th>
<th>Latent Heat Flux Sensitivity</th>
</tr>
</thead>
<tbody>
<tr>
<td>$u&lt;3$</td>
<td>2.1</td>
<td>4.2</td>
</tr>
<tr>
<td>$3&lt;u&lt;5$</td>
<td>5.2</td>
<td>6.7</td>
</tr>
<tr>
<td>$u&gt;5$</td>
<td>5.3</td>
<td>10.0</td>
</tr>
</tbody>
</table>

Table 5.1: Sensible and latent heat flux sensitivities derived from the Nauru’99 field data (left of columns), COARE data (middle of columns), and ECMWF data (right of columns) expressed as W/(m$^2$*sec*K).

<table>
<thead>
<tr>
<th>Wind Speed (m/s)</th>
<th>Sensible Heat Flux Sensitivity</th>
<th>Latent Heat Flux Sensitivity</th>
</tr>
</thead>
<tbody>
<tr>
<td>$1$</td>
<td>1.5</td>
<td>5.30</td>
</tr>
<tr>
<td>$2$</td>
<td>3</td>
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<td>13.9</td>
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<td>$8$</td>
<td>13</td>
<td>27.8</td>
</tr>
<tr>
<td>$9$</td>
<td>15.5</td>
<td>31.30</td>
</tr>
</tbody>
</table>

Table 5.2: Sensible and latent heat flux sensitivities derived from the bulk formulas (see Section 3.1.3), expressed as W/(m$^2$*sec*K). The latent heat flux sensitivities here are conservative in their calculations, and may double, or triple, under certain environmental conditions (air-skin $\Delta T$ becomes more positive, high wind speed and low humidity).
the latent heat flux may become a negative or positive feedback. If the air-skin $\Delta T$ becomes more negative through an increase in the skin temperature, the latent heat flux will release more ocean heat to the atmosphere, thereby lowering the air-skin $\Delta T$ and acting as a negative feedback. However, as summarized in the top half of Fig 5.4, there may be a second order latent heat flux feedback that results from the atmospheric boundary layer losing heat due to the nature of the release of the latent heat energy with height. The latent heat release profile for the TOGA COARE region was found to have a net cooling effect for the lowest 1 km of the atmosphere (Tao, Lang et al. 2000). This would imply a cooling near the surface, which would act as a positive feedback that causes the air-skin $\Delta T$ to be more negative. The magnitude of this second order feedback is not very large, using a basic calculation from Tao, Lang et al it is estimated to be less than 10 W/(m$^2$) (2000).

Still considering a ‘more negative’ initial disturbance to the air-skin $\Delta T$, within the latent heat flux and air-skin $\Delta T$ there is an additional second order feedback, also noted in Fig. 5.4. This feedback occurs as the latent heat is released into the atmosphere at height as the water vapor condenses. This release of latent heat may warm the atmosphere from 2 km to 10 km height up to 6 K per day (Tao, Lang et al. 2000). For downwards longwave emission, that means this feedback may reach strengths of up to 43 W/m$^2$. This exceeds the values listed in Table 6.2, because those values were calculated using conservative environmental conditions, and may double or triple in value under certain situations. This will be explained more completely in the next paragraph. In addition, prior to condensation this increased water vapor may cause the air to be at or above
saturation, and thus promote cloud formation. The clouds would act to both increase the amount of downward longwave radiation and to reduce the amount of solar radiation reaching the surface. The magnitude of this second order feedback is of importance to the balance of air-skin $\Delta T$ cumulative feedback loops, but obtaining this magnitude is beyond the scope of the work presented here. If the initial disturbance to the air-skin $\Delta T$ causes the value to become less negative through a decrease in the skin temperature, then the latent heat flux acts as a positive feedback, with the previously mentioned second order feedbacks becoming negative.

The magnitude of the latent heat flux varies widely depending on numerous oceanic and atmospheric conditions. Based on the bulk aerodynamic formulas, the humidity and wind speed will have the largest impacts on the latent heat flux. The field data from the Nauru’99 cruise shows a correlation between the air-skin temperature difference and the latent heat flux, as do the COARE model simulations. The ECMWF data show no correlation between the air-skin temperature difference and latent heat flux when binned by wind speed. The exact method used by ECMWF to calculate the latent heat flux is not clearly described. It has been previously shown that the latent heat flux coefficient has a dependence on the air-skin temperature difference (Kara, Hurlburt et al. 2004). The transfer coefficient is also quite dependent on the wind speed (see Fig. 5.5), meaning the latent heat flux has a primary and second-order dependence on the wind speed. For a wind speed of 1 m/s, the latent heat flux coefficient decreases dramatically as the air-skin $\Delta T$ approaches zero. In comparison, at winds of 6 m/s the transfer coefficient does not decrease as quickly. This effect is seen in the COARE model simulation of the latent heat
flux (Fig. 4.5). The latent heat flux values calculated in Table 5.2 were calculated for mean conditions observed during the Nauru’99 cruise. During peak conditions of wind speed and humidity, those values for the latent heat flux may double, or even triple.

The net feedback from the sensitive, latent, and longwave fluxes given a disturbance that makes the air-skin ΔT more negative will be a dampening of the initial perturbation. As shown in Fig. 5.4, the primary feedbacks for such a disturbance are all negative, with the second order feedbacks related to the latent heat flux being less dominant than the primary feedback. This is why, following a period of intense solar heating, the skin layer will return to its mean temperature within a few hours. A very basic calculation shows that for a boundary layer height of 600m and an initial disturbance that makes the air-skin ΔT more negative by 1 K, the flux feedbacks (at winds of 5 m/s) from the sensitive, longwave, and latent heat flux will restore the air-skin ΔT within 11 hours.

The net effect from the sensitive, latent, and longwave fluxes given a disturbance that makes the air-skin ΔT more positive may be negative or positive feedback. In the bulk formula calculations, the latent heat flux dominates the feedback loops beginning at winds of 4 m/s, assuming mean humidity levels from the Nauru’99 cruise. When winds are at or below 4 m/s, the sensitive and longwave negative feedbacks are greater than the latent heat flux positive feedback, and the air-skin ΔT is stabilized around ∼1 K. However, the field data from the Nauru’99 cruise show the latent heat flux becoming dominant around winds of 7 m/s. The second-order feedbacks from the latent heat flux are not considered in the bulk formula, which may account for the differences in the critical values of net positive and net negative feedbacks. The discrepancy between the
critical values may also be attributed to the time scale and spatial scales of the release of latent heat. How much of the latent heat is released into the lower atmosphere, and when, will certainly have a dramatic impact on the net sensitivity of the heat fluxes to the air-skin $\Delta T$. It should be noted that the conditions aboard the Nauru’99 cruise that lead to a net positive feedback loop make up a mere 5% of the total data. The remaining 95% of the data feature conditions that lead to net negative feedback loops, stabilizing the air-skin $\Delta T$ at its mean value of -1 K.

The exchange coefficients for the latent and sensible heat flux measurements have very little reason to change drastically due to geographic location. The primary change from one location to another may be attributed to different atmospheric stability, wind, humidity, or temperature levels, but these changes are accounted for within the flux formulas. The geographic changes that may affect the exchange coefficients are changes in gravity, which varies by up to 0.5% from the equator to the poles, and the thickness of the atmospheric boundary layer. An assumed atmospheric boundary layer thickness of 600m is reasonable in most areas (Kara, Hurlburt et al. 2004). This suggests why the air-skin $\Delta T$ distributions are unvarying with respect to season, longitude, and latitude.
Figure 5.5: The Latent heat flux coefficient, $C_L$, plotted against the air-skin temperature difference for wind speeds of 1 m/s (left) and 6 m/s (right) (Kara, Hurlburt et al. 2004)
Chapter 6

Conclusions and Future Work

Acquiring the Air-Sea $\Delta T$ from Infrared Interferometers gives more accurate data than traditional measurements. The real time calibration and carefully chosen spectral windows allow for an accuracy of better than 0.1 K. Through investigating the air-skin $\Delta T$ measurements from a wide array of oceanic and atmospheric conditions, a global $\sim$1 K mean of the value is observed, which is stabilized through a sensitive balance of ocean and atmospheric variables that work in a series of feedback loops. Empirical descriptions
of the radiometric $\Delta T$ measurement anomalies exist, and have been explained by reasonable oceanic and atmospheric interaction processes. Following the anomalies due to atmospheric fronts and solar radiation/low wind speeds, the air-skin $\Delta T$ is returned to its mean value of $-1$ K through feedback loops involving the heat fluxes between the ocean and lower atmosphere. The sensible and longwave fluxes act as negative feedbacks, stabilizing the air-skin $\Delta T$ around its mean value. The latent heat flux may act as a negative or positive feedback, with second-order feedbacks involving the release of latent heat throughout the atmosphere and conditions that encourage cloud formation. The net effect of the sensible, latent, and longwave fluxes is negative for the prevalent oceanic and atmospheric conditions, and these account for 95% of the data from the Nauru’99 cruise. At high enough wind speeds and humidity values, the feedback from the latent heat flux may become large enough to dominate the other fluxes, and this results in a net positive feedback, allowing the air-skin $\Delta T$ to deviate further from its mean value. Future work will involve more cruise data with joint efforts to obtain reliable SST and heat flux data. It would also be of interest to better characterize the heat, gas, and momentum flux sensitivities to determine additional feedback loops regarding the $-1$ K value and narrow distribution of the air-sea $\Delta T$ observed in the radiometric measurements.

The effects of rain are of interest when investigating the air-skin $\Delta T$ and potential feedback loops. Rain falling through the atmosphere exchanges heat both as it falls and as it contacts the surface. The heat flux, $E$, is given by:

$$E = RC_w \Delta T_{ro}$$  \hspace{1cm} (9)
where $R$ is the rain rate, $C_w$ is the heat capacity of water, and $\Delta T_{ro}$ is the temperature difference between the rain and ocean water. As a droplet of rain falls, its temperature is ordinarily close to the wet bulb temperature, and hence cooler than the atmospheric and sea surface temperatures in tropical regions. Upon contact with the surface, this implies a heat loss by the ocean (Gosnell, Fairall et al. 1995). Within the intertropical convergence zone, observations indicate that the sensible heat flux related to rain may account for 40% of the net surface heat flux while it is raining (Flament and Sawyer 1995). During rainfall, the sensible heat flux is a large component of the net surface heat flux, with contributions ranging from -65.0 W/m² to -204 W/m². This range equates to approximately 15%-60% of the net heat flux during a time of rainfall. Observations by Anderson and Hinton et al. (1998) show the rain heat flux is significant on longer time scales in the warm-pool region, with a pronounced balance between high precipitation rates and surface cooling. During a four month observation period, the rain heat flux was -2.8 W/m² (ocean cooling) amounting to 15% of the net surface heat flux (Anderson, Hinton et al. 1998). The M-AERI is unable to make useful measurements during periods of precipitation, thus quantifying the effect of rainfall within the air-skin $\Delta T$ feedback loops is beyond the scope of this research.

While the diurnal cycle certainly affects the air-skin $\Delta T$, it is still unknown if the SST variation due to diurnal heating affects atmospheric physics in a significant way. The diurnal variation in the skin temperature has been shown to affect precipitation and surface fluxes variation on an intraseasonal scale. One numerical experiment demonstrated that significant differences in cloudiness and humidity may be attributed to
the diurnal variation in SST; however the physical processes responsible for the observations are still unclear (Li, Yu et al. 2001). It has also been shown to modify the vertical profile of the air temperature, the vertical profile of humidity, and sea-breeze calculations (Kawai and Wada 2007). The diurnal cycle of SST is expected to have some impact on the atmosphere, and efforts to simulate the diurnal variation in the in the upper ocean in a global climate model have already begun. This requires fine vertical and temporal resolution, and the necessary parameterization for shortwave extinction and trapping depth to simulate the large increase in SST due to the diurnal cycle (Kawai and Wada 2007). Some parameterization schemes and that allow for the realistic representations of diurnal SST variation have been suggested, and investigations are currently underway to better characterize the physics of the atmosphere and diurnal SST variation (e.g. Schiller and Godfrey 2005).
Bibliography


