The Response of the Ocean Thermal Skin Layer to Air-Sea Surface Heat Fluxes.

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UNIVERSITY OF MIAMI

THE RESPONSE OF THE OCEAN THERMAL SKIN LAYER TO AIR-SEA SURFACE HEAT FLUXES

By
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A DISSERTATION

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THE RESPONSE OF THE OCEAN THERMAL SKIN LAYER TO AIR-SEA SURFACE HEAT FLUXES

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There is much evidence that the ocean is heating as a result of an increase in concentrations of greenhouse gases (GHGs) in the atmosphere from human activities. GHGs absorb infrared radiation and re-emit infrared radiation back to the ocean’s surface which is subsequently absorbed. However, the incoming infrared radiation is absorbed within the top micrometers of the ocean’s surface which is where the thermal skin layer exists. Thus the incident infrared radiation does not directly heat the upper few meters of the ocean. We are therefore motivated to investigate the physical mechanism between the absorption of infrared radiation and its effect on heat transfer at the air-sea boundary. The hypothesis is that since heat lost through the air-sea interface is controlled by the thermal skin layer, which is directly influenced by the absorption and emission of infrared radiation, the heat flow through the thermal skin layer adjusts to maintain the surface heat loss, assuming the surface heat loss does not vary, and thus modulates the upper ocean heat content.

This hypothesis is investigated through utilizing clouds to represent an increase in incoming longwave radiation and analyzing retrieved thermal skin layer vertical temperature profiles from a shipboard infrared spectrometer from two research cruises. The data are limited to night-time, no precipitation and low winds of less than 2 m/s to remove
effects of solar radiation, wind-driven shear and possibilities of thermal skin layer disruption. The results show independence of the turbulent fluxes and emitted radiation on the incident radiative fluxes which rules out the immediate release of heat from the absorption of the cloud infrared irradiance back into the atmosphere through processes such as evaporation and increase infrared emission. Furthermore, independence was confirmed between the incoming and outgoing radiative flux which implies the heat sink for upward flowing heat at the air-sea interface is more-or-less fixed. The surplus energy, from absorbing increasing levels of infrared radiation, is found to adjust the curvature of the thermal skin layer such that there is a smaller gradient at the interface between the thermal skin layer and the mixed layer beneath. The vertical conduction of heat from the mixed layer to the surface is therefore hindered while the additional energy within the thermal skin layer is supporting the gradient changes of the skin layer’s temperature profile. This results in heat beneath the thermal skin layer, which is a product of the absorption of solar radiation during the day, to be retained and cause an increase in upper ocean heat content.

The accuracy of four published skin layer models were evaluated by comparison with the field results. The results show a need to include radiative effects, which are currently absent, in such models as they do not replicate the findings from the field data and do not elucidate the effects of the absorption of infrared radiation.
To my husband, Yousi, and daughter, Zoey - your endless love and unselfish support has been instrumental in my inspirations and motivations.

To my parents - thank you for your love, encouragement and guidance. All that I am or hope to be, I owe to the both you.
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Chapter 1

Introduction

The ocean’s role in the climate system is important, covering about 70 % of the Earth’s surface and containing 97 % of all the water on Earth. The large specific heat of water and storage capacity of carbon dioxide (as opposed to the atmosphere) means that the ocean acts as a heat and carbon sink, modulating the surrounding air temperatures and atmospheric carbon dioxide (Bigg et al., 2003). The timescales for this vast amount of energy stored in the ocean can be of order of seconds to centuries depending on the depth and its interaction with the atmosphere. The ocean’s energy influx to the atmosphere fuels weather patterns and the climate system. Thus, the ocean’s forcing of the atmosphere plays a pivotal role in the Earth’s energy budget, and understanding ocean-atmosphere coupling is vital in studies of the ocean, atmosphere, and the climate system.

Due to the fact that the ocean is a large heat sink, the observed increase in upper ocean heat content (OHC) is thus a concern as this would lead to an upset of the natural energy budget of the Earth’s climate system. Ocean warming has been shown to account for over 90% of the increase in energy accumulated in the climate system between years 1971 and 2010 (IPCC, 2014). Evidence of a consistent increase in upper OHC (Levitus et al., 2012; IPCC, 2014) over the past decade is shown in fig. 1.1
This trend is constructed from data obtained from the World Ocean Database 2009 (Boyer et al., 2009), Argo profiling floats and bathythermograph data from year 1955 - 2010. Two-thirds of the increase in OHC occurs in the upper 700 m from the ocean surface at a rate of 0.27 W/m$^2$ over the World Ocean and a volume mean warming of 0.18 °C while the remaining one-third is accounted for in the upper 700-2000 m. The total calculated volume mean warming is 0.09 °C at a rate of 0.39 W/m$^2$ over the World Ocean.

![Figure 1.1: Time series for the World Ocean of ocean heat content (1022 J) for the 0-2000 m (red) and 700-2000 m (black) layers based on running pentadal (five-year) analyses. Reference period is 1955-2006. Each pentadal estimate is plotted at the midpoint of the 5-year period. The vertical bars represent +/- 2*standard estimate about the pentadal estimate for the 0-2000 m estimates and the grey-shaded area represent +/- 2*standard estimate about the pentadal estimate for the 0-700 m estimates. The blue bar chart at the bottom represents the percentage of one-degree squares (globally) that have at least four pentadal one-degree square anomaly values used in their computation at 700 m depth. Blue line is the same as for the bar chart but for 2000 m depth (Levitus et al., 2012).](image-url)

It is understood from the greenhouse effect that the increase in atmospheric greenhouse gases (GHG’s) would result in an increase in the absorption and re-emission of lower frequency infrared (IR) thermal radiation which subsequently warms the Earth’s
The latest Intergovernmental Panel on Climate Change (IPCC) report has stated that since the pre-industrial era, there has been an increase in atmospheric concentrations of anthropogenic greenhouse gases (GHG), namely carbon dioxide (CO$_2$), methane (CH$_4$) and nitrous oxide (N$_2$O) (IPCC, 2014). This is shown in fig. 1.2, extracted from IPCC (2014), and indicates an increase in GHG emissions at a rate of 1.3 %/year from year 1970-2000 and at a rate of 2.2 %/year from year 2000-2010. The effect of this increase of GHG’s on the climate system has a 95-100% probability of causing the currently observed unprecedented warming of the climate since the mid-20th century. Emissions of CO$_2$ from the burning of fossil fuels and industrial processes are most disconcerting, contributing to about 78 % of the total anthropogenic GHG increase from 1970 to 2010.

![Figure 1.2: Total annual anthropogenic greenhouse gas emissions (gigatonne of CO$_2$-equivalent per year, GtCO$_2$-eq/yr) for the period 1970 to 2010 by gases: CO$_2$ from fossil fuel combustion and industrial processes; CO$_2$ from Forestry and Other Land Use (FOLU); methane (CH$_4$); nitrous oxide (N$_2$O); fluorinated gases covered under the Kyoto Protocol (F-gases). Extracted from IPCC (2014).](image)

However, it is not clear how the greenhouse effect directly affects the ocean’s heat uptake in the upper 700 m of the ocean. This is because the penetration depth of IR
radiation in water is within sub-millimeters scales (fig. 1.3). Thus implying that the increase in IR radiation due to increases in GHG’s would only be absorbed within the top millimeter of the ocean surface and does not directly influence the temperatures beyond these depths.

Figure 1.3: Plot of penetration depth versus wavenumber. Red dotted line indicates $v = 500 \, \text{cm}^{-1}$ and $3000 \, \text{cm}^{-1}$. Penetration depth obtained from $I(z) = I_0 e^{-\alpha z}$ (eq. 1a) with $R_{img}$ obtained from Bertie and Lan (1996).

Therefore, to understand the mechanism contributing to this increase in OHC, it is crucial to obtain in-depth knowledge of the thermodynamics involved in air-sea interaction studies. The majority of the interfacial heat fluxes flows from the ocean to the atmosphere (surface cooling) thus the sea surface temperature (SST) plays a very important role in controlling the amount of heat released from the ocean into the atmosphere as it acts as the surface boundary condition for the atmosphere. SST’s in this dissertation are more accurately known as the skin SST because it is essentially the temperature at the air-sea boundary which holds the largest weighting in determining the flux rate at this boundary. Fairall et al. (1996) mentioned that skin SST values must be accurate to $\pm 0.2 \, \text{K}$ to obtained accuracy values of $10 \, \text{W/m}^2$ in heat balance estimations. Furthermore, many established climate models requires consistent and
accurate skin SST values for climate change projections. For example, Ohring et al. (2005) and ISSTST (2010) have stated the need for an absolute accuracy of $< 0.1$ K and stability of $< 0.04$ K/decade for satellite derived skin SST values for applications in climate studies and monitoring. Satellite skin SST measurements provide continuous global data but lack the required accuracy of 0.1 K, with an accuracy of $\sim 0.23$ K for the Advanced Along-Track Scanning Radiometer (AATSR) (Embury et al., 2012), while in-situ measurements lack the temporal and spatial sampling and continuity, as well as rarely achieving the required accuracy. There is thus a need to improve the accuracy of satellite skin SST measurements. One major source of error is the variability of the cool-skin effect (illustrated schematically in fig. 1.4) which occurs because of the heat flow from the ocean to the atmosphere (e.g. McAlister and McLeish (1969); Harris et al. (1995); Donlon et al. (2002); Minnett et al. (2011)). Correcting this error requires a thorough understanding of the physics of the sub-surface layer which involves the thermal skin layer, the electromagnetic skin layer, and the viscous sublayer. The differences in these layers will be further described in the subsequent sub-sections.

1.1 The thermal skin layer (TSL).

Of the three skin layers mentioned above, we are particularly interested in the thermal skin layer as this layer exists directly on the aqueous side of the air-sea interface, extending to only about a tenth of a millimeter from the top of the ocean’s surface (Katsaros et al., 1977). Thus, this layer serves as a bottleneck for the transfer of heat at the air-sea interface. The direction of flow of heat is mostly from the ocean to the atmosphere thus implying that the surface temperature is cooler than the temperature just below the underlying surface waters. This temperature difference ranges from $\sim 0.1$ K during high wind speeds of $> 7$ m/s to $\sim 0.6$ K during low wind conditions of
<2.5 m/s (Donlon et al., 2002). Fig. 1.4, extracted from Gentemann and Minnett (2008), shows a cartoon of the TSL during night-time or high wind daytime conditions (left plot) and low wind daytime conditions (right plot). The TSL rides on top of the well-mixed layer and stratified layer depending on the conditions. As such, the TSL is almost always present, with occurrences of momentary disruption caused by wave breaking. Its creation and destruction are also influenced by rainfall, however, the TSL disrupted by such processes is able to restore itself very quickly in a matter of seconds (Jessup et al., 1997).

![Diagram of TSL](image)

Figure 1.4: A: Night-time or daytime with high wind conditions. B: Daytime with low wind conditions. (Gentemann and Minnett, 2008)

It is thus important that the TSL be considered during the analysis of any processes involving sea surface temperatures and interfacial heat fluxes. This is because the energy flow down the TSL’s temperature gradient sustains the latent, sensible and net infrared (IR) radiative energy losses by means of molecular exchanges (McAlister and McLeish, 1969; Jessup et al., 1997; Minnett et al., 2011). Therefore, a strong temperature gradient is established due to the poor efficiency of heat transfer by molecular conduction as illustrated in fig. 1.4.
The mean TSL’s properties are not influenced by processes such as turbulence or convection as the TSL is embedded in the viscous sublayer which will be described further in Section 1.3 (Veron et al., 2011): turbulence occurs within the ocean’s mixed layer which lies beneath the TSL, while convection is driven by heat flowing from the ocean surface and occurs within the atmospheric and oceanic boundary layers. The instantaneous skin layer thickness and temperature gradients are also modulated by sea surface renewal effects, which describes the continuous replacement of surface water parcels with water parcels from the bulk of the water due to turbulent eddies (Soloviev and Schlüssel, 1994; Veron et al., 2011). However, this is not a concern as we consider mean conditions, with spatial and time averages over scales longer than renewal events, as for our TSL study we will be using measurements from a ship-board passive remote sensing radiometer which has a footprint of approximately 1 m$^2$, and integration periods of over one minute (Wong and Minnett, 2016b). The TSL’s presented in this dissertation are therefore an average situation where the effects of surface renewals are not resolved.

The largest difficulty in understanding the TSL is in obtaining accurate in-situ measurements for analysis of its temperature gradient. This is because the TSL’s very thin features require highly accurate instrumentation capable of very fine-scale measurements. To date, there are no known in-situ temperature profilers with the ability to provide measurements in the required high resolution of sub-millimeter scales. The highest resolution profiler known is the Skin Depth Experimental Profiler (SkinDeEP) (Ward et al., 2004) has a resolution of 0.5 mm. Other in-situ attempts to measure the TSL temperature profile using mechanically invasive probes which physically disrupt the skin SST, and pose a high-risk in altering the TSL gradient (e.g. Katsaros et al., 1977; Ward et al., 2004). Thus, passive remote sensing techniques are ideal since these sources of errors would be eliminated. However, satellite-derived SST retrievals are unsuitable because of the large uncertainties involved and lack the
accuracy required in our analysis. For example, as mentioned above, the AATSR (Embury et al., 2012) has an accuracy of \( \sim 0.23 \) K (with continued improvements as noted in the Ph.D. thesis of Embury (Embury, 2014) and is not suitable for our problem as the temperature difference within the TSL ranges from 0.1 K to 0.6 K (Donlon et al., 2002). Thus, it is better to use a ship-board passive remote sensing radiometer which provides the accuracy and precision required for studies of the TSL. This instrument will be introduced in Section 2.1.1 and a retrieval technique which exploits properties of the electromagnetic (EM) skin layer to obtain the TSL temperature profile through the instrument’s spectral measurements would be presented in Section 3.2.

### 1.2 The electromagnetic (EM) skin layer.

The EM skin layer exists due to EM properties of water which control the absorption and emission of IR radiation and therefore control heat losses and gains by IR radiation at the air-sea boundary. Fig. 1.3 shows the penetration depth within the IR band of \( \sim 0.01 \) cm\(^{-1} \) - 4000 cm\(^{-1} \) (\( \sim 2.5 \) µm - 10000 µm wavelengths) with 2 red dotted vertical lines indicating the range 500 cm\(^{-1} \) - 3000 cm\(^{-1} \) (\( \sim 3.33 \) µm - 20 µm) which is the wavenumber range of the IR spectrometer used in our studies (to be described in Section 2.1.1). The penetration depth or propagation pathlength of emitted thermal radiation is deduced from Beer-Lambert’s law:

\[
I(z) = I_0 e^{-\alpha z}
\]  

(1a)

where \( I(z) \) is the radiant intensity at depth \( z \), \( I_0 \) is the radiant intensity at \( z = 0 \) and \( \alpha \) is the absorption coefficient determined from the imaginary component of the refractive index of water, \( R_{img} \) \( (\alpha = 4\pi v R_{img}, \text{where } v = \text{wavenumber in cm}^{-1} ) \). The penetration depth, \( D_p \), is defined as \( 1/\alpha \), which is indicative of the depth at which the incident radiant intensity falls to 1/e of its original value.
The penetration depth may also be regarded as the emission depth assuming local thermodynamic equilibrium. Thus from fig. 1.3, it is noted that the majority of the emissions originate from depths of < 0.1 mm, within the TSL. This property is the basis of IR spectrometers on ships, aircrafts and satellites as the radiation measured by these instruments originate from this layer thereby allowing us to obtain skin temperature readings.

However, because the ship-board spectroradiometer used in our studies has a 55° sea surface view angle (Section 2.1.1), corrections on \( D_p \) must be applied by taking into account the angle of refraction at the interface, \( \theta_r \), such that \( D_p = \frac{1}{\alpha} \cos \theta_r \). \( \theta_r \) is calculated using Snell’s law which takes into account the real component of the refractive index of water assuming that the view angle of the spectrometer is at 55° (Hanafin, 2002).

Ignoring the complicating factors of the disruption of the TSL by rain drops or the absorptive effects of solar radiation in the TSL, the exponential rate of emission and absorption of IR radiation is defined by eq. 1a and therefore leads to a highly non-linear TSL temperature profile as opposed to a linear temperature profile which has been assumed by many prior studies of the TSL (e.g. Saunders, 1967; Fairall et al., 1996). The curvature of the TSL temperature profile is expected to be smaller closer to the air-sea interface as the heat loss by IR emission increases (McAlister and McLeish, 1969).

### 1.3 The viscous sublayer.

The viscous sublayer is known to be a very thin region adjacent to the air-sea interface in which turbulent velocity fluctuations are suppressed by viscosity. The flow within this layer is therefore intermittently laminar given the presence of sea surface renewal events, however, as we are considering a mean viscous sublayer, turbulence due to sea
surface renewal events will be averaged out and we only need to account for the viscous shear stresses. The viscous sublayer is typically of orders of millimeters (Veron et al., 2011), and is not to be confused with the TSL which is thinner, whose thickness is of orders of sub-millimeters scales. The TSL and EM skin layer are therefore embedded in the viscous sublayer (Veron et al., 2011). The largest difference between the viscous sublayer and TSL is that the TSL includes both viscous and radiative effects thus resulting in a non-linear temperature profile whereas the viscous sublayer only takes into account viscous processes therefore a linear temperature profile is expected.

Studies of the viscous sublayer at the air-sea interface are largely geared towards obtaining the temperature difference or depth of the viscous sublayer by using well known theories which govern molecular heat flux transfer rates and assuming a fully viscous flow close to the boundary layer. One of the earliest studies was by Saunders (1967) who presented a purely theoretical result using the heat flux equation and through the use of dimensional argument, deduced that the thickness of the viscous sublayer is only determined by viscous stresses, kinematic viscosity and water density. This theory has since been improved with technological advances which have allowed for finer and more accurate temperature measurements to validate his model against in-situ measurements (Fairall et al., 1996; Kent et al., 1996; Zhang and Zhang, 2012). The improved models build on Saunders’ model by taking into account different viscous sublayer thickness scales (e.g. Kolmogorov and Batchelor microscales) and comparison with in-situ measurements allows for better approximations of the non-dimensional numbers in the Saunders model by considering buoyancy and shear forces, and diurnal variability in attempts to obtain suitable parameterized equations for the depth or temperature difference within the viscous thermal sublayer.

Another set of theories developed of the viscous sublayer takes into account sea surface renewal effects (e.g. Katsaros et al., 1977; Liu et al., 1979; Soloviev and Schlüssel, 1994; Veron et al., 2011). Surface renewal theory looks into describing the
renewal rate of a surface water parcel, due to turbulent eddies, through the use of a probability density function to obtain the depth and temperature difference within the viscous sublayer. The Kolmogorov time scale is typically used as it gives the smallest time scales in turbulent flow (Brutsaert, 1975). There has also been continued improvements in estimating the renewal rate by including processes such as shear stresses, free and forced convection and surface wave breaking (Soloviev and Schlüssel, 1994).

To reiterate, radiative effects are not accounted for in these models, therefore the gradient within the viscous thermal sublayer is constant (linear temperature profile) while in reality the TSL’s temperature profile is non-linear. Studies of thermal variations within the viscous boundary have been much more extensive as compared to the TSL, thus it will be ideal to provide an analysis of some viscous layer models with the TSL retrievals.

1.4 Objectives and motivation.

The objective of this research is to understand and provide an explanation of how an increase in levels of anthropogenic greenhouse gases in the atmosphere, which raises the amounts of incident longwave radiation on the ocean surface, causes the upper OHC to increase. This conundrum arises because it is known that the incident longwave radiation only penetrates the top submillimeter scales from the ocean surface and does not directly heat the layers beyond. Furthermore, at submillimeter scales, the mechanism for the transport of heat is through molecular conduction and not by turbulence or convection. Given the sign of the vertical temperature gradient in the TSL, the heat from the absorption of longwave radiation will be conducted to the sea surface. This raises even more questions about the cause of the observed increase in upper OHC as it suggests that all heat due to the absorption of increased longwave
radiation should be concentrated in the upper submillimeter from the interface. We hypothesize that variations in the temperature gradient in the thermal skin layer (TSL), which exists within the top submillimeter of the ocean surface and is directly affected by the absorption and emission of longwave radiation, modulates the amount of heat flow within the air-sea interface. Thus, any changes in the gradient of the TSL due to variations in the absorbed longwave radiation will provide an explanation on the mechanism as to how the OHC increases through the retention of heat from the absorption of solar radiation within the bulk of the ocean.

To address this hypothesis, we have utilized spectral measurements from a shipboard passive remote sensing infrared radiometer to obtain fine-scale measurements of the TSL vertical temperature profile. The drawback of using a shipboard instrumentation in this study is the lack of consistent long term measurements with high spatial scales necessary for the detection of anthropogenic effects on the levels of GHG’s. We have thus exploited signals obtained from cloud longwave radiative forcing as a surrogate for the effects of GHG’s on the ocean’s surface.

Cloud radiative forcing is used as a replacement for the effect of GHG’s because they occur quickly compared to the effects of increasing levels of GHG’s. Furthermore, using a doubling of CO$_2$ concentrations, a typical forcing in climate models, only gives a radiative signal increase of 4 W/m$^2$ (Ramanathan et al., 1979) which is small compared to a cloud radiative forcing which can provide a signal of $\sim$200 W/m$^2$. Thus the shorter time scales and higher radiative signal provided by cloud forcing makes the problem tractable. However, it is important to establish the differences in radiative signatures between cloud and GHG forcing, including the measured longwave radiation provided by broadband infrared radiometers as these differences will directly affect the variations in the TSL.
This dissertation therefore aims to answer the following questions to address the main objectives:

1. Develop a technique to retrieve the TSL vertical temperature profile at a desired resolution and accuracy such that analysis may be made with changes in cloud-driven radiative forcing.

2. Assess how well does cloud cover act as a surrogate for the effects of increasing levels of GHG’s and are incoming longwave measurements obtained from a broadband infrared radiometer data representative of the increase in GHG’s.

3. Investigate the relationship between the air-sea interfacial fluxes with retrieved properties of the TSL.

4. Assess the accuracy of currently published viscous layer models in representing our results obtained from the research cruises.

The dissertation is organized as follows, Chapter 2 describes the instrumentation used to obtain the incoming longwave radiation and IR spectrum for the retrieval of the TSL vertical profile. The research cruises and quality controls used for the field data are also discussed in this chapter. Chapter 3 presents the technique used to retrieve the TSL profile and justifies the use of cloud cover as a surrogate for GHG effects. This chapter therefore addresses points 1 and 2. Included in Chapter 3 are the processing techniques used to derive the interfacial turbulent fluxes and sea surface emissivity values which are required in the retrieval of the TSL. Chapter 4 addresses point 3 and shows the correlations between the interfacial heat fluxes with the retrieved temperature differences and thickness of the TSL obtained from two research cruises. A mechanism of how heat within the bulk of the ocean is retained based on the correlations observed with the radiative fluxes and TSL will be described. Point 4 is discussed in Chapter 5 which presents an investigation of four published TSL
models. This chapter establishes any similarities or issues between the four models and compares the results with field data results in Chapter 4. Finally, Chapter 6 concludes this dissertation, including recommendations for future research.
Chapter 2

Instrumentation and field data

This chapter introduces the instruments used in data collection in two field campaigns, the Nauru 1999 (NAURU99) and African Monsoon Multidisciplinary Analysis 2006 (AMMA06) campaigns. Both are described along with an outline of the quality controls applied to the data presented in Section 2.3.

2.1 Instrumentation.

In this section, we introduce two main instruments used in data collection for our analysis. The first is the Marine-Atmospheric Emitted Radiance Interferometer (M-AERI) which is used to obtain the sea surface emission spectra and the second is the Eppley Precision Infrared Radiometer (PIR) which measures the incoming longwave radiation.

2.1.1 Marine-Atmospheric Emitted Radiance Interferometer.

The description of the Marine-Atmospheric Emitted Radiance Interferometer (M-AERI) is extracted from Wong and Minnett (2016b). The M-AERI is a sea-going, well-calibrated, Fourier Transform Infrared (FTIR) Interferometer (Minnett et al.,
and was developed from the Atmospheric Emitted Radiance Interferometer (AERI) (Knuteson et al., 2004) at the Space Science and Engineering Center, University of Wisconsin-Madison for the Department of Energy’s Atmospheric Radiation Measurement Program (Stokes and Schwartz, 1994). The M-AERI measures radiances (units: mW m$^{-2}$ sr$^{-1}$ (cm$^{-1}$)$^{-1}$) emitted in the wavenumber range 500-3000 cm$^{-1}$ ($\sim$3-20 µm in wavelength) using two infrared (IR) detectors to attain this wide spectral range. It has an effective spectral resolution of 0.5 cm$^{-1}$ which resolves many gaseous absorption and emission lines in the atmosphere. Minnett et al. (2001) describe the details of its operation, accuracy and applications and a brief summary will be given in this section.

The M-AERI is mounted at the bow of the ship where its view of the sea surface is ahead of the bow wave of the ship. The M-AERI uses a scan mirror that cycles through a sequence of scene views consisting firstly of the upwelling radiance from the sea surface and the downwelling atmospheric radiances measured at complementary angles with typical at-sea deployment measured incidence angles, $\theta_i$, of 55°, as shown in fig. 2.1, followed by a zenith measurement of the downwelling radiation. This set of scene views is sandwiched between two calibration sequences consisting of measurements of emission from two blackbody (BB) cavities mounted with their axes at 60° and 120° to the vertical, with the upper BB cavity maintained at a temperature of 60°C and the lower BB cavity floating at ambient air temperature. Radiance measurements are averaged over 45-s intervals consisting of 45 independent interferograms and the accuracy of the derived brightness temperature (BT) are determined to be < 0.02 K at 20°C and < 0.04 K at 30°C with a signal-to-noise (SNR) ratio range of 135-3135 at the short wavelengths and 1135-5400 at the long wavelengths. Typically skin SST measurements are derived at a wavelength of 7.7 µm which corresponds to radiance values of $\sim$55 mW m$^{-2}$ sr$^{-1}$ (cm$^{-1}$)$^{-1}$ and a SNR of $\sim$1800 for a 45 s average which is achieved by cooling the two detectors to $\sim$77 K (boiling point of liquid nitrogen) by
a Stirling Cycle cooler. The M-AERI is covered by a tarpaulin during heavy rain or sea-spray to keep the scan mirror dry as emission from water droplets on the mirror would contaminate the measurements. A full measurement sequence of the scene mirror takes about 10 mins.

Figure 2.1: Schematic of viewing geometry of the M-AERI obtained from Hanafin (2002).

The schematic of the M-AERI’s viewing geometry is shown in fig. 2.1 and eq. 2a, described by Hanafin (2002) and Minnett et al. (2001), shows the observed upwelling radiance, $R_{sea}(v, \theta)$, measured by the M-AERI while viewing the sea surface, and consists of the spectral radiation emitted at the sea surface, $B(v, SST_{skin})$, the downwelling atmospheric emission, $R_{sky}(v, \theta)$, reflected at the sea surface and the component of atmospheric emission from the layer below the level of the instrument at height $h$, $R_h(v, \theta)$, which includes both direct and reflected emission from the sea surface attenuated by the atmosphere between the surface and height $h$. $v$ is wavenumber and $\theta$ is the emission angle referenced to zenith. $\epsilon(v, \theta)$ is the sea surface emissivity.

$$R_{sea}(v) = \epsilon(v, \theta)B(v, SST_{skin}) + (1 - \epsilon(v, \theta))R_{sky}(v, \theta) + R_h(v, \theta) \quad (2a)$$

Since the M-AERI is generally mounted on a ship at a height of a few meters, $R_h(v, \theta)$ is very small and will be neglected in the analysis used here. Similarly, the
atmospheric attenuation along the path-lengths from the sea surface to the M-AERI can be neglected. Rewriting eq. 2a without \( R_h(v, \theta) \) and solving for the skin sea surface temperature, \( SST_{\text{skin}} \) gives:

\[
SST_{\text{skin}} = B^{-1}\{R_{\text{sea}}(v, \theta) - (1 - \epsilon(v, \theta))R_{\text{sky}}(v, \theta)/\epsilon(v, \theta)\}
\] (2b)

where \( B^{-1} \) is the inverse Planck function. Eq. 2b is used to perform the atmospheric correction and to retrieve the skin SST value from M-AERI measurements of spectral radiances, \( R_{\text{sea}} \) and \( R_{\text{sky}} \).

The top panel of fig. 2.2 shows a sample radiance spectrum from the M-AERI taken during night-time and cloud-free conditions across the full wavenumber range of 500-3000 cm\(^{-1}\). The bottom panel shows the radiances converted into BT using Planck’s function. The blue lines indicate the upwelling radiance from the sea surface and exhibit a nearly smooth curve similar to a Planck distribution with the exception of random spikes observed at frequencies where the signal is masked by thermal noise (e.g. 1500-1700 cm\(^{-1}\)) due to the atmosphere being insufficiently transmissive for the internal BB calibration to function well. It is the small deviations from Planck’s function at constant temperature that results from the spectral dependence of the emission depth that conveys the information on the temperature gradient in the thermal skin layer that we seek. From the bottom panel, the ‘drop’ in the atmospheric BT values at frequencies of 700-1300 cm\(^{-1}\) and 2000-2900 cm\(^{-1}\) indicates the presence of atmospheric windows, where the atmosphere is sufficiently transmissive to allow the M-AERI to measure radiances emitted from higher in the atmosphere. Away from these atmospheric spectral transmission windows, the atmosphere is less transmissive and the measured temperature from the M-AERI is warmer because it is sensing radiation emitted from gases lower in the atmosphere.
Figure 2.2: Spectra from M-AERI. Top panel shows the measured radiances while the bottom panel shows the calculated equivalent brightness temperatures (BT) from Planck’s function. The blue line indicates the sea surface spectra, the red line indicates the atmospheric spectra at a -55° from zenith, while the black line indicates the atmospheric spectra at zenith. Evidence of atmospheric window regions are shown by the colder sky BT values. The warmer spikes in these regions correspond to absorption and emission of atmospheric gases. Spike features in other wavelength regions (e.g. 1400 - 1700 cm⁻¹) indicate a short atmospheric pathlength which means the calibration measurements are not clean and the data should not be used.

2.1.2 Eppley Precision Infrared Radiometer (PIR).

To measure the incoming longwave radiation, LWᵢᵣ, a gimbaled Eppley Precision Infrared Radiometer or PIR is installed on the top of the main mast for the field programs along with the Portable Radiation Package (PRP) which includes a secondary PIR installed on the top foremast of the ships.

The Eppley PIR is developed by the Eppley Laboratory, Inc. and receives IR radiation in a range of approximately 3.5 to 50 μm (200 - 2857 cm⁻¹ in wavenumber), capturing the majority of the IR spectral band (~3.5 - 100 μm). The Eppley PIR senses the IR radiation and creates an output voltage which is divided by the Instrument’s
Sensitivity (approximately $4 \mu V/Wm^{-2}$) to obtain a flux value. This computed flux is further corrected for the PIR case’s outgoing radiation by subtracting the flux generated by the case temperature, $T_{\text{case}}$, using Stefan Boltzmann’s law (case flux = $5.6704 \cdot 10^{-8} \cdot T_{\text{case}}^4$). The Eppley PIR is also calibrated in the Eppley Laboratory using a reference blackbody which consists of a water-bath controlled, hemispherical, low temperature (0 - 50$^\circ$C) source. The calibration results are obtained from two distinct temperature values and are supplemented by comparing with witness standards. The instrument samples at 1 Hz.

![Eppley Precision Infrared Radiometer](www.eppleylab.com/instrumentation/precision_infrared_radiometer.htm)

### 2.2 Field programs.

This section describes the two field programs considered in our data analyses. The M-AERI was installed on both cruises and both field programs are chosen because of the relatively low wind speeds ($< 10$ m/s) and mostly clear skies with little or no precipitation. The low wind speeds and precipitation ensures plentiful readings in which there will be no disruption of the thermal skin layer by breaking waves. Sufficient clear sky readings are important as we rely on these data to determine the spectral emissivity which utilizes the atmospheric absorption features reflected at the sea surface (which will be masked by low and mid-level clouds) for its derivation (Section 3.2). Installation photographs of the M-AERI on both cruises are given in fig 2.4 while the cruise tracks are plotted in fig. 2.5.
2.2.1 Nauru 1999 field program.

The Nauru 1999 (NAURU99) was a joint program between the Department of Energy’s Atmospheric Radiation Measurement program, the National Oceanic and Atmospheric Administration (NOAA) and the Japan Marine Science and Technology center. It was conducted in the tropical western Pacific Ocean in June and July 1999. The RV Mirai left Yokohama, Japan on June 5 and steamed south to Nauru, with a 24 hour stop in Chuuk, Federated States of Micronesia on June 13. The Intensive Observation Period (IOP) took place near the island of Nauru from June 17 to July 4, 1999, and had a regular 3-hour schedule, steaming slowly into the wind for one hour for eddy correlation flux measurements then steaming back to the original position and staying
on station for an hour. The eddy correlation flux measurements were compared with turbulent flux measurements derived from the COAREv3.0 algorithm as obtained from a report from Fairall (1999). The IOP was characterized by fair weather with light to moderate winds with air temperatures ranging from 299 K - 302 K and SSTs from 301 K - 303 K. The wind speed ranged from 0 m/s - 9 m/s with a relative humidity of 65 % - 82 % and air pressure range of 1008 mbar - 1014 mbar.

2.2.2 African Monsoon Multidisciplinary Analysis (AMMA) 2006 field program.

Measurements taken by the M-AERI during the AMMA 2006 cruise (AMMA06) of the NOAA ship Ronald H. Brown (RHB) took place in the tropical Atlantic Ocean from May 28 to July 14, 2006. AMMA06 was a coordinated international project conducted to improve knowledge and understanding of the West African Monsoon, its
variability and impacts. There were 2 legs: Leg 1 was from May 28 to June 17 where the RHB left San Juan, Puerto Rico, heading south-east with the majority of the leg spent off West Africa. Leg 2 was from June 22 to July 14, leaving Recife, Brazil and returning to Charleston, South Carolina. Measurements from the M-AERI began on May 28 and ended on July 14, 2006. A gap from June 17 - June 22 corresponds to a port call in Recife, Brazil while other larger time gaps found in M-AERI data were due to instrument failure or periods of bad weather. The skin SST temperature, measured at a wavelength of $\sim 7.7 \mu m$ ranged from 295 K to 305 K.

The ranges of meteorological variables were: air temperatures 295 K to 302 K, air pressure readings 1010 hPa to 1030 hPa, relative humidity 50\% to 80\% with wind speeds 0 m/s to 20 m/s.

### 2.3 Quality controls.

For our analysis, only night-time data (19:30 - 07:30 h local time for NAURU99 and 20:00 - 07:00 h local time for AMMA06) with low winds ($< 10$ m/s) were used to avoid possible contamination of the measurements by reflected and scattered solar radiation and to ensure the thermal skin layer was rarely disrupted. The first part of this section shows the minimal effect of the absorptivity of solar radiation on the thermal skin layer. This is to show that the absorbed solar radiation effects on the TSL during the day are negligible as compared to the absorption of IR from the atmosphere. The second part discusses the choice of a wind speed limit of 10 m/s chosen for our analyses.
2.3.1 Influence of solar radiation on the thermal skin layer.

The absorptivity of solar radiation in the upper 10 mm of the ocean is analyzed to show that the influence of solar radiation is minimal within the thermal skin layer. Eq. 2c shows the attenuation of solar radiation flux at depth, \( SR(z) = -SW(z)/(c_p \rho) \), given the solar radiative flux just beneath the ocean surface, \( SR(0) \).

\[
SR(z) = SR(0) \sum_{i=1}^{9} a_i \exp(-\beta_i z)
\]

(2c)

\( a_i \) are the weights corresponding to the spectrally distributed absorption coefficients, \( \beta_i \). The values of the coefficients are listed in table 2.1. For \( \beta_i \) at 0.2 - 0.6 \( \mu m \), we have listed the coefficient for Type I waters (clearest water type) as it is noted that the absorption within this wavelength is influenced by the turbidity of the water (Jerlov, 1968; Soloviev and Schlüssel, 1996). The importance of the turbidity of water is not as significant at other wavelengths due to a strong increase in \( \beta_i \) with wavelength.

<table>
<thead>
<tr>
<th>wavelength (( \mu m ))</th>
<th>i</th>
<th>( a_i )</th>
<th>( \beta_i ) (m(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.2-0.6</td>
<td>1</td>
<td>0.2370</td>
<td>0.066 (Type I)</td>
</tr>
<tr>
<td>0.6-0.9</td>
<td>2</td>
<td>0.3600</td>
<td>0.4405</td>
</tr>
<tr>
<td>0.9-1.2</td>
<td>3</td>
<td>0.1790</td>
<td>31.75</td>
</tr>
<tr>
<td>1.2-1.5</td>
<td>4</td>
<td>0.0870</td>
<td>182.5</td>
</tr>
<tr>
<td>1.5-1.8</td>
<td>5</td>
<td>0.0800</td>
<td>1201</td>
</tr>
<tr>
<td>1.8-2.1</td>
<td>6</td>
<td>0.0246</td>
<td>7937</td>
</tr>
<tr>
<td>2.1-2.4</td>
<td>7</td>
<td>0.0250</td>
<td>3195</td>
</tr>
<tr>
<td>2.4-2.7</td>
<td>8</td>
<td>0.0070</td>
<td>12790</td>
</tr>
<tr>
<td>2.7-3.0</td>
<td>9</td>
<td>0.0004</td>
<td>69440</td>
</tr>
</tbody>
</table>

Table 2.1: Values of irradiance absorption coefficient, \( \beta_i \), and spectral weighting coefficients, \( a_i \), according to Paulson and Simpson (1977). For wavelengths 0.2 - 0.6 \( \mu m \), the turbidity of water is important (Jerlov, 1968; Soloviev and Schlüssel, 1996) and we have stated the irradiance absorption coefficient for Type I waters.

Fig. 2.6 shows the decay of solar irradiance calculated from eq. 2c to a depth of 10 mm assuming an incident solar radiation of 1000 W/m\(^2\). Assuming the clearest type I waters and with a tropical sun overhead, the amount of solar radiation absorbed
within the upper 0.1 mm was calculated to be 37 W/m² (3.7 %) which is insignificant compared to the energy from IR atmospheric emission absorbed in a similar depth and thus the effect of solar heating on the thermal skin layer is therefore negligible. The absorption rate increases exponentially with depth and at a depth of 1 mm, 132 W/m² (13.2 %) is absorbed while at 1 cm, 261 W/m² (26.1 %) of solar radiation is absorbed which is significant. However, our focus is on the top 0.1 mm from the sea surface which shows insignificant absorptivity of solar radiation rates as compared to IR.

![Graph](image)

Figure 2.6: Decay of solar irradiance assuming an incidence radiation of 1000 W/m² calculated from eq. 2c with $a_i$ and $\beta_i$ from table 2.1. Red dotted horizontal line indicates a depth of 0.1 mm while the vertical line corresponds to a value of 963 W/m². Blue dotted horizontal line indicates a depth of 1 mm while the vertical line corresponds to a value of 868 W/m².
2.3.2 Distribution of winds and its effect on the thermal skin layer.

Winds disrupt the TSL when they are strong enough to create breaking waves which typically occur at wind speeds above 7 m/s. Jessup et al. (1997) showed this momentary disruption through analysis of video and IR images of a breaking wave in the open ocean and found that the time for the TSL to re-establish itself in the wakes of the breaking waves is correlated with the energy dissipation rate. The recovery rate of the TSL was found to be less than a second and the field results were supported by laboratory measurements where breaking waves ranging from spilling to plunging were mechanically generated. This affirms that the TSL is almost always present even during winds greater than 7 m/s except at regions where the wave breaking occurs.

Fig 2.7 shows individual histogram plots of NAURU99 and AMMA06 in red, along with a dashed-dotted blue line and dashed-dotted black line overlaid to indicate the probability distribution function (PDF) and Weibull fit of 18 years of wind speed data constructed from the Special Sensing Microwave Imager (SSM/I) data version 6 from January 1, 1988 through December 31, 2006. The PDF and Weibull fit are both obtained from Woods et al. (2014) within the region denoted in fig 2.8 by the red square and red diamond for NAURU99 and AMMA06 respectively. Plotted along in fig 2.8 are the cruise tracks of NAURU99 and AMMA06. The distribution of wind speed data of both NAURU99 and AMMA06 agrees well with the PDF and Weibull fit. For AMMA06, we notice a bimodal histogram with a peak at about 7 m/s and a second smaller peak at about 12 m/s. The first peak agrees well with the PDF and Weibull fit, while the second smaller peak at 12 m/s is explained by a group of higher winds observed during the first half of leg 1 of AMMA06 between dates 26th May to 5th June 2006.

For the dataset below, imposing a wind speed threshold of 7 m/s implies that we would be representing 74.94 % of winds within the NAURU99 cruise and 52.58 % of
winds within the AMMA06 cruise while the SSM/I data shows that 82.34\% and 34.68\% over the equatorial Pacific and mid-latitude Atlantic Ocean would be represented respectively.

Figure 2.7: (a) Histogram of measurements for NAURU99 cruise. Black dashed-dotted line represents the PDF of 18 years of measurements over the region denoted by the red square in fig 2.8, constructed using the Special Sensing Microwave Imager (SSM/I) data version 6 while the blue dashed-dotted line represents the Weibull fit. (b) Histogram of measurements for AMMA06 cruise. Black dashed-dotted line represents the PDF of 18 years of measurements over the region denoted by the red diamond in fig 2.8, constructed from SSM/I data version 6 while the blue-dashed-dotted line represents the Weibull fit. SSM/I wind speed distributions are obtained from Woods et al. (2014).
Figure 2.8: Plot of mean wind speed over the entire ocean constructed from 18 years of measurements from SSM/I data version 6. Solid black line: Plot of NAURU99 cruise tracks. Dotted black line: Plot of AMMA06 cruise tracks. Red square: Region where wind speed measurements were obtained from SSM/I data version 6 for comparison with NAURU99 wind speed data. Red diamond: Region where wind speed measurements were obtained from SSM/I data version 6 for comparison with AMMA06 wind speed data.

Further analysis of SSM/I data over the entire ocean is shown in fig. 2.9 where the PDF and cumulative distribution function (CDF) are plotted (Woods et al., 2014) together with the combined wind speed for NAURU99 and AMMA06 during the night and day in red. Similar to the analysis made for winds within the individual cruise regions, if only winds below 7 m/s are to be accounted for, this would represent about 47.13 % of winds over the Worlds Oceans from the SSM/I data and about 81.04 % of winds for NAURU99 and AMMA06.

The above percentages show that a threshold of 7 m/s is able to represent the majority of winds for our cruise data and almost half of the World’s wind distribution is accounted for. From fig. 2.8, a large fraction of wind speeds below 7 m/s occurs
within the equatorial region and increases to an average of 10 m/s at the poles. Analysis of winds below 7 m/s are therefore highly representative of the equatorial and mid-latitude regions which comprises of the Pacific, Atlantic and Indian Ocean and accounts for the main bulk of the World’s Oceans. If the threshold is to be extended to 10 m/s, the percentage of winds accounted for would be 76.52 % and 94.08 % for SSM/I and the two field programs respectively (fig. 2.9) and our analysis would be able to include regions of the Antarctic and Artic Oceans. For our data analysis in the subsequent chapters, a threshold of 10 m/s has been applied simply to illustrate the relationships observed at higher winds, however the focus of our analysis is on lower winds of < 7 m/s.

Figure 2.9: Plot of the probability distribution function (top) and cumulative distribution function (bottom) for wind speed data collected from NAURU 1999 and AMMA 2006 and 18 years histogram of measurements over the entire ocean constructed using the Special Sensing Microwave Imager (SSM/I) data version 6. SSM/I wind speed distributions are obtained from Woods et al. (2014).
Chapter 3

Methods

This chapter presents an overview of the principles and methods used behind the techniques and considerations made in our analysis. We first describe the derivation of the sea surface spectral emissivity, $\epsilon$, from a variance minimizing technique which is essential for atmospheric corrections on the M-AERI’s sea surface emission spectra.

Next, we present a retrieval of the TSL vertical profile using hyperspectral radiance measurements from the M-AERI. Following this, we calculate the turbulent fluxes through the use of the bulk-aerodynamic formulae and describe a set of polynomial functions used to determine the exchange coefficients. The last section discusses the pros and cons of using cloud cover to investigate the response of the TSL profile with incoming longwave radiation and its analogy to the effect of increasing levels of greenhouse gases.

3.1 Deriving the sea surface spectral emissivity.

The sea surface emissivity varies with wavenumber and numerous environmental conditions such as water temperature, salinity, the emission angle at the sea surface, the wind-speed dependence of the tilts of facets of the sea surface and surfactants. It is an important parameter used in the reflection correction for M-AERI spectral
measurements before applying the methodology described in Section 3.2. This section describes a technique to derive the sea surface emissivity values from M-AERI data and discuss the sensitivity of the emissivity values to BTs.

### 3.1.1 The Variance Minimizing Technique (VarMinT).

Hanafin (2002) demonstrated a robust piecewise linear variance minimizing technique to derive spectral emissivity values from M-AERI spectral data and is termed VarMinT. With reference to eq. 3a, the VarMinT works by taking a wavenumber segment within the radiance spectrum, $R_{water}$ and $R_{sky}$, and calculates the variance of the brightness temperature, $B(v, T_{skin})$, obtained for different emissivity, $\epsilon(v)$ values. The $\epsilon(v)$ corresponding to the smallest variance calculated is taken to be $\epsilon(v)$ across the wavenumber segment and assumed to be constant over this range. Following Hanafin (2002), the VarMinT was applied to each spectrum at 3 different wavenumber segments ($5 \text{ cm}^{-1}$, $7.5 \text{ cm}^{-1}$, $10 \text{ cm}^{-1}$). The $\epsilon(v)$ calculated for each wavenumber segment are subsequently averaged into $10 \text{ cm}^{-1}$ intervals such that at least 4 estimates occur in each $10 \text{ cm}^{-1}$ average and are passed through a digital box-car low-pass filter to produce a smooth spectral emissivity.

$$\epsilon(v) = \frac{R_{water} - R_{sky}}{B(v, T_{skin}) - R_{sky}}$$  \(3a\)

The assumptions for the VarMinT are that the sky radiance, $R_{sky}$, is not correlated with the sea surface emissivity spectrum, $\epsilon(v)$, and $\epsilon(v)$ is a smooth function with wavenumber. Fig. 3.1 shows a sample emissivity plot derived from a M-AERI spectrum taken from the Nauru 1999 cruise together with emissivity values calculated by Filipiak et al. (2008) at a $55^\circ$ view angle, $3 \text{ m/s}$ wind speed and at a water temperature of $300 \text{ K}$ with a salinity of $35 \text{ psu}$. We observe slight discrepancies between our derived emissivity values using the VarMinT with Filipak’s data which may be due to differences in the
environmental conditions in which the dataset was obtained including the fact that Filipiak’s data are obtained from laboratory experiments and are therefore in a very controlled environment. Otherwise, the VarMinT method is robust and is very close to Filipiak’s data.

![Graph](image)

Figure 3.1: Spectral emissivity values calculated for a NAURU 1999 cruise spectra. Blue line: 5 cm\(^{-1}\) bins; red line: 7.5 cm\(^{-1}\) bins; green line: 10 cm\(^{-1}\) bins; black dotted line: averaged and low-pass filtered emissivity values; magenta line: emissivity values obtained from Filipiak et al. (2008)

3.1.2 Emissivity sensitivity test.

Here we describe a sensitivity test to relate changes in \(\epsilon(v)\) to the changes in BT. For this test, the \(\epsilon(v)\) spectrum, as derived above, was decreased by a constant value of 0.01 and the BT was calculated. A ratio of the changes in BT to \(\epsilon(v)\), i.e. \(S_e = \frac{\Delta(BT)}{\Delta\epsilon(v)}\), as a function of wavenumber is shown in fig. 3.2. Drawn in the same figure is a blue dotted reference line of \(S_e = 20\), indicating where a 0.01 change in \(\epsilon\) would result in a 0.2 K change in temperature. Above this line, small changes in \(\epsilon\) would result in larger
BT variations. From Planck’s law, the percentage change in radiance resulting from a percentage change in BT, $S_{BT} = \frac{\Delta B(v,T)}{\Delta BT/B_T}$, increases with increasing wavenumber in the IR region. Therefore, changes in $\epsilon$ affect the changes in radiance and we expect $S_{\epsilon}$ to increase with increasing wavenumber. This is shown within the wavenumber range of 800-1000 cm$^{-1}$ as a rapid radiance drop (120 - 50 mW sr$^{-1}$ m$^{-2}$ (cm$^{-1}$)$^{-1}$) is observed. However, $S_{\epsilon}$ does not continue to increase steadily because of a plateau in radiances from 1000 - 1200 cm$^{-1}$ at $\sim$50 mW sr$^{-1}$ m$^{-2}$ (cm$^{-1}$)$^{-1}$ and 2640 - 2800 cm$^{-1}$ at $\sim$0.15 mW sr$^{-1}$ m$^{-2}$ (cm$^{-1}$)$^{-1}$. Hence, the variability in $S_{\epsilon}$ is not largely due to the magnitude of the downwelling sky radiance, but is instead dominated by the variability of $\epsilon$. This is deduced from observing fig. 3.1 that at 2640 - 2800 cm$^{-1}$, $\epsilon \approx$ 0.95 while at 1000 - 1200 cm$^{-1}$, $\epsilon \approx$ 0.97, thereby implying that a given variation of $\epsilon$ would have a larger effect on BT for wavenumbers with higher $\epsilon$ (eq. 3a).

Plotted on fig. 3.2 is the downwelling sky radiance in green. Note that regions in which the downwelling sky radiance ‘peaks’, the gradient experiences a ‘trough’. This is because a peak of the downwelling sky radiance results in the contrast between the upwelling sea radiance and downwelling sky radiance to be smaller resulting in a poorer estimate of $\epsilon$, (the denominator in eq. 3a approaches zero). The ‘peaks’ occur due to changes in the atmospheric transmissivity spectrum and is caused by the gaseous components of the cloud-free atmosphere. We would therefore observe larger oscillations at deeper depths in the BT with depth profile (eg. fig. 3.7) because the measured downwelling sky radiance spectrum is more variable at higher wavenumbers (fig. 3.2). This also implies that there is more variability in the contrast required for the VarMinT approach to produce $\epsilon$. Thus the oscillations, or large variability, are likely to be artifacts of the process, and are not physical, underscoring the need to fit a smooth curve to the temperature with depth profile to attain the first-guess profile in the TSVD retrieval process.
Change in emissivity = 0.01. Change in BT = $C_2v/\log(c_1v^3 + 1) - C_2v/\log(c_2v^3 + 1)$ where $R_1 = R_{sea} - R_{sky}(1 - \epsilon)/\epsilon$ and $R_2 = R_{sea} - R_{sky}(1 - (\epsilon - 0.01))/\epsilon - 0.01$)

### 3.2 Retrieving the thermal skin layer’s vertical profile.

Details pertaining to the retrieval of the thermal skin layer’s vertical profile are described and demonstrated with synthetic data in Wong and Minnett (2016a). Wong and Minnett (2016b) demonstrated the same technique on field data with a comparison to the results obtained from synthetic data to show the feasibility of the retrieval technique. This section is largely extracted from Wong and Minnett (2016a) and will describe the retrieval technique and assumptions on the thermal skin layer’s profile in order to obtain a physically reasonable temperature profile.
3.2.1 The inverse problem.

At a scene mirror angle of 235° from nadir, the M-AERI is effectively sensing the total amount of radiation emitted from the sea surface, $R_{sea}$, including the reflected sky radiance, $(1 - \epsilon)R_{sky}$, where $\epsilon$ is the sea surface emissivity. Details in obtaining $\epsilon$ were described in Section 3.1. We are interested in $R_{sea}$ as it contains information regarding radiation emitted from each underlying water layer from the sea surface and can therefore be used in the retrieval of the thermal skin layer profiles. It has been established by Wong (2013) and McKeown et al. (1995) that $R_{sea}$ is a summation of the radiance emitted from each infinitesimal layer attenuated by the intervening layer from the sea surface with an attenuation factor defined by Beer-Lambert’s law:

$$R_{sea}(v) = -\int_{0}^{\infty} B(v, T(-z)) \frac{d(e^{-\alpha(-z)})}{dz}dz$$

(3b)

$B(v, T(z))$ is Planck’s function with $T(z)$ being the vertical temperature profile of interest. $\frac{d(e^{-\alpha(-z)})}{dz}$ is the weighting function given by Beer-Lambert’s law. The vertical ordinate, $z$, is defined positive upwards with $z = 0$ being the ocean surface, thus all values of $z$ are negative since we are analyzing depths beneath the ocean surface and the intensity of radiation decreases with negative depth.

If $T(z)$ is linear, the problem is trivial and solving eq. 3b analytically will result in a linear solution. However, $T(z)$ is highly non-linear (e.g. Saunders, 1967; Liu et al., 1979) therefore posing a problem in solving eq. 3b as the equation is ill-conditioned (Eyre, 1987; Rodgers, 2004). An ill-conditioned problem means that the outcome is highly sensitive to the accuracy of the measurements, $R_{sea}(v)$, as small errors in the measurements would result in large errors in the results.

To overcome this issue, there are numerous techniques termed regularization methods which constrain the solution through the introduction of an additional criterion such that there will be convergence towards the true solution. One such
regularization method is known as the Truncated Singular Value Decomposition (TSVD) and has been shown to be able to perform a retrieval of the thermal skin layer profile using radiance spectra obtained from the M-AERI (Wong, 2013; Wong and Minnett, 2016a,b). The next section describes the formulation of the inverse problem before application of the TSVD method.

3.2.2 The linearized model.

Before applying the TSVD technique to the inverse problem, a numerical integration of eq. 3b using a 100-point Gauss-Legendre quadrature rule with an interval of integration from -0.04 cm to 0 cm is performed. This allows for 20 unevenly spaced nodes to be within depths of -0.01 cm to 0 cm. Extending the interval of integration to a depth of -0.04 cm ensures that the entire source region of the emitted IR radiation is included in the integral.

A Taylor expansion is subsequently performed assuming that we know a set of first-guess solutions: $T_{1}^{fg}, T_{2}^{fg}, ..., T_{n}^{fg}$ at $z_{1}, z_{2}, ..., z_{n}$ which also gives rise to $I^{fg}(v_{j})$ through Planck’s law, where subscript $j = 1, 2, ..., m$ and $m$ is the number of wavenumber values. $I^{fg}(v_{j})$ therefore represents the radiance spectrum of the first-guess temperature profile. This results in:

$$R_{sea}(v_{j}) - I^{fg}(v_{j}) = \sum_{i=1}^{n} \frac{\partial I(v_{j})}{\partial T_{i}} |_{fg} (T_{i}^{fg} - T_{i})$$

(3c)

The superscript $fg$ denotes the first-guess profiles. We are therefore minimizing the difference between the measurements and the first-guess profile at every wavenumber $v_{j}$. The Jacobian matrix, $A_{ji} = \frac{\partial I(v_{j})}{\partial T_{i}} |_{fg}$, has a high condition number of $\sim 10^{21}$ which is indicative of the difficulty in obtaining a physical solution as the inversion of the problem would be contaminated by errors.
The TSVD technique is subsequently applied to eq. 3c. A singular value decomposition is performed on the Jacobian matrix, $A$, decomposing it into three matrices $U$, $S$ and $V$ (i.e. $A = USV^T$). An example of the decay of the singular values of $A$ which is contained in the diagonal matrix $S$, is shown in fig. 3.3 along with a red dotted-dashed line which represents the M-AERI’s noise level of 0.01% and a red solid line which is indicative of the machine epsilon:

![Image of singular values decay](image)

**Figure 3.3:** Logarithmic plot of the singular values of the discretized inverse problem, eq. 3c. The red dashed line indicates the noise level of 0.0001 (0.01%). The red solid line indicates the level of the machine epsilon ($\epsilon = 2.2204 \cdot 10^{-16}$).

Based on the M-AERI’s noise level of 0.01%, it is apparent that the first 6 singular values should be kept and the remaining singular values discarded as they are much smaller than the error level of the M-AERI. Using more than 6 singular values would mean that the inversion would be taking into significance the noise in the measurements and this would result in unphysical retrievals.
Eq. 3c can therefore be solved by inverting $A$ such that $A^{-1} = V S_p^{-1} U^T$, where $p=6$ is the truncation parameter from fig. 3.3. Eq. 3d shows how $T_i$ may be obtained given knowledge of the remaining parameters in the equation.

$$T_i - T_i^{fg} = A^{-1}[R_{sea}(v_j) - I^{fg}(v_j)]$$

(3d)

### 3.2.3 Determining the first-guess temperature profile.

We will now address how the first-guess profile, $T_i^{fg}$ and $I^{fg}(v_j)$, is obtained. It is ideal for the first-guess profile to have a similar structure to the expected retrieved profile as it will aid in the convergence through the minimization of $R_{sea}(v_j)$ and $I^{fg}(v_j)$. A bulk surface flux model by Liu et al. (1979) which made partial use of the surface renewal theory to derive temperature profiles in the viscous sublayer has been adopted. Liu et al. (1979) solved for $T(z, t)$ using the heat diffusion equation, $\frac{\partial T}{\partial t} = \kappa \frac{\partial^2 T}{\partial z^2}$, where $\kappa$ is the thermal conductivity of water, under 3 different boundary conditions: (1) Forced convection with a constant skin and sub-skin temperature; (2) Free convection with a constant temperature difference; (3) Free convection with a constant heat flux. Case 3 was adopted here (eq. 3e) as the first-guess model because this is the most realistic of open ocean conditions. Furthermore, since attempts are being made to retrieve the temperature profile of the thermal skin layer, it is preferable not to make any assumptions about the temperature values.

$$\frac{T - T_b}{T_s - T_b} = \pi^{0.5} 6i^3 \text{erfc}(\frac{2z}{3\pi^{0.5}\delta_c})$$

(3e)

where $6i^3 \text{erfc}(x) = \frac{(1+x^2)e^{-x^2}}{\sqrt{\pi}} - (1.5 + x^2)xe\text{rfc}(x)$ and erfc($x$) is a complementary error function defined as 1-erf($x$). $\delta_c = \kappa \frac{T_s - T_b}{Q_0}$ is a scaling depth with $T_s$ and $T_b$ being the surface and sub-skin temperature respectively and $Q_0$ represents the purely conductive heat flux averaged over time or space.
To obtain a set of good first guess values from the M-AERI, a simple direct mapping technique previously demonstrated by McKeown et al. (1995) and Hanafin (2002) is used. This mapping technique utilizes knowledge of the emission depth with wavenumber relationship in fig. 1.3, \( D_p(v) \), and maps the brightness temperature (BT) spectrum derived from the M-AERI directly to the emission depth with respect to \( v \) to obtain a vertical temperature profile. Eq. 3e is fitted to this temperature profile to produce a smooth continuous curve that is used as the first-guess profile for the TSVD technique. A sample of the first-guess profile is plotted as a dotted-dashed blue line in fig. 3.4.

### 3.2.4 Synthetic data results.

The TSVD method was tested firstly on synthetic data with and without the addition of noise (Wong and Minnett, 2016a) and subsequently on field data (Wong and Minnett, 2016b). Comparison between the field data results with the noise-added synthetic results showed much similarity in the need to either average many spectra (\( \sim 300 \)) or to decrease the spectral resolution by binning individual spectrum into 11 wavenumber intervals such that a physical retrieval may be obtained. Averaging many spectra is not ideal as it means that we are effectively retrieving one thermal skin temperature profile over a course of \( \sim 5 \) nights which is not useful for future scientific studies and will not be discussed here (refer to Wong and Minnett (2016b)). Therefore, the spectral resolution was lowered to 11 wavenumber intervals such that individual temperature profiles may be retrieved for every set of measurements. The choice of 11 wavenumber intervals was chosen by analyzing synthetic data injected with noise equivalent to the M-AERI’s noise equivalent delta radiance (NE\( \Delta \)L) specification.

From the discussion in Section 3.1 and fig. 3.2, it was observed that noise in the radiance values results in BT values at higher wavenumbers (2640-2800 cm\(^{-1} \)) to have a much larger variation due to the higher BT-to-radiance sensitivity. As a result, the
radiance values at higher wavenumbers (2640-2800 cm$^{-1}$) were averaged into one bin, while the lower wavenumber range (800-1200 cm$^{-1}$) was averaged into 10 bins at 40 cm$^{-1}$ intervals.

The limits of the first-guess profile’s subskin temperature, $T^{fg}_b$, and skin temperature, $T^{fg}_s$, was found to play a significant role in the retrieval process. $T^{fg}_b$ was observed to be at a cooler temperature compared to the true subskin temperature value. The reason for this is due to the decay in the absorbed radiation by Beer-Lambert’s law which means that less radiation is absorbed at deeper depths. Similarly, when the M-AERI is sensing radiation emitted at deeper depths, the magnitude of the radiation measured is smaller than its original value while the radiation measured at the sea surface by the M-AERI is representative of the radiation emitted due to the sea surface’s temperature. As a result, when the attenuation rates are not taken into consideration, as in the case of obtaining the first-guess profile where temperatures are simply obtained by Planck’s function, a lower temperature value is calculated at deeper depths. The remainder of this section would show the results obtained from tests which were run on synthetic data and describe how an iterative process was set-up to obtain suitable values of $T^{fg}_b$ and $T^{fg}_s$.

A sample profile, $T_{original}(z)$, was generated from eq. 3e with $T_s = 301$ K, $T_b = 301.3$ K, $Q_0 = 50$ W/m$^2$ and the simulated radiance spectrum, $R_{sea}(v)$ was generated. The objective here in using synthetic data is to use the TSVD technique to try and obtain $T_{original}(z)$ and to evaluate any discrepancies between the retrieved profile and $T_{original}(z)$. A first-guess profile was obtained by direct mapping using $D_p$ as described in Section 3.2.3.

Fig. 3.4 shows the results of using the TSVD technique on $T_{original}(z)$ at different truncation levels of $p = 3, 4, 5$ and 6. Overshoots in temperature are seen when the truncation level, $p$, is larger than 3 which are almost certainly unphysical. The iterated solution’s subskin temperature is defined to be $T^{result}_b$ and for discussion purposes,
we have defined $T_{\text{result}}^b$ to occur at $z = -0.04$ cm. The solution shows temperature inversions at depths of about -0.005 cm to -0.01 cm at $p = 3$ and 4, and the inversions become more pronounced at $p \geq 4$. This shows that invoking the direct mapping of emission depth to BT is insufficient to allow the solutions to converge to a realistic profile. We define the temperature profile to be realistic when the profile is monotonic and no inversions are observed. The iterated solutions closest to $T_{\text{original}}(z)$ ($p = 3$ and 4) still results in temperature inversion artifacts. It is also observed that $T_{\text{result}}^b$ converges to the first-guess’ profile temperature value at the same depth, $T_{\text{fg}}^b$, thereby resulting in $T_{\text{result}}^b$ to be lower than the desired value. The observed temperature inversions near the surface are likely a result of the iterated solution compensating for the lower $T_{\text{result}}^b$.

![Figure 3.4: Plot of temperature with depth. Green: brightness temperature obtained from M-AERI radiances assuming the M-AERI is measuring $T_{\text{original}}(z)$, mapped to emission depth; black solid line: $T_{\text{original}}(z)$; blue dotted line: $T_{\text{fg}}(z)$ which is a complementary error function fitted to the data points shown in green; red lines: the iterated result at $p = 3, 4, 5$ and 6.](image-url)
It is clear that $T_{fg}^s$ and $T_{fg}^b$ play a very important role in the convergence of the solution and must be of close proximity to $T_s$ and $T_b$ of $T_{original}(z)$ to avoid unphysical artifacts. The importance of a first guess that is close to the final solution is a common requirement for many ill-conditioned inversion problems especially when multiple solutions are possible. To test this, $T_{fg}^b$ was incrementally increased by 0.01 K to higher temperatures until no temperature inversions were observed in the retrieved solution and it was found that this occurred at $T_b = T_{fg}^b + 0.07$. This resulted in the solution converging to $T_{original}(z)$ (fig. 3.5) and re-affirms the need for a good estimation of the limits of the first-guess profile, $T_{fg}^s$ and $T_{fg}^b$. The iterative method therefore accommodates the process of obtaining a suitable $T_{fg}(z)$ such that the retrieval by the TSVD method converges to a realistic solution.

Figure 3.5: Left hand plot: Result without iteration to obtain suitable limits for $T_{fg}(z)$. Right hand plot: Iterated result with $T_{fg}(z) = T_{fg}(z) + 0.07$. Green: brightness temperature obtained from M-AERI radiances assuming the M-AERI is measuring $T_{original}(z)$, plotted to emission depth; black solid line: $T_{original}(z)$; blue dotted line: $T_{fg}(z)$ which is a complementary error function fitted to the green; red dotted line: the iterated result at $p = 6$.

Next, $T_{original}(z)$ was injected with noise according to the noise equivalent delta radiance (NE$\Delta$L) properties of the M-AERI. The M-AERI has a NE$\Delta$L normal distribution of 0.2 mW sr$^{-1}$ m$^{-2}$(cm$^{-1}$)$^{-1}$ and 0.015 mW sr$^{-1}$ m$^{-2}$(cm$^{-1}$)$^{-1}$ at wavenumbers 670-1400 cm$^{-1}$ and 2000-2600 cm$^{-1}$ respectively (Minnett et al., 2001). Fig. 3.6 shows that noise in the radiance values causes the BT values to have a much larger variation
at higher wavenumbers due to the higher BT-to-radiance sensitivity in this range. Furthermore, the process of adjusting $T^f_g$ has resulted in an over-estimation of $T_b$ due to the persistence of temperature inversions. However, there is still convergence in the upper 0.02 mm of the retrieved profile with a $T_{\text{iterated}}-T_{\text{original}}$ value of < 0.1 K. The retrieved profile is affected by the large BT variation in the deeper water layer of > 0.05 mm, resulting in the solution failing to converge sometimes to the physical profile.

Figure 3.6: Left plot: Plot of temperature versus depth. Green: brightness temperature obtained from simulated M-AERI radiance spectrum assuming the M-AERI measures $T_{\text{original}}(z)$ with noise added to the radiance spectrum; black line: $T_{\text{original}}(z)$ without noise; blue line which is underneath the red line: $T_f^g(z)$ which is a complementary error function fitted to the green with $T_b = T^f_g + A$, where $A$ is a constant; red solid line: the iterated result at p=2. Right plot: Temperature difference between $T_{\text{original}}(z)$ and the iterated results $T_{\text{iterated}}$ at p=2.

To reduce the noise in $T_{\text{original}}(z)$, radiance values were averaged for every 40 cm$^{-1}$ wavenumber interval at the lower wavenumber range of 800-1200 cm$^{-1}$ and due to the large variability seen at the higher wavenumber range of 2640-2800 cm$^{-1}$, we averaged this entire range into one bin. This gives 10 bins at the lower wavenumbers and 1 bin at the higher wavenumbers (fig. 3.7).
The TSVD iterative approach was applied to 100 noise-injected spectra, where each spectrum is averaged to the 11 wavenumber intervals stated above. The truncation parameter was observed to produce smooth continuous curves similar to the desired profile when $p = 2$. Fig. 3.8 shows the retrieved profile of 5 individual runs. The results show the convergence of the retrieved profile to the initial, specified profile but with some slight discrepancies mainly in the $T_b$ value which may again be attributed to the higher variability of BT at deeper depths. Despite the variation in $T_b$, a smooth profile can be retrieved with a curvature similar to the original profile for all 100 runs. Of the 100 simulation runs, the mean retrieved $T_b$ and $T_s$ is 301.25 K and 300.97 K with a standard deviation of 0.023 K and 0.020 K respectively. The average of the difference in radiance between the iterated result and the original profile at the averaged lower wavenumber of $\sim$ 2720 cm$^{-1}$ is 0.0145 mW sr$^{-1}$ m$^{-2}$(cm$^{-1}$)$^{-1}$ with a standard deviation of 0.00252 mW sr$^{-1}$ m$^{-2}$(cm$^{-1}$)$^{-1}$ while at the higher wavenumbers, the average is 0.00818 mW sr$^{-1}$ m$^{-2}$(cm$^{-1}$)$^{-1}$ with a standard deviation of 0.000606 mW sr$^{-1}$ m$^{-2}$(cm$^{-1}$)$^{-1}$. These radiance differences are consistently $\leq$ 0.06 mW sr$^{-1}$ m$^{-2}$(cm$^{-1}$)$^{-1}$ and $\leq$ 0.01 mW sr$^{-1}$ m$^{-2}$(cm$^{-1}$)$^{-1}$ which translates to a radiance difference percentage of 0.1% and 2.5% and a BT error percentage of $\leq$ 0.016% and $\leq$ 0.11% calculated at 300 K at the lower and higher wavenumber ranges respectively. Fig. 3.9 shows a flowchart of the final algorithm.

Typical skin SST retrieved from measurements of spectrometers, such as the M-AERI, are performed at a wavelength of 7.7 $\mu$m ($= 1300$ cm$^{-1}$) (Minnett et al., 2001). From the above percentages, the TSVD method has the potential to obtain temperature retrieval accuracies of $\leq$ 0.048 K at this wavelength. This is better than the desired accuracy of 0.1 K and shows promise for obtaining accurate skin SST measurements from spectrometers, such as M-AERI, or the validation of satellite skin SST retrievals.
Figure 3.7: Sample dataset of temperature versus depth with noise added to the radiance spectrum of $T_{\text{original}}(z)$. Black stars represent the averaging of the radiance values into wavenumber intervals of 40 cm$^{-1}$ for the shallowest BT values and 1 averaged point for the deeper BT. Black line represents the best fit complementary error function through the 11 points.

Figure 3.8: Five sample retrievals with noise added to the radiance spectrum of $T_{\text{original}}(z)$. Green stars: 11 averaged points from the noisy BT plot; black line: $T_{\text{original}}(z)$ without noise; blue line: $T^{fg}(z)$ which is a complementary error function fitted to the green crosses with $T_b = T^{fg}_b + A$ where $A$ is a constant; dashed red line: iterated result at p=2.
Figure 3.9: Flowchart of final algorithm for the retrieval of the thermal skin layer from hyperspectral measurements obtained from an interferometer using an iterative TSVD method (Wong and Minnett, 2016a).
3.2.5 Extracting properties from the retrieved thermal skin layer profile.

To analyze the dependences of the retrieved thermal skin layer profile on air-sea fluxes (described in Chapter 4), we have extracted and defined a number of variables from the retrieved profiles which will be used in the analyses. These variables are listed below and fig. 3.10 shows a cartoon of the definition of these variables.

(i) $SST_{skin}$ — Temperature at $z = 0$ mm.

(ii) $SST_{subskin}$ — Temperature at lower boundary of TSL, where the TSL profile transitions to the isothermal mixed layer.

(iii) $SST_{0.1mm}$ — Temperature at the maximum penetration depth of the EM skin layer within wavenumbers $500 - 3000$ cm$^{-1}$.

(iv) $SST_{5m}$ — Temperature measured by a thermosalingraph at $z = 5$ m.

(v) $\Delta T_{skin-5m}$ — Temperature difference between $SST_{skin}$ and $SST_{5m}$.

(vi) $\Delta T_{skin-0.1mm}$ — Temperature difference between $SST_{skin}$ and $SST_{0.1mm}$.

(vii) $\frac{\Delta T_{0.1mm}}{\Delta T_{5m}}$ — Ratio of $\Delta T_{skin-0.1mm}$ versus $\Delta T_{skin-5m}$. 
3.3 Calculation of turbulent fluxes.

To derive the turbulent heat fluxes of sensible (SH) and latent heat (LH), we use the following bulk aerodynamic formulas:

\[
LH = C_L L_v \rho_a U_{10} (q_a - q_s) \tag{3f}
\]

\[
SH = C_S C_p \rho_a U_{10} (T_a - T_s) \tag{3g}
\]

where \(C_L\) and \(C_S\) are the exchange coefficients for LH and SH over the ocean respectively, \(L_v\) is the latent heat of vaporization (\(\sim 2.5 \times 10^6\) J kg\(^{-1}\)), \(\rho_a\) is the density of air,
$C_p$ is the specific heat capacity of air, $q_a - q_s$ is the difference in the specific humidity of air and the saturated specific humidity, $T_a - T_s$ is the temperature difference between the air and sea surface.

Stability-dependent polynomial functions derived by Kara et al. (2005) are used to determine the exchange coefficients, $C_L$ and $C_S$. The polynomial functions are parameterizations of the exchange coefficients derived from the Coupled Ocean-Atmosphere Response Experiment (COARE) bulk algorithm (version 3.0) (Fairall et al., 2003). Kara et al. (2005) modified the COARE (v3.0) to produce $C_L$ at various $T_a - T_s$ and $U_{10}$ intervals over the global ocean. They also stated that the roughness length for temperature is assumed to be equivalent to the roughness length for humidity in the COARE (v3.0) algorithm such that the moisture, $C_L$, and heat transfer, $C_S$, coefficients are equal. Kara et al. (2005) exchange coefficients are termed the Naval Research Laboratory Air-Sea Exchange Coefficients (NASEC) and are suitable for our studies as they enable the use of the skin SST parameter in the turbulent flux calculations over a wide range of atmospheric conditions.

Kara et al. (2005) derived $C_L$ for 3 different ranges of $T_a - T_s$: $-8.00^\circ C \leq T_a - T_s < -0.75^\circ C$, $-0.75^\circ C \leq T_a - T_s \leq 0.75^\circ C$ and $0.75^\circ C < T_a - T_s \leq 7.00^\circ C$. Eq. 3h below shows $C_L$ derived for the range $-0.75^\circ C \leq T_a - T_s \leq 0.75^\circ C$ as all our data lie within this range while fig. 3.11 shows the $C_L$ values derived for our dataset.

$$C_L = C_{L0}U_{10} + C_{L1}U_{10}(T_a - T_s)$$ (3h)

For $U_{10} \leq 5$ m/s

$$C_{L0} = -0.01056U_{10}^2 + 0.09743U_{10} + 0.858$$

$$C_{L1} = -0.07706U_{10}^2 + 0.7345U_{10} - 1.927$$
For $U_{10} \geq 5$ m/s

$$C_{L0} = -0.0000216U_{10}^2 + 0.00961U_{10} + 1.023$$

$$C_{L1} = -5.048U_{10}^{-2} + 0.2048U_{10}^{-1} - 0.00393$$

Figure 3.11: Derived $C_L$ values from Kara et al. (2005) for the NAURU 1999 and AMMA 2006 dataset.

### 3.4 Cloud cover as a surrogate for increasing levels of greenhouse gases.

Detecting changes in the downwelling infrared irradiance resulting from increases in anthropogenic greenhouse gases (GHG’s) is difficult since consistent datasets with long timeframes are required to produce a significantly detectable signal. The M-AERI has only about two decades of measurements at sporadic locations, which limits our analysis of increasing GHG emissions as the downwelling longwave signal will be very
small. Thus we have exploited the use of variability of clouds overhead during M-AERI readings as a surrogate for an increase in atmospheric IR emission due to rising levels of GHG. As this is an indirect approach to the problem, there are pros and cons to the technique of using clouds as the atmospheric forcing. These advantages and disadvantages will be discussed in this section.

Under the presence of clouds, there is an increase in IR emission as the base of low lying clouds radiate at higher temperatures compared to those of the upper atmosphere - an increase in cloud forcing has a positive effect on the LW$_{in}$ radiation. Fig. 3.12 shows the atmospheric spectrum when the M-AERI is sensing radiation at zenith. When no clouds are present overhead (black line), we note the presence of an atmospheric window (where the atmosphere is more transmissive) from $\sim$800 - 1250 cm$^{-1}$. When a cloud is overhead (blue line), the M-AERI senses radiation emitted from the cloud base which occurs at a higher temperature. This results in a closure of the atmospheric window creating a ‘Planckian’ shaped spectrum and a higher IR emission.

The advantage in using cloud cover as a surrogate of GHG’s is the high signal in LW$_{in}$ produced in short time intervals thereby making the problem more tractable. The LW$_{in}$ forcing due to a passing cloud is much stronger (about an increase of 200 W/m$^2$) as compared to a typical forcing of 3.7 W/m$^2$ anticipated for a doubling of CO$_2$ (Ramanathan et al., 1979; IPCC, 2014). This higher signal enables us to detect any instantaneous subtle response in the curvature of the TSL due to a sudden increase in LW$_{in}$ and therefore permits an approach to address the problem.

One main issue which needs to be addressed is the difference in spectral composition between the downwelling infrared spectrum due to the presence or absence of clouds and contributions from each GHG. Shown in fig. 3.13 are the absorption spectra of four main GHG’s (H$_2$O, CO$_2$, O$_3$, CH$_4$) and their net effect on a radiative spectrum calculated at 300 K while fig. 3.14 shows the transmissivity spectrum of the four main
GHG’s. A temperature of 300 K was chosen as it is the approximate average SST temperature measured during the field cruises. Although the main anthropogenic GHG contributor is CO₂, both natural and anthropogenic GHG’s are included to show the similarities and differences in spectral signatures.

Comparing fig. 3.12 with 3.13, we note that the atmospheric window in fig. 3.12 corresponds largely to the absorptivity spectrum of H₂O which is the largest natural GHG. This is expected since clouds are largely made up of water vapor. Therefore, the radiative signal due to an increase in cloud cover is comparable to large increases in H₂O concentrations in the atmosphere. We can thus understand the feasibility of using radiation from clouds by comprehending differences between the H₂O absorption spectrum with other GHG’s.

The first spectral difference noted between figs. 3.12 and 3.13 is the absence of data below 500 cm⁻¹ in the M-AERI spectra. In this region, we observe that H₂O
Figure 3.13: Absorption spectrum of 4 different greenhouse gases. The black line shows the spectrum calculated using Planck’s function at 300 K. The blue lines show the contributions of selected greenhouse gases at 300 K. Absorptivity data is obtained from the high-resolution transmission absorption database (HITRAN) (Rothman et al., 2013).

has an absorptivity of $\sim 100\%$ ($\sim 0\%$ transmittance) while $\text{CO}_2$, $\text{O}_3$ and $\text{CH}_4$ shows the reverse (low absorptivity and high transmittance). Changes in atmospheric $\text{H}_2\text{O}$ would therefore be most sensitive to the downwelling infrared spectrum within this wavenumber range. This implies that from 0 - 500 cm$^{-1}$, the atmospheric window due to $\text{H}_2\text{O}$ is dependent on the humidity in the atmosphere. The field data in this study are concentrated within the tropics, with a typical humidity of 75 % (50 - 80 % for AMMA06 and 65 - 82 % for NAURU99). Thus, the absence of data from 0 - 500 cm$^{-1}$ within M-AERI spectra is not of major concern due to the high humidity with relatively small variability.

The next most prominent differences in spectral composition are the narrow absorption bands in $\text{CO}_2$, $\text{O}_3$ and $\text{CH}_4$ at approximately 520 - 800 cm$^{-1}$ and 2200 - 2400 cm$^{-1}$ for $\text{CO}_2$, 560 - 880 cm$^{-1}$ and 920 - 1240 cm$^{-1}$ for $\text{O}_3$, and 1080 - 1490 cm$^{-1}$.
Figure 3.14: Transmissivity spectrum of 4 different greenhouse gases. Transmissivity data is obtained from the high-resolution transmission absorption database (HITRAN) (Rothman et al., 2013).

for CH₄ where the transmissivity approaches zero implying that the atmosphere is essentially opaque at these wavenumbers. H₂O misses these absorption windows with a transmissivity approaching 1. This means that in these bands, when an increase in CO₂, O₃, or CH₄ occurs, there will be an increase in absorption and emission of radiation resulting in an increase in the downwelling radiation within these wavenumber ranges, whereas this effect will not be observed if there is an increase in H₂O or cloud cover. Replotting fig. 1.3 to show the emission depth range for the spectral windows of CO₂, O₃, CH₄ and H₂O (fig. 3.15), we observe that for CO₂, O₃ and CH₄, the emission depth corresponds to the upper 0.02 mm. We therefore expect to observe changes to the distribution of heat within the upper 0.02 mm of the TSL which would otherwise not be observed from an increase in H₂O or presence of clouds.

It is thus important to assess how the narrower spectral windows for CO₂, O₃ and CH₄ would modify the response of the TSL’s gradient as compared to increasing clouds
or H₂O concentration. However, to fully understand these changes would require a full radiative modelling with the TSL which would extend beyond the scope of this thesis. Therefore, to address this issue, we perform a simple analysis by considering CO₂ spectra simulated by a line-by-line radiative transfer model (LBLRTM) with cloudy and clear sky spectra. It should be noted that in this simulation, we are only assuming a beam of radiation impinging onto the sea surface from zenith. This simulation is performed as follows:

(i) Generate the following atmospheric spectra (fig. 3.16)

(a) Clear sky spectrum (denoted 1*CO₂) using temperature and humidity values from the tropical western pacific region and assuming US Standard atmosphere for profiles of CO₂, O₃, N₂O, CO, CH₄ and O₂.

(b) Cloudy sky atmospheric spectrum.

(c) 2*CO₂ spectrum from the clear sky spectrum.
(d) 3*CO$_2$ spectrum from the clear sky spectrum.

(ii) Using Beer’s Law, generate a heating rate profile (fig. 3.17). This is performed by calculating at every depth, the total radiation absorbed within each layer by integrating the radiation absorbed within each layer with respect to the M-AERI’s wavenumber range. (i.e. $\sum_{v=500cm^{-1}}^{3000cm^{-1}} \delta I(v)(v_{i+1} - v_i)$ mWm$^{-2}$sr$^{-1}$, where $v_i$ is the wavenumber at depth $i$, $\delta I(v)_i = I_0(v)e^{-\alpha(v)z_i} - I_0(v)e^{-\alpha(v)z_{i+1}}$ and $I_0(v)$ is the surface incident radiation at $v$. The depth resolution used is 0.01 mm.)

(iii) Compare the heating rate profiles by subtracting the heating rate profiles generated through i(b), i(c) and i(d) with i(a).

(iv) Analyze the difference in heating rate profiles (fig. 3.18) with the differences in atmospheric spectra (fig. 3.19).

From figs. 3.16 and 3.19, we note that the difference between cloudy and clear sky spectra has a much higher magnitude and broader spectral range compared to the difference between a 2*CO$_2$ or 3*CO$_2$ and clear sky spectra. Clouds generate a radianc difference of approximately 40 mW m$^{-2}$ sr$^{-1}$ (cm$^{-1}$)$^{-1}$, about 8 times that of the 3*CO$_2$ spectrum ($\sim$ 5 mW m$^{-2}$ sr$^{-1}$ (cm$^{-1}$)$^{-1}$). Thus, we expect to see a much larger change in the heating rates for cloud forcing which is shown clearly in fig 3.18. The top plot of fig. 3.18 shows that the cloud forcing produces a total difference in absorbed radiation of $\sim$ 9000 mW m$^{-2}$ sr$^{-1}$ at the surface whereas 3*CO$_2$ forcing only gives $\sim$ 500 mW m$^{-2}$ sr$^{-1}$. The bottom plot of fig. 3.18 shows clearly that even with 3*CO$_2$ forcing, only $< 5.5\%$ of the heating rate signal from cloud forcing is represented. In addition, CO$_2$ forcing is observed to vary the top 0.01 mm of the TSL while the effect of cloud forcing extends much deeper (to about 0.09 mm). This is expected due to the more effective closure of the spectral window in the cloudy spectrum, and as shown by the magenta dotted line in fig. 3.15, will affect the entire emission depth.
Figure 3.16: Atmospheric spectra generated for (1) clear sky ($1^*\text{CO}_2$) (blue line) (2) $2^*\text{CO}_2$ (red line) (3) $3^*\text{CO}_2$ (green dotted line) (4) cloudy sky (black line) using temperature and humidity values from the tropical western pacific region and assuming US Standard atmosphere for profiles of CO$_2$, O$_3$, N$_2$O, CO, CH$_4$ and O$_2$ from the line-by-line radiative transfer model (LBLRTM).

range whereas CO$_2$ (red dotted line is fig. 3.15) is localized within the top 0.01 mm. Thus, by utilizing cloud forcing (as opposed to CO$_2$ forcing), we expect the amplified downwelling longwave signal to magnify the temperature variations in the TSL and to affect the TSL at deeper depths such that changes in the TSL profile is detectable through the retrieval of the TSL profile using M-AERI sea surface emission spectra.

Despite the small signal produced by an increase in CO$_2$, however, it is also important to consider the effects of an increase in water vapor due to the warming of the atmosphere through a doubling of CO$_2$. With Earth’s warming, the saturation vapor pressure increases exponentially from the Clausius-Clapeyron equation: $e_s = e_0 e^{\frac{L_v}{R_v} (\frac{1}{T} - \frac{1}{T_0})}$, where $e_s$ is the saturation vapor pressure at $T$, $e_0$ is the vapor pressure at temperature $T_0$, $L_v$ is the latent heat of vaporization and $R_v = 8.314$ J/mol K. This means that at warmer temperatures, more water vapor can be present
The radiance heating rate, \[ \sum \delta I(v)(v_{i+1} - v_i) \times 10^4 \text{ mW m}^{-2} \text{Sr}^{-1} \times 10^4 \], is calculated as a function of depth (cm).

Figure 3.17: Heating rate profiles calculated for a cloudy atmospheric spectrum (black line), a clear sky spectrum (blue line), 2*CO\(_2\) spectrum (dotted red line), 3*CO\(_2\) spectrum (dotted green line).

In the atmosphere without reaching saturation. Thus, even though climate models introduce a typical forcing of doubling CO\(_2\) concentrations (Ramanathan et al. (1979)) in the atmosphere, which is very small as compared to cloud forcing, to analyze its effects on global warming, the increase in anthropogenic GHG’s would also increase the capacity of the atmosphere to hold more water vapor in the lower troposphere, the most abundant GHG in the atmosphere, generating a positive feedback.

From figs. 3.12 and 3.13, it is shown that the atmospheric window closed by clouds corresponds to the H\(_2\)O atmospheric window. This indicates that an increase in water vapor would produce a very similar heating rate profile as compared to that produced by cloud forcing. The most significant difference would be the magnitude of the water vapor forcing which will be much smaller than that due to the presence of a cloud. Manabe and Wetherald (1975) have simulated through the use of a general circulation model (Manabe (1969)) that the tropospheric relative humidity increases
Figure 3.18: Top: Difference in heating rate profiles from fig. 3.17. Bottom: Ratio of the difference in heating rate profiles.
by approximately 2.5% and with an increase in the surface air temperature $\sim$3 K in the tropics when forced with a doubling of CO$_2$. An atmospheric spectrum with a 2.5% increase in water vapor is simulated from the same clear sky spectrum in fig. 3.16 using the LBLRTM and is denoted by 1.025*H$_2$O in the top plot of fig. 3.20. Calculating the difference in heating rate between the clear sky and 1.025*H$_2$O spectra, the maximum difference in incident infrared radiation at the surface is $< 300$ mWm$^{-2}$sr$^{-1}$ (bottom plot of fig. 3.20). In terms of magnitude, this is much smaller than that produced by cloud forcing but is slightly larger than that produced by doubling CO$_2$ (heating rate difference at $z = 0$ cm for 2*CO$_2$, 3*CO$_2$ and 1.025*H$_2$O are 280 mWm$^{-2}$sr$^{-1}$, 475 mWm$^{-2}$sr$^{-1}$ and 289 mWm$^{-2}$sr$^{-1}$ respectively). The heating rate differences are also shown to extend to depths similar to that produced by cloud forcing. Both these results are expected given that the radiance spectra between clear sky and 1.025*H$_2$O spectra does not vary as greatly as compared to a cloudy and clear sky spectra (top plot of fig. 3.20).
In summary, our analysis of utilizing variations in cloud cover is seen to be an adequate, but imperfect, tool for studying the consequences of increasing concentrations of GHG’s. The lack of data below 500 cm$^{-1}$ is not of great concern given that we are analyzing data over the tropics. The difference in atmospheric windows between cloud forcing, CO$_2$ and H$_2$O forcing results in large differences in magnitude of the simulated heating rate profiles. However, this large difference is advantageous in our analysis with the TSL. This is because the large heating rates produced by cloud forcing will give a much more easily detectable signal in the TSL and allow us to analyze the response of the TSL to increasing levels of longwave radiation. It should also be noted that the doubling of CO$_2$ is an example to place the increase in GHG concentrations in perspective. We should be considering the total effect of GHG’s and from the analysis of increasing concentrations of the two most abundant GHG’s, CO$_2$ and H$_2$O, it is shown that the combined effect is synonymous to cloud forcings in terms of the decay of the heating rates with depth.
Figure 3.20: Top: Atmospheric spectra generated for (1) clear sky (1*CO$_2$) (blue line) (2) 1.025*CO$_2$ (red dotted line) using temperature and humidity values from the tropical western pacific region and assuming US Standard atmosphere for profiles of CO$_2$, O$_3$, N$_2$O, CO, CH$_4$ and O$_2$ from the line-by-line radiative transfer model (LBLRTM). Bottom: Difference in heating rate profiles between a clear sky spectrum (1*CO$_2$) and a 1.025*H$_2$O spectrum.
Chapter 4

Analysis of data obtained from NAURU99 and AMMA06 field programs

This chapter presents a detailed analysis of the NAURU99 and AMMA06 cruise datasets. We first discuss the inherent dependencies between the surface heat fluxes and wind speed within both datasets in Section 4.1. Following which, Section 4.2 provides analyses on the properties of the retrieved TSL profiles obtained from the TSVD method with wind speed. Section 4.3 discusses the dependences observed with surface fluxes and ties in the findings with a heat retention mechanism which answers to the question of how increasing levels of longwave radiation would affect the TSL such that there will be an increase in ocean heat content. Finally, Section 4.4 concludes this chapter’s findings.

4.1 Observations of interfacial heat fluxes and wind speed from the NAURU99 and AMMA06 field programs.

This section describes the observed dependencies between the heat fluxes with wind speed, $U_{10}$, obtained during the two cruises. The paragraphs below discuss the general
relationships found between the net flux, incoming and net longwave radiative fluxes, and sensible and latent heat turbulent fluxes with $U_{10}$. As our focus is on the radiative fluxes, Section 4.1.1 describes correlation analyses found between sensible heat (SH) and latent heat (LH) changes with incoming longwave ($LW_{in}$) and outgoing longwave ($LW_{out}$). Section 4.1.2 shows changes in net longwave ($LW_{net}$) with $LW_{in}$ and Section 4.1.3 describes the dependences found between net flux (Q) with $LW_{in}$ and $LW_{net}$.

The fluxes are defined negative for an upward flow (from the ocean to the atmosphere). The net flux, $Q$, is defined as the sum of net longwave radiation, latent and sensible heat fluxes (eq. 4a); incident solar radiation is excluded given that day-time data are omitted as discussed in Section 2.3.1. The net longwave radiation is the difference between the incoming and outgoing longwave radiation (eq. 4b):

$$Q = LW_{in} - LW_{out} - LH - SH,$$ (4a)

where $LW_{out} = \epsilon \sigma T^4$, $\sigma = 5.6710^8 W m^2 K^4$, $\epsilon =$ sea surface emissivity

$$LW_{net} = LW_{in} - LW_{out}$$ (4b)

An initial comparison between fluxes and wind speed, $U_{10}$ (fig. 4.1), shows a correlation between $Q$, LH and SH with $U_{10}$ with an $R^2$ of 0.68, 0.77 and 0.43 respectively. This is expected given that LH and SH have a direct wind speed dependence as expressed in the bulk-aerodynamic formulae (eq. 3f and 3g) and from the Kara et al. (2005) polynomial functions (eq. 3h). Since only night-time data are being analyzed, the magnitude of $Q$ is largely defined by LH thereby explaining the correlation between $Q$ and $U_{10}$. We do not expect to observe a correlation between $U_{10}$ and the radiative fluxes, $LW_{net}$ or $LW_{in}$ as depicted in fig. 4.1, with $R^2$ values of 0.01 and 0.07 respectively. The following subsections discuss the dependences observed between $U_{10}$, SH, LH with $LW_{net}$ and $LW_{in}$. 
Figure 4.1: Plot of fluxes (W/m²) versus wind speed (m/s) from AMMA06 and NAURU99 cruise. $R^2$ calculated for $Q$, $LH$, $SH$, $LW_{in}$ and $LW_{net}$ are 0.68, 0.77, 0.43, 0.07 and 0.01 respectively.

4.1.1 Relationships between the turbulent fluxes (SH and LH) with the radiative fluxes ($LW_{in}$ and $LW_{out}$).

In this section, we expect to observe a decoupling between the turbulent and radiative fluxes as this would mean that an increase in $LW_{in}$ is not immediately returned to the atmosphere by turbulent exchanges. The independence observed would therefore verify our hypothesis of the TSL responding to $LW_{in}$ changes which results in heat from absorbed solar radiation in the mixed layer to remain.
Plots of SH and LH with LW\textsubscript{in} and LW\textsubscript{out} are given in fig. 4.2. No correlation is found between these variables and we observe the similar $U_{10}$ dependence with SH and LH which has been discussed and observed in fig. 4.1. LW\textsubscript{out} shows two groups of values, one centered at approximately -480 W/m\textsuperscript{2} while the other group ranges from about -435 to -470 W/m\textsuperscript{2}. This is because from Stefan-Boltzmann Law (eq. 4a), there is a direct proportionality between LW\textsubscript{out} and the fourth power to the sea surface temperature (SST); the values centered about -480 W/m\textsuperscript{2} correspond to those taken from the NAURU99 cruise in the tropical Pacific Ocean, while the other group of LW\textsubscript{out} values corresponds to the AMMA06 cruise in the subtropical to tropical Atlantic Ocean.

We can expect this lack of correlation between SH, LH with LW\textsubscript{in}, LW\textsubscript{out} given that LH and SH are turbulent fluxes where the energy released (or absorbed) is due to phase transitions or from temperature differences. The emission or absorption of radiative fluxes is instead dependent on the temperature of the atmosphere (LW\textsubscript{in}) and of the surface (LW\textsubscript{out}) and its emission/absorptivity value is a function of wavelength. Thus, variations in longwave radiation should not be directly affected by LH and SH and verifies our hypothesis.

To further ensure that the hypothesis is well-established, we also need to ensure that there is no coupling between the effect of clouds and turbulent fluxes, such as increased winds due to the presence of convective clouds resulting in an increase of the turbulent fluxes. Plots of the rate of change of turbulent fluxes with the rate of change of LW\textsubscript{in} and LW\textsubscript{out} are given in fig. 4.3. No correlation is found between these parameters thereby ensuring that the turbulent heat loss does not change with increased LW\textsubscript{in}. This also indicates that the heat generated in the TSL due to the absorption of IR is not immediately returned to the atmosphere by LH or SH. There is thus an increase in the rate of energy production in the skin layer, and a fixed sink of heat at the top of the skin layer.
Due to the lack of correlation between the radiative fluxes and LH and SH, we can therefore focus our analysis of the TSL on the radiative fluxes. This is because we expect any changes to the TSL to be largely influenced by the absorption and emission of IR radiation rather than turbulence since molecular diffusion dominates the vertical transport within the TSL and turbulence is suppressed.

Figure 4.2: Top two plots: Plot of SH and LH with LW\textsubscript{in}. Bottom two plots: Plot of SH and LH with LW\textsubscript{out}. Plot shows a cluster of points centered about -480 W/m\textsuperscript{2} which corresponds to the NAURU99 cruise and a spread of points from -435 W/m\textsuperscript{2} to 470 W/m\textsuperscript{2} corresponding to the AMMA06 cruise. The color bar represents wind speed at 10 m, $U_{10}$. 
Figure 4.3: Top two plots: Rate of change of SH and LH with the rate of change of \( LW_{in} \). Bottom two plots: Rate of change of SH and LH with the rate of change of \( LW_{out} \).
4.1.2 The relationship between incoming (LW$_{in}$) and net long-wave LW$_{net}$ fluxes.

In this section, we would like to explore whether changes in LW$_{in}$ would affect LW$_{out}$ or LW$_{net}$. This is to support our hypothesis that the absorption of increase IR radiation is not immediately compensated by the release of heat back into the atmosphere through LW$_{out}$. Fig. 4.4 shows a plot of LW$_{net}$ versus LW$_{in}$ for all wind speeds. The colors represent $SST_{skin}$ and open circles indicate data from AMMA06 while crosses represent data from NAURU99. Regression lines ($LW_{net} = G \cdot LW_{in} + H = LW_{in} - LW_{out}$) are calculated for every 1 K $SST_{skin}$ interval from 295 to 302 K and three of the regression lines are shown in fig. 4.4 denoted by the blue (295-296 K), green (298-299 K) and black (301-302 K) solid lines. If a strong linear relationship is observed between LW$_{in}$ and LW$_{net}$, LW$_{out}$ can therefore be regarded as a constant irrespective of LW$_{in}$. A table of the regressed lines’ coefficients is given in table 4.1.

![Figure 4.4: Plot of LW$_{in}$ versus LW$_{net}$ for all wind speeds. The color bar denotes $SST_{skin}$ and shows that at $SST_{skin}$ temperatures, the relation between LW$_{net}$ and LW$_{in}$ is inversely correlated with an average gradient of 0.97. Blue, green and black solid lines are regressed to the points between 295 - 296 K, 298 - 299 K and 300 - 301 K respectively. Crosses represents data from NAURU99, open circles represent data from AMMA06.](image)
<table>
<thead>
<tr>
<th>SST_{skin} (K)</th>
<th>Gradient (W/m^2/K)</th>
<th>Intercept (W/m^2)</th>
</tr>
</thead>
<tbody>
<tr>
<td>295-296</td>
<td>1.022</td>
<td>444</td>
</tr>
<tr>
<td>296-297</td>
<td>1.014</td>
<td>443</td>
</tr>
<tr>
<td>297-298</td>
<td>0.923</td>
<td>414</td>
</tr>
<tr>
<td>298-299</td>
<td>0.960</td>
<td>433</td>
</tr>
<tr>
<td>299-300</td>
<td>0.979</td>
<td>448</td>
</tr>
<tr>
<td>300-301</td>
<td>0.956</td>
<td>444</td>
</tr>
<tr>
<td>301-302</td>
<td>0.952</td>
<td>461</td>
</tr>
</tbody>
</table>

Table 4.1: Table indicating the gradient, G and LW_{net} intercept, H (i.e. LW_{net} = G*LW_{in} + H) of the regressed lines according to data of the SST_{skin} interval.

From table 4.1, the average gradient, G, is 0.97 and is found to not be significantly different from a gradient of 1 but significantly different from a gradient of 0. This implies that a unit increase in the magnitude of LW_{in} would result in a unit decrease in the magnitude of LW_{net}. This signifies that there is no change in the retrieved SST_{skin} with LW_{in} which also means that LW_{out} is not influenced by the increase in LW_{in}. Because we are considering temperature changes within the TSL and not the bulk of the ocean, the unobservable change in SST_{skin} is likely due to the large thermal capacity of sea water (~3850 J/(Kg K)) despite observing a large increase in LW_{in}. The colored striations in fig. 4.4 further indicates the dependence of LW_{net} on SST_{skin} because of the fourth power dependence with LW_{out} (eq. 4a).

For our current dataset and majority of open ocean conditions at low winds of < 10 m/s, LW_{out} is always higher than LW_{in}, which is expected given that the SST is always higher than the overlying air temperature. This gives a negative LW_{net} which indicates a net upward longwave flux from the ocean to the atmosphere. Thus, an increase in the magnitude of LW_{in} would result in a decrease in the magnitude of LW_{net} since LW_{out} has very little variability (due to the small variability of SST_{skin} as shown in fig. 4.4). The LW_{in} variability, besides being highly variable due to the presence or absence of clouds, is also dependent on the concentration of atmospheric gases that interact with the IR radiation, of which the most important is water vapor. The amount of water vapor content depends on the relative humidity and air
temperature (Clausius-Clapeyron equation in Section 3.4) and therefore has much variability, with the relative humidity having a higher influence on the water vapor content than the air temperature. For example, from table 4.2, using the mean values obtained from the NAURU99 and AMMA06 cruise, with a relative humidity of 75% and air temperature of 299.7829 K, a saturation vapor pressure, $e_s$, of 3.4893 kPa is calculated ($e_s = 0.6108 \times e^{(17.27T_{air}/(T_{air}+237.3))}$). Assuming a ±10% change in relative humidity (67.5% and 82.5%), this corresponds to a ±10% change in vapor pressure, $e_0$. If we calculate the change in temperature corresponding to a similar change in $e_0$ (using the Clausius-Clapeyron equation), the temperature change is about ±8%.

<table>
<thead>
<tr>
<th>RH (%)</th>
<th>$e_0 = RH/100% \times e_s$ (kPa)</th>
<th>$T_0$ (K) ($\frac{e_s}{e_0} = \exp\left(\frac{L_v R_v}{R_v} \left(\frac{1}{T_0} - \frac{1}{T_s}\right)\right)$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>67.5</td>
<td>2.3553</td>
<td>211.6150</td>
</tr>
<tr>
<td>75</td>
<td>2.6170</td>
<td>229.7269</td>
</tr>
<tr>
<td>82.5</td>
<td>2.8787</td>
<td>249.0061</td>
</tr>
</tbody>
</table>

Table 4.2: Table showing the calculated $e_0$ and $T_0$ at different RH. $e_0$ is calculated using the equation: $e_0 = RH/100\% \times e_s$. $T_0$ is calculated using the equation $\frac{e_s}{e_0} = \exp\left(\frac{L_v R_v}{R_v} \left(\frac{1}{T_0} - \frac{1}{T_s}\right)\right)$.

From the observed result of a less negative $LW_{net}$ corresponding to a more positive $LW_{in}$ in fig. 4.4, we would therefore expect the net flux, $Q$, to decrease accordingly assuming the other turbulent fluxes have little or no variability. This relationship is shown in fig. 4.5 and will be further explored in the next section along with a discussion on the dependences between $Q$; $LW_{net}$ and $LW_{in}$. A less negative $LW_{net}$ also implies that there will be a surplus of energy within the TSL from the absorbed radiation. The additional energy has altered the curvature of the mean TSL profile such that the amount of heat being released back into the atmosphere from beneath the TSL is lowered, therefore providing a mechanism for energy from absorbed sunlight to be retained in the bulk of the ocean. The decrease in magnitude of $LW_{net}$ in fig.
4.4 therefore provides us with the first indication of how the absorption of increasing amounts of LW_in would result in a rise in OHC despite the longwave absorption depths not exceeding a few microns.

### 4.1.3 Dependences between net flux (Q) with incoming (LW_in) and net longwave (LW_net) fluxes.

Although fig. 4.1 shows no correlation between U_10 with LW_in or LW_net, the dependency between U_10 and Q is significant, through the U_10 influence on the turbulent fluxes. This calls for the need to stratify the data to identify the wind effects. Therefore, to examine the dependence on winds, the data have been segregated into 4 wind speed regimes (U_10 < 2 m/s, 2 m/s < U_10 < 4 m/s, 4 m/s < U_10 < 6 m/s and 6 m/s < U_10 < 10 m/s). Subsequent analysis and discussions will be based on these 4 wind speed intervals.

Fig. 4.5 shows the dependences of Q on LW_in (top 4 plots) and LW_net (bottom 4 plots) segregated by the four wind speed intervals. The top 4 plots of Q versus LW_in, show a significant correlation at the lowest wind speed interval of U_10 < 2 m/s (R^2 = 0.70) where a more positive LW_in is associated with a less negative Q. This correlation decreases progressively as the wind speed becomes higher (R^2 = 0.04 at 6 m/s < U_10 < 10 m/s). Similarly, the bottom four plots of Q versus LW_net, indicate a very strong correlation is observed at U_10 < 2 m/s (R^2 = 0.94) where a less negative LW_net results in a less negative Q. This correlation also becomes less significant as the wind speed increases (R^2 = 0.28 at 6 m/s < U_10 < 10 m/s).

The less significant correlations at higher winds can be explained by the larger influence of wind speed on Q. From fig. 4.1, at U_10 > 4 m/s, the magnitude of Q is greater than 100 W/m² which has well exceeded that of LW_net (< 100 W/m²). As a result, the main component of Q is LH (ignoring SH because the magnitude of SH is small) which is heavily dependent on U_10 and thus causes the correlation between
Q and LW\textsubscript{in} and LW\textsubscript{net} to diminish because the effects of LH exceed those of LW\textsubscript{net}. Furthermore, because the data are segregated into U\textsubscript{10} intervals to separate the wind speed effects, it is clear that as the wind speed decreases, the increase in $R^2$ (from 0.04 to 0.7 for Q versus LW\textsubscript{in} and from 0.28 to 0.94 for Q versus LW\textsubscript{net}) signifies that the longwave radiation plays a much more significant role in determining the fluxes at the air-sea interface, as is to be expected. The calculated p-values are $9.86 \times 10^{-16}$ and $3.95 \times 10^{-7}$ for Q versus LW\textsubscript{in} at $U_{10} < 2$ m/s and $2$ m/s $< U_{10} < 4$ m/s respectively and for Q versus LW\textsubscript{net}, the p-values are calculated to be less than $10^{-20}$ for all four wind speed regimes thereby showing that the linear regressions are significant at a 5\% significance level. Thus, subsequent analysis should be focused on the lower U\textsubscript{10} bins of 0-2 m/s and 2-4 m/s to allow for a better analysis of the radiative effects on the TSL.

At $U_{10} < 2$ m/s, we observe that the increase in magnitude of LW\textsubscript{in} results in the magnitude of Q and LW\textsubscript{net} to decrease (fig. 4.5) which is expected (as discussed in Section 4.1.2). This provides evidence that less heat escapes from the ocean to the atmosphere when there is are greater amounts of LW\textsubscript{in}. The gradient of the regressed line between Q and LW\textsubscript{net} is of 1.04 (not significantly different from 1), indicates that a given unit change in LW\textsubscript{net} corresponds to a similar change in Q. This again tells us that at low winds, the variations in Q are only due to the variations in LW\textsubscript{net} and are minimally influenced by LH and SH. Evidence of variations of the retrieved thermal gradients of the TSL are presented and discussed in the next section and are linked to relationships found with radiative heat fluxes.
Figure 4.5: Plots of Q with LW\textsubscript{in} and LW\textsubscript{net} segregated into four wind speed regimes. The colorbar denotes U\textsubscript{10} in m/s. (a) Plot of Q with LW\textsubscript{in}. P-values calculated at 95% confidence level for net flux with LW\textsubscript{in} at U\textsubscript{10} < 2 m/s, 2 m/s < U\textsubscript{10} < 4 m/s, 4 m/s < U\textsubscript{10} < 6 m/s, 6 m/s < U\textsubscript{10} < 10 m/s are \textit{9.86*10}^{-16}, \textit{3.95*10}^{-7}, \textit{0.909}, \textit{0.0005} respectively while gradients calculated are 1.48, 0.801, 0.018 and -0.58 respectively. (b) Plot of Q with LW\textsubscript{net}. P-values calculated at 95% confidence level for net flux with LW\textsubscript{net} at U\textsubscript{10} < 2 m/s, 2 m/s < U\textsubscript{10} < 4 m/s, 4 m/s < U\textsubscript{10} < 6 m/s, 6 m/s < U\textsubscript{10} < 10 m/s are \textit{3.91*10}^{-35}, \textit{7.80*10}^{-42}, \textit{2.23*10}^{-31}, \textit{3.00*10}^{-21} respectively while gradients calculated are 1.04, 1.34, 1.58 and 1.69 respectively.
4.2 Dependences between wind speeds on properties of the thermal skin layer profile.

In this section, we first establish the relationships between $\Delta T_{\text{skin}-5m}$ and $\Delta T_{\text{skin}-0.1mm}$ with wind speed (Section 4.2.1). Section 4.2.2 shows the correlations found between $\Delta T_{0.1mm}/\Delta T_{5m}$ with wind speed.

4.2.1 Dependences between $\Delta T_{\text{skin}-5m}$ and $\Delta T_{\text{skin}-0.1mm}$ with wind speed.

Fig. 4.6 shows a plot of $\Delta T_{\text{skin}-5m}$ and $\Delta T_{\text{skin}-0.1mm}$ with $U_{10}$. As a refresher, $\Delta T_{\text{skin}-5m}$ is defined to be the temperature difference between the retrieved temperature at $z \approx 0$ and the temperature measured by a thermosalinograph at 5 m, while $\Delta T_{\text{skin}-0.1mm}$ is defined as the temperature difference between the retrieved temperature at $z \approx 0$ and at $z = -0.1$ mm, which is taken as the temperature at the deepest depth of the M-AERI TSL profile. The magenta triangle line shows the Donlon et al. (2002) parameterization of $\Delta T_{\text{skin}-5m}$ against $U_{10}$ based on a study using measurements from 6 research cruises from multiple investigators in which the skin temperature was defined as the temperature measured by sea-going radiometers, of several different designs, and the temperature at $\sim 5$ m depths were measured by thermosalinographs on the ships. The line with the green triangles shows the Minnett et al. (2011) parameterization of $\Delta T_{\text{skin}-5cm}$ (note the units) against $U_{10}$ based on field measurements from a research cruise off New Zealand with $\Delta T_{\text{skin}-5cm}$ defined to be the temperature difference between the ocean’s skin temperature measured by a M-AERI and the temperature at 5 cm depth measured by a thermometer on a small surface-following float tethered close to the footprint on the sea surface of the M-AERI field-of-view. A least squares approximation to our field data, binned to every 1 m/s, using the same equation Donlon and Minnett used for their parameterizations:
\( \Delta T = A + B \exp(-U_{10}/C) \), was performed and is denoted by the blue and red lines along with the mean and ±1 standard deviation represented by the blue and red triangles and error bars in fig. 4.6.

We compare coefficients A, B and C, between the parameterizations with \( \Delta T_{\text{skin}-5m} \) (solid blue line) and \( \Delta T_{\text{skin}-0.1mm} \) (solid red line) with \( U_{10} \) and with Donlon’s and Minnett’s parameterization. Coefficient A, which signifies the asymptotic value when \( U_{10} \) is large, is similar (at a value of -0.2 K) for both \( \Delta T_{\text{skin}-5m} \) and \( \Delta T_{\text{skin}-0.1mm} \) meaning that at high wind speeds, both \( \Delta T_{\text{skin}-5m} \) and \( \Delta T_{\text{skin}-0.1mm} \) tends to -0.2 K. This also indicates that there is close agreement between SST\(_{0.1mm}\) and SST\(_{5m}\) which implies that at high winds, the TSL profile is essentially isothermal between depths of 0.1 mm to 5 m thereby suggesting that the TSL is thinner than the deepest emission depth of the EM skin layer as SST\(_{0.1mm}\) is measuring the value of SST\(_{\text{subskin}}\). This result is expected under higher wind conditions as the wind driven shear would erode the TSL resulting in the skin layer to be thinner (e.g. Saunders (1967); Soloviev and Schlüssel (1994)).

Coefficient B defines the curvature of the least squares fit and it is shown that coefficient B varies greatly depending on the dataset. Donlon has a value of 0.3 which is close to our derived B value of 0.24 for \( \Delta T_{\text{skin}-5m} \) thus giving confidence in the \( \Delta T_{\text{skin}-5m} \) measurements. Minnett shows a much higher curvature value of 0.724 which may be attributed to the fact that the measurements are taken at higher latitudes during the end of the summer months (March-April 2004) resulting in a greater temperature difference within the skin layer due to stronger insolation and weaker winds. Furthermore, the temperature difference is obtained between depths of the ocean’s surface and 5 cm rather than at 5 m which provides a more accurate representation of the temperature difference within the TSL. Therefore, we expect the value of B from Minnett et al. (2011) to correspond closely to the derived B of \( \Delta T_{\text{skin}-0.1mm} \). However, this is not the case, as B = 0.025 for \( \Delta T_{\text{skin}-0.1mm} \) meaning
that the retrieved $\Delta T_{\text{skin-0.1mm}}$ is independent of wind speed. This deviation will be addressed in the next few paragraphs.

Coefficient C, as suggested by Donlon et al. (2002), is indicative of a threshold value that suggests the transition from free convective and molecular heat transfer processes (at lower $U_{10}$) to wind-induced shear-driven turbulent heat transfer (at higher $U_{10}$). Donlon mentioned that their derived value of 3.7 m/s “marks the onset of ubiquitous small waves on the sea surface” and noted the significant increase in $\Delta T$ at $U_{10} < 2$ m/s which is similarly seen in our data of $\Delta T_{\text{skin-5m}}$. Our derived C value of 3.5 m/s for $\Delta T_{\text{skin-5m}}$ is lower than Donlon’s C value and may again simply be due to the latitudinal variations between Donlon’s data and ours. The values of C in our data are in more agreement with Donlon’s and Minnett’s, ranging from 2 m/s to 3.7 m/s.

Under night-time, low wind conditions, the diurnal thermocline is eroded by convection driven by surface heat loss leaving behind a turbulent isothermal mixed layer beneath the TSL. Thus, we would expect SST$_{5m}$ and SST$_{0.1mm}$ to be of similar values or have a constant ratio provided SST$_{0.1mm}$ is representative of SST$_{\text{subskin}}$. However, as shown clearly in fig. 4.6, there is an increase in deviation between $\Delta T_{\text{skin-0.1mm}}$ and $\Delta T_{\text{skin-5m}}$ as $U_{10} < 3.5$ m/s. This indicates that SST$_{0.1mm}$ is not measuring SST$_{\text{subskin}}$ and implies that the TSL is thicker than the EM skin layer thereby limiting the retrieval of SST$_{\text{subskin}}$. 
Figure 4.6: Scatter plot of $\Delta T_{\text{skin-0.1mm}}$ (top: grey +) and $\Delta T_{\text{skin-5m}}$ (bottom: solid grey dots) and versus $U_{10}$. Red line represents least squares fit to the mean (red triangles) of $\Delta T_{\text{skin-0.1mm}}$. Blue line represents least squares fit to the mean (blue triangles) of $\Delta T_{\text{skin-5m}}$. Both averages are calculated at every 1 m/s interval and the blue and red error bars corresponds to $\pm$ 1 standard deviation. Magenta triangle line denotes Donlon et al. (2002) parameterization while green triangle line denotes Minnett et al. (2011) parameterization.

To illustrate the wind driven changes in the TSL through the observed dependences between $\Delta T_{\text{skin-0.1mm}}$, $\Delta T_{\text{skin-5m}}$ and $U_{10}$, a cartoon of the TSL profile under high and low wind conditions is drawn in fig. 4.7. The cartoon indicates, assuming $\text{SST}_{5m}$ is representative of $\text{SST}_{\text{subskin}}$, how at higher winds, the erosion of the TSL which leads to a thinner TSL with a smaller temperature difference allows $\text{SST}_{\text{subskin}}$ to be retrieved as the subskin depth occurs within 0.1 mm. As opposed to a thicker TSL at low winds which would result in a discrepancy in our retrieved $\text{SST}_{0.1mm}$ with the measured $\text{SST}_{5m}$. 

$$
\Delta T_{\text{skin-0.1mm}} = -0.2 - 0.24 \exp(-\frac{U_{10}}{3.5})
$$

$$
\Delta T_{\text{skin-5m}} = -0.21 - 0.025 \exp(-\frac{U_{10}}{2})
$$

$$
\text{Donlon2002: } -0.14 - 0.3 \exp(-\frac{U_{10}}{3.7})
$$

$$
\text{Minnett2011: } -0.13 - 0.724 \exp(-\frac{U_{10}}{2.86})
$$
Figure 4.7: Cartoon of TSL profile under night-time conditions. Solid black line represents TSL profile under high winds (> 3.5 m/s). Dotted black line represents TSL profile with under low winds (< 3.5 m/s). Red star denotes subskin temperature, \( \text{SST}_{\text{subskin}} \). Blue star denotes skin temperature, \( \text{SST}_{\text{skin}} \). The vertical scale is non-linear.

### 4.2.2 Dependence between \( \Delta T_{0.1\text{mm}}/\Delta T_{5\text{m}} \) with wind speed.

In this section we analyze the dependence of the ratio \( \Delta T_{0.1\text{mm}}/\Delta T_{5\text{m}} \) on the surface wind, \( U_{10} \). As our retrievals are limited to EM skin layer emission depths (\( \sim 0.1 \text{ mm} \) within wavenumbers 500 - 3000 cm\(^{-1} \) (fig. 1.3)), if the thickness of the TSL exceeds this emission depth, the retrieved \( \text{SST}_{0.1\text{mm}} \) is not representative of \( \text{SST}_{\text{subskin}} \). Thus we have introduced a temperature ratio, \( \Delta T_{0.1\text{mm}}/\Delta T_{5\text{m}} \), the value of which is related to the thickness of the TSL. We assume a turbulent isothermal mixed layer beneath the TSL such that \( \text{SST}_{\text{subskin}} \approx \text{SST}_{5\text{m}} \) under night-time conditions. A smaller ratio means a larger difference between \( \text{SST}_{0.1\text{mm}} \) and \( \text{SST}_{5\text{m}} \) which implies that more of the vertical temperature gradient lies below the emission depth thereby indicating a thicker TSL.
Fig. 4.8 shows the plot of $\Delta T_{0.1mm}/\Delta T_{5m}$ versus $U_{10}$ with the red dots representative of bins of 0.5 m/s. $\Delta T_{0.1mm}$ starts deviating from $\Delta T_{5m}$ at about 3.5 m/s and approaches a value of $\approx 0.5$ below 2 m/s. The threshold of 2 m/s coincides with the parameterized coefficient $C$ between $\Delta T_{0.1mm}$ and $U_{10}$ while the 3.5 m/s threshold coincides with the parameterized coefficient $C$ between $\Delta T_{0.1mm}$ and $U_{10}$ (fig. 4.6). This suggests that as winds increase from 2 m/s to 3.5 m/s, wind effects start exceeding the molecular effects within the TSL, eroding and thinning the TSL through increased levels of turbulence below thereby resulting in $\Delta T_{0.1mm}$ approaching $\Delta T_{5m}$. For wind speeds below 2 m/s, molecular effects dominate and a thicker TSL exists. Thus, to analyze the response of the TSL to radiative effects, we should analyze data with winds under 2 m/s such that the effects of wind-driven shear will be reduced and the focus can be on effects due to interfacial fluxes and molecular heat transfer processes. The threshold of 2 m/s also agrees with what was observed and discussed in Section 4.1.3 between Q and LW$_{in}$ and LW$_{net}$.

It is unfortunate that SST$_{0.1mm}$, does not fully capture the TSL temperature difference at low winds. However, the agreement between $\Delta T_{0.1mm}$ and $\Delta T_{5m}$ at high winds provides evidence to support existing TSL models that the thickness of the TSL is inversely proportional to wind speed (e.g Saunders (1967); Soloviev and Schlüssel (1994); Fairall et al. (1996)). Chapter 5 will provide an in-depth analysis of these various models.
4.3 Dependences between surface fluxes on properties of the thermal skin layer profile.

In this section, only data with $U_{10} < 2$ m/s are discussed. Plots with $U_{10} > 2$ m/s are shown in Appendix A. The reason for presenting analyses at $U_{10} < 2$ m/s is that, as mentioned in the previous section 4.2.2, the 2 m/s threshold is observed to be when the effect of winds on the TSL exceeds the molecular conductive effects. Moreover, it has been established in Section 4.1.3 that $Q$ consists largely of the radiative fluxes at low winds of $< 2$ m/s which would also allow the isolation of turbulent flux effects. Appendix A also shows that for $U_{10} > 2$ m/s, no correlations are found with heat fluxes.

As established in Section 4.1.1, the turbulent fluxes are independent of the radiative fluxes and no coupling was found between increased winds due to convective clouds
and the turbulent fluxes. Thus, we only analyze the TSL properties with $LW_{in}$, the incoming longwave measured by the PIR, $LW_{net}$ and an additional variable, $LW_{in@zenith}$ (units: W/sr/m$^2$) which is defined as the integral of the M-AERI’s measured zenith spectrum. $LW_{in@zenith}$ is analyzed because the M-AERI has a much higher precision than the PIR. Furthermore, our retrievals of the TSL are in relation to M-AERI spectra. Thus the advantage of using $LW_{in@zenith}$ is to ensure that our comparison between the retrievals and the incoming IR is consistent, without external variables possibly influencing the IR readings. As opposed to using $LW_{in}$ which, being a product of the PIR, is less precise and measures a much wider angle of IR radiation therefore being more susceptible to the presence of clouds out of the M-AERI’s field of view and may result in difficulty in obtaining a signal between the TSL and IR radiative changes. Details pertaining to the advantages and disadvantages of using $LW_{in@zenith}$ rather than $LW_{in}$ will be further discussed in Section 4.3.2.

In Section 4.3.1, we first establish the relationships between $\Delta T_{skin-5m}$, $\Delta T_{skin-0.1mm}$ and $\Delta T_{0.1mm}/\Delta T_{5m}$ with $LW_{in}$, $LW_{in@zenith}$ and $LW_{net}$. Section 4.3.2 presents a discussion on the difference in correlations observed between $LW_{in}$ and $LW_{in@zenith}$.

### 4.3.1 Dependence of $\Delta T_{skin-5m}$, $\Delta T_{skin-0.1mm}$ and $\Delta T_{0.1mm}/\Delta T_{5m}$ on radiative fluxes.

In this section, we address the dependencies found between $\Delta T_{skin-5m}$, $\Delta T_{skin-0.1mm}$ and $\Delta T_{0.1mm}/\Delta T_{5m}$ with $LW_{in}$, $LW_{in@zenith}$ and $LW_{net}$ (figs. 4.9, 4.10, 4.11).
Figure 4.9: Top left: $\Delta T_{\text{skin-5m}}$ versus $LW_{\text{in}}$. Top right: $\Delta T_{\text{skin-5m}}$ versus $LW_{\text{net}}$. Bottom left: $\Delta T_{\text{skin-5m}}$ versus $LW_{\text{in@zenith}}$. Blue dots show the scatter. Solid black dots represent points binned to every 5 Wm$^{-2}$ of $LW_{\text{in}}$ and $LW_{\text{net}}$ and 2 Wsr$^{-1}$m$^{-2}$ of $LW_{\text{in@zenith}}$ with ±1 standard deviation. Black solid line denotes the regressed line to the solid black dots.

Figure 4.10: Top left: $\Delta T_{\text{skin-0.1mm}}$ versus $LW_{\text{in}}$. Top right: $\Delta T_{\text{skin-0.1mm}}$ versus $LW_{\text{net}}$. Bottom left: $\Delta T_{\text{skin-0.1mm}}$ versus $LW_{\text{in@zenith}}$. Blue dots show the scatter. Solid black dots represent points binned to every 5 Wm$^{-2}$ of $LW_{\text{in}}$ and $LW_{\text{net}}$ and 2 Wsr$^{-1}$m$^{-2}$ of $LW_{\text{in@zenith}}$ with ±1 standard deviation. Black solid line denotes the regressed line to the solid black dots.
The first conclusion from fig. 4.9 is the lack of correlation between $\Delta T_{\text{skin}-5\text{m}}$ with the radiative fluxes. This is an important result because it indicates that $\Delta T_{\text{skin}-5\text{m}}$ is independent of the radiative fluxes. Although we should not expect to observe any relationship between $\Delta T_{\text{skin}-5\text{m}}$ and the radiative fluxes because the 5 m depth has well exceeded the radiative emission depth in the longwave regime, however as mentioned above, there exists the isothermal mixed layer beneath the TSL during low wind, night-time conditions, thus SST$_{5\text{m}}$ may be representative of SST$_{\text{subskin}}$. This independence observed therefore allows us to conclude that changes in radiative fluxes do not affect the temperature difference within the TSL. Our result is further supported by previously published Donlon and Robinson (1997) whom did not find any dependence between the amount of cloud cover with $\Delta T$, which was defined to be the temperature difference between the bulk (measured by the SeaBird thermosalinograph...
at 5.5 m) and skin sea surface temperature (measured by the Satellites International Limited STR-100 radiometer).

From figs. 4.10 and 4.11, we observe significant correlations (using a t-test at a 5% significance level) between $\Delta T_{\text{skin}-0.1\text{mm}}$ and $\Delta T_{0.1\text{mm}}/\Delta T_{5\text{m}}$ with $\text{LW}_{\text{in@zenith}}$. It is observed that as $\text{LW}_{\text{in@zenith}}$ increases, $\Delta T_{\text{skin}-0.1\text{mm}}$ becomes more negative while $\Delta T_{0.1\text{mm}}/\Delta T_{5\text{m}}$ approaches unity. This observation, together with the independence found with $\Delta T_{\text{skin}-5\text{m}}$, implies that there is an adjustment of the gradient of the TSL rather than variations in the temperature difference due to increased longwave radiation from the closure of the cloud atmospheric window.

The response of the TSL profile due to an increased cloud forcing is illustrated in fig. 4.12. From the observed results as stated in the previous paragraph, we expect the TSL profile to adjust its curvature from the dotted black line to the solid black line in fig. 4.12. This curvature change agrees with our observations - $\Delta T_{\text{skin}-0.1\text{mm}}$ increases, $\Delta T_{0.1\text{mm}}$ approaches $\Delta T_{5\text{m}}$ and there is no change in $\Delta T_{\text{skin}-5\text{m}}$. 
With reference to fig. 4.12, as LW\textsubscript{in\textregistered\text{zenith}} increases, the TSL profile varies from a lower to higher gradient near the interface and higher to lower gradient at subskin depths. The adjustment of the TSL profile to a lower gradient at subskin depths implies that the release of heat from beneath the TSL decreases. In other words, there is a hindrance to the effective removal of heat from the mixed layer to the atmosphere.

This finding also ties in with our explanation from Section 4.1.2 - under low winds, there exists a surplus of energy within the TSL as an increase in LW\textsubscript{in} occurs, LW\textsubscript{net} and Q becomes less negative (fig. 4.4 and 4.5). This additional absorbed heat due to an increase in LW\textsubscript{in} modifies the TSL profile from the dotted to solid black line in fig. 4.12 to aid the removal of this excess heat, stored within the TSL, back into the atmosphere. As a result, the effective removal of heat from the mixed layer to the atmosphere is inhibited, thus decreasing the magnitude of LW\textsubscript{net} and retaining the heat beneath the TSL which is largely derived from the absorption of solar radiation.
This in turn results in an increase in the upper ocean heat content. Because the temperature difference within the TSL does not change, we therefore do not observe any dependency with the turbulent fluxes as discussed in Section 4.1.1.

4.3.2 The differences observed between LW\textsubscript{in} and LW\textsubscript{in@zenith}.

LW\textsubscript{in@zenith} is derived from irradiance measurements from the very well calibrated M-AERI’s narrow beam directed at zenith and covers a wavelength interval ranging from 500-3000 cm\textsuperscript{-1} while LW\textsubscript{in} consists of measurements from the less well calibrated PIR and is the product of integrating over a hemisphere from 200-2500 cm\textsuperscript{-1} (4-50µm). The differences between LW\textsubscript{in@zenith} and LW\textsubscript{in} are that the incident radiant energy is integrated over different incidence angles and spectral ranges. The two variables are plotted in fig. 4.13.

![Figure 4.13: Scatter plot of LW\textsubscript{in@zenith} (Wsr\textsuperscript{-1}m\textsuperscript{-2}) against LW\textsubscript{in} (Wm\textsuperscript{-2}). Black solid line denotes regressed line to the scatter plot.](image)

We first address the difference in the integration wavenumber range. Using a 300 K spectrum calculated by Planck’s law as a reference and integrating the spectrum from
500-3000 cm$^{-1}$ and 200-2500 cm$^{-1}$ at zenith, we obtain a flux value of 107.8 Wsr$^{-1}$m$^{-2}$ and 141.3 Wsr$^{-1}$m$^{-2}$ respectively. This gives a ratio of 0.76 which means that if we omit the spectral distribution when calculating the flux value, LW$_{in\@zenith}$ represents three quarters of the PIR’s measurement if the PIR measures radiation from zenith. This spectral difference between the M-AERI and PIR is not expected to affect the resulting dependencies with the TSL because the atmospheric window closure due to cloud forcing is entirely captured within the M-AERI spectra. Furthermore, as discussed in Section 3.4, the absence of data below 500 cm$^{-1}$ is not of concern in our study area which is concentrated within the tropics (high humidity with relatively small variability) and implies small changes in atmospheric water vapor thereby not affecting the water vapor window below 500 cm$^{-1}$.

The small difference between the flux values calculated above by integrating the spectrum from 500-3000 cm$^{-1}$ and 200-2500 cm$^{-1}$ at zenith suggests that majority of the differences observed between the variable LW$_{in}$ and LW$_{in\@zenith}$ is due to the difference in acceptance angle. The PIR’s hemispheric response means that its measurement includes radiation arriving with non-zero zenith angles. At low winds, when the TSL is thicker, the angular distribution plays a much more important role in the TSL’s response and the inclusion of incoming longwave at oblique angles may have masked any signals we had hoped to observe with the TSL properties. For example, clouds not within the M-AERI’s zenith view but within the PIR’s view angles would result in a high LW$_{in}$ value but the TSL’s response may be smaller (fig. 4.14) as compared to the presence of an overhead cloud. To demonstrate the effect of the angular distribution of incident radiation on the TSL heating rates, the effects of three incident beams at different zenith angles, assuming a source of radiation at 300 K are shown in fig. 4.14. This plot clearly shows that radiation impinging on the sea surface at oblique angles has a smaller effect than at zenith and is in fact proportionate to the cosine of the incident angle. This provides a simple snapshot
of the effect of the angular distribution of incident radiation on the TSL; a more comprehensive analysis would require full radiative transfer modelling that is beyond the scope of this thesis. Furthermore, the percentage of variability of $LW_{in}$ ($\sim (80 \text{ W/m}^2)/(400 \text{ W/m}^2) \times 100\% = 20\%$) is lower than $LW_{in@zenith}$ ($\sim (25 \text{ W/sr/m}^2)/(90 \text{ W/sr/m}^2) \times 100\% = 28\%$) which may increase the difficulty in observing correlations with the retrieved TSL parameters. By considering the parameter $LW_{in@zenith}$, we are able to isolate our findings and explanations on the response of the TSL to that being forced by longwave radiation coming in at zenith.

![Figure 4.14: Heating rates calculated from a 300 K incident ray of radiation at 0 degrees (black line), 30 degrees (blue line) and 60 degrees (red line).](image)

**4.4 Discussion and conclusion.**

In summary, our dataset consists of two cruises in the tropics held during the summer months and through the analysis of night-time data with winds less than 10 m/s, we first confirmed that the turbulent fluxes (LH and SH) are independent of the radiative fluxes and no coupling was found between cloud effects and the turbulent
fluxes, nor was there dependencies observed between \( \text{LW}_{\text{in}} \) and \( \text{LW}_{\text{out}} \). Establishing these independencies are important because it allows us to focus our analysis on the radiative fluxes and supports our hypothesis of the TSL influencing the heat flow at the interface as it tells us the absorbed IR radiation is not immediately returned to the atmosphere through the upward fluxes of LH, SH and \( \text{LW}_{\text{out}} \).

The importance of wind effects on the fluxes and the TSL were also established from analyzing \( Q \) versus \( \text{LW}_{\text{in}} \) and \( \text{LW}_{\text{net}} \) at different \( U_{10} \) intervals (fig. 4.5) and \( \Delta T_{\text{skin}-5m} \) and \( \Delta T_{\text{skin}-0.1mm} \) versus \( U_{10} \) (fig. 4.6). Wind effects need to be considered and removed during our analysis because we are interested in the molecular influence due to the absorption and emission of radiative fluxes on the TSL and not the effect of shear stress from winds. It was found that the analysis needs to be further constrained to data at very low winds of \(< 2 \) m/s in order to observe significant correlations between the properties of the TSL and radiative fluxes. Above this threshold, wind shear erodes the viscous sublayer which in turn causes the TSL to be thinner. The effects of winds on \( \Delta T_{\text{skin}-5m} \) and the thickness of the TSL agrees with previously published literature (e.g. Saunders (1967); Donlon et al. (2002); Minnett et al. (2011)) and it is fortunate that the retrieved TSL profiles are able to capture the subskin temperature at high winds since \( \Delta T_{\text{skin}-0.1mm} \approx \Delta T_{\text{skin}-5m} \). This is because the agreement between \( \Delta T_{\text{skin}-0.1mm} \) and \( \Delta T_{\text{skin}-5m} \) indicates to us that the retrieved \( \Delta T_{\text{skin}-0.1mm} \) are accurate thus removing the possibility of an issue with the retrieved \( \Delta T_{\text{skin}-0.1mm} \) at low winds. Furthermore, this agreement shows proof that the TSL is thinning. Because our retrievals are limited by the emission depth of the EM skin layer thereby restricting the retrievals to the top \( \sim 0.1 \) mm, the disagreement between \( \Delta T_{\text{skin}-0.1mm} \) and \( \Delta T_{\text{skin}-5m} \) at low winds tells us that the subskin temperature has very likely occurred beyond a depth of 0.1 mm. As such, we have introduced a ratio \( \Delta T_{0.1mm}/\Delta T_{5m} \) as an indicator of the fraction of the total temperature change across the TSL that is sampled by the M-AERI profile retrievals.
Under very low winds (< 2 m/s), evidence of the storage of upper ocean heat content is first shown in the less negative LW$_{\text{net}}$ and Q when there is a more positive LW$_{\text{in}}$ (fig. 4.5). Because less upward heat flux is associated with an increased absorption of incoming longwave, there is therefore a surplus of energy generated within the TSL. This additional energy cannot be conducted into the ocean beneath the viscous skin layer as that would require conduction up a temperature gradient. Neither do we expect the surplus energy to entrain from the TSL to the mixed layer through surface renewal events as a mean TSL profile in which the signatures of sea surface renewal events are averaged out is considered in this dissertation.

Instead, the surplus of energy has been shown to adjust the curvature of the TSL (Section 4.3) such that the gradient at the bottom boundary of the TSL adjusts from a higher to lower gradient while at the interface the gradient adjusts from a lower to higher gradient. This was established through the lack in correlations between LW$_{\text{in}}$ with $\Delta T_{\text{skin-5m}}$ and SST$_{\text{skin}}$ (fig. 4.4) which tells us that the absorption of LW$_{\text{in}}$ is independent of $\Delta T_{\text{skin-subskin}}$ and the correlations observed between LW$_{\text{in}@\text{zenith}}$ with $\Delta T_{\text{skin-0.1mm}}$ and $\Delta T_{0.1mm}/\Delta T_{5m}$ which shows a lowered thickness of the TSL.

To put into perspective the impact of the points discussed above, fig. 4.15 illustrates the distribution of heat within the TSL, with and without the presence of clouds. The flux values in fig. 4.15 are the average values obtained from both research cruise. Fig. 4.15 shows that since the outgoing fluxes of LW$_{\text{out}}$, LH and SH does not change with increased LW$_{\text{in}}$, the additional heat from the increased radiative forcing has ‘replaced’ part of the heat flux input from the mixed layer resulting in the contribution of heat from the mixed layer to be lowered and is now supporting the TSL temperature gradient. This decrease is again evident from the correlations observed between LW$_{\text{in}@\text{zenith}}$ with $\Delta T_{\text{skin-0.1mm}}$ and $\Delta T_{0.1mm}/\Delta T_{5m}$ as it shows that the absorption of increased longwave adjusts the TSL such that a lower gradient occurs at subskin depths thereby hindering the heat flow at the bottom boundary of the TSL. The
overall outgoing net flux at the interface which consists of LW\textsubscript{in}-LW\textsubscript{out}-SH-LH is also lowered. Fig. 4.15 shows clearly how less heat is supplied from the mixed layer to the TSL.

![Diagram showing heat balance of TSL with and without clouds]

Figure 4.15: Heat balance of the TSL with and without the presence of clouds.

It is also noted that dependencies were observed with LW\textsubscript{in@zenith} rather than LW\textsubscript{in}. The reason for the independent findings with LW\textsubscript{in} is likely due to the inaccuracy of the PIR as opposed to M-AERI spectral readings as discussed in Section 4.3.2. As the TSL is thicker at low winds, the addition of obliquely impinging radiation on the sea surface has likely obscured the signal we were expecting to observe with the TSL profile. The significant correlations observed with LW\textsubscript{in@zenith} tells us that it is important to assess the distribution of the LW\textsubscript{in} in our analysis given that the changes in the TSL are small.

To conclude, tying in all findings presented in this chapter provides an explanation of the mechanism for retaining upper ocean heat content as the incident infrared radiation increases. The absorption of increased LW has been observed to support the curvature change of the TSL, with a higher gradient forming at the interface and a lower gradient at subskin depths. The lowered gradient at subskin depths impedes
the release of heat stored within the mixed layer beneath the TSL. The energy from the absorption of $\text{LW}_m$ is therefore cycled back into the atmosphere, and because the heat sink at the interface does not change, this means that less heat from the mixed layer contributes to the release of heat at the interface. The energy beneath the TSL, which is due to the absorption of solar radiation during the day, is therefore retained thereby causing an increase in upper ocean heat content.
Chapter 5

Models of the thermal skin layer

In this chapter, an analysis of simulated temperature differences, $\Delta T_{model}$, and viscous layer depths, $\delta_z^{(model)}$, from four different skin layer models is presented. These four models do not take into account the radiative effects within the skin layer and concentrate on the skin layer’s dependences with interfacial heat transfer and surface wind drag. Thus, we refer to these models as viscous layer models and is the sublayer where turbulent velocity fluctuations are suppressed by viscosity. This chapter aims to explore how well the models are able to emulate the observations and whether the dependencies found between the models and surface fluxes support our hypothesis. We expect to observe similar trends between $\Delta T_{model}$ and $\Delta T_{skin-5m}$ given that the models verify their output using $\Delta T_{skin-5m}$ data and any differences to be explained by the lack of radiative effects in the models. Section 5.1 provides a description of the models. Section 5.2 explores the relationships between $\Delta T_{model}$ and $\delta_z^{(model)}$ derived from the models with wind speed. A similar analysis is performed in Section 5.3 but with the surface fluxes, focusing on the radiative fluxes. Finally, we present conclusions of this chapter in Section 5.4.
5.1 Description of published viscous skin layer models.

The four viscous skin layer models that will be explored are based on the heat conduction equation:

\[
\Delta T_{\text{model}} = \frac{Q_0 \delta_z(\text{model})}{\rho_w c_p \kappa}
\]

where \(Q_0\) is the net flux consisting of the sensible, latent and net longwave radiative heat flux from the ocean to the atmosphere, \(\rho_w\) is the density of seawater (1025 Kg/m\(^3\)), \(c_p\) is the specific heat of water (4190 J/Kg K), \(\kappa\) is the thermal diffusivity of water (m\(^2\)/s) and \(\delta_z(\text{model})\) is the thickness of the viscous sublayer in which the TSL is embedded (Chapter 1). The focuses of the models are to obtain appropriate expressions for \(\delta_z(\text{model})\) or \(\Delta T_{\text{model}}\) based on various small-scale theories, namely surface renewal processes and dissipation. The four models are briefly described below and tabulated in table 5.1:

Saunders (1967), (hereinafter Saunders1967) was one of the first to obtain an analytical solution for \(\Delta T_{\text{model}}\). His simple yet elegant solution takes into the account that \(\delta_z(\text{model})\) is determined by viscous stresses and using a dimensional argument along with Fourier’s law of thermal conduction, is able to deduce a solution for \(\Delta T_{\text{model}}\) across the “cool skin” layer. This model is only applicable to higher winds of “perhaps > 2 m/s” where the Richardson number, \(R_i\), (ratio of buoyancy to the shear effects) is small when calculated to a depth of 1 m from the ocean surface. To take into account free convective conditions \((R_i > 1\) which implies a dead calm situation), he referred to studies of heat loss from horizontal surfaces or between parallel horizontal surfaces by free convection and presented an entirely different formulation of \(\Delta T_{\text{model}}\).

Soloviev and Schlüssel (1994), (hereinafter SS1994) utilized surface renewal theory by considering renewal time of fluid elements adjacent to the sea surface. These
fluid elements are constantly replaced by fluid from beneath the viscous layer. The fluid element’s contact time with the interface increases as the roughness length scale increases. SS1994 analyzed 3 different wind speed regimes: buoyancy (low wind speeds), shear (moderate wind speeds), and breaking waves (high wind speeds), and provided a parameterization of the renewal time scales. They also showed that the opposing effects of winds and heat fluxes balance each other out at about wind speeds of 4-5 m/s. For SS1994, we have used an updated set of empirical constants from Soloviev (2007) that are based on new laboratory data and more recent field programs.

Fairall et al. (1996), (hereinafter Fairall1996) is based on Saunder’s model but took one step further by taking into account shear and buoyancy effects. They used Kolmogorov microscales, which describes the smallest velocity length scale in a flow (i.e. the smallest turbulent eddies), to estimate $\delta_z^{(model)}$ by assuming both parameters are proportional to each other. This is performed by scaling the rate of dissipation of turbulent kinetic energy, $\epsilon$, $\left( \delta_z^{(model)} \propto \left( \frac{v_w^3}{\epsilon} \right)^{1/4} \right)$ to 2 regimes: a shear-generated turbulence regime which is proportional to the friction velocity in water, $u_{*w}$, and a convective regime that is proportional to surface cooling thereby including buoyancy effects, $Q_b$. The 2 regimes were combined into a single expression and they concluded that shear and convective effects are comparable when the modified Saunder’s coefficient, $\lambda_2 = 4.8$.

Zhang and Zhang (2012), (hereinafter ZZ2012) is also based on Saunder’s model but made use of Batchelor microscales for their estimation of $\delta_z^{(model)}$, $\left( \delta_z^{(model)} \propto \left( \frac{v_w \kappa^2}{\epsilon} \right)^{1/4} \right)$. Batchelor microscales describe the smallest scalar length of flow and is dependent on the molecular thermal diffusivity of water, $\kappa$, as opposed to Kolmogorov length scales which are dependent on viscosity, $v_w$. Thus, the Batchelor scales are smaller than Kolmogorov scales and indicates the region where diffusion takes place. They derived expressions for the turbulent dissipation rates for 2 regimes: free convection ($R_i > 1$) where buoyancy effects dominate and forced convection ($R_i < 1$) in which
the dissipation rates are determined by the frictional velocity of water, $u_{*w}$.

A summary of the models for $\Delta T$ are given in Table 5.1:

| Saunders (1967) | $\Delta T_{model} = \frac{Q_0 \lambda_1 v_{sw}}{v_{sw} u_{*w}}$ | Saunders coefficient, $\lambda_1 = 7$
|-----------------|--------------------------------------------------|----------------------------------|
| For Ri < 1      | $\Delta T_{model} = \left( \frac{Q_0}{A^\gamma (\kappa v_{sw})} \right)^{\frac{1}{3}}$ | $\tau = \rho_w u_{*w}^2$
|                 |                                                   | Frictional velocity of water, $u_{*w}$ |
|                 |                                                   | $v_{sw}$ Kinematic viscosity of water |
|                 |                                                   | $A = 0.2$ Gravitational acceleration, $g$
|                 |                                                   | $\alpha$ Coefficient of thermal expansion |
|                 |                                                   | $\gamma$ Thermal conductivity of water |
|                 |                                                   | $\kappa$ Thermal diffusivity of water |
|                 | $\Delta T_{model} = \frac{Q_0}{A^\gamma (\kappa v_{sw})}^{\frac{1}{3}}$ |                                                   |
|                 |                                                   |                                                   |
| Soloviev and Schlüssel (1994) | $\Delta T_{model} = \Lambda_0 \sqrt{P_r} \left( 1 + \frac{K_{e_{cr}}}{K_{e_{cr}}} \right)^{0.5} \left( 1 + \frac{Q_0}{u_{*w}^4} \right)$ | Critical surface richardson number, $R_f_{cr} = 0.0213$
|                 |                                                   | Critical Keulegan number, $K_{e_{cr}} = 0.18$
|                 |                                                   | Prandtl number, $P_r = \frac{v_{sw}}{\kappa}$
|                 |                                                   | Keulegan number, $K_e = \frac{u_{*w}^3}{g v_{sw}^2}$
|                 |                                                   | $\Lambda_0 = 7.4$
|                 |                                                   | Surface richardson number, $R_f_0 = \frac{\alpha g v_{sw}^4}{\rho_w c_p u_{*w}^4}$
|                 |                                                   | $LH + SH + LW_{net} + \frac{\beta S_0 c_p}{\alpha L_e} LH$ |
|                 |                                                   |                                                   |
| Fairall et al. (1996) | $\Delta T_{model} = \frac{Q_0 \lambda_2 v_{sw}}{\gamma u_{*w}^4}$, where $\lambda_2 = 6(1 + (\frac{Q_0 2^{4.9} g \rho_w c_p v_{sw}^3}{u_{*w}^4})^{\frac{3}{2}})^{\frac{1}{3}}$ | Virtual surface cooling that includes buoyancy effects of salinity due to evaporation, $Q_b = Q_0 + \left( \frac{0.026 c_e}{\alpha L_e} \right) Q LH$
|                 |                                                   | Latent heat of vaporization, $L_e$ |
\[
\Delta T_{\text{model}} = \frac{Q_0 \delta_z(\text{model})}{\rho_w c_p \kappa_T}, \quad \delta_z(\text{model}) = \left(\frac{v_w \kappa_T^2}{\varepsilon}\right)^{\frac{3}{2}}
\]

Table 5.1: Published models of the temperature difference across the thermal skin layer.

5.2 Dependence of simulated \( \Delta T_{\text{model}} \) and \( \delta_z(\text{model}) \) on wind speed.

This section discusses the dependence of the simulated temperature differences, \( \Delta T_{\text{model}} \), and depth of the viscous layer, \( \delta_z(\text{model}) \), derived from the four models on winds, \( U_{10} \). The observed wind speed and heat fluxes were input into the models to generate \( \Delta T_{\text{model}} \) and \( \delta_z(\text{model}) \). A comparison between the parameterized \( \Delta T_{\text{model}} \) with \( U_{10} \) and the parameterized temperature differences obtained from the field data, \( \Delta T_{\text{skin}-5m} \) and \( \Delta T_{\text{skin}-0.1mm} \) with \( U_{10} \) will be included. We expect to observe similarities between the simulated \( \Delta T_{\text{model}} \) with the measured \( \Delta T_{\text{skin}-5m} \) given that the model outputs were previously verified against available \( \Delta T_{\text{skin}-5m} \) measurements.

5.2.1 \( \Delta T_{\text{model}} \) dependence on wind speed.

Fig. 5.1 shows plots of \( \Delta T_{\text{model}} \) with \( U_{10} \) including the parameterization for \( \Delta T_{\text{skin}-0.1mm} \) and \( \Delta T_{\text{skin}-5m} \) represented by the red and green inverted triangle lines respectively. \( \Delta T_{\text{model}} \) was binned into \( U_{10} \) bin intervals of 1 m/s (magenta stars) and a least squares
fit in the form of $\Delta T = A + Be^{(-U_{10}/C)}$ (similar to Donlon et al.’s (2002) and Minnett et al.’s 2011 expression for their parameterization of $\Delta T_{\text{skin}-5m}$ versus $U_{10}$) is drawn through the mean values (magenta line). All four $\Delta T_{\text{model}}$ behave in an expected manner - showing a sharp decrease with increasing $U_{10}$ at very low winds and approaching an asymptote within the moderate to high wind speed range. Fig. 5.2 and 5.3 shows $\Delta T_{\text{model}}$ versus $\Delta T_{\text{skin}-5m}$ and $\Delta T_{\text{skin}-0.1mm}$ respectively with a solid back line of $y=x$ drawn.

![Graphs showing parameterization of $\Delta T_{\text{model}}$](image)

Figure 5.1: Plot of $\Delta T_{\text{model}}$ (K) with $U_{10}$ (m/s). Inverted triangle red line denotes the parameterization $\Delta T_{\text{skin}-0.1mm} = -0.21 - 0.025 \times e^{(-U_{10}/2)}$. Inverted triangle green line denotes the parameterization $\Delta T_{\text{skin}-5m} = -0.2 - 0.24 \times e^{(-U_{10}/3.5)}$. Magenta stars denotes the mean values of $\Delta T_{\text{model}}$ binned into 1 m/s intervals and the magenta line is the least squares fit to the magenta stars.

Comparing the parameterized curves in fig. 5.1, we first note that ZZ2012 agrees remarkably well throughout the entire wind speed range with the parameterized field data of $\Delta T_{\text{skin}-5m}$. This is not surprising and is very encouraging, given the improvement of instruments over the years for verification of model results with field data, and the continued development of the viscous layer models. Fig 5.2 also shows
Figure 5.2: Plot of $\Delta T_{\text{model}}$ (K) with $\Delta T_{\text{skin-5m}}$ (K). Solid black line denotes $y=x$. Black solid dots represent data binned to every 0.1 K of $\Delta T_{\text{skin-5m}}$ with error bars denoting a 95% confidence interval. Percentage of data above line $y=x$: 54.1%, 70.2%, 96.2% and 41.2% for Saunders1967, Fairall1996, SS1994 and ZZ2012 respectively.

Figure 5.3: Plot of $\Delta T_{\text{model}}$ (K) with $\Delta T_{\text{skin-0.1mm}}$ (K). Solid black line denotes $y=x$. Black solid dots represent data binned to every 0.1 K of $\Delta T_{\text{skin-0.1mm}}$ with error bars denoting a 95% confidence interval. Percentage of data above line $y=x$: 32.3%, 39.4%, 69.5% and 27.8% for Saunders1967, Fairall1996, SS1994 and ZZ2012 respectively.
ZZ2012 to be one of the models which best matches the observed $\Delta T_{\text{skin}-5m}$ where 41.2% of the scatter lies above the line $y=x$ and the errorbars show the scatter to be most ‘contained’ about the line $y=x$.

Saunders1967 and Fairall1996 both show good agreement from moderate to high wind speeds when comparing the parameterized curves in fig. 5.1. The difference between Saunders1967 and Fairall1996 occurs at low winds of $\sim<2$ m/s. The large deviation observed at low winds with Saunders1967 is not representative of $\Delta T_{\text{skin}-5m}$ nor $\Delta T_{\text{skin}-0.1mm}$ under very light winds ($<2$ m/s) because his model focuses on the derivation of $\Delta T_{\text{model}}$ from moderate to higher winds. Fairall1996 who introduced the modified Saunders coefficient to rectify the calculated $\Delta T_{\text{model}}$ at very low winds from the Saunders1967 model shows an improvement in the calculated $\Delta T_{\text{model}}$ but unfortunately we still observe a deviation from the parameterized $\Delta T_{\text{skin}-5m}$. It should be noted that the Saunders (or modified Saunders) coefficient is subjective given that it is based on statistical evaluation with $\Delta T_{\text{skin}-5m}$ measurements on field data under different environmental conditions (e.g. Fairall1996 uses data obtained from the Tropical Ocean-Global Atmosphere (TOGA) Coupled Ocean-Atmosphere Response Experiment (COARE) Webster and Lukas (1992)), thus it may likely be that the modified Saunders coefficient used for the NAURU99 and AMMA06 field data is not suitable thereby resulting in the discrepancy seen.

It is interesting to note that the parameterized graphs of Fairall1996 and $\Delta T_{\text{skin}-0.1mm}$ show good agreement. However, as discussed in Chapter 4, $\Delta T_{\text{skin}-0.1mm}$ at low winds may not be representative of the subskin temperature therefore the agreement with Fairall1996 is not convincing. Furthermore, fig 5.3 shows that all the models (including Fairall1996) do not agree well with $\Delta T_{\text{skin}-0.1mm}$. There have been known issues with Fairall1996’s model. For example, Fairall1996’s model showed smaller $\text{SST}_{\text{skin}}$ values (fig. 10 of Fairall et al. (1996)) after correcting for the cool-skin effect when they compared the modelled results to an IR radiometer and thermosalinograph reading.
at 2 m depth, which should not be the case. Yet, on another set of IR radiometer skin SST and thermosalinograph data onboard a different ship, they observe that their corrected SST\textsubscript{skin} values are higher than the measured data thereby implying flaws within their model. Despite this, their model spurred the development of better viscous sublayer models such as ZZ2012.

Comparing the parameterized curves of SS1994 with $\Delta T_{\text{skin}-5m}$, good agreement is observed, however the SS1994 values are shown to be $\sim 0.2$ K smaller than the measured $\Delta T_{\text{skin}-5m}$. This is the only model analyzed which uses surface renewal theories to calculate $\Delta T_{\text{model}}$ and is the only model which shows lower values of $\Delta T_{\text{model}}$ as compared to $\Delta T_{\text{skin}-5m}$ and $\Delta T_{\text{skin}-0.1\text{mm}}$. Thus, the probable reason for the $\sim 0.2$ K offset is that the surface renewal theories are underestimating the temperature difference. Although one may argue that it is not ideal to perform comparisons with $\Delta T_{\text{skin}-5m}$ when analyzing the thermal viscous sublayer because between the subskin depth and a depth of 5 m, despite good convective mixing, there are many other contributions which may affect the viscous sublayer such as the logarithmic boundary layer. However, it is important to keep in mind that the $\Delta T_{\text{skin}-5m}$ and $\Delta T_{\text{skin}-0.1\text{mm}}$ are averaged spatially and temporally as opposed to an instantaneous measurement. Thus, we do not expect to resolve sea surface renewal events in the measured $\Delta T_{\text{skin}-5m}$ and $\Delta T_{\text{skin}-0.1\text{mm}}$ values.

Fig. 5.2 also shows that majority of the scatter lies above the line $y=x$ which implies that most of the calculated $\Delta T_{\text{model}}$ are underestimating $\Delta T_{\text{skin}-5m}$. One potential and important issue which may result in this underestimation has been addressed by Kent et al. (1996) who performed a comparison between four different skin layer models, of which two are the Saunders1967 and SS1994 model, using their cruise data from the subtropical Atlantic Ocean during June 1992. Their readings consisted of day and night-time data of a radiometer operating at 11 $\mu$m and a thermistor dragged from the ship at $\sim 0.1$ m to 1 m depth. Kent et al. (1996) observed similar smaller calculated
\( \Delta T \) values than the measured \( \Delta T \), with \( \sim 0.1 \) K offset with the SS1994 model while better agreement was observed with the Saunders1967 model. They stated that using the ratio of the densities of water to air to estimate the friction velocity of water, \( u_{\text{w}}^* \), from the friction velocity of air, \( u_{\text{a}}^* \) may be an issue. This is because all the energy from wind stresses in the air are assumed to be transferred to the ocean surface layer to generate turbulence and the existence of other terms such as horizontal advection flows or form drag are ignored. If these effects are included, the estimated value of \( u_{\text{w}}^* \) will be reduced and result in a higher \( \Delta T_{\text{model}} \) since \( \Delta T_{\text{model}} \propto (u_{\text{w}}^*)^{-1} \) thereby allowing the values of \( \Delta T_{\text{model}} \) to be more realistic compared to \( \Delta T_{\text{skin} - 0.1 \text{mm}} \) and \( \Delta T_{\text{skin} - 5 \text{m}} \). Thus, the models themselves may require further adjustments through the inclusion of physics of small-scale physical phenomena which are not included.

Kent et al. (1996) also showed that at low winds of \(< 2 \) m/s, none of the models agreed with their data. The performance of such viscous sublayer models has also been known to be less than ideal in their calculation of \( \Delta T \) at low wind speeds (Soloviev, 2007). It is therefore not ideal to analyze \( \Delta T_{\text{model}} \) to answer to the objective of this thesis which is to evaluate the effects of radiative flux on heat transfer at the air-sea interface. This is because as established in Chapter 4, in order to isolate the heat transfer mechanisms at the interface to radiative fluxes, it is best to analyze data at very low winds of \(< 2 \) m/s. Nevertheless, it is still useful to compare observations with the calculated modelled results which would aid in further improvements of the viscous layer models.

Overall, the relationship observed between \( \Delta T_{\text{model}} \) with \( U_{10} \) is expected as the models summarized in Table 5.1 all exhibit an inverse dependency between \( \Delta T_{\text{model}} \) with the friction velocity of water, \( u_{\text{w}}^* \). The observed trend of \( \Delta T_{\text{model}} \) with \( U_{10} \) also agrees with that observed from the parameterization of \( \Delta T_{\text{skin} - 5 \text{m}} \) of our field data (inverted triangle green line).
Kent et al. (1996) conclude that a Saunders type model, but with larger values of the Saunders coefficient, is useful. Similarly, our comparison with the calculated and measured temperature differences with winds shows that there is better agreement between the Saunders based models, Saunders1967, Fairall1996 and ZZ2012 with $\Delta T_{\text{skin-5m}}$. Unfortunately, none of the models are able to replicate the temperature differences observed at low winds which is the focus of this thesis. ZZ2012 showed the best results with respect to $\Delta T_{\text{model}}$ at low winds, however there is still a large amount of scatter observed. This tells us that continued improvement is much needed in modelling $\Delta T_{\text{model}}$ of the viscous sublayer at low winds. To do so would require deeper insight into the basis of each model’s underlying physics and a larger dataset for verification. We must also be more rigorous in the definition of the depth of the measurements. Despite this, it is still very encouraging to note the very similar trend between $\Delta T_{\text{skin-5m}}$ and $\Delta T_{\text{model}}$, which suggests to us that the theories are valid but may simply require modification of some of the parameterized coefficients in the models.

5.2.2 Dependence of $\delta_z(\text{model})$ on wind speed.

Fig 5.4 shows a plot of the modelled TSL thickness, $\delta_z(\text{model})$, with $U_{10}$. The models show the expected thicker viscous sublayer at low winds, decreasing in thickness with increasing $U_{10}$ until it approaches an asymptote value of $\sim 0.25$ mm. Note that the asymptote value of $\delta_z(\text{model})$ is larger than the deepest emission depth of the EM skin layer of $\sim 0.1$ mm (bottom plot of fig. 5.4). This shows that the theories which define the viscous sublayer confirms that the TSL and EM skin layer is embedded within the viscous sublayer as mentioned in Chapter 1.

From moderate to high winds ($> 2$ m/s), $\delta_z(\text{model})$ is observed to agree well between the 3 models (Saunders1967, Fairall1996 and ZZ2012) while SS1994 shows a thinner layer. Comparison with $\Delta T_{0.1mm}/\Delta T_{5m}$ in fig. 5.5 simply shows the expected thinning
of the viscous sublayer and TSL. This expected thinning of the viscous sublayer is due to the increase in turbulence beneath the ocean surface as the surface wind stress erodes the viscous sublayers to its minimum depth scale where the turbulent eddies are constrained and viscosity dominates.

ZZ2012 also stated that their use of the Batchelor micro-scale implies that \( \delta_z(\text{model}) \propto \sqrt{v_w} \) as compared to \( \delta_z(\text{model}) \propto v_w^{1/3} \) for Fairall1996, where \( v_w \) is the kinematic viscosity of water. This explains the larger values compared to those of Fairall1996 and Saunders1967. The thinnest \( \delta_z(\text{model}) \) are calculated using the SS1994 model and shows that the use of surface renewal theories would enable us to resolve much thinner layers as compared to the use of small-scale theories such as Kolmogorov or Batchelor scales.

At low wind speeds (< 2 m/s), the increase in thickness at low winds agrees with our observations - discrepancies between \( \Delta T_{\text{skin}-0.1mm} \) and \( \Delta T_{\text{skin}-5m} \) at low winds was attributed to an increase in thickness of the TSL such that more of the thermal structure of the TSL lies below the IR emission depth resulting in \( \Delta T_{\text{skin}-0.1mm} < \Delta T_{\text{skin}-5m} \). Fig. 5.6 further illustrates this agreement, showing that the thicker \( \delta_z(\text{model}) \) corresponds to all \( \Delta T_{0.1mm}/\Delta T_{5m} < 1 \) at \( U_{10} < 2 \) m/s and as \( U_{10} \) increases, the decrease in \( \delta_z(\text{model}) \) is observed to correspond to \( \Delta T_{0.1mm}/\Delta T_{5m} \) centering about 1.

Unfortunately, we are unable to establish a relationship between the ratio \( \Delta T_{0.1mm}/\Delta T_{5m} \) with \( \delta_z(\text{model}) \) due to the high scatter observed in fig. 5.6. Furthermore, we do not have in-situ depth measurements of the TSL to verify both \( \Delta T_{0.1mm}/\Delta T_{5m} \) and \( \delta_z(\text{model}) \). It is therefore difficult to justify the validity of the theories the models have on the TSL depth as it is unknown which model best represent the ‘true’ TSL depth. The only conclusive conclusion is that both models and observations show a thickening of the viscous sublayer as winds decrease until a point of very low winds (\( \sim 2 \) m/s) where the viscous sublayer depth is observed to asymptote or approach a maximum
value. It is interesting to note that the plot of $\Delta T_{0.1mm}/\Delta T_{5m}$ versus wind speed (fig. 4.8) shows a similar asymptotic effect at low winds as observed in fig. 5.4 with SS1994 and Fairall1996. This implies an increase in the temperature gradient of the viscous sublayer ($\Delta T_{model}/\delta z_{(model)}$) since $\Delta T_{model}$ and $\Delta T_{skin-5m}$ were observed to increase (figs. 5.1, 4.6) at low winds. The increase in the gradient ($\Delta T_{model}/\delta z_{(model)}$) of the viscous sublayer as $U_{10}$ decreases below 2 m/s confirms that the extraction of heat from the ocean to the atmosphere is increasingly governed by molecular conduction as opposed to convective transfer.

Figure 5.4: Top: Plot of $\delta z_{(model)}$ (cm) with $U_{10}$ (m/s). Bottom: Magnification of the top plot of $\delta z_{(model)}$ from 0 to 0.4 cm. Red dotted line denotes $\delta z_{(model)} = 0.025$ cm.
Figure 5.5: Plot of $\delta_{z_{(model)}}$ (cm) with $\Delta T_{0.1mm}/\Delta T_{5m}$ with the data binned to 1 m/s $U_{10}$ bins. The colorbar denotes $U_{10}$.

Figure 5.6: Plot of $\delta_{z_{(model)}}$ (cm) with $\Delta T_{0.1mm}/\Delta T_{5m}$. The colorbar denotes $U_{10}$. 
5.3 Thermal skin layer model prediction \((\Delta T_{\text{model}}\) and \(\delta z(\text{model})\)) dependence on observed interfacial fluxes.

This section provides an analysis of \(\Delta T_{\text{model}}\) and \(\delta z(\text{model})\) with the observed interfacial radiative fluxes. Discussions of the effects of turbulent fluxes will be limited as the relationship between latent heat and \(\Delta T_{\text{model}}\) was observed to be very similar to that observed with \(Q\) while the sensible heat dependences are inconclusive, perhaps as the sensible heat range is very small \((< 10 \text{ W/m}^2)\).

5.3.1 Dependence between \(\Delta T_{\text{model}}\) with observed surface radiative fluxes.

Analyzing the dependency of \(\Delta T_{\text{model}}\) with \(Q\) (fig. 5.7), we once again note the relationship between \(U_{10}\) and \(Q\) - higher winds correspond to higher net fluxes and vice versa. This also corresponds to lower \(\Delta T_{\text{model}}\) values. At higher \(U_{10}\) and \(Q\), the increase in turbulence due to winds results in the erosion of the viscous sublayer from below, thus we observe a smaller and relatively constant \(\Delta T_{\text{model}}\).

Another interesting observation in fig. 5.7 is the gradient of the individual color bands, which denote \(U_{10}\) values. Taking into account a wind speed interval of, for example 0-2 m/s (fig. 5.8), which is represented by the dark blue points, \(Q\) is observed to decrease with \(\Delta T_{\text{model}}\). In other words, by considering the regime of almost non-existent shear effects which results in an undisturbed ocean surface, a clear direct correlation is observed between \(Q\) and \(\Delta T_{\text{model}}\). Fig. 5.8 clearly illustrates this relationship between \(\Delta T_{\text{model}}\) and \(Q\) at \(U_{10} < 2 \text{ m/s}\). With the exception of Saunders1967, the proportionate relationship between \(Q\) and \(\Delta T_{\text{model}}\) is expected as all models implicitly show \(\Delta T_{\text{model}} \propto Q\) if wind effects are ignored.
Figs. 5.7 and 5.8 therefore show clearly the opposing effect $Q$ and $U_{10}$ has on $\Delta T_{model}$ as higher $U_{10}$ results in a lower $\Delta T_{model}$ and a higher $Q$ results in a higher $\Delta T_{model}$ (when $U_{10}$ effects are minimized). This reconfirms the need to segregate our data into wind speed intervals so that our analysis can be focused on the molecular effects and the TSL.

Figure 5.7: $\Delta T_{model}$ (K) with net flux, $Q$ (W/m$^2$). The colorbar denotes $U_{10}$ (m/s). Filled circles represent data from NAURU99. Hollow circles represent data from AMMA06.
Figure 5.8: Plot of $\Delta T_{\text{model}}$ (K) with net flux, $Q$ (W/m$^2$) at $U_{10} < 2$ m/s. $R^2$ values calculated for models Saunders1967, Fairall1996, SS1994 and ZZ2012 are 0.013, 0.99, 0.92 and 0.4 respectively.

Continuing the analysis of the effects between the radiative fluxes and $\Delta T_{\text{model}}$, we analyze $LW_{\text{in}}$, $LW_{\text{net}}$ and $LW_{\text{in}@\text{zenith}}$ with $\Delta T_{\text{model}}$. Once again, we do not expect to observe any dependences between $\Delta T_{\text{model}}$ with the radiative fluxes from moderate to high winds because the turbulent wind shear stresses exceed the IR effects on $\Delta T_{\text{model}}$. Appendix B illustrates this point with plots of $LW_{\text{in}}$, $LW_{\text{net}}$ and $LW_{\text{in}@\text{zenith}}$ with $\Delta T_{\text{model}}$ for all winds. Thus, we focus our analysis on data at $U_{10} < 2$ m/s (figs. 5.9, 5.10, 5.11). Fig. 5.12 shows the plot between the turbulent fluxes and $\Delta T_{\text{model}}$ for $U_{10} < 2$ m/s.
Figure 5.9: Plot of $\Delta T_{\text{model}}$ (K) with $LW_{\text{in}}$ (W/m$^2$) at $U_{10} < 2$ m/s. R$^2$ values calculated for models Saunders1967, Fairall1996, SS1994 and ZZ2012 are 0.13, 0.74, 0.76 and 0.54 respectively.

Figure 5.10: Plot of $\Delta T_{\text{model}}$ (K) with $LW_{\text{net}}$ (W/m$^2$) at $U_{10} < 2$ m/s. R$^2$ values calculated for models Saunders1967, Fairall1996, SS1994 and ZZ2012 are 0.09, 0.97, 0.98 and 0.60 respectively.
Figure 5.11: Plot of $\Delta T_{\text{model}}$ (K) with LW$_{\text{in@zenith}}$ (W/sr/m$^2$) at $U_{10} < 2$ m/s. $R^2$ values calculated for models Saunders1967, Fairall1996, SS1994 and ZZ2012 are 0.09, 0.84, 0.84 and 0.52 respectively.

Figure 5.12: Plot of $\Delta T_{\text{model}}$ (K) with LH+SH (W/m$^2$) at $U_{10} < 2$ m/s. $R^2$ values calculated for models Saunders1967, Fairall1996, SS1994 and ZZ2012 are 0.49, 0.06, 0.0064 and 0.18 respectively.

Significant correlations are observed between $\Delta T_{\text{model}}$ and the radiative fluxes for SS1994, Fairall1996 and ZZ2012 models under conditions of $U_{10} < 2$ m/s. Saunders1969
does not show significant correlations which will be dismissed as it has been established that his model is not a good representation of the viscous sublayer under very low wind speed conditions and the focus of his paper was to formulate $\Delta T_{\text{model}}$ of the TSL from moderate to higher winds. Contrary to the findings presented in Chapter 4 where a direct correlation was only established between $\Delta T_{\text{skin} - 0.1 \text{mm}}$ and $LW_{\text{in} @ \text{zenith}}$, and we concluded that the temperature difference between the skin and subskin layers, (i.e. $\Delta T_{\text{skin} - 5 \text{m}}$) are independent of longwave radiation, the models show $\Delta T_{\text{model}}$ to decrease with an increase in IR flux. Although this result does not agree with the field data analysis, the negative correlation established from figs. 5.9 to 5.11 is expected given that strong negative correlations of $Q$ with $LW_{\text{in}}$ (fig. 4.5) and positive correlations of $Q$ with $\Delta T_{\text{model}}$ (fig. 5.7) were observed. These results also agree with our initial expectations of $\Delta T_{\text{model}}$’s response to longwave as more IR is absorbed closer to the interface than at depth thereby raising the temperature of the sea surface more so than at the subskin depth layers resulting in a smaller temperature difference within the TSL. The difference in the results obtained from the models and observations is very likely due to the absence of radiative laws in the models as the models simply include the effects of radiative fluxes in the total heat flux parameter without considering the exponential decay of the absorbed or emitted radiative fluxes. The findings from the analysis of $\Delta T_{\text{model}}$ with longwave therefore tells us that under free convection and at high molecular heat conduction rates, even with the absence of radiative effects, the theories governing the models do indeed show that the viscous sublayer’s temperature gradient is dependent on the radiative fluxes. The dependency is such that a smaller viscous sublayer gradient is formed as the incoming radiative flux increase and also indicates the net flux to decrease thereby suggesting retaining of heat flux in the bulk of the ocean. However, this result does not explain the non-linear response of the TSL with the radiative fluxes which was otherwise explained through the field results.
The independency between $\Delta T_{model}$ and the turbulent fluxes (fig. 5.12), on the other hand, agrees with the field data correlations established between $\Delta T_{skin-5m}$ and $\Delta T_{skin-subskin}$ with the turbulent fluxes. This independence is important as it isolates the heat exchange mechanism between the viscous sublayer with the atmosphere and ocean’s mixed layer to be solely due to radiative processes under minimal wind shear influences.

The results from figs 5.9 to 5.12 therefore show that the decrease in the viscous sublayer’s temperature gradient established through the analysis of the calculated $\Delta T_{model}$ implies less heat is transferred to the atmosphere through conduction, while increasing heat transfer through molecular mechanisms which requires a temperature gradient.

5.3.2 Dependence of $\delta_z(model)$ on observed radiative surface fluxes.

In this section, $\delta_z(model)$ will be analyzed with the radiative fluxes ($LW_{in}$, $LW_{net}$ and $LW_{in@zenith}$) and compared with observational results. Appendix B shows plots of $\delta_z(model)$ against $LW_{in}$, $LW_{net}$ and $LW_{in@zenith}$ for all wind speeds and does not exhibit any correlation. This is expected as established previously, and our focus is on low winds ($U_{10} < 2$ m/s) given the wind speed dependence observed with the surface fluxes. Plots of $\delta_z(model)$ against the radiative fluxes for $U_{10} < 2$ m/s are shown in figs. 5.13, 5.14 and 5.15 with the calculated regressed coefficients given in table 5.2.
Figure 5.13: Plot of $\delta_z$ (model) (cm) versus $LW_{in}$ (W m$^{-2}$). Black dotted line represents the regressed line. $R^2$ values are provided in table 5.2.

Figure 5.14: Plot of $\delta_z$ (model) (cm) versus $LW_{net}$ (W m$^{-2}$). Black dotted line represents the regressed line. $R^2$ values are provided in table 5.2.
Figure 5.15: Plot of $\delta_z^{\text{(model)}}$ (cm) versus LW_{in@zenith} (W sr$^{-1}$ m$^{-2}$). Black dotted line represents the regressed line. $R^2$ values are provided in table 5.2.

Table 5.2: Table of $R^2$ values calculated from figs. 5.13, 5.14 and 5.15.

<table>
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<td>0.64</td>
<td>0.32</td>
<td>0.07</td>
</tr>
</tbody>
</table>

From figs 5.13, 5.14, 5.15 and table 5.2, only Fairall1996 shows significant correlations with the incident radiative flux. Saunders1967, SS1994 and ZZ2012 do not show relationships between the radiative flux and calculated model depth. As mentioned previously, there are no data to verify the accuracy of $\delta_z^{\text{(model)}}$; therefore it is unclear which model, if any, is a good representation of the viscous sublayer depth. The regressed lines however do show the correlations to be opposite to what was observed in Chapter 4. In Chapter 4, it was established that as LW_{in@zenith} increases and LW_{net} decreases, $\Delta T_{0.1mm}/\Delta T_{5m}$ increases which implies there is a concentration of the vertical temperature gradient near the interface. Figs. 5.13, 5.14 and 5.15 imply $\delta_z^{\text{(model)}}$
to thicken, rather than to thin with respect to increasing longwave incident radiation. In theory, a thicker layer viscous sublayer is expected with increasing longwave as the erosion of the viscous sublayer due to turbulence subsides. Recalling that the TSL is embedded within the viscous sublayer as described in Chapter 1, the increase in \( \delta_z(\text{model}) \) combined with the observed result with \( \Delta T_{0.1\text{mm}}/\Delta T_{5\text{m}} \) which indicates more of the TSL’s temperature profile lies near the interface, therefore suggests that the increase in incoming longwave would result in the lower boundary of the TSL to increasingly be much smaller from the viscous sublayer’s depth. It is noted that the derived TSL profiles together with the models considers a mean vertical temperature profile averaged through \( \sim 1 \text{ m}^2 \) and \( \sim 1 \text{ min} \), therefore this analysis does not take into account instantaneous events such as potential sea surface renewal events on the TSL or viscous sublayer. The deviation between the thickness of the TSL and viscous sublayer would thus result in heat within the mixed layer being less effectively transported through the viscous sublayer, down the TSL temperature gradient.

5.4 Discussion and conclusion.

To summarize, the relationships established and observed between the derived \( \Delta T_{\text{model}} \) and \( \delta_z(\text{model}) \) with winds and fluxes are expected. For \( \Delta T_{\text{model}} \), its decrease as winds increase agrees well with the field data from moderate to high winds except for a slight underestimation of \( \Delta T_{\text{skin}-5\text{m}} \) which is a known issue as found by Kent et al. (1996). The issue was attributed to the inaccuracy of the estimation of the friction velocity of water from that of air as this assumes all the energy from wind stresses in the air are transferred to the ocean surface. At low winds of < 2 m/s, none of the models are able to represent \( \Delta T_{\text{skin}-5\text{m}} \) adequately. This is not surprising as Kent et al. (1996) and Soloviev (2007) have noted similar results under low wind conditions. This indicates that the theories governing the models are currently unable to capture the
molecular effects of the viscous sublayer. This may be due to the absence of radiative absorption and emission in the models, and which play an important role in defining the viscous sublayer’s temperature gradient at low winds. Because the models are unable to capture the field data at low winds, yet as shown in Chapter 4 our analysis in this thesis needs to be performed at winds < 2 m/s to meet the dissertation’s objectives, the subsequent analyses of $\Delta T_{\text{model}}$ with fluxes does not necessarily explain the response of the TSL to longwave fluxes. Despite this, this chapter proceeds to explore the relationships between $\Delta T_{\text{model}}$ and $\delta_z(\text{model})$ with longwave fluxes at low winds, keeping in mind that the models are representing the linear temperature profile of the viscous sublayer as opposed to the desired non-linear temperature profile of the TSL.

At winds < 2 m/s, $\Delta T_{\text{model}}$ is observed to decrease and $\delta_z(\text{model})$ is observed to increase with increasing LW$_{in}$. Given that the models explicitly shows $\Delta T_{\text{model}}$ to be directly correlated with $Q$ (fig. 5.8), $\delta_z(\text{model})$ to be inversely correlated to $Q$ and results in Chapter 4 show $Q$ to be negatively correlated to LW$_{in}$ at $U_{10} < 2$ m/s (fig. 4.5), these relationships are therefore expected. Although these results do not agree with field observations where $\Delta T_{\text{skin}-5m}$ was observed to be independent of LW$_{in} @ \text{zenith}$ and $\Delta T_{0,1mm}/\Delta T_{5m}$ was observed to increase with LW$_{in} @ \text{zenith}$ (indicating an increase of the vertical temperature gradient in the TSL near the interface), the results between the models and field observations may be combined to explain the warming of the bulk of the ocean. As there is more incoming longwave radiation, the curvature of the TSL is modified such that more of the temperature gradient being concentrated closer to the surface while the viscous sublayer is observed to thicken with a decrease in its temperature gradient. The result of these gradient changes hinders the flow of heat from the mixed layer into the viscous sublayer as the rate of conduction is lowered by the lower viscous sublayer’s temperature gradient. Furthermore, the deviation of the bottom boundary of the TSL from that of the viscous sublayer makes it even more
difficult for the heat to traverse through the TSL and escape into the atmosphere.

The models only consider the importance between wind shear and buoyancy effects on the viscous sublayer and are therefore unable to resolve the nonlinear radiative effects observed in the TSL. Despite this, we observe the ZZ2012 model to best represent the observations based on comparing $\Delta T_{\text{model}}$ with $\Delta T_{\text{skin} - 5m}$ from figs. 5.1 and 5.2 as majority of the scatter plot is observed to be centered about the line $y = x$. In terms of the depth of the viscous layer, $\delta z_{(\text{model})}$, none of the models are observed to agree well with the ratio $\Delta T_{0.1mm} / \Delta T_{5m}$ (fig. 5.6). We are therefore unable to establish any relationship between these two variables. Furthermore, as there are no in-situ depth measurements available, it is difficult to assess the validity of $\delta z_{(\text{model})}$.

The difference between the field observations and model results therefore highlights the importance of including radiative effects in the models under low wind speed conditions. Without the inclusion of the radiative effects, the models are unable to describe and elucidate the influence of the increase in radiation on the TSL.
Chapter 6

Conclusions and future work

6.1 Conclusions.

The increasing trends of the warming of upper ocean heat content (OHC) associated with an increase in greenhouse gases is well known, with much supporting evidence in the literature (e.g. Levitus et al. (2012); IPCC (2014)). However, no known explanation of how this increase in longwave radiation on the sea surface causes the rise in upper OHC has yet been established. This work presents a first and one of its kind theory in bridging the two observed trends - a hypothesis for the indirect heating of the ocean is presented whereby the vertical temperature gradient of the mean TSL is modified by the absorption of IR from the atmosphere which results in heat below the skin layers being retained. This work also shows that the TSL profile, otherwise often overlooked due its small scale, plays a crucial role in heat exchanges at the air-sea interface.

The main finding presented in this dissertation is that the absorption of the increased longwave radiation alters the TSL such that the temperature gradient at the interface increases while the temperature gradient at the bottom boundary of the TSL (between the TSL and mixed layer) decreases. This gradient change in the TSL
is supported by the additional energy obtained from the absorption of the increase in incoming longwave radiation and hinders the release of heat from below the TSL back into the atmosphere. Therefore, heat beneath the TSL which is largely a product of the absorption of solar radiation during the day remains. This theory was established and justified through meeting the four objectives presented in Chapter 1, each of which will be summarized in the paragraphs below.

First, in Sections 3.1 and 3.2 a technique which utilizes IR spectra from a passive shipboard radiometer to obtain the vertical temperature profile of the TSL is presented. The ability to obtain the TSL profile has opened up the possibility of analyzing temperature structures at submillimeter scales. However, as with all instrumentation, there exist limitations which hinder scientific progress and must be worked around. The major limitation of this technique which affects our objectives is that the penetration/emission depth of the IR frequencies is limited to the top 0.1 mm below the water surface. This means that if the TSL extends deeper than 0.1 mm, the retrievals would not be able to capture the entire TSL profile. This issue was evident in Section 4.2 where deviations between $\Delta T_{\text{skin}-5\,\text{m}}$ and $\Delta T_{\text{skin}-0.1\,\text{mm}}$ occurred at low winds. To go about this issue, a ratio of the temperature differences, $\Delta T_{0.1\,\text{mm}}/\Delta T_{5\,\text{m}}$, was therefore used to indicate the fraction of the TSL temperature difference captured in the thickness of the electromagnetic skin layer.

Secondly, as it is not possible to obtain sufficiently long time-scale TSL retrievals using the M-AERI for the analysis of increasing levels of GHG’s, the radiative effect of passing clouds has been used as a surrogate in this dissertation. Even if M-AERI data are available over a few decades, the $\sim 4$ W/m$^2$ signal produced from doubling CO$_2$ is small which increases the difficulty in addressing the problem. The pros and cons of using cloud cover has been discussed in Section 3.4. In summary, cloud forcing produces a much higher radiative signal over short time-scales which allows the problem to be more tractable. The main concern is the different spectral compositions
between GHG and cloud forcings. This was addressed through comparing the heating rates with depth for the various GHG’s, cloudy and clear sky spectra and was shown to work towards our advantage as the heating rate signal was shown to be similar but amplified with cloud forcing. The conclusion is that cloud forcing is an adequate tool within our area of study in addressing our main objective of understanding consequences of increasing concentrations of GHG’s.

Spectral differences between the derived flux from M-AERI spectra and the PIR readings were also observed and addressed in Section 4.3.2. The derived flux from M-AERI spectra, $LW_{in@zenith}$, is a more accurate indication of the amount of flux impinging on the spot of the sea surface where the TSL retrievals were obtained. This is because of the lack of accuracy in the PIR’s measurements which may have masked any signals which we had hoped to observe between the PIR readings and the TSL properties.

In response to the third objective, which is also the main aim of this dissertation, there are some important relationships derived between the surface fluxes and TSL. Firstly, the turbulent fluxes are independent of changes in the radiative fluxes which shows the processes involving these two types of fluxes do not immediately and directly affect one another. Secondly, the incoming and outgoing longwave fluxes are also found to be independent. This tells us that the absorbed IR radiation is not immediately returned to the atmosphere through LH, SH and LW$_{out}$ thus there is a heat sink at the air-sea interface that does not respond to changes in LW$_{in}$. Thirdly, wind effects must be taken into consideration and we chose to analyze data at very low winds ($< 2 \text{ m/s}$) in order to reveal the signals resulting from the absorption and emission of radiative fluxes. The retrieved $\Delta T_{\text{skin-0.1mm}}$ is limited by the emission depth of the EM skin layer and thus there was a need to use the ratio $\Delta T_{0.1mm}/\Delta T_{5m}$ as an indicator to the depth of the TSL.
Storage of upper OHC was shown to occur through firstly the lowering of LW_{net} with an increase in LW_{in} at low winds, signifying that less heat is removed from the ocean as LW_{in} increases. Next, the independence observed between $\Delta T_{\text{skin}-5m}$ and LW_{in} and positive correlations between $LW_{in@\text{zenith}}$ with $\Delta T_{\text{skin}-0.1mm}$ and $\Delta T_{0.1mm}/\Delta T_{5m}$ tells us that the absorption of longwave radiation adjusts the TSL such that a lower gradient occurs at subskin depths while a higher gradient occurs at the aqueous side of the air-sea interface. This supports the hypothesis that an increase in IR flux results in the storage of heat in the bulk of the ocean as the lowered temperature gradient at subskin depths means the flow of heat from the mixed layer to the TSL decreases. Instead, the additional energy from the absorbed IR is used to support the curvature change of the TSL to maintain the surface heat loss. As a result, less heat from the mixed layer originating from the absorption of solar radiation in the visible part of the electromagnetic spectrum during the day is supplied to the release of heat at the interface.

In Chapter 5, 4 different published viscous sublayer models were analyzed and the results were compared against the field observations. It was concluded that none of the models are able to adequately simulate the variable $\Delta T_{\text{skin}-5m}$ under very low wind conditions of $< 2$ m/s. The main reason that the models have failed to capture this temperature difference is because the focus of the models is on wind-driven shear and buoyancy, which works well from moderate to higher wind speed conditions. However, at very low winds, radiative effects dominates and the absence of these absorption and emissive radiative effects has therefore resulted in the models’ inability to replicate $\Delta T_{\text{skin}-5m}$. Different results were obtained from analyzing $\Delta T_{\text{model}}$ and LW_{in} as compared to the field observations - the increase in LW_{in} was found to decrease $\Delta T_{\text{model}}$ as opposed to the independence found in the field observations. Despite the disagreement, the $\Delta T_{\text{model}}$ responses to LW_{in} were expected and by keeping in mind that the models are representative of the temperature gradient of the viscous
In this section we discuss some future work which will further improve our understanding of the physics of the interactions between the TSL and radiative fluxes. There are a few noteworthy issues which have hindered our progress in attaining desired results.
The first is the issue of the IR penetration depth which does not extend beyond 0.1 mm and has resulted in the retrieved temperature values to be restricted to the top 0.1 mm. The physics of the IR absorption spectra in water cannot be altered and has to be worked around. It would be desirable to obtain in-situ temperature readings at sub-millimeter scales which would complement the temperature retrievals obtained from the M-AERI. Micro temperature profilers such as the Skin Depth Experimenal Profiler (SkinDeEP) instrument developed by Ward et al. (2004) may be used for such in-situ measurements. Performing similar analyses in this dissertation with M-AERI retrievals and SkinDeEP readings would not only provide more insight to TSL variations with the surface fluxes, but also help in verification of the TSL retrievals.

This brings us to the next proposed work which aims to help further support the theory established in this dissertation: It would be ideal to perform a controlled experiment as was described by Hanafin (2002) where the M-AERI measurements were interpreted in terms of a linear temperature profile of the surface from a pool of water with controlled temperatures and surface heat fluxes. A controlled experiment would allow for the environmental parameters to be better constrained and reduce uncertainties arising from poorly determined factors in the at-sea measurements.

Of course, it is important to analyze more data, at various locations (i.e. at different latitudes or at coastal areas and at higher winds during daytime) such that the relationships between the TSL and surface fluxes established in this dissertation may be improved. Analysis at higher winds during night time would greatly complement our current findings, but to do so would require very accurate and precise instrumentation to obtain an observable signal. Daytime data under low wind conditions is also an area of interest, however the analysis would be difficult given the influence of solar effects. Other environmental factors which may affect the TSL, such as the presence of surfactants, has not been explicitly considered. However, as our analysis involves
field data, the effects from such naturally occurring factors are implicit in our results as surfactants are ubiquitous at the sea surface (Tsai and Liu (2003)).

The currently published models have been shown to be inadequate at producing $\Delta T$ values at very low winds. This issue needs to addressed and tells us one must be careful when using such models. The suggestion proposed in this dissertation is to include radiative effects at low winds as it has been shown that through field data analysis, the longwave radiation constitutes the majority of the net flux and thus should not be ignored. There is therefore a continued need to refine the viscous sublayer models with a focus on low wind conditions. To aid the improvement of these models, more field data should also be obtained to complement and justify the theories which define the models. Furthermore, the current ability to retrieve the temperature profile within the upper 0.1 mm of the ocean’s surface suggests that the viscous sublayer models need to be re-assessed to better represent the top sub-millimeter temperature profile as currently the models have been relying on temperature differences between the skin sea surface temperature and temperatures obtained at centimeter or even meter depth scales. Doing so would improve the understanding of the TSL and potential push the boundaries of our current understanding of small-scale air-sea interactions.
Appendix A

Plots of $\Delta T_{\text{skin}-0.1\text{mm}}$ and $\Delta T_{\text{skin}-5\text{m}}$ against $\text{LW}_{\text{in}}$, $\text{LW}_{\text{net}}$ and $\text{LW}_{\text{in}@\text{zenith}}$.

Figure A.1: Plots of $\Delta T_{\text{skin}-0.1\text{mm}}$ (K) versus $\text{LW}_{\text{in}}$ (Wm$^{-2}$). Black dots are mean $\Delta T_{\text{skin}-0.1\text{mm}}$ binned to every 5 Wm$^{-2}$ of $\text{LW}_{\text{in}}$ with ±1 standard deviation.
Figure A.2: Plots of $\Delta T_{\text{skin-5m}}$ (K) versus LW$_{\text{in}}$ (Wm$^{-2}$). Black dots are mean $\Delta T_{\text{skin-5m}}$ binned to every 5 Wm$^{-2}$ of LW$_{\text{in}}$ with ±1 standard deviation.

Figure A.3: Plots of $\Delta T_{\text{skin-0.1mm}}$ (K) versus LW$_{\text{net}}$ (Wm$^{-2}$). Black dots are mean $\Delta T_{\text{skin-0.1mm}}$ binned to every 5 Wm$^{-2}$ of LW$_{\text{net}}$ with ±1 standard deviation.
Figure A.4: Plots of $\Delta T_{\text{skin}-5m}$ (K) versus LW$_{\text{net}}$ (W m$^{-2}$). Black dots are mean $\Delta T_{\text{skin}-5m}$ binned to every 5 W m$^{-2}$ of LW$_{\text{net}}$ with $\pm$1 standard deviation.

Figure A.5: Plots of $\Delta T_{\text{skin}-0.1mm}$ (K) versus LW$_{\text{in@zenith}}$ (W sr$^{-1}$ m$^{-2}$). Black dots are mean $\Delta T_{\text{skin}-0.1mm}$ binned to every 2 W sr$^{-1}$ m$^{-2}$ of LW$_{\text{in@zenith}}$ with $\pm$1 standard deviation.
Figure A.6: Plots of $\Delta T_{\text{skin-5m}}$ (K) versus $\text{LW}_{\text{in@zenith}}$ (W sr$^{-1}$ m$^{-2}$). Black dots are mean $\Delta T_{\text{skin-5m}}$ binned to every 2 W sr$^{-1}$ m$^{-2}$ of $\text{LW}_{\text{in@zenith}}$ with ±1 standard deviation.
Appendix B

Plots of $\Delta T_{model}$ and $\delta_z$ (model) against $LW_{in}$, $LW_{net}$ and $LW_{in@zenith}$.

Figure B.1: Plots of $\Delta T_{model}$ (K) versus $LW_{in}$ (W/m$^2$).
Figure B.2: Plots of $\Delta T_{\text{model}}$ (K) versus LW$_{\text{net}}$ (W/m$^2$).

Figure B.3: Plots of $\Delta T_{\text{model}}$ (K) versus LW$_{\text{in}@\text{zenith}}$ (W/sr/m$^2$).
Figure B.4: Plots of $\delta_z^{(model)}$ (cm) versus $LW_{in}$ (W/m$^2$).

Figure B.5: Plots of $\delta_z^{(model)}$ (cm) versus $LW_{net}$ (W/m$^2$).
Figure B.6: Plots of $\delta_z$ (model) (cm) versus LW$_{\text{in@zenith}}$ (W/sr/m$^2$).
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