A MODEL FOR THE ESTIMATION OF THE SURFACE FLUXES OF
MOMENTUM, HEAT, AND MOISTURE OF THE CLOUD
TOPPED MARINE ATMOSPHERIC BOUNDARY LAYER
FROM SATELLITE MEASURABLE PARAMETERS

by

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ABSTRACT

This thesis develops a model for the estimation of the surface fluxes of momentum, heat, and moisture of the cloud topped marine atmospheric boundary layer by use of satellite remotely sensed parameters. The parameters chosen for the problem are the integrated liquid water content, \( q_{li} \), the integrated water vapor content, \( q_{vi} \), the cloud top temperature, the sea surface temperature, and either a measure of the 10 meter neutral wind speed or the friction velocity at the surface.

Under the assumptions of a horizontally homogeneous, well-mixed boundary layer, the model calculates the equivalent potential temperature and total water profiles of the boundary layer along with the boundary layer height from inputs of \( q_{li} \), \( q_{vi} \), and cloud top temperature. These values, along with the 10m neutral wind speed or friction velocity and the sea surface temperature, are then used to estimate the surface fluxes using the methods of Stage (1979) and Liu, Katsaros, and Businger (1979).

The development of a scheme to parameterize the integrated water vapor outside of the boundary layer for
the cases of cold air outbreak and California coastal stratus is presented. The scheme involves the mean profiles of relative humidity and temperature along with the mean jump of potential temperature through the inversion to calculate the integrated water vapor. In turn, this leads to a method of calculating the radiative temperature flux at the cloud top by use of Staley and Jurica's (1970; 1972) radiation model.

Sensitivity studies are presented showing the potential accuracy of the technique. The model's expected accuracies with existing satellite technology are 24% for the heat flux and 82% for the moisture flux. However, these expected errors will be reduced with improvement of the satellite parameter retrieval algorithms.

In conclusion, with improvements of current technology, the model will provide reasonable estimates of the surface fluxes. The model is the only one of its kind currently available to estimate the surface fluxes of momentum, heat, and moisture of the cloud topped marine atmospheric boundary layer from satellite measurable parameters.
DEDICATION

This thesis and the work that went into it was made possible by the incredible patience shown towards me by my fiancee, Lori, and my family, and it is to these people whom I dedicate this.
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CHAPTER 1: INTRODUCTION

1.1 PURPOSE

The purpose of this thesis is to develop a model for the estimation of the surface fluxes of momentum, heat, and moisture by use of satellite remotely sensed parameters of the cloud-topped marine atmospheric boundary layer.

The thesis first introduces the problem. Second, the equations used to solve the problem are shown. Third, the computational form of the model is presented. Fourth, the results of the test runs of the model are given. Last, the thesis conclusions are drawn.

The satellite parameters chosen for this problem are the integrated liquid water content of the atmosphere, the integrated water vapor content of the atmosphere, the cloud top temperature, the sea surface temperature, and either a measurement of the 10 meter neutral wind speed or the friction velocity at the surface. This set of parameters is not the only possible set, but the technology exists today to remotely sense this set; hence,
it was chosen for this study. The current technology to measure this set is discussed later in this chapter, but first the importance of this study is considered.

1.2 SCIENTIFIC IMPORTANCE OF COLD-AIR OUTBREAK

Modelling of the air-sea surface fluxes of momentum, heat, and moisture by use of remotely sensed satellite parameters is of importance due to the effect these fluxes have on the regional and global weather.

Cold air outbreak has a pronounced effect on the western sections of midlatitude oceans off the east coasts of Asia and North America. In these regions, high winds, large air-sea temperature differences, and relatively dry air combine to give heat and moisture fluxes which are much greater than oceanic means. These fluxes over the warm water currents of the Kuroshio and Gulf Stream are known to have profound effects on cyclogenesis and storm intensification, particularly on the development of wintertime cyclones from Taiwan and Cape Hatteras lows (Chou and Atlas, 1982; Agee and Howley, 1977).

Recently, a method has been suggested by Chou and Atlas (1982) and Stage (1983) to estimate the heat and moisture surface fluxes in the region between the shore
and the edge of a cloud bank during cold air outbreak, by remotely sensed parameters. This method uses the boundary layer model developed by Stage and Businger (1981 a,b) for cold air outbreak. Further research is needed; however, to extend the estimation of the fluxes into the cloudy region. Hence, the reason for this study is the need of a model to estimate the surface fluxes in the cloudy region.

1.3 THE MODEL

The model developed in this study is capable of estimating the surface fluxes of momentum, heat, and moisture during a cold air outbreak by use of remotely sensed parameters of the marine atmospheric boundary layer. This model is based on the models of Stage (1979) and Liu, et al. (1979). It assumes unstable boundary layer conditions and uses the Businger diabatic profiles. Because of the assumption of unstable conditions, the model is not valid in neutral or stable conditions. However, the large air-sea temperature differences common in cold air outbreak or the large cloud top radiative temperature fluxes seen in California coastal stratus cases both produce the well-mixed boundary layer
conditions needed for the model to run. The model is also limited by the presence of fog. Although the model has been revised to include the presence of advection fog, it is still unclear as to how the equations which calculate the surface fluxes react in foggy conditions. Hence, fog is an area in which further study is necessary.

1.4 SATELLITE TECHNOLOGY

The technology exists today to measure the parameters: cloud top temperature, integrated water vapor content, integrated liquid water content, sea surface temperature, and either a 10m neutral wind speed or a friction velocity for use in this model to estimate the surface fluxes.

Cloud top temperature may be obtained from infrared satellite measurements, such as the infrared sensors on the GOES east and west satellites. The other parameters may be calculated by use of a scanning multichannel microwave radiometer (SMMR) such as the one carried aboard SEASAT or the one currently aboard NIMBUS 7.

The SMMR instrument's primary purpose is to measure sea surface temperature and to provide a measurement of the surface wind speed. However, SMMR also provides

As one can see, with existing technology the parameters needed for this model may be estimated from satellite measurements. In the next chapter, the model development is discussed.
CHAPTER 2: MODEL DEVELOPMENT

2.1 INTRODUCTION

The model development is discussed in this chapter. As stated in the first chapter, the purpose of the model is to estimate the surface fluxes of momentum, heat, and moisture. One method of accomplishing this task is to find an estimate of the variables \( \Theta_b, q_L, \) and \( z_b \) (See Figure 2.1.1). This may be done with satellite measurements of \( \Theta_b, q_{L1}, \) and \( q_{V1}, \) and assuming an unstable, well mixed boundary layer with \( \Theta_b \) and \( q_L \) being conserved quantities (See Figure 2.1.2). Then, with an estimate of the neutral wind speed or the friction velocity and the sea surface temperature, the surface fluxes may be found by using the bulk aerodynamic formulas (See Figure 2.1.3).

First in this chapter, some basic variables and thermodynamics are introduced. Second, the derivation of equations for the model are shown. Third, the model equations are inverted. Fourth, the equations used to estimate the surface fluxes are shown. Fifth, the
Figure 2.1.1. Atmospheric Cross Section with Oversized Marine Atmospheric Boundary Layer. Arrows pointing down indicate satellite measurable parameters, arrows pointing up indicate fluxes.
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Figure 2.1.3. Model Flowchart Showing Steps Necessary to Estimate the Fluxes
parameterization of the liquid and water vapor contents outside of the boundary layer is presented. Last, a method to calculate the radiative temperature flux at cloud top is discussed.

2.2 BASIC VARIABLES AND THERMODYNAMICS

The basic variables used in this paper are now introduced.

\[ T_d \]  Dewpoint temperature  
\[ \Theta \]  Potential temperature  
\[ \Theta_0 \]  Sea surface temperature  
\[ \Theta_b \]  Cloud top temperature  
\[ q_t \]  Total water mixing ratio  
\[ q_{vb} \]  Water vapor mixing ratio at cloud top  
\[ q_{lb} \]  Liquid water mixing ratio at cloud top  
\[ z_b \]  Cloud top height  
\[ z_c \]  Cloud base height  
\[ z_s \]  Saturation level height  
\[ k \]  von Karman's constant  
\[ \epsilon \]  Ratio of molecular weights of water and air

The basic thermodynamic properties are now shown. The equivalent potential temperature may be approximated
\[ \theta_e = \theta + \left( \frac{L_v}{C_p} \right) q_v \]  

(2.2.1)

where \( L_v \) is the latent heat of evaporation for water and \( C_p \) is the specific heat of air at constant pressure. The total water mixing ratio is defined as

\[ q_e = q_v + q_l \]  

(2.2.2)

The Clausius-Clapeyron equation may be written as

\[ de = \left[ \frac{eL_v \varepsilon}{(T_d^2) R} \right] dT_d \]  

(2.2.3)

where \( e \) is the water vapor partial pressure and \( \varepsilon \) is taken as 0.622. The water vapor partial pressure is related to the mixing ratio by

\[ q_v = \frac{\varepsilon}{p} e \]  

(2.2.4)

where \( p \) = atmospheric pressure. The differential of (2.2.4) gives
Substituting the hydrostatic equation \( \frac{dp}{dz} = -\rho g \), the perfect gas law, \( p = \rho RT \), and (2.2.3) into (2.2.5) gives

\[
\left( \frac{dq}{q} \right) = \left( \frac{dv}{v} \right) - \left( \frac{dp}{p} \right) \quad (2.2.5)
\]

\[
\left( \frac{dq}{q} \right) = \left( \frac{L_v}{(T_d^2 R)} \right) dT_d + \left( \frac{g}{(R T)} \right) dz \quad (2.2.6)
\]

Now, (2.2.6) may be integrated if one knows the saturation mixing ratio, \( q_r \), corresponding to some dewpoint temperature, \( T_d \), allowing the defining of two functions:

\[
Q(T) = q_r \exp \left( \left( -\frac{L_v}{R} \right) \left( 1/T - 1/T_d \right) \right) \quad (2.2.7)
\]

and

\[
T_{dew}(q) = \left[ \left( \frac{R}{L_v} \right) \ln \left( \frac{q}{q_r} \right) + 1/T_d \right]^{-1} \quad (2.2.8)
\]

Thus, \( Q(T) \) is the saturation mixing ratio for air at \( z=0 \), with temperature, \( T \); and \( T_{dew}(q) \) is the dewpoint for air at \( z=0 \), with water vapor mixing ratio, \( q \). These two functions are inverses; therefore,
In order to evaluate these functions, Stage (1979) obtained saturation vapor pressures from the Smithsonian Meteorological Tables. He chose values to give the best fit for the temperature range between 0 and 10 °C. This gives the function as

\[ Q(T) = 5.3542 \times 10^{-3} \exp\left(-5399.286 \left(1/T - 3.59505 \times 10^{-3}\right)\right) \]  

(2.2.10)

and

\[ T_{dew}(q) = \frac{5399.286}{[21.0640 - \ln(q \times 10^3)]} \]  

(2.2.11)

where \( T \) and \( T_{dew} \) are in Kelvin and \( q \) and \( Q \) are in g/kg.

The liquid water potential temperature may be defined as

\[ \Theta_1 = \Theta_e - \left(L_v/C_p\right)q_t \]  

(2.2.12)

and since \( \Theta_e \) and \( q_t \) are conserved in non-precipitating, no
ice, parcel motion, $\Theta_1$ is also a conserved quantity. Below the saturation level, $q_1=0$, and $\Theta_1=\Theta$. The saturation level or lifting condensation level is expressed as

$$z_s=(\Theta_1-T_d)/((\gamma-(gT)/(L_v E)))$$

(2.2.13)

where $\delta$ is the dry adiabatic lapse rate and $(gT)/(L_v E)$ may be expressed as the lapse rate of dewpoint. The last thermodynamic property to be introduced is the virtual potential temperature, $\Theta_v$.

$$\Theta_v=\Theta(1+(1/\varepsilon-1)q_v-q_1)$$

(2.2.14)

Virtual potential temperature introduces the effects of both the buoyancy of water vapor and liquid water drag.

2.3 DERIVATION OF MODEL'S EQUATIONS

The equations used in this model were developed for the case of cold air outbreak. The marine atmospheric boundary layer (MABL) is assumed to be cloud topped and well mixed with $\Theta_0$ and $q_1$ being conserved quantities. The
cloud base is assumed to be equal to the saturation level and that all motions in the cloud layer are saturated. The cloud top is assumed to be equivalent to the base of the inversion layer.

A method is needed to determine values for the cloud top temperature, $\Theta_b$, the integrated liquid water content, $q_{li}$, and the integrated water vapor content, $q_{vi}$, of the marine boundary layer. The marine boundary layer model developed by Stage (1979) provides such a method. Stage incorporates a routine to handle saturated partial behavior in which inputs of $\Theta_e$ and $z_b$ produce values of $\Theta_b$ and $q_v$ for a saturated parcel. Hence, with inputs of $\Theta_e$, $q_t$, and $z_b$, it is possible to find $\Theta_b$, $q_{li}$, and $q_{vi}$.

A method is now shown in which the cloud top temperature may be estimated by use of an iterative method with initial inputs of $\Theta_e$, $q_t$, and $z_b$, and an initial guess for $\Theta_b$.

Cloud top temperature may be found by use of the following equations.

$$
T_{db} = \Theta_b - \left( \frac{\delta - (gT)}{L_v \epsilon} \right) z_b
$$

(2.3.1)

The water vapor mixing ratio at cloud top may be found by using (2.2.9b) and (2.3.1) where
From (2.3.2) it follows that

\[ dq_v = dQ/dT[(\delta_b - (gT/(L_v \xi)))dz_b] \]  

The finite difference form of (2.3.3) is

\[ dQ/dT = (Q(T_{db} + \Delta T) - q_v)/\Delta T \]  

where \( \Delta T \) is a small increment. The quantity \( \delta \theta_b \) is expressed as

\[ \delta \theta_b = [\theta_e - (\theta_b + (L_v/C_p)q_v)/[1+(L_v/C_p)dQ/dT] \]  

Therefore, by making an initial guess at \( \theta_b \), one may use equations (2.3.1) to (2.3.5) and

\[ \theta_{b_{new}} = \theta_{b_{old}} + \delta \theta_b \]

one may iterate numerically until a suitably small \( \delta \theta_b \) is reached, thus providing \( \theta_b \) and \( q_v \).
The liquid water mixing ratio at cloud top may be found from

\[ q_{lb} = q_t - q_{vb} \]  \hspace{1cm} (2.3.6)

In order to solve for the integrated liquid water content, \( q_{li} \), the depth of the cloud layer, \( d_c \) must be found. Using equations (2.2.12), for \( \Theta_1 \), and (2.2.13), for \( z_s \), and assuming that \( z_s \), the saturation level, is equal to \( z_c \), the cloud base level

\[ d_c = z_b - z_c \]  \hspace{1cm} (2.3.7)

Below cloud base, \( q_l = 0 \). The liquid water mixing ratio is assumed to increase linearly with height from \( z_c \) to \( z_b \), and then is assumed to be equal to zero above the boundary layer (See Figure 2.3.1).

Thus the \( q_l \) profile in the cloud layer forms a triangle, and the integrated liquid water content may be found by

\[ q_{li} = 0.5q_{lb}d_c \]  \hspace{1cm} (2.3.8)

The integrated total water content, \( q_{ti} \), of the boundary
Figure 2.3.1. Profile of $q_1$ in Marine Boundary Layer
layer is expressed as

\[ q_{ti} = \int_{0}^{z_b} q_t \, dz \]

Since \( q_t \) is well mixed in the boundary layer

\[ q_{ti} = q_t \times z_b \quad (2.3.9) \]

Hence the final quantity, the integrated water vapor content, may be found

\[ q_{vi} = q_{ti} - q_{li} \quad (2.3.10) \]

Now that the quantities, \( \Theta_b, q_{li}, \) and \( q_{vi} \) have been derived, the problem may be inverted.

2.4 INVERSION OF EQUATIONS

The previous section showed a method in which with inputs of \( \Theta_s, q_t, \) and \( z_b \) were used to find the satellite parameters \( \Theta_b, q_{li}, \) and \( q_{vi}. \) In order to compute the boundary layer profiles from these satellite parameters a numerical method must be found to invert the set of
equations in the last section. By inverting the equations in Section 2.3 allows one to take the satellite measurable data and calculate \( z_b \) and the boundary layer profiles of \( \Theta_e \) and \( q_t \). To begin, the satellite measurable parameters, \( q_{li} \), \( q_v \), and \( \Theta_b \) are used as input. Then one may calculate

\[
q_{ti} = q_v + q_{li}
\]  

(2.4.1)

It is also possible to express \( q_{ti} \) as

\[
q_{ti} = \int_{z_0}^{z_{top}} q_t \, dz + \int_{0}^{z_b} q_t \, dz
\]

Taking \( \int_{z_0}^{z_{top}} q_t \, dz \) as a known quantity, this will be discussed in Section 2.6, and taking \( q_t \) constant below \( z_b \) one may write

\[
q_{ti} = \int_{z_0}^{z_{top}} q_t \, dz + q_t \, z_b
\]

where

\[
q_t \, z_b = q_{ti} - \int_{z_0}^{z_{top}} q_t \, dz
\]  

(2.4.2)

We now have the quantity \( q_t \, z_b \) from which \( q_t \) and \( z_b \) may
be found by guessing $z_b$ and solving for $q_t$. Then, by use of the following set of equations and a *regula falsi* method, it is possible to estimate the variables $\Theta_e$, $q_t$, and $z_b$ which are needed in the flux calculations.

By guessing $z_b$ and solving

$$q_t = \frac{q_t x z_b}{z_b} \quad (2.4.3)$$

one is left with $q_t$ and $z_b$. From equation (2.3.1) one gets

$$T_{db} = \Theta_b - \left( \frac{\gamma - (qT)/(L_v \epsilon)}{\gamma} \right) z_b \quad (2.4.4)$$

The water vapor mixing ratio may be found by

$$q_{vb} = Q(T_{db}) \quad (2.4.5)$$

Where $Q(T)$ is the function described by equation (2.2.10). The equivalent potential temperature, $\Theta_e$, is found from equation (2.2.1)

$$\Theta_e = \Theta_b + \frac{L_v}{C_p} q_{vb} \quad (2.4.6)$$
The liquid water mixing ratio at cloud top is

\[ q_{1b} = q_t - q_{vb} \quad (2.4.7) \]

The dewpoint just above the sea surface may be found from

\[ T_{do} = T_{dew}(q_t) \quad (2.4.8) \]

Where \( T_{dew}(q) \) is the function expressed by equation (2.2.11). The liquid water potential temperature comes from equation (2.2.12)

\[ \theta_1 = \theta + (L_v / c_p) q_t \quad (2.4.9) \]

The saturation level may be calculated using equation (2.2.13)

\[ z_s = (\theta_1 - T_{do}) / (\gamma - (qT) / (L_v \epsilon)) \quad (2.4.10) \]

If \( z_s \) is greater than \( z_b \), no cloud is formed in the model, and the model ends. If \( z_s \) is less than zero, then fog is present below cloud level. In the case of fog, \( q_{11} \), is no longer proportional to the area of the triangle shown in Figure 2.3.1. The trapezoidal rule can now be applied as
Figure 2.4.1. Profile of $q_1$ in Marine Boundary Layer with Fog
can be seen in Figure 2.4.1.

The value of \( q_1 \) at the cloud base is called \( q_{1c} \). Under no fog conditions, \( q_{1c} = 0 \). In the case of fog, the cloud base is at the surface and \( q_{1c} \) is not equal to zero. Note that the term "fog" refers to advection fog. If fog is present in the boundary layer, \( q_{1c} \) may be found from the input values of \( \Theta_s \), \( q_t \), and \( z_s \) by the same iterative method shown in Section 2.3.

The dewpoint temperature at \( z_s \) may be found by

\[
T_{ds} = \Theta_s - \left( \gamma - (gT)/(L_v) \right) z_s
\]  

(2.4.11)

where a value for \( \Theta_s \) has been guessed. The water vapor mixing ratio at saturation level is

\[
q_{vs} = Q(T_{ds})
\]  

(2.4.12)

where \( Q(T) \) is from equation (2.2.10). A small increment, \( \Delta T \), is now introduced in which

\[
T_{ds\ new} = T_{ds\ old} + \Delta T
\]  

(2.4.13)

The mixing ratio of \( T_{ds\ new} \) is
found by the same method as (2.2.11). The finite difference form, equation (2.3.4), is now employed

\[
d\frac{Q}{dT} = \left( q(T_{\text{dnew}}) - q_{\text{vs}} \right) / \Delta T
\]  
(2.4.15)

From equation (2.3.5)

\[
\delta \theta = \left[ \theta_e - \left( \theta_s + (L_v/C_p)q_{\text{vs}} \right) / \left( 1 + (L_v/C_p)dQ/dT \right) \right]
\]  
(2.4.16)

If \( \delta \theta_s \) is less than some tolerance level than \( q_{1C} \) may be found by

\[
q_{1C} = q_t - q_{\text{vs}}
\]  
(2.4.17)

Otherwise, equations (2.4.11) to (2.4.16) are repeated with

\[
\theta_{\text{dnew}} = \theta_{\text{old}} + \delta \theta_s
\]  
(2.4.18)

until \( q_{1C} \) is found.

If \( 0 < z_c < z_b \) than a cloud is formed in the model and no
fog is present. Then, the triangle rule illustrated by Figure 2.3.1 applies. The equation used to handle both fog and no fog cases is

\[ q_{1i} = 0.5d_c(q_{1b} + q_{1c}) \]  

(2.4.19)

The difference between the value of \( q_{1i} \) calculated in (2.4.19) and the value of \( q_{1i} \) measured by the satellite is then checked to see if it is below an acceptable tolerance value. If it is not, then a \emph{regula falsi} routine is used to calculate a new value of \( z_b \) and equations (2.4.3) to (2.4.19) are repeated until an acceptable value is achieved. Once this occurs, the boundary layer profiles of \( \Theta_e \) and \( q_t \) are found. It is then possible to begin the calculations to derive the surface fluxes of momentum, heat, and moisture, which is the topic of the next section.

2.5 CALCULATION OF SURFACE FLUXES

The calculation of the surface fluxes of momentum, heat, and moisture may now be done. As discussed in the first chapter, satellite measurements of both a neutral
wind speed at some height and a friction velocity may be calculated by means of empirical formulas. Because the debate continues on which is a more appropriate measurement in its correspondence to the measured backscatter, I have chosen to allow the use of either in the model. With the addition of satellite measured sea surface temperature, the surface fluxes may be found.

By following the method of Liu, et al. (1979), the surface fluxes may be estimated by first finding \( u_* \) and \( z_0 \). Then, the roughness Reynolds number is computed from which \( z_t \) and \( z_q \) are found. With the addition of the sea surface temperature and mixing ratio, the surface fluxes may be found by using the bulk aerodynamic formulas and the Businger-Dyer unstable diabatic profiles.

For input involving a neutral stability wind, Kondo's (1975) relation is used

\[
C_{dn} = p + q(u_{10n})^r
\]  

(2.5.1)

Where \( p \), \( q \), and \( r \) are constants which vary for wind speed velocities between 0.3 and 50.0 m/s. Kondo's relation was chosen because of its ability to cover large ranges of wind speed. This allows the model to incorporate cases such as California coastal stratus which do not exhibit
the large air-sea temperature differences common for cold air outbreaks; thus, creating a near neutral atmospheric stability.

Equation (2.5.1) may be expressed as

\[
\left( \frac{u_*}{u_{10n}} \right)^2 = p + q \left[ \frac{(u_*/k) \ln(10/z_0)}{u_{10n}} \right]^{-1}
\]  

(2.5.2)

where I have used

\[
C_{dn} = \left( \frac{u_*}{u_{10n}} \right)^2
\]  

(2.5.3a)

\[
u_{10n} = \left( \frac{u_*/k}{\ln(10/z_0)} \right)
\]  

(2.5.3b)

One may calculate the friction velocity, \(u_*\), by

\[
u_* = (C_{dn} u_{10n}^2)^{1/3}
\]  

(2.5.4)

The roughness length, \(z_0\), may be found from the expression for \(u_{10n}\) in equation (2.5.3b).

\[
z_0 = 10 \left[ \exp\left( \frac{(u_{10n}k)}{u_*} \right) \right]^{-1}
\]  

(2.5.5)

For input involving satellite derived friction
velocity, Charnock's (1955) relation is used to find the roughness length. The equation is

\[ z_0 = a(u_*)^2/g \]  \hspace{1cm} (2.5.6)

where \( g \) is the gravitational acceleration and \( a \) is Charnock's constant. Wu (1969) in his study of sea surface roughness, found \( a=0.0156 \) to be a reasonable value for a large range of oceanic data.

Following the method of Liu, et al. (1979) the roughness Reynolds number is expressed as

\[ R_r = z_0 u_*/\nu \]  \hspace{1cm} (2.5.7)

where \( \nu \) is the kinematic viscosity.

According to Liu, et al. (1979), the variables \( z_t \) and \( z_q \) may be expressed in terms of the roughness Reynolds number by

\[ z_t = (a_1 R_r^{-a_2} \nu)/u_* \]  \hspace{1cm} (2.5.8)

and
\[ z_q = (a_2 R_r^{0.2} \chi) / u_* \quad (2.5.9) \]

where the coefficients \( a_1, b_1, a_2, \) and \( b_2 \) for different ranges of \( R_r \) are shown in Table 2.5.1, taken from Liu, et al. (1979).

The distribution of velocity, temperature, and humidity for the surface layer, outside of the region where molecular effects dominate, are governed by the Businger profiles

\[ (u-u_0)/u_* = [\ln(z/z_0) - \Psi_u]/k \quad (2.5.10) \]

\[ (\theta-\theta_0)/\theta_* = [\ln(z/z_t) - \Psi_t]/\alpha_h k \quad (2.5.11) \]

\[ (q-q_0)/q_* = [\ln(z/z_q) - \Psi_q]/\alpha_e k \quad (2.5.12) \]

The diabatic terms, \( \Psi_u, \Psi_t, \) and \( \Psi_q \) are given by

\[ \Psi_u = 2 \ln([1+(1-\beta_u(z_0/L))^{1/a_u}]/2) + \ln([1+(1-\beta_u(z_0/L))^{1/a_u}]^{1/2}) \]

\[ -2 \tan^{-1}[(1-\beta_u(z_0/L))^{1/a_u}] + \Psi/2 \quad (2.5.13) \]

\[ \Psi_t = 2 \ln([1+(1-\beta_t(z_0/L))^{1/a_t}]/2) \quad (2.5.14) \]
<table>
<thead>
<tr>
<th>$R_r$</th>
<th>$a_1$</th>
<th>$b_1$</th>
<th>$a_2$</th>
<th>$b_2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>0. - 0.11</td>
<td>0.177</td>
<td>0.</td>
<td>0.292</td>
<td>0.</td>
</tr>
<tr>
<td>0.11 - 0.825</td>
<td>1.376</td>
<td>0.929</td>
<td>1.808</td>
<td>0.826</td>
</tr>
<tr>
<td>0.825 - 3.0</td>
<td>1.026</td>
<td>-0.599</td>
<td>1.393</td>
<td>-0.528</td>
</tr>
<tr>
<td>3.0 - 10.0</td>
<td>1.625</td>
<td>-1.018</td>
<td>1.956</td>
<td>-0.870</td>
</tr>
<tr>
<td>10.0 - 30.0</td>
<td>4.661</td>
<td>-1.475</td>
<td>4.994</td>
<td>-1.297</td>
</tr>
<tr>
<td>30.0 - 100.0</td>
<td>34.904</td>
<td>-2.067</td>
<td>30.790</td>
<td>-1.845</td>
</tr>
</tbody>
</table>
and

$$\psi_q = 2 \ln \left( \frac{1}{1 + (1 - \beta_q (z_q/L)))^{1/3}} \right) / 2 $$

(2.5.15)

respectively. Liu, et al. (1979) used \( k = 0.4, \)

\( 1/(\alpha_{hk}) = 1/(\alpha_{sk}) = 2.2, \) and \( \beta_u = \beta_t = \beta = 16. \)

To include the effects of the buoyancy of water vapor, the Obukhov length, \( L, \) may be expressed in terms of the virtual potential temperature flux

$$L = \left( \Theta_v u^* \right) / (g k \Theta_v *)$$

(2.5.16)

where \( \Theta_v \) comes from equation (2.2.14)

$$\Theta_v = \Theta_1 \left( 1 + (1/\varepsilon - 1) q_t \right)$$

(2.5.17)

and \( \Theta_v * \) is expressed as

$$\Theta_v * = \Theta_* \left( 1 - (1/\varepsilon - 1) q_t \right) + (1/\varepsilon - 1) \Theta q_t$$

(2.5.18)

The profiles (2.5.10) to (2.5.12) are formulated in terms of mean quantities for the boundary layer. These mean values can be associated with the surface flux profiles. Near the surface, mechanical shear dominates
and large gradients may exist in the mean profiles. Higher up, buoyancy begins to dominate and the shears gradually vanish as the profiles approach their mixed layer mean values.

If a matching height, $h$, is introduced it is possible to solve the problem of matching the mean mixed layer profiles with the surface layer profiles (Wyngaard, et al., 1974). Let $h$ be the height at which the profiles (2.5.10)-(2.5.12) produce the mixed layer mean values. Then one may write

\[ \ln(h/z_0) - \Psi_u(h/L) = \ln(-h/L) - \Psi_u(-h/L) - \ln(-z_0/L) \]

\[ \ln(h/z_t) - \Psi_t(h/L) = \ln(-h/L) - \Psi_t(-h/L) - \ln(-z_t/L) \quad (2.5.19) \]

\[ \ln(h/z_q) - \Psi_q(h/L) = \ln(-h/L) - \Psi_q(-h/L) - \ln(-z_q/L) \]

The height at which the gradients become vanishingly small is determined by the relative importance of shear and buoyant production of turbulence kinetic energy and thus $h$ should be proportional to $-L$ (Wyngaard, et al., 1974). Therefore,
\[ K_u = \ln(-h/L) - \Psi_u(-h/L) \]

\[ K_t = \ln(-h/L) - \Psi_t(-h/L) \]  
(2.5.20)

\[ K_q = \ln(-h/L) - \Psi_q(-h/L) \]

are constants and the profiles (2.5.10)-(2.5.12) may be expressed as

\[ (u-u_0)/u^* = [K_u - \ln(-z_0/L)]/k \]  
(2.5.21)

\[ (\Theta - \Theta_0)/\Theta^* = [K_t - \ln(-z_t/L)]/\alpha_h k \]  
(2.5.22)

\[ (q-q_0)/q^* = [K_q - \ln(-z_q/L)]/\alpha_e k \]  
(2.5.23)

Wyngaard, et al. (1974) took \(-h/L = -10.0\), thus giving \(K_u = -0.2\) and \(K_t = K_q = -1.02\).

In order to solve equations (2.5.21)-(2.5.23) the Obukhov length, \(L\), must be found. Equations (2.5.16)-(2.5.18) show that it is necessary to know \(\Theta^*\) and \(q^*\) in order to do this problem. If an initial guess of \(\Theta_v^*\) is given and equation (2.5.17) is used to solve for \(\Theta_v\), the Obukhov length may be determined by equation (2.5.16). The parameter \(\Theta_e^*\) may be found from equation (2.5.22) and
the satellite measured sea surface temperature $\Theta_0$.

$$\Theta_* = (\Theta_1 - \Theta_0) / \left[ \alpha_h k (K_c - \ln(-z_t/L)) \right]$$  \hspace{1cm} (2.5.24)

The sea surface value of the water vapor mixing ratio, $q_0$, may be found by using the function $Q(T)$ expressed in equation (2.2.10).

$$q_0 = Q(\Theta_0)$$  \hspace{1cm} (2.5.25)

Assuming no liquid water or ice is present in the mixed layer where the mean profiles exist, $q_t$ may be used as the value of the water vapor mixing ratio of the mixed layer. Thus, $q_*$ may be found from equation (2.5.23).

$$q_* = (q_t - q_0) / \left[ \alpha_s k (K_q - \ln(-z_q/L)) \right]$$  \hspace{1cm} (2.5.26)

A new value of $\Theta_{v*}$ may now be calculated using equation (2.5.18). The difference between the guessed value and the calculated value of $\Theta_{v*}$ may be found

$$\Delta \Theta_{v*} = \Theta_{v*guess} - \Theta_{v*calc}$$  \hspace{1cm} (2.5.27)
When $\Delta \theta_{v*}$ reaches a small value approximately equal to zero, $\theta_{v*}$ has been found, and therefore, the values of $\theta_*$ and $q_*$ are known.

The surface fluxes associated with the profiles expressed in equations (2.5.21)-(2.5.23) may be written as

$$\overline{u'w'}_0 = u_*^2$$  \hspace{1cm} (2.5.28)

$$\overline{w'\theta'}_0 = -u_* \theta_*$$  \hspace{1cm} (2.5.29)

$$\overline{w'q'_t}_0 = -u_* q_*$$  \hspace{1cm} (2.5.30)

where $\overline{u'w'}_0$ is the surface flux of momentum, $\overline{w'\theta'}_0$ is the surface flux of sensible heat, and $\overline{w'q'_t}_0$ is the surface flux of moisture. Since $u_*$, $\theta_*$, and $q_*$ are now known, the fluxes may be calculated from these equations.

The method listed above for determining the fluxes assumes unstable conditions. Therefore, a limit must be placed on the calculations for when conditions approach near neutral stability.

As conditions approach near neutral stability, the relation $z_0/L$ goes to zero since $L$ goes to plus or minus infinity. The author chose to limit $z_0/L$ to be
\[(z_0/L)_{\text{min}} \leq 1.0 \times 10^{-7}\]  \hspace{1cm} (2.5.31)

By taking a typical value of \(z_0\) over the ocean as \(2.0 \times 10^{-4}\) meters (Miyake, et al., 1970), this gives an upper limit of the Obukhov length of

\[-L = 2000\] meters

The maximum height of \(z_b\) in the model is 2000 meters. In order for the Obukhov length not to exceed \(z_b\), the assumption expressed in equation (2.5.31) appears to be a valid lower limit for near neutral stability conditions.

2.6 PARAMETERIZING WATER VAPOR ABOVE THE MABL

A method to parameterize the water vapor content above the MABL is now presented. An analysis is needed in which the profile of \(q_f\) above the boundary layer may be found. From this profile, the integrated water vapor content may be found since \(q_f\) is assumed to be zero above \(z_b\). A climatological profile is the easiest means available of accomplishing the analysis. Schemes for both cold air outbreak and California coastal stratus are given
in this section.

2.6.a Cold Air Outbreak

The profile for cold air outbreak was derived from the NASA Mesoscale Air-Sea EXchange (MASEX) project. In order to find the profile of $q_t$ above the boundary layer, profiles of temperature, relative humidity, and mixing ratio were studied. The profiles of temperature and relative humidity were chosen to find the profile of $q_t$. The mixing ratio profile was discarded because the relative humidity profile seemed more readily adaptable to a wide range of situations.

Given temperature and relative humidity, one may find $q_t$. So that the profiles studied in MASEX may be made relevant to other cases, the lapse rates of temperature and relative humidity were found. In addition, the mean jump of temperature from the bottom of the inversion to the top of the inversion was calculated along with the mean relative humidity at the inversion top. By using these mean values of $\Delta \Theta_\tau$ and RH, and the lapse rates of $\Theta$ and RH, the profile above the boundary layer of $q_t$ may be found.

The mean values of $\Delta \Theta_\tau$ and RH are 3 °C and 45%,
respectively. The lapse rates of RH and $\Theta$ are 8 %/km and 6 °C/km, respectively. By adding $\Delta T$ to the cloud top temperature and integrating over 10m steps using the lapse rates listed above, $q_v$ above the MABL may be found.

2.6.6 California Coastal Stratus

Similarly, a profile for California coastal stratus was computed. Analysis of the profiles found in the studies of Neiburger, et al. (1961) and Meitin (1975) yielded the mean values for this case. Values of $\Delta T = 7$ °C, RH=35%, lapse rate of temperature=6 °C/km, and lapse rate of RH=8 %/km were extracted from the data.

2.7 RADIATIVE FLUX

Given the profiles above the boundary layer, the radiative flux at cloud top may be found by using the flux emissivity method of Staley and Juric (1970;1972). Since the model assumes no cloud above the boundary layer, the flux at the top of the atmosphere is equal to zero.

The optical path length, $u(z)$, is given by the integration of the water vapor mixing ratio profile.
In equation (2.7.1), \( z_{\text{top}} \) refers to the highest sound level available. The flux emissivity \( \varepsilon(u) \) is found by the interpolation of the values tabulated by Staley and Jurica; thus, giving the downward radiative temperature flux as

\[
F = \left( \frac{\sigma}{\rho c_p} \right) \int_{z_0}^{z_{\text{top}}} T^4 d\varepsilon(u(z))
\]  

(2.7.2)

since the atmosphere is assumed to be cloud free above the MABL.

The method described above, or any other similar radiation model, will yield a value for the radiative temperature flux. Due to the uncertainties involved when using mean profiles in place of actual profiles, the radiative temperature fluxes calculated by the model are not yet reliable. When a better method of finding the profile of water vapor above the MABL is found, the radiative flux method listed above will be useful.
CHAPTER 3: COMPUTATIONAL FORM OF THE MODEL

3.1 INTRODUCTION

The computational form of the model is discussed in this chapter. The equations for the model are given in Chapter 2; hence, this chapter references the equations instead of repeating them. This chapter shows the steps needed to produce estimates of the surface fluxes from the input data.

3.2 THE MODEL

In order to estimate the surface fluxes of momentum, heat, and moisture, the satellite inputs of cloud top temperature, integrated liquid water content, and integrated water vapor content must be converted into $z_b$ and the boundary layer profiles of $\Theta_e$ and $q_t$. First a guess of the boundary layer height is given. Then, the integrated water vapor content outside of the boundary layer is found by the method in Section 2.6. Then, the method given in Section 2.4 is used to calculate $\Theta_e$, $q_t$. 

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and \( z_b \). The integrated liquid water is calculated from the values found in Section 2.4. This calculated value is then compared to the satellite measured value. If the difference between the measured and calculated values of \( q_{11} \) is approximately equal to zero, the model continues. Otherwise, a *regula falsi* routine is used to obtain a new value of the boundary layer height, and the process above is repeated until the difference between the measured and calculated \( q_{11} \)'s is approximately equal to zero.

Upon finding the values of \( \Theta_\ast, q_t, \) and \( z_b \), the model continues. If a neutral wind speed is provided, the model calculates the friction velocity and roughness length by the method shown in equations (2.5.1) to (2.5.5). On the other hand, if a friction velocity is given, the roughness length is calculated by equation (2.5.6). With the values of \( u_* \) and \( z_0 \), and following the method of Liu, et al. (1979), a roughness Reynolds number may be computed. This is accomplished by equations (2.5.7) to (2.5.9).

The last satellite measured value, the sea surface temperature, is now used. Values of \( \Theta_\ast \) and \( q_\ast \) must be found so that the fluxes may be computed. From \( \Theta_\ast \) and \( q_t \), \( \Theta_1 \) is found. The quantity \( (\Theta_1 - \Theta_0) \) is used for the computation of \( \Theta_\ast \). The quantity \( (q_t - q_0) \), where \( q_0 \) is found from the sea surface temperature, is used for
calculating \( q^* \). With an initial guess of \( \Theta_{v^*} \), equations (2.5.16) to (2.5.18) and equations (2.5.24) to (2.5.25) are employed to calculate \( \Theta_* \) and \( q_* \). Then \( \Theta_{v^*} \) is computed by equation (2.5.26) and compared to the value of the original guess of \( \Theta_{v^*} \) by equation (2.5.27). If the difference is not approximately equal to zero, a regular falsi routine is employed to compute a new \( \Theta_{v^*} \). The process above is repeated until the difference is approximately equal to zero, at which time the model continues.

Finally, with values for \( u_* \), \( \Theta_* \), and \( q_* \), the surface fluxes of momentum, heat and moisture may be estimated. Equations (2.5.28) to (2.5.30) are used to accomplish this task.

A note about the computations listed above. Original test runs of the model showed that more than one solution existed for each combination of input parameters. Numerically, these solutions were valid; however, these solutions were physically unrealistic due to the extremely high dewpoint temperatures and low mixing ratios encountered. In order to "force" the model to converge to the "realistic" answer the following was done. An upper limit of 2000m for \( z_0 \) was chosen, since the majority of
the physically unrealistic solutions were at $z_b$ values greater than 2000m. To handle the unrealistic solutions that occurred below 2000m, the initial guesses of $z_b$ for the regula falsi routine were set at 1200 and 100m. Since no double solutions were seen to exist below 1200m, any solutions with $z_b$ less than this height will be accurate. For solutions between 1200 and 2000m, where double solutions might occur, a very restrictive tolerance was placed on $q_{11}$. Runs of the model showed that $q_{11}$ was extremely sensitive to changes in height. By requiring a near perfect convergence of the measured and calculated integrated liquid water contents, the problem of a possible double solution in the 1200 to 2000m range seems to have been solved.

In this chapter, the computational form of the model has been discussed. The steps needed to estimate the surface fluxes from the original satellite data have been shown. The next chapter deals with the model's results for its test runs.
CHAPTER 4: MODEL RESULTS

4.1 INTRODUCTION

The results of the test runs of the model are discussed in this chapter. First, the applications of the model are presented. Next, the percent variation of the output parameters with variation of the input parameters is shown. Next, the expected accuracy of the model with respect to current accuracies of satellite measurements is presented. Last, the strengths and weaknesses of the model are discussed.

4.2 MODEL APPLICATIONS

The model was originally developed for the case of cold air outbreaks; however, it may also be applied to California coastal stratus. The model assumptions of a horizontally homogeneous, cloud-topped boundary layer with little or no liquid water above the boundary layer apply to both cases mentioned above.

In order to test the model's capabilities, the
routine described in Section 2.3 was run over a large range of values for \( \Theta_b \), \( q_t \), and \( z_b \). The predicted values of \( \Theta_b \), \( q_{li} \), and \( q_{vi} \) from this routine were then used as input for the model. For the cases in which the routine in Section 2.3 predicted clouds and/or clouds and fog, the model computed values of \( \Theta_b \), \( q_t \), and \( z_b \) which agreed with the variables inputted into the routine. However, for cases in which no clouds were predicted by the routine, the model would not converge to an answer. The routine in Section 2.3 was then run for typical ranges of values for the case of cold air outbreak with \( \Theta_b \) ranging from -30 to 10 °C, \( q_t \) ranging from 1.0 to 5.0 g/kg, and \( z_b \) ranging from 500 to 2000 meters. The routines predicted values of \( \Theta_b \), \( q_{li} \), and \( q_{vi} \) were then used as the range of values to compute the sensitivities of the output parameters described in the next section.

4.3 VARIATION OF INPUT AND OUTPUT PARAMETERS

The percent variation of the heat and moisture fluxes with variation of \( u_{10n} \), \( u_* \), \( \Theta_b \), \( \Theta_0 \), \( q_{li} \), and \( q_{vi} \) is now presented. The model was tested over the range of values listed above plus the ranges for \( u_{10n} \) of 5 to 20 m/s, for \( u_* \) of .1 to .7 m/s, and for \( \Theta_0 \) of 5 to 20 °C. The
figures presented in this chapter are from one test range of values for the model. The input variables for the cases shown here were held constant, except when they were the variable being tested, at \( u_{10} = 10 \text{ m/s}, \Theta_b = -15 \degree\text{C}, \Theta_0 = 10 \degree\text{C}, q_{11} = 300 \text{ m g/kg}, \) and \( q_{v1} = 900 \text{ m g/kg}. \)

The 10m neutral wind speed was tested over the range of 5 to 14 m/s in this case. \( u_{10} \) is related to the surface fluxes through \( u^*, z_0, z_t, \) and \( z_q \). The results of the variation of both the heat and moisture fluxes with respect to \( u_{10} \) is nearly linear. Analysis shows that with a +/-10% variation in \( u_{10} \) resulting in a +/-10% variation in both the heat and moisture fluxes.

The friction velocity was tested next with a range of .2 to .65 m/s in this case. The friction velocity is directly related to the fluxes as well as through \( z_0, z_t, \) and \( z_q \). The results of the tests showed the variation is, as in the case of \( u_{10} \), nearly linear. Analysis of the results shows that a +/-10% variation in \( u^* \) produces a +/-9% variation in both the heat and moisture fluxes.

Next, the cloud top temperature was tested over a range of -15 to -10.5 \degree\text{C} in this case. The cloud top temperature is related to the heat and moisture fluxes through \( \Theta_1 \) and \( q_c \). Results for the variation of the heat
flux may be seen in Figure 4.3.1.

As seen in Figure 4.3.1, the heat flux decreases for increasing $\theta_b$. The change in slope of the curve is due to the presence of fog predicted by the model. Because of the uncertainty of the model's evaluation of fluxes under foggy conditions, only cases in which no fog is present were considered for error determinations. The error for the no fog portion of this case as well as in all other cases is for a $\pm 1^\circ C$ change in $\theta_b$, the heat flux varies by $\pm 4\%$. Figure 4.3.2 shows the variation of the moisture flux with $\theta_b$.

As can be seen in the figure, the moisture flux also decreases with increasing $\theta_b$. The change in slope of the curve seen in the figure corresponds with the formation of fog. The results for a $\pm 1^\circ C$ change in $\theta_b$, in the no fog cases, are a $\pm 4.5\%$ variation in the moisture flux.

Sea surface temperature was the next variable tested. The range in this case was from 5 to 14 $^\circ C$. Sea surface temperature effects the heat flux directly and the moisture flux through $q_0$. Figures 4.3.3 and 4.3.4 show the variations of heat and moisture fluxes, respectively.

As shown in the figures, both the heat and moisture fluxes increase with increasing $\theta_0$. For a $\pm 1^\circ C$ change of $\theta_0$, the heat flux varies $\pm 3\%$, and the moisture flux
Figure 4.3.1. Variation of Heat Flux with Cloud Top Temperature
Figure 4.3.2. Variation of Moisture Flux with Cloud Top Temperature
Figure 4.3.3. Variation of Heat Flux with Sea Surface Temperature
Figure 4.3.4. Variation of Moisture Flux with Sea Surface Temperature
varies +/-7%.

Next, the integrated liquid water content was tested over the range of 300 to 480 m g/kg, for this case. The integrated liquid water content affects the heat flux through $\Theta_1$ and the moisture flux through $q_t$. Figure 4.3.5 shows the variation of the heat flux, and Figure 4.3.6 shows the variation of the moisture flux.

As can be seen in Figure 4.3.5, the heat flux increases with increasing $q_{11}$. Analysis shows that for a +/-100 m g/kg change in $q_{11}$, the heat flux changes +/-0.25%. However, in Figure 4.3.6, one may see that the moisture flux decreases with increasing $q_{11}$. A +/-100m g/kg change in $q_{11}$ produces a +/-3.5% change in the moisture flux.

Figures 4.3.7 and 4.3.8 show the variation of the heat and moisture fluxes with varying integrated water vapor content. The integrated water vapor was varied from 720 to 900 m g/kg, for this case. The heat and moisture fluxes are affected by $q_{v1}$ through $\Theta_1$ and $q_t$, respectively.

As shown in the figures, the heat flux decreases and the moisture flux increases with increasing $q_{v1}$. Analysis shows that for a +/-100m g/kg change in $q_{v1}$, the heat flux
Figure 4.3.5. Variation of Heat Flux with Integrated Liquid Water Content
Figure 4.3.6. Variation of Moisture Flux with Integrated Liquid Water Content
Figure 4.3.7: Variation of Heat Flux with Integrated Water Vapor Content
Figure 4.3.8. Variation of Moisture Flux with Integrated Water Vapor Content.
varies $-/+/0.6\%$ and the moisture flux varies $+/-/5\%$.

### 4.4 EXPECTED MODEL ACCURACIES WITH SATELLITE DATA

The expected model accuracies with current satellite technology is now presented. Through infrared measurements, the cloud top temperature may be found within $+/-/2\, ^\circ\mathrm{C}$. The accuracies of the sea surface temperature, the $10\text{m}$ neutral wind speed, and the friction velocity are $+/-/1.5\, ^\circ\mathrm{C}$, $+/-/2\text{m/s}$ or $20\%$, and $+/-/20\%$, respectively (Satellite Microwave Remote Sensing, edited by T.D. Allan, 1983). These values were obtained from results of SEABAT and NIMBUS 7 data analysis. The expected retrieval errors of the integrated liquid and water vapor contents for the SMMR package on NIMBUS 5 are $+/-/650\text{g/kg}$ and $+/-/1500\text{g/kg}$, respectively (Chang and Wilheit, 1979).

By applying these accuracies to the percent variations of the model discussed in the previous section, the expected accuracy of the model may be found. Since the difference of the variations with respect to $u_{10n}$ and $u_*$ is $1\%$, the larger of the two errors will be used.

Given a $20\%$ error in satellite derived $u_{10n}$, the model will produce a $20\%$ error in both the heat and
moisture fluxes. A 2 °C error in $\Theta_b$ produces a 8% error in the heat flux and a 9% error in the moisture flux. An error of 1.5 °C in the sea surface temperature gives a 4.5% error in the heat flux and an 6.75% error in the moisture flux. A 650 m g/kg error in $q_{li}$ produces a 1.6% error in the heat flux and a 22.75% error in the moisture flux. An error of 1500 m g/kg for $q_{vi}$ gives a 9% error in the heat flux and a 75% error in the moisture flux. Taking these errors to be independent of each other, it is possible to compute the expected r.m.s. error of the model. For the errors listed above for each parameter, the expected overall r.m.s. error of the model is 24% for the heat flux and 82% for the moisture flux (see Table 4.4.1). Although these errors are quite high, it must be understood that the satellite technology which allows the measurements of the necessary parameters is still in its infancy. The major errors in the model may be attributed to the calculations of $q_{li}$ and $q_{vi}$. Future improvement in the algorithms used to find these parameters would significantly improve this model's estimates of the surface fluxes.
Table 4.4.1. Individual Expected Errors due to Each Satellite Parameter and Total Expected Model Error

<table>
<thead>
<tr>
<th>PARAMETER</th>
<th>SATELLITE RMS ERRORS</th>
<th>HEAT</th>
<th>MOISTURE</th>
</tr>
</thead>
<tbody>
<tr>
<td>$u_{10n}$</td>
<td>20%</td>
<td>20%</td>
<td>20%</td>
</tr>
<tr>
<td>$\Theta_b$</td>
<td>2.0 °C</td>
<td>8%</td>
<td>9%</td>
</tr>
<tr>
<td>$\Theta_0$</td>
<td>1.5 °C</td>
<td>4.5%</td>
<td>6.75%</td>
</tr>
<tr>
<td>$q_{l1}$</td>
<td>650 m g/kg</td>
<td>1.6%</td>
<td>22.75%</td>
</tr>
<tr>
<td>$q_{v1}$</td>
<td>1500 m g/kg</td>
<td>9%</td>
<td>75%</td>
</tr>
</tbody>
</table>

Total Expected R.M.S. Errors: 24%  62%
4.5 STRENGTHS AND WEAKNESSES

The model's strengths and weaknesses are now discussed. The model handles well in cloudy and cloudy with fog conditions. The model is extremely quick, and, given reasonable input accuracies, will produce reasonable estimates of the surface fluxes. As far as the author can tell, the model is the only one currently available which will provide estimates of the surface fluxes for a cloud topped MABL from satellite measurable parameters.

On the other hand, the model is not capable of calculating the surface fluxes when a cloud is not present. The calculation of the fluxes is also unclear when gaps in the clouds are present. If the gaps are large enough, the calculation of the cloud top temperature becomes difficult, and the profiles assumed in the model break down. The model's assumption that the cloud top is also the base of the inversion is not always correct. The cloud may extend into the inversion layer; however, due to the uncertainty of how far into the cloud the infrared cloud top temperature is actually being measured, this weakness may be corrected. The presence of fog below the cloud layer also presents a problem due to the
uncertainty of how the model's equations react to fog. The main weakness of the model is the parameterization of the water vapor above the boundary layer. The mean values chosen are not representative of all possible cases, and further work in this area is needed.
In conclusion, this thesis has produced a viable model to estimate the surface fluxes from satellite measurable parameters. The model applies to cases of cold air outbreak, California coastal stratus, or to any cloud topped MABL which has a sufficiently strong heat flux or cloud top radiative cooling to cause the potential temperature and total water profiles of the layer to be well-mixed. With improvements of the current satellite technology, this model will provide accurate estimates of the surface fluxes. The model is relatively simple and extremely quick running. The model is also the only one of its kind currently available to estimate the surface fluxes of momentum, heat, and moisture of the cloud topped marine atmospheric boundary layer from satellite measurable parameters.

In light of this thesis, some areas of further research are suggested. More work is necessary to find a method of parameterizing the water vapor content above the MABL. More research is also necessary for testing the model for the case of California coastal stratus to see
how well it reacts under nearly neutral conditions. The evaluation of the surface fluxes under foggy conditions and how well the model handles them is another area for study. The problem of gaps in the cloud deck and a method to handle this is an area for further thought. Also, other possible sets of parameters should be examined to see if a better method than the one presented may be found.
REFERENCES


