UTC-M MESOSCALE ATMOSPHERIC MODEL IMPROVEMENTS – MPI PARALLELIZATION & SATELLITE BASED ASSIMILATION OF WATER AND LAND SURFACE TEMPERATURES

by

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The objective of this thesis was to improve the UTC-M Florida Tech Mesoscale Atmospheric Boundary Layer Model (Bostater, Gimond, Uhlhorn, McNally, 2000). First, a parameterization of land surface temperature is achieved with the coupling of a radiation sub model at the surface and a thermal inertia parameterization from satellite data. Second, an increase in resolution from 10km x 10km to 1km x 1km was applied. The increase in resolution increased the computer time, and therefore the model code was parallelized to run on a 48 node Beowulf supercomputer.

The thermal subroutine used in this research is based on Bostater and McNally (1996), McNally (1997) and was inserted in the UTC-M mesoscale model to improve the parameterization of land and ocean surface temperatures using
MODIS satellite data. The land and ocean surface temperatures are calculated through the atmospheric transfer model of Gregg and Carder (1990) coupled with a thermal inertia model due to the research of Gannon (1978) and Sobrino et al. (1991, 1994, 1999). The thermal model subroutine enables an estimation of the relation between solar radiation and land surface characteristics. These model subroutines hopefully allow a better representation of the change in surface temperatures for different land surfaces during a 24-hour period simulation. Thus, better representation of the sea breeze phenomena is predicted. The thermal model subroutine uses satellite data such as surface reflectance, emissivity, ozone burden, and angstrom exponent from MODIS and brightness temperatures from AVHRR. The thermal model subroutine outputs data including total downwelling radiation, upwelling short and longwave radiation, net energy budget, and thermal inertia and allows for a better understanding of land and ocean surface temperature changes during a 24-hour sea breeze simulation.

Model outputs suggest the importance of an accurate representation of land surface temperatures. The close relationship between land surface temperature differences over different land uses over the 1 km model domain is shown in the atmospheric sea breeze circulation. Low-level atmospheric convergence / divergence are shown to be enhanced over areas of sharp temperatures gradients that coincide over sea-land margins of the Indian River Lagoon.
The insertion of a thermal model subroutine and increased grid resolution to 1 km necessitated the parallelization of the UTC-M mesoscale model. After profiling the model, the most time consuming subroutine was parallelized. These were the horizontal advection subroutine and the thermal subroutine. A substantial gain of 3 hours was possible with the use of 16 processors. Future improvements could be achieved with additional parallelization of the model’s finite difference calculations.

Future improvements are needed to better represent the sea breeze phenomenon. A parameterization of latent and sensible heat in the net budget equation for calculation of land and ocean surface temperatures would be useful. An accurate representation of the dynamic state of the sea-surface with a more representative ocean current field over the Gulf Stream could also improve model simulations by providing more realistic bottom boundary and initial conditions over water. The next step should be nesting the UTC-M model with synoptic weather forecast models. In this thesis research, conservation of thermal energy was not conducted, thus limiting the results. Such an analysis will have to be conducted in order to validate the thermal model developed before it can be used for practical applications.
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List of Symbols

The International System (SI) of units is used exclusively for purpose of this research and the m.k.s. (meter, kilogram, second) system is applied. All symbols are defined in terms of the fundamental units. Dimensions of the fundamental physical quantities are expressed as follows:

L  Length expressed in units of meters (m).
M  Mass expressed in units of kilograms (kg).
T, t  Time expressed in units of seconds (s).
K  Temperature expressed in units of Kelvin (K).

θ  Instantaneous potential temperature (K).
θ o  Minimum value for a diurnal cycle (K).
θ* (t)  Time dependent departure of temperature from the minimum value (K).

u, v, w  Zonal (positive eastward), meridional (positive northward), vertical (positive upward) Cartesian components of velocity (m s-1).
q  Specific humidity (g kg-1).
R  Universal gas constant (kg m2 s-2 mol-1).
ρ  Density (kg m-3).
P, p  Pressure (kg m-1 s-2).
\( g \)
Gravity (m \( s^{-2} \)).

\( f \)
Coriolis parameter (s\(^{-1} \)).

\( u_* \)
Friction velocity (m \( s^{-1} \)).

\( K \)
Eddy-diffusivity coefficient

\( \theta_* \)
Scale potential temperature (K).

\( q_* \)
Scale specific humidity (g kg\(^{-1} \)).

\( \vec{u}(z_0), \vec{v}(z_0) \)
Zero over land, and are equal to the ocean current speed over water.

\( k \)
Von Karman constant (dimensionless).

\( \Psi_{M,H,Q} \)
Dimensionless wind shear, temperature and humidity gradient respectively.

\( Q(\lambda) \)
Net radiation at the surface (W m\(^{-2} \)).

\( K(\lambda) \uparrow \)
Upwelling reflected shortwave (solar) radiation (W m\(^{-2} \)).

\( K(\lambda) \downarrow \)
Downwelling shortwave radiation transmitted through the air (Wm\(^{-2} \)).

\( I(\lambda) \uparrow \)
Longwave (infrared, IR) radiation emitted up (W m\(^{-2} \)).

\( I(\lambda) \downarrow \)
Longwave diffuse IR radiation down (W m\(^{-2} \)).

\( E_{diff}^d (\lambda) \)
Diffuse downwelling solar irradiance (W m\(^{-2} \) nm\(^{-1} \)).

\( E_{dir}^d (\lambda) \)
Direct downwelling solar irradiance (Wm\(^{-2} \)nm\(^{-1} \)).

\( \nu \)
Dimensionless ratio of indirect to direct sunlight.

\( S_o (\lambda) \)
Mean extraterrestrial irradiance corrected for earth-sun distance and orbital eccentricity (Wm\(^{-2} \)nm\(^{-1} \)).

\( H_0 \)
Solar constant (W m\(^{-2} \)).

\( D \)
Earth-sun distance factor (dimensionless).

xx
\[ \theta \] Solar zenith angle (Wm-2nm-1).

\[ T_r(\lambda) \] Transmittance after Rayleigh scattering (dimensionless).

\[ T_a(\lambda) \] Transmittance after aerosol scattering (dimensionless).

\[ T_{oc}(\lambda) \] Transmittance after ozone, oxygen and water vapor absorption (dimensionless).

\[ T_s(\lambda) \] Transmittance after adsorption by uniformly mixed gases (nitrogen and oxygen) (dimensionless).

\[ I_r(\lambda) \] Diffuse component of irradiance arising from Rayleigh scattering after molecular absorption (W m-2 nm-1).

\[ I_a(\lambda) \] Diffuse component of aerosol scattering after molecular absorption (Wm-2nm-1).

\[ I_g(\lambda) \] Diffuse component of irradiance arising from ground-air multiple interactions (W m-2 nm-1).

\[ a(\lambda) \] Albedo or surface reflectance of the surface (dimensionless).

\[ \varepsilon(\lambda) \] Emissivity of the surface (dimensionless).

\[ \sigma \] Stephan-Boltzmann constant (W m-2 K-4).

\[ T \] Temperature of the surface (Kelvin).

\[ E \] Total emissive power of a surface (W m-2).

\[ E_b \] Total emissive power of an ideally radiating surface (blackbody) at the same temperature (W m-2).

\[ \delta \] Solar declination (degrees).

\[ n \] Julian day of the year.

\[ E_{qt} \] Equation of time correction (hours).
$H_s$  True solar time (hours).

$H_{ls}$  Local standard time (hours).

$L_{sm}$  Longitude of the standard meridian of the time zone (degrees).

$L_{og}$  Longitude of the grid point (degrees).

$h$  Hour angle about solar noon (degrees).

$H_{sr}$  Time of sunrise after midnight (hours).

$\theta$  Zenith angle (degrees).

$\phi$  Latitude of the grid point (degrees).

$M(\theta)$  Non-dimensional atmospheric path length.

$M'$  Path length corrected for nonstandard atmospheric pressure.

$P_o$  Standard atmospheric pressure (101.325 kPa).

$M_{oz}$  Ozone path length.

$D$  Extraterrestrial irradiance corrected for earth-sun distance (Wm$^{-2}$).

$\varphi$  Day angle (radians).

$\lambda$  Wavelength (m).

$W$  Precipitable water vapor (cm) in a vertical path.

$a_w$  Water vapor absorption coefficient (dimensionless).

$H_{oz}$  Ozone scale height (km).

$a_{oz}$  Ozone absorption coefficient (km$^{-1}$).

$a_u$  Combination of an absorption coefficient and gaseous amount (dimensionless).

$\tau_a$  Aerosol transmittance coefficient (dimensionless).

$\beta$  Turbidity non-dimensional coefficient from the aerosol
concentration.

\( \alpha \)
Angstrom non-dimensional exponent.

\( H_a \)
Aerosol scale height (km-1)

\( c_a(550) \)
Aerosol extinction coefficient at 550 nm (km-1).

\( \tau_a(550) \)
Aerosol optical thickness at 550 nm (dimensionless).

\( V \)
Visibility (km).

\( T_{a a \lambda} \)
Transmittance term for aerosol absorption (dimensionless).

\( T_{a s \lambda} \)
Transmittance term for aerosol scattering (dimensionless).

\( F_s \)
Fraction of the aerosol scatter downward.

\( r_{g \lambda} \)
Ground albedo as a function of wavelength (dimensionless).

\( r_{s \lambda} \)
Sky reflectivity (dimensionless).

\( T'_{a s \lambda} \)
Primed transmittance terms (dimensionless).

\( F'_s \)
Primed fraction of the aerosol scatter downward (dimensionless).

\( P \)
Thermal inertia (1 TIU = 1 W s1/2 m-2 K-1 = cal s-1/2 cm-2 oC-1).

\( K \)
Thermal conductivity (J m-1 s-1 K-1).

\( c \)
Specific heat of the material (J kg-1 K-1).

\( A_n \)
Fourier coefficients of \( f \).

\( B_n \)
Fourier coefficients of \( f \).

\( C_t \)
Atmospheric transmittance (dimensionless).

\( R_{\text{earth}} \)
Earth emitted radiation.

\( R_{\text{sky}} \)
Downward longwave sky radiation.

\( H \)
Sensible heat flux to the atmosphere.

\( LE \)
Latent heat flux to the atmosphere.
Linearization coefficients of the boundary condition.
Angular velocity of rotation of the Earth (rad s⁻¹).
Day-night surface temperature difference.
Time (hours) of diurnal and nocturnal satellite passes.
Phase angle (radians).
Brightness temperatures in channel 4 and 5 of AVHRR.
Latitude of the region studied (radians).
Digital number with a solar-zenith angle of 0 degrees.
Digital number with an angle of 0 degrees.
Central wave number of channel filter (cm⁻¹).
Digital number from AVHRR channel 3, 4 or 5.
Constants appended from the AVHRR file.
Weight of a nearest neighbor.
Distance of nearest neighbor from point of interest.
Initial vertical soil temperature profile (°C).
Incremental flux acting over \( d \gamma \) (cal cm⁻² s⁻¹).
Diffusivity (cm² s⁻¹).
Same as the thermal inertia (TI).
Temperature at the next time step \( n + 1 \) (Kelvin).
Initial temperature at the current time step (Kelvin).
Net flux across the air-ground interface, which is considered constant over \( \Delta t \) (cal cm⁻² s⁻¹).
Time step (seconds).
Grid cell size (meters).
<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>$C$</td>
<td>Speed of the fastest wave permitted by the equations (m/s).</td>
</tr>
<tr>
<td>$\overline{U}$</td>
<td>Average wind speed (m/s).</td>
</tr>
<tr>
<td>$c$</td>
<td>Specific heat of air at constant pressure.</td>
</tr>
<tr>
<td>$z_o$</td>
<td>Surface roughness (meters).</td>
</tr>
<tr>
<td>$z$</td>
<td>Height of wind and air temperature sensors (meters).</td>
</tr>
<tr>
<td>$L$</td>
<td>Latent heat of vaporization (J/kg).</td>
</tr>
<tr>
<td>$\overline{h_a}$</td>
<td>Average humidity of the air (g/m³).</td>
</tr>
<tr>
<td>$h_g$</td>
<td>Average humidity at the ground surface (g/m³).</td>
</tr>
</tbody>
</table>
### List of Abbreviations

<table>
<thead>
<tr>
<th>Abbreviation</th>
<th>Full Form</th>
</tr>
</thead>
<tbody>
<tr>
<td>MM5</td>
<td>Fifth-generation NCAR / Penn State Mesoscale Model</td>
</tr>
<tr>
<td>NGM</td>
<td>Nested Grid Model</td>
</tr>
<tr>
<td>NORAPS6</td>
<td>Navy Operational regional Atmospheric prediction System version 6</td>
</tr>
<tr>
<td>RAMS</td>
<td>Regional Atmospheric Modeling System</td>
</tr>
<tr>
<td>RWM</td>
<td>Relocatable Window Model</td>
</tr>
<tr>
<td>SST</td>
<td>Sea Surface Temperatures</td>
</tr>
<tr>
<td>UTC-M</td>
<td>U Turbulent Closure – Methodology</td>
</tr>
<tr>
<td>NCEP</td>
<td>National Centers for Environmental Predictions</td>
</tr>
<tr>
<td>NOAA</td>
<td>National Oceanic and Atmospheric Administration</td>
</tr>
<tr>
<td>AVHRR</td>
<td>Advanced Very High Resolution Radiometer</td>
</tr>
<tr>
<td>EDT</td>
<td>Eastern Daylight Time</td>
</tr>
<tr>
<td>LST</td>
<td>Local Standard Time</td>
</tr>
<tr>
<td>Z</td>
<td>Zulu</td>
</tr>
<tr>
<td>GIS</td>
<td>Geographic Information Systems</td>
</tr>
<tr>
<td>SGI</td>
<td>Silicon Graphics Incorporated</td>
</tr>
<tr>
<td>MIPS</td>
<td>Million Instructions Per Second</td>
</tr>
<tr>
<td>MODIS</td>
<td>Moderate Resolution Imaging Spectroradiometer</td>
</tr>
<tr>
<td>Acronym</td>
<td>Description</td>
</tr>
<tr>
<td>---------</td>
<td>-------------</td>
</tr>
<tr>
<td>IBM</td>
<td>International Business Machines</td>
</tr>
<tr>
<td>MPI</td>
<td>Message Passage Interface</td>
</tr>
<tr>
<td>ENVI</td>
<td>The Environment for Visualizing Images</td>
</tr>
<tr>
<td>EOS</td>
<td>Earth Observing System</td>
</tr>
<tr>
<td>IR</td>
<td>Infrared</td>
</tr>
<tr>
<td>EOS-AM-1</td>
<td>Earth Observation System AM-1 spacecraft</td>
</tr>
<tr>
<td>ESE</td>
<td>Earth Sciences Enterprise (NASA)</td>
</tr>
<tr>
<td>ASTER</td>
<td>Advanced Spaceborne Thermal Emission and Reflection Radiometer</td>
</tr>
<tr>
<td>CERES</td>
<td>Clouds and the Earth's Radiant Energy System</td>
</tr>
<tr>
<td>MISR</td>
<td>Multi-angle Imaging Spectro-Radiometer</td>
</tr>
<tr>
<td>MOPITT</td>
<td>Measurements of Pollution in the Troposphere</td>
</tr>
<tr>
<td>BRDF</td>
<td>Bi-directional Reflectance Distribution Function</td>
</tr>
<tr>
<td>ISIN</td>
<td>Integerized Sinusoidal Projection</td>
</tr>
<tr>
<td>EDC</td>
<td>EROS Data Center</td>
</tr>
<tr>
<td>DAAC</td>
<td>Distributed Active Archive Center</td>
</tr>
<tr>
<td>DU</td>
<td>Dobson Units</td>
</tr>
<tr>
<td>STP</td>
<td>Standard Temperature and Pressure</td>
</tr>
<tr>
<td>TIU</td>
<td>Thermal Inertia Unit</td>
</tr>
<tr>
<td>SAA</td>
<td>Satellite Active Archive</td>
</tr>
<tr>
<td>NESDIS</td>
<td>National Environmental Satellite, Data, and Information Service</td>
</tr>
</tbody>
</table>
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I would like to express my appreciation to my family and friends who have supported me all the way through this adventure. I would like to thank my advisor Dr. Charles Bostater, who gave me the opportunity to do this research and be a teacher assistant. I believe both activities made me a more knowledgeable and a more mature person. I also want to thank my committee members, Dr. Charles Fulton who made me discover the world of parallel processing and Dr. Eric Thosteson for serving as one of my committee member.

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CHAPTER 1

INTRODUCTION AND BACKGROUND

1.1 General Background

To provide reliable information on the diurnal evolution of the planetary boundary layer and its dynamic characteristics, parameterization of land surface processes is needed in order to predict vertical divergence or convergence in a mesoscale model (Xiu and Pleim, 2001). Numerical atmospheric models have demonstrated the potential effects of landscape characteristics on climate. For instance, McCumber (1980) applied a mesoscale model to evaluate the effect of vegetation on the development of the summer sea breeze over south Florida. This study showed that sharp horizontal changes in the character and type of vegetation cover would induce significant mesoscale circulation. Several other authors have demonstrated the impact of land surface and vegetation on mesoscale circulation (Garrett 1982, Yamada 1982, Anthes 1984). The landscape and land use strongly impact atmospheric circulations and atmospheric boundary layer vertical structure. Thus parameterization of the earth’s surface is one of the more important aspects of mesoscale atmospheric modeling and is needed to insure that land surface temperatures and temperatures flux from the land to air are included in successful simulations with mesoscale models as demonstrated using the UTC-M model by
Many mesoscale models have been developed in the past 30 years, and scientists keep improving them to better simulate and forecast weather. The MM5 (the fifth-generation NCAR / Penn State Mesoscale Model) is one of the most popular mesoscale models. The MM5 is used for studies involving convective systems, fronts, land-sea breeze, mountain-valley circulation, and urban heat islands. Another popular model is the ETA model, which uses a mathematical coordinate system that takes into account topographical features such as mountains. The NGM model (Nested Grid Model) is a short-range model that forecasts variables such as temperature at various levels of the atmosphere, amount of precipitation, position of upper level troughs and ridges, and the position of surface high and low pressure areas. The Navy Operational regional Atmospheric prediction System version 6 (NORAPS6) is a regional-scale, primitive equation, hydrostatic model that uses a split-explicit time integration scheme to predict dynamic and thermodynamic variables (Hodur, 1987). NORAPS focuses on ocean area forecasts (i.e. NORAPS6 models sea ice). The Regional Atmospheric Modeling System (RAMS) is a regional-scale primitive equation model that can be configured hydrostatically or non-hydrostatically (Pielke, 1992). RAMS applications revolve around micro to mesoscale atmospheric weather forecasting (cloud scale to extra tropical cyclone scale). The Relocatable Window Model (RWM) is a regional-scale primitive equation model that uses a quasi-
Lagrangian advection scheme to predict u and v wind components, potential
temperature, surface pressure, and specific humidity (Englehart et al., 1993). A
complete comparison of mesoscale models can be found in table A1.

In 1996, Bostater et al. developed a UTC-M mesoscale numerical model to
simulate convergence and divergence at the land-water margin (Bostater, Gimond,
Uhlhorn, McNally, 2000). Conclusions have shown the importance of an accurate
representation of the dynamic state of the sea-surface for modeling of coastal
atmospheric mesoscale phenomena (Bostater, Huddleston and King, 2002). Over
the Space Coast of Central Florida, low-level atmospheric convergence and
divergence is enhanced over areas of sharp sea surface temperature (SST)
gradients. The results described the application of a three-dimensional mesoscale
model to a complex air-land-sea margin coastal atmosphere where there are large
lateral sea-surface temperature gradients as well as land-sea margin thermal
gradients. The above marine boundary layer of the numerical model was initialized
using sea-surface temperatures derived from high-resolution (~ 1.1 km) AVHRR
infrared imagery (NOAA, 1990). The spatial resolution of the initial conditions
allowed for mesoscale ocean features such as the west-wall of the Gulf Stream and
warm or cold-core eddies to influence turbulent fluxes of heat and moisture from
ocean to atmosphere (see figures 1.1-1.3) through the initial conditions.
However, many physical processes important in the atmosphere were not included such as: 1) inclusion of latent energy, 2) a radiative transfer scheme 3) a detailed parameterization of cloud physics. In 2002, the UTC-M model was modified for the inclusion of latent heat was completed by assuming that adding water vapor at lower levels has the same effect as heating at these levels (Bostater, et al. 2002). Air rich in water vapor is lighter or less dense than dry air. This is because a given volume of air contains a set number of molecules of the mixture of the various atmospheric constituents. Thus the lighter water (H$_2$O) molecules displace heavier nitrogen (N$_2$) or oxygen (O$_2$) molecules resulting in a lighter mixture. Water vapor also contains the potential for latent heat release within the lower portion of the atmosphere as the air expands and cools during ascent.
Figure 1.1 UTC-M model grid showing the 10x10 km grid cells from the original model domain from approximately Daytona Beach, Fl. to West Palm Beach, Florida. This model domain covers the North-South region of the Indian River Lagoon. The Eastern boundary of the grid lies within the eastern side of the Gulf Stream current in the Atlantic Ocean off the Florida Space coast near NASA Kennedy Space Center and lies within the St. Johns River Basin at the western boundary. The dashed line shows the long. axis location of the longitudinal axis of the Gulf Stream current (Bostater, Gimond, Uhlhorn, McNally, 2000). See Appendix B for the new 2 km and 1 km grids used in this research with the axis of the Gulf Stream for the dates of the simulation used in this research.
Figure 1.2 Boundary layer (30 m) winds in ms$^{-1}$ (see arrow) and potential temperature (dashed lines) in K, 6hrs into simulation using grided SST data Sea surface temperature derived from AVHRR imagery for 7-31-96 (19Z) (Bostater, Gimond, Uhlhorn, McNally, 2000).
Figure 1.3 Convergence field (m/s/deg) at the model 30-meter altitude after the model has run 12 hours (July 31, 1800 LDT). The dashed lines indicate the convergence field and indicates low-level convergence over the coastline. This suggests the vertical advection and increase in motion and cloud developments could be expected to occur in the intensified convergence areas long this land-water margin (Bostater, Gimond, Uhlhorn, McNally, 2000).
Another limitation of the original UTC-M model was an accurate representation of the land surface temperature change during the day. The surface temperature over land was assumed to be represented in the UTC-M model as:

\[
\theta = \theta_0 + \theta''(t),
\]

(1.1)

where:

\(\theta\) = the instantaneous potential temperature in Kelvin at the surface of the earth \((z=0)\),

\(\theta_0\) = the minimum value for a diurnal cycle,

\(\theta''(t)\) = the time dependent departure of temperature from the minimum value.

The minimum value is defined to occur at 10 Z (sunrise during the summer months in Central Florida).

\(t\) = the time in seconds.

Uhlhorn (1996) estimated the instantaneous surface temperature for Central Florida for July 16\(^{th}\) 1996 from:

\[
\theta(t) = 6.57 \times 10^{-28} t^6 - 2.354 \times 10^{-22} t^5 + 3.138 \times 10^{-17} t^4 - 1.854 \times 10^{-12} t^3 \\
+ 4.027 \times 10^{-8} t^2 + 9.573 \times 10^{-5} t + 2.976 \times 10^2.
\]

(1.2)

The above equation was a best fit of the hourly-observed temperatures reported by the National Weather Service at Melbourne, FL on 16 July 1996. The marine boundary layer (air-sea interface) of the UTC-M model was initialized using sea-surface temperatures derived from high-resolution (1.1 km x 1.1 km) AVHRR infrared imagery. It was proposed that the UTC-M mesoscale model could
be also improved by utilization of land surface boundary conditions especially, if the resolution of the grid cells were on the order of 1 km as shown by Bostater, Gimond, Huddleston and King (2002) and Bostater, King and Huddleston (2003).

One approach to parameterization of land surface temperatures can be obtained with a radiation parameterization coupled with a thermal inertia parameterization of land use types. In order to determine accurate land surface temperatures, the concept first demonstrated used by Gannon (1978) is applied in this research, using satellite data inputs. The approach also relies heavily on the excellent work of Sobrino (Sobrino et al. 1991, Sobrino et al. 1994 and Sobrino 1999) and Cracknell and Vaughan (1992).

Image processing and Geographic Information Systems (GIS) have allowed a substantial advancement in the ease of gathering and interpretation of global and geospatial regional land surface data sets that are necessary inputs to models. GIS will continue to be used to store and help create new data sets (data layers or themes) from different sources of data with different projections and spatial resolutions. In addition it is anticipated that future satellite data will be extensively used to provide necessary input variables such as albedo and emissivity from MODIS and brightness temperatures from AVHRR imagery.
Previously, the UTC-M model (Bostater, et al. 2000) was run on a 10 km x 10 km grid. At this scale, the micro convergence and divergence initiated in the Banana River or Mosquito Lagoon, for example, could not be adequately represented. A finer grid size of 1 km x 1 km was hypothesized to be appropriate to better understand the micro-convergence and divergence affected by these river/lagoon water bodies and their role on the space coast, Florida sea-breeze.

Increasing the spatial resolution of the model computational domain increases the computational time of a sea breeze cycle; therefore parallelization of the computer code is applied in this research to the model and run on the Florida Tech, IBM Beowulf, 48 node Linux parallel computer system. The Beowulf supercomputer is a high-performance massively parallel computer. The Beowulf cluster at Florida Tech is a 48-node IBM system, comprised of 47 computer nodes and 1 head node. Parallel applications programmed with the message-passing library are run on the Beowulf. MPI software systems allow one to write message-passing parallel programs that run on a cluster, in FORTRAN and C. MPI stands for Message Passing Interface. MPI is the standard for multicomputer and cluster message passing introduced by the Message-Passing Interface Forum in April 1994. The goal of MPI is to develop a widely used standard for writing message-passing programs. As such the interface attempts to establish a practical, portable, efficient, and flexible standard for message passing. Florida Tech currently
supports MPI on the IBM Beowulf System as well as a laboratory of SGI workstations.

The improved Florida Tech mesoscale model was profiled to obtain an accurate set of statistics on every subroutine and function calls. Then, the parallelization work focused on the subroutine that utilize the most significant computational time. These were the horizontal advection subroutine and the thermal subroutine.

Then, the parallelized code was profiled again and compared to the serial program. Tests were performed with 1 to 25 processors to observe any time reduction in running the parallelized code. A complete description of the profiling and parallelization is described in chapter 5.

The advantage of MPI is that it can accomplish certain kinds of work much faster than a single computer working alone. Up to this point, the UTC-M Florida Tech model had been written as a serial, or sequential, program. That is to say, the instructions inside the program have been carried out step-by-step, one at a time, on a single Silicon Graphics Inc. MIPS processor system.
The goal of this project is to allow Blue, an IBM MPI based supercomputer, to execute the program and to run it simultaneously, or “in parallel”, on a specified number of processors.

Only certain kinds of work can be run profitably (that is, faster) on a parallel processing supercomputer. As a general rule a task that can be run profitably on a parallel supercomputer will generally not run any faster on one unless it is designed and written to take advantage of the unique parallel environment. Very little commercial software has yet been written that is designed a-priori to run in a parallel environment.

This research is unique and novel in that it uses GIS, remote sensing and parallel processing techniques to improve the Florida Tech UTC-M Atmospheric Planetary Boundary Layer Mesoscale model.

The goal of this research has thus been divided into 3 distinct areas of research. First, the parameterization of the land surface processes, (2) profiling and parallelization of the code for use on a MPI Linux cluster and (3) sensitivity analysis of the model results.
1.2 General Description of the UTC-M model

The Florida Tech UTC-M atmospheric sea breeze model (Bostater et al. 2000, 2002, 2003) is a 3-dimensional, “primitive equation” model. The primitive equations are the system of equations that govern large-scale atmospheric motions and represent the synthesis of the ideal gas law, the first law of thermodynamics the Newton’s second law for the horizontal component of the motion and the conservation of mass.

1.2.1 General Equations

The primitive equations comprise 4 prognostic and 3 diagnostic equations. The 4 prognostic equations (equation 1.3 – 1.6) are the horizontal components of the momentum equation, the thermodynamic energy equation and the humidity equation. The 3 diagnostic equations (equations 1.7 – 1.9) are the continuity equation and the hydrostatic equation of state as reproduced below.

\[
\begin{align*} 
\frac{\partial \tilde{u}}{\partial t} + \frac{\partial u}{\partial x} + \frac{v \cdot \partial u}{\partial y} + \frac{w \cdot \partial u}{\partial z} &= f \nu - \frac{1}{\rho} \frac{\partial p}{\partial x} + \left( \frac{\partial u' u'}{\partial x} + \frac{\partial u' v'}{\partial y} + \frac{\partial u' w'}{\partial z} \right), \\
\frac{\partial \tilde{v}}{\partial t} + \frac{\partial v}{\partial x} + \frac{v \cdot \partial v}{\partial y} + \frac{w \cdot \partial v}{\partial z} &= -f u - \frac{1}{\rho} \frac{\partial p}{\partial y} + \left( \frac{\partial v' u'}{\partial x} + \frac{\partial v' v'}{\partial y} + \frac{\partial v' w'}{\partial z} \right), \\
\frac{\partial \tilde{\theta}}{\partial t} + \frac{\partial \theta}{\partial x} + \frac{v \cdot \partial \theta}{\partial y} + \frac{w \cdot \partial \theta}{\partial z} &= \frac{\partial \theta' u'}{\partial x} + \frac{\partial \theta' v'}{\partial y} + \frac{\partial \theta' w'}{\partial z}.
\end{align*}
\]
\[ \frac{\partial \bar{q}}{\partial t} + u \frac{\partial \bar{q}}{\partial x} + v \frac{\partial \bar{q}}{\partial y} + w \frac{\partial \bar{q}}{\partial z} = \frac{\partial (\bar{q}'u')}{\partial x} + \frac{\partial (\bar{q}'v')}{\partial y} + \frac{\partial (\bar{q}'w')}{\partial z}, \]  
(1.6)

\[ \rho_s \left( \frac{\partial \bar{u}}{\partial x} + \frac{\partial \bar{v}}{\partial y} \right) + \frac{\partial (\rho_s \bar{w})}{\partial z} = 0, \]  
(1.7)

\[ \frac{\partial \bar{p}}{\partial z} = -\rho_s g, \]  
(1.8)

\[ \bar{p} = \rho_s \bar{T}. \]  
(1.9)

These 4 prognostic equations are written for a turbulent fluid. The separation of the mean motion and the turbulent perturbations is accomplished through the Reynolds decomposition. The results of the Reynolds decomposition contain unknown turbulent flux terms. These terms are parameterized through a turbulent closure parameterization where the prognostic expressions are used to determine the mean variables and the turbulent fluxes estimated based on observations. The flux terms are assumed to be proportional to the local gradient of the mean variables (Uhlhorn, 1996). For the case of x-directed momentum, the flux term with the proportionality constant \( K \) is:

\[ \bar{u}'u' = K \frac{\partial \bar{u}}{\partial x}, \]  
(1.10)

where:

\( K \) = the parameterization relating the turbulent fluctuations to the gradient of the mean flow. It is often called the eddy-diffusivity coefficient.
1.2.2 Surface Data Inputs

Sea-level pressure and dewpoint temperature are obtained from 10 reporting stations across the East Central Florida area (Table 1.1). The pressure and dewpoint observations are interpolated using an exponentially weighted Kriging method. Sea and land surface initial values are determined from the thermal model (see chapter 4). In the vertical domain, upper air balloons soundings are used for measurements of wind speed and direction, temperature, and dewpoint temperature.

Table 1.1 Station information used in the model initialization process. $P_{sfc}$ is the sea level pressure in millibars and $T_d$ is the dewpoint temperature in degrees Celsius.

<table>
<thead>
<tr>
<th>Station</th>
<th>Identifier</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Data Obtained</th>
</tr>
</thead>
<tbody>
<tr>
<td>Daytona Bch.</td>
<td>DAB</td>
<td>29.18</td>
<td>-81.06</td>
<td>$P_{sfc}$, $T_d$</td>
</tr>
<tr>
<td>Gainesville</td>
<td>GNV</td>
<td>29.69</td>
<td>-80.61</td>
<td>$P_{sfc}$, $T_d$</td>
</tr>
<tr>
<td>Orlando (Intl)</td>
<td>MCO</td>
<td>28.43</td>
<td>81.32</td>
<td>$P_{sfc}$, $T_d$</td>
</tr>
<tr>
<td>W. Palm Bch.</td>
<td>PBI</td>
<td>26.68</td>
<td>-80.1</td>
<td>$P_{sfc}$, $T_d$</td>
</tr>
<tr>
<td>Vero Bch.</td>
<td>VRB</td>
<td>27.66</td>
<td>-80.42</td>
<td>$P_{sfc}$, $T_d$</td>
</tr>
<tr>
<td>Patrick A.F.B.</td>
<td>COF</td>
<td>28.24</td>
<td>-80.61</td>
<td>$P_{sfc}$, $T_d$</td>
</tr>
<tr>
<td>Titusville</td>
<td>TTS</td>
<td>28.61</td>
<td>-80.81</td>
<td>$P_{sfc}$, $T_d$</td>
</tr>
<tr>
<td>St. Augustine</td>
<td>SAUF1</td>
<td>29.90</td>
<td>-80.20</td>
<td>$P_{sfc}$, $T_d$</td>
</tr>
<tr>
<td>Buoy 9</td>
<td>41009</td>
<td>28.5</td>
<td>-80.20</td>
<td>$P_{sfc}$</td>
</tr>
<tr>
<td>Buoy 10</td>
<td>41010</td>
<td>28.9</td>
<td>-78.5</td>
<td>$P_{sfc}$</td>
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</table>

1.2.3 Numerical Model Domain

The numerical domain can be divided into 2 sections, the horizontal and vertical domain. According to Hsu (1988), the sea breeze phenomenon has a horizontal extend of 200 km or less. Vertical scales can extend from tens of meters to the depth of the troposphere. The model grid covers the east coast of Central
Florida with one-half of the grid over land and the other half over the ocean. (see figure 1.1). The grid spacing used in the original model is 10 km and contained a total of 25 x 25 points. In the improved model, the grid spacing is 1 km and contains 200 x 200 points. The spatial coverage of 200 km x 200 km is assumed to be enough to capture the location of the Gulf Stream. Open-type boundary conditions are used for the lateral boundary conditions where the method is achieved by definition of a zero gradient of all variables at the boundaries:

$$\frac{\partial \phi}{\partial x}, \frac{\partial \phi}{\partial y} = 0.$$  \hspace{1cm} (1.11)

The vertical domain extends from the earth’s surface to an altitude of 10 km. The domain is divided into sixteen log-linearly space levels. The grid resolution decreases from the earth’s surface up to the top of the free atmosphere. The vertical structure of the atmosphere can be expressed through the set of equations seen below:

Surface (constant flux) layer:

$$\bar{u}(z) = \bar{u}(z_0) + \left(\frac{u_*}{k}\right) \left[ \ln \left( \frac{z}{z_0} \right) - \Psi_M \right] \cos \alpha , \hspace{1cm} (1.12)$$

$$\bar{v}(z) = \bar{v}(z_0) + \left(\frac{u_*}{k}\right) \left[ \ln \left( \frac{z}{z_0} \right) - \Psi_M \right] \sin \alpha , \hspace{1cm} (1.13)$$

$$\bar{\theta}(z) = \bar{\theta}(z_0) + 0.74 \theta \left[ \ln \left( \frac{z}{z_0} \right) - \Psi_H \right] , \hspace{1cm} (1.14)$$
\( \bar{q}(z) = \bar{q}(z_0) + 0.74q_o \left[ \ln \left( \frac{z}{z_0} \right) - \Psi_q \right] \), \hspace{1cm} (1.15)

where:
\( \bar{u}(z_0), \bar{v}(z_0) = \) zero over land, and are equal to the ocean current speed over water.

**Transition Layer:**
\[
\begin{align*}
\frac{\partial \bar{u}}{\partial t} + u \frac{\partial \bar{u}}{\partial x} + v \frac{\partial \bar{u}}{\partial y} + w \frac{\partial \bar{u}}{\partial z} &= f \bar{v} - \frac{1}{\rho_o} \frac{\partial \bar{p}}{\partial x} + \frac{\partial (\bar{u}' \bar{u}')}{\partial x} + \frac{\partial (\bar{u}' \bar{v}')}{\partial y} + \frac{\partial (\bar{u}' \bar{w}')}{\partial z} , \\
\frac{\partial \bar{v}}{\partial t} + u \frac{\partial \bar{v}}{\partial x} + v \frac{\partial \bar{v}}{\partial y} + w \frac{\partial \bar{v}}{\partial z} &= -f \bar{u} - \frac{1}{\rho_o} \frac{\partial \bar{p}}{\partial y} + \frac{\partial (\bar{v}' \bar{u}')}{\partial x} + \frac{\partial (\bar{v}' \bar{v}')}{\partial y} + \frac{\partial (\bar{v}' \bar{w}')}{\partial z} , \\
\frac{\partial \bar{\theta}}{\partial t} + u \frac{\partial \bar{\theta}}{\partial x} + v \frac{\partial \bar{\theta}}{\partial y} + w \frac{\partial \bar{\theta}}{\partial z} &= \frac{\partial (\bar{u}' \bar{\theta}')}{\partial x} + \frac{\partial (\bar{v}' \bar{\theta}')}{\partial y} + \frac{\partial (\bar{w}' \bar{\theta}')}{\partial z} , \\
\frac{\partial \bar{q}}{\partial t} + u \frac{\partial \bar{q}}{\partial x} + v \frac{\partial \bar{q}}{\partial y} + w \frac{\partial \bar{q}}{\partial z} &= \frac{\partial (\bar{u}' \bar{q}')}{\partial x} + \frac{\partial (\bar{v}' \bar{q}')}{\partial y} + \frac{\partial (\bar{w}' \bar{q}')}{\partial z} ,
\end{align*}
\] \hspace{1cm} (1.16-1.19)

**Free-atmosphere:**
\[
\begin{align*}
\frac{\partial \bar{u}}{\partial t} + u \frac{\partial \bar{u}}{\partial x} + v \frac{\partial \bar{u}}{\partial y} + w \frac{\partial \bar{u}}{\partial z} &= f \bar{v} - \frac{1}{\rho_o} \frac{\partial \bar{p}}{\partial x} , \\
\frac{\partial \bar{v}}{\partial t} + u \frac{\partial \bar{v}}{\partial x} + v \frac{\partial \bar{v}}{\partial y} + w \frac{\partial \bar{v}}{\partial z} &= -f \bar{u} - \frac{1}{\rho_o} \frac{\partial \bar{p}}{\partial y} , \\
\frac{\partial \bar{\theta}}{\partial t} + u \frac{\partial \bar{\theta}}{\partial x} + v \frac{\partial \bar{\theta}}{\partial y} + w \frac{\partial \bar{\theta}}{\partial z} &= 0 , \\
\frac{\partial \bar{q}}{\partial t} + u \frac{\partial \bar{q}}{\partial x} + v \frac{\partial \bar{q}}{\partial y} + w \frac{\partial \bar{q}}{\partial z} &= 0 ,
\end{align*}
\] \hspace{1cm} (1.20-1.23)

46
where:

- $u, v, w$ = the zonal (positive eastward), meridional (positive northward), vertical (positive upward) Cartesian components of velocity ($\text{m s}^{-1}$),
- $\theta$ = the potential temperature in Kelvin,
- $q$ = the specific humidity (g kg$^{-1}$),
- $u_*$ = the friction velocity (m s$^{-1}$),
- $\theta_*$ = the scale potential temperature (K),
- $q_*$ = the scale specific humidity (g kg$^{-1}$),
- $k$ = the Von Karman constant (dimensionless),
- $\Psi_{M,H,Q}$ = the dimensionless wind shear, temperature and humidity gradient respectively.

The vertical structure of the atmosphere in the UTC-M Mesoscale model is shown below:

![Vertical Model Domain](image)

Figure 1.4 Vertical Model Domain. The UTC-M Mesoscale model consists of sixteen grid levels and three physical layers (Uhlhorn, 1996).

### 1.2.4 Methods of Solution

All prognostic equations described in above sections are non-linear partial differential equations. In order to solve these equations, numerical methods are required to seek solutions that analytical methods cannot solve. Finite differences
are the numerical methods used in the UTC-M Mesoscale model. The forward-time/centered-space (FTCS) finite difference approximation for the u-wind field at time $\Delta t$ in the future is:

$$u_{j}^{n+1} = -\frac{u_{j}^{n}\Delta t}{2\Delta x} \left( u_{j+1}^{n} - u_{j-1}^{n} \right) + u_{j}^{n}.$$  \hspace{1cm} (1.24)

This method is however numerically unstable and can be corrected with a staggered-leapfrog finite difference method where information at a future time step (n+1) is calculated from the (n-1) time step. This method was applied for the estimation of horizontal advection. More information on the UTC-M Mesoscale model can be found in Eric Uhlhorn’s thesis (Uhlhorn, 1996).

A flow chart of the Mesoscale Model can be found in figures 1.5 and 1.6.

The next chapter discusses the net surface radiation sub model of the thermal model. Chapter three presents the thermal inertia sub model of the thermal model. Techniques to estimate land surface temperatures are discussed in chapter four. Parallelization and data insertion techniques are explained in chapter five. Finally the results of the thermal model and the UTC-M mesoscale model results are described in chapter six. Results in simulation are discussed in chapter seven. A figure summarizing the complete steps of the research can be found in figure 1.7.
Figure 1.5 Flow chart summarizing the first 2 steps in the UTC-M atmospheric planetary boundary layer mesoscale model: the initialization of the model and the different model runs.
Figure 1.6 Flow chart summarizing the last step in the UTC-M atmospheric planetary boundary layer mesoscale model: the main model program with the outputs.
Figure 1.7 Flow chart describing the complete project. From the creation of the input files with the use of remote sensing techniques such as ENVI and MODIS reprojection tool but also with the use of GIS techniques such as ArcView to the insertion of the thermal sub-model to the UTC-M model and finally the parallelization with the Beowulf IBM-cluster.
CHAPTER 2

THERMAL SUBROUTINE PART 1

NET SURFACE RADIATION SIMULATION

2.1 Earth Surface Radiation

The Earth's climate is governed from the outside by solar energy, which itself depends on both the power radiated by the Sun and the Earth's position in relation to the Sun. As it is at a very high temperature, around 6000 °K, most of the sun’s energy is radiated as short waves (from 0.3 to 0.7 µm), in the visible and near infrared (from 0.7 to 3 µm), (Iqbal, 1983).

The original aspect of this research is that land surface radiation is calculated for 541 bands (439 bands in the shortwave range and 102 in the longwave range). Most research calculates radiation estimation at one wavelength, which is not as representative as using the entire spectrum. Atmospheric temperature plays a key role in the energy balance of our planet, which is heated by the absorption of solar radiation and cooled by the emission of thermal IR radiation. In order to include radiative effects in simulations of the mesoscale atmosphere, radiation should be grouped into 3 types:

Short-wave radiation (0.295µm – 0.860µm).
Mid-infrared and infrared radiation (0.861\(\mu\)m – 3\(\mu\)m).

Longwave radiation (11\(\mu\)m - 12\(\mu\)m), range for emissivity estimation.

The reason why the shortwave and the longwave radiation can be handled as two separate wavelength dependent components is explained below. The net radiation can be divided into four components:

\[
Q(\lambda) = K(\lambda) \uparrow + K(\lambda) \downarrow + I(\lambda) \uparrow + I(\lambda) \downarrow, \quad (2.1)
\]

where:

\(K(\lambda) \uparrow\) = the upwelling reflected shortwave (solar) radiation (W m\(^{-2}\)),

\(K(\lambda) \downarrow\) = the downwelling shortwave radiation transmitted through the air (W m\(^{-2}\)),

\(I(\lambda) \uparrow\) = the longwave (infrared, IR) radiation emitted up (W m\(^{-2}\)),

\(I(\lambda) \downarrow\) = the longwave diffuse IR radiation down (W m\(^{-2}\)).

Figures 2.1 and 2.2 illustrate equation (2.1).
2.2 Downwelling Radiation

2.2.1 The Solar Radiation

Solar radiation absorbed at the earth’s surface and in the atmosphere is the initial source of energy causing atmospheric motions. A parameterization is needed for a better understanding of the effects of radiation on the atmospheric circulation. The downward solar radiation is at its maximum from 0.45-0.50 \( \mu m \), and 99% of the total downwelling energy is between 0.3 \( \mu m \) and 3.8 \( \mu m \) (LeNoble, 1993). The remaining 1% is below 0.3 \( \mu m \), which is not included in the model developed here.
The downward irradiance coming from the sun is mostly short-wave (peaks at 485nm) radiation and is composed of 2 components:

- Direct radiation.
- Diffuse radiation.

The scattering effects of air molecules and aerosols generate the diffuse radiation. For a complete coverage of the spectrum, the direct and diffuse irradiance on a surface normal to the direction of the sun at ground level will be estimated for each channel or wavelength ($\lambda$) ranging from 0.2 $\mu$m to 3 $\mu$m.

The model we use parameterizes the downward radiation and was discussed by Bird and Riordan (1986) and completed by Gregg and Carder (1990), and modified by Bostater et al (1995) for radiative transfer for hyperspectral remote sensing applications over water. The latter model used for applications over water, parameterizes the diffuse radiation as a simple proportion of the direct flux according to the equation below (McNally, 1997; Lamb, 1995; U.S. Dept., 1978):

$$E_{\text{diff}}^{\lambda} = \nu E_{\text{dir}}^{\lambda}, \quad (2.2)$$

where:

$E_{\text{diff}}^{\lambda} = \nu E_{\text{dir}}^{\lambda}$ = the diffuse downwelling solar irradiance (W m$^{-2}$ nm$^{-1}$),

$\nu$ = the dimensionless ratio of indirect to direct sunlight,

$E_{\text{dir}}^{\lambda}$ = the direct downwelling solar irradiance (W m$^{-2}$ nm$^{-1}$).
Table 2.1 Dimensionless ratio of indirect to direct sunlight from 300nm to 3000nm. The values between 835nm and 3000nm are identical. Values used in McNally’s thesis (McNally, 1997).

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<th>ν</th>
<th>Wave</th>
<th>ν</th>
<th>Wave</th>
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2.2.2 Direct Radiation

The downward direct irradiance is given (Bird and Riordan, 1986) as:

\[ E_{\text{dir}}^d (\lambda) = S_o (\lambda) \cos \theta T_r (\lambda) T_a (\lambda) T_{oz} (\lambda) T_u (\lambda) T_w (\lambda), \]  

where:

- \( E_{\text{dir}}^d (\lambda) \) = the direct downwelling solar irradiance (W m\(^{-2}\) nm\(^{-1}\)),
- \( \theta \) = the solar zenith angle (Wm\(^{-2}\)nm\(^{-1}\)),
- \( T_r (\lambda) \) = the transmittance after Rayleigh scattering (dimensionless),
- \( T_a (\lambda) \) = the transmittance after aerosol scattering (dimensionless),
- \( T_{oz} (\lambda) \) = the transmittance after ozone absorption (dimensionless),
- \( T_u (\lambda) \) = the transmittance after adsorption by uniformly mixed gases (nitrogen and oxygen) (dimensionless),
- \( T_w (\lambda) \) = the transmittance after water vapor absorption (dimensionless),
- \( S_o (\lambda) = H_o D \) = the mean extraterrestrial irradiance corrected for earth-sun distance and orbital eccentricity (W m\(^{-2}\) nm\(^{-1}\)),

where:

- \( H_o \) = the solar constant (W m\(^{-2}\)),
- \( D \) = the earth-sun distance factor (dimensionless), (see section 2.4.8).

2.2.3 Diffuse Radiation

The downward diffuse irradiance is given as (Bird, 1986):
\[ E_{\text{diff}}^d (\lambda) = I_r (\lambda) I_a (\lambda) I_g (\lambda), \quad (2.4) \]

where:

\( E_{\text{diff}}^d (\lambda) = \) the diffuse downwelling solar irradiance (W m\(^{-2}\) nm\(^{-1}\)),

\( I_r (\lambda) = \) the diffuse component of irradiance arising from Rayleigh scattering after molecular absorption (W m\(^{-2}\) nm\(^{-1}\)),

\( I_a (\lambda) = \) the diffuse component of aerosol scattering after molecular absorption (W m\(^{-2}\) nm\(^{-1}\)),

\( I_g (\lambda) = \) the diffuse component of irradiance arising from ground-air multiple interactions (W m\(^{-2}\) nm\(^{-1}\)).

A complete description of all terms for the direct and diffuse radiation calculation can be found in section 2.4.

### 2.2.4 Total Radiation

The total (shortwave) downwelling irradiance is equal to the direct and diffuse irradiance coming from the sun.

\[ K(\lambda) \downarrow = E_{\text{diff}}^d (\lambda) + E_{\text{dir}}^d (\lambda). \quad (2.5) \]

All terms are explained in equations 2.1, 2.3, 2.4.

### 2.2.5 Longwave Downwelling Radiation

The downward longwave radiation \( I(\lambda) \downarrow \) is very difficult to calculate because radiative flux divergence equations should be integrated vertically as
explained by Stull, 1988. In this radiation sub-model, we consider a cloud free environment and thus one layer is considered. This is a limitation of the technique used but the assumption is made to demonstrate the technique. Consequently, we can assume downward longwave radiation to be spatially constant but time dependent over a 24 hours simulation. The Florida Solar Energy Center located in Cocoa, FL maintains meteorological instrumentation. One instrument collects the longwave downward radiation and a graph of the variation of the downward longwave radiation during October 23rd 2002 can be found in figure 2.3.
Figure 2.3 Downwelling longwave radiation measured by the Florida Solar Energy Center from Oct 23 2002 00:00:00 to Oct 24 2002 00:00:00 located in Cocoa, FL (28.4N, 80.8W).
2.3 Upwelling Radiation

Radiation landscape characteristics of the underlying land surface are exceptionally important to determine the radiative interaction between the atmosphere and the land surface. This refers to the amount of solar radiation reflected by the surface also called the land surface albedo.

Accurate estimation of the surface emissivity is also essential to estimate longwave infrared radiation. Emissivity is the process that estimates the amount of longwave radiation reemitted by the land surface. In the radiation energy budget, the upwelling radiation is composed of both the reflected shortwave and the emitted longwave radiation.

Both ground reflectance (albedo) and emissivity were obtained from MODIS satellite imagery. Before explaining in details how shortwave and longwave radiation are estimated, a brief section on MODIS will precede.

2.3.1 MODIS Description

MODIS (or Moderate Resolution Imaging Spectroradiometer) is the key instrument aboard the Terra satellite. The Terra (formally known as EOS-AM-1) satellite is the flagship of EOS. It provides global data on the state of the atmosphere, land, and oceans, as well as their interactions with solar radiation and with one another. Terra was successfully launched on December 18, 1999. The
Terra mission is part of NASA's Earth Sciences Enterprise (ESE). The Terra spacecraft features following instruments:

- ASTER - Advanced Spaceborne Thermal Emission and Reflection Radiometer.
- CERES - Clouds and the Earth's Radiant Energy System.
- MISR - Multi-angle Imaging Spectro-Radiometer.
- MODIS - Moderate-resolution Imaging Spectroradiometer.
- MOPITT - Measurements of Pollution in the Troposphere.

A well-designed web site describes all MODIS products, algorithms and satellite specifications (MODIS, 2003).

Terra MODIS is viewing the entire Earth's surface every day, acquiring data in 36 spectral bands, or groups of wavelengths (see table 2.2).
Table 2.2 MODIS Technical Specifications and spatial resolutions.

<table>
<thead>
<tr>
<th>Primary Use</th>
<th>Band</th>
<th>Bandwidth</th>
<th>Spectral Radiance</th>
<th>Required SNR</th>
</tr>
</thead>
<tbody>
<tr>
<td>Land/Cloud/Aerosols Boundaries</td>
<td>1</td>
<td>620 – 670</td>
<td>21.8</td>
<td>128</td>
</tr>
<tr>
<td>Land/Cloud/Aerosols Properties</td>
<td>2</td>
<td>841 – 876</td>
<td>24.7</td>
<td>201</td>
</tr>
<tr>
<td></td>
<td>3</td>
<td>459 – 479</td>
<td>35.3</td>
<td>243</td>
</tr>
<tr>
<td></td>
<td>4</td>
<td>545 – 565</td>
<td>29.0</td>
<td>228</td>
</tr>
<tr>
<td></td>
<td>5</td>
<td>1230 – 1250</td>
<td>5.4</td>
<td>74</td>
</tr>
<tr>
<td></td>
<td>6</td>
<td>1628 – 1652</td>
<td>7.3</td>
<td>275</td>
</tr>
<tr>
<td></td>
<td>7</td>
<td>2105 – 2155</td>
<td>1.0</td>
<td>110</td>
</tr>
<tr>
<td>Ocean Color/Phytoplankton/Biogeochemistry</td>
<td>8</td>
<td>405 – 420</td>
<td>44.9</td>
<td>880</td>
</tr>
<tr>
<td></td>
<td>9</td>
<td>438 – 448</td>
<td>41.9</td>
<td>838</td>
</tr>
<tr>
<td></td>
<td>10</td>
<td>483 – 493</td>
<td>32.1</td>
<td>802</td>
</tr>
<tr>
<td></td>
<td>11</td>
<td>526 – 536</td>
<td>27.9</td>
<td>754</td>
</tr>
<tr>
<td></td>
<td>12</td>
<td>546 – 556</td>
<td>21.0</td>
<td>750</td>
</tr>
<tr>
<td></td>
<td>13</td>
<td>662 – 672</td>
<td>9.5</td>
<td>910</td>
</tr>
<tr>
<td></td>
<td>14</td>
<td>673 – 683</td>
<td>8.7</td>
<td>1087</td>
</tr>
<tr>
<td></td>
<td>15</td>
<td>743 – 753</td>
<td>10.2</td>
<td>586</td>
</tr>
<tr>
<td></td>
<td>16</td>
<td>862 – 877</td>
<td>6.2</td>
<td>516</td>
</tr>
<tr>
<td>Atmospheric/Water Vapor</td>
<td>17</td>
<td>890 – 920</td>
<td>10.0</td>
<td>167</td>
</tr>
<tr>
<td></td>
<td>18</td>
<td>931 – 941</td>
<td>3.6</td>
<td>57</td>
</tr>
<tr>
<td></td>
<td>19</td>
<td>915 – 965</td>
<td>15.0</td>
<td>250</td>
</tr>
<tr>
<td>Surface/Cloud Temperature</td>
<td>20</td>
<td>3.660 - 3.840</td>
<td>0.45(300K)</td>
<td>0.05</td>
</tr>
<tr>
<td></td>
<td>21</td>
<td>3.929 - 3.989</td>
<td>2.38(335K)</td>
<td>2.00</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
<td>---</td>
</tr>
<tr>
<td>22</td>
<td>3.929 - 3.989</td>
<td>0.67(300K)</td>
<td>0.07</td>
<td></td>
</tr>
<tr>
<td>23</td>
<td>4.020 - 4.080</td>
<td>0.79(300K)</td>
<td>0.07</td>
<td></td>
</tr>
<tr>
<td>Atmospheric Temperature</td>
<td>24</td>
<td>4.433 - 4.498</td>
<td>0.17(250K)</td>
<td>0.25</td>
</tr>
<tr>
<td>Temperature</td>
<td>25</td>
<td>4.482 - 4.549</td>
<td>0.59(275K)</td>
<td>0.25</td>
</tr>
<tr>
<td>Cirrus Clouds</td>
<td>26</td>
<td>1.360 - 1.390</td>
<td>6.00</td>
<td>150(SNR)</td>
</tr>
<tr>
<td>Water Vapor</td>
<td>27</td>
<td>6.535 - 6.895</td>
<td>1.16(240K)</td>
<td>0.25</td>
</tr>
<tr>
<td>Cloud Properties</td>
<td>28</td>
<td>7.175 - 7.475</td>
<td>2.18(250K)</td>
<td>0.25</td>
</tr>
<tr>
<td>Ozone</td>
<td>29</td>
<td>8.400 - 8.700</td>
<td>9.58(300K)</td>
<td>0.05</td>
</tr>
<tr>
<td>Surface/Cloud Temperature</td>
<td>30</td>
<td>9.580 - 9.880</td>
<td>3.69(250K)</td>
<td>0.25</td>
</tr>
<tr>
<td>Ozone</td>
<td>31</td>
<td>10.780 - 11.280</td>
<td>9.55(300K)</td>
<td>0.05</td>
</tr>
<tr>
<td>Temperature</td>
<td>32</td>
<td>11.770 - 12.270</td>
<td>8.94(300K)</td>
<td>0.05</td>
</tr>
<tr>
<td>Cloud Top Altitude</td>
<td>33</td>
<td>13.185 - 13.485</td>
<td>4.52(260K)</td>
<td>0.25</td>
</tr>
<tr>
<td>Altitude</td>
<td>34</td>
<td>13.485 - 13.785</td>
<td>3.76(250K)</td>
<td>0.25</td>
</tr>
<tr>
<td></td>
<td>35</td>
<td>13.785 - 14.085</td>
<td>3.11(240K)</td>
<td>0.25</td>
</tr>
<tr>
<td></td>
<td>36</td>
<td>14.085 - 14.385</td>
<td>2.08(220K)</td>
<td>0.35</td>
</tr>
</tbody>
</table>

1 Bands 1 to 19 are in nm; Bands 20 to 36 are in µm
2 Spectral Radiance values are \( \text{W/m}^2 \text{-µm-sr} \)
3 SNR = Signal-to-noise ratio
4 NE(\(\Delta T\)) = Noise-equivalent temperature difference

Note: Performance goal is 30-40% better than required

MODIS data is divided in several levels briefly described in table 2.3.

Table 2.3 Description of all MODIS data levels (Riggs et al. 2003).

<table>
<thead>
<tr>
<th>MODIS Levels</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Level 1B (L1B)</td>
<td>Swath (scene) of MODIS data geolocated to latitude and longitude centers of 1 km resolution pixels.</td>
</tr>
<tr>
<td>Level 2 (L2)</td>
<td>A geophysical product that remains in latitude and longitude orientation of L1B.</td>
</tr>
<tr>
<td>Level 2 gridded (L2G)</td>
<td>A gridded format of a map projection. At L2G the data products are referred to as tiles, each tile being a piece, e.g. 10° x 10° area, of a map projection. L2 data products are gridded into L2G tiles by mapping the L2 pixels into cells of a tile in the map projection grid. The L2G algorithm creates a gridded product necessary for the level 3 products.</td>
</tr>
<tr>
<td>Level 3 (L3)</td>
<td>A geophysical product that has been temporally and or spatially manipulated, and is in a gridded map projection format and comes as a tile of the global grid.</td>
</tr>
</tbody>
</table>
2.3.1.1 MODIS 09 L2 – Ground Reflectance (albedo)

The MODIS 09 L2 surface reflectance product is a seven-band product computed from the MODIS Level 1B land bands 1, 2, 3, 4, 5, 6, and 7 (centered at 648 nm, 858 nm, 470 nm, 555 nm, 1240 nm, 1640 nm, and 2130 nm, respectively, see table 2.1). The product (Level 1B) is an estimate of the surface spectral reflectance for each band, as it would have been measured at ground level if there were no atmospheric scattering or absorption. The correction scheme includes corrections for the effect of atmospheric gases, aerosol, and thin cirrus clouds. The albedo values are a directional hemispherical reflectance (black sky albedo) obtained by integrating the BRDF (Bi-directional Reflectance Distribution Function) over the exitance hemisphere for a single irradiance direction and a bihemispherical reflectance (white sky albedo) obtained by integrating the BRDF over all viewing and irradiance directions (Wan, 1999). In the original model, Bird used a constant value for ground albedo of 0.2. This research is using MODIS albedo data to show the importance of surface reflectance determination in order to accurately estimate land surface temperature.

2.3.1.2 MODIS 11 L2 – Emissivity

In our application, emissivity depends on the land matter and emissivity values are extracted also from MODIS satellite data. The Aqua MOD11 product is used and contains Level 2 and 3 LST (Land Surface Temperatures) where
emissivity are retrieved from Aqua MODIS data at spatial resolutions of 1 km over global land surfaces under clear-sky conditions. Emissivities are estimated by the classification-based emissivity method (Snyder and Wan, 1998) according to land cover types in the pixel determined by the input data in quarterly Land Cover (MOD12Q1) and daily Snow Cover (MOD10_L2). Snyder and Wan used angular reflectance and emissivity to convert the radiance of a pixel as measured from space to land-surface temperature. The latter authors used a kernel model to develop a look-up table for the MODIS land-surface temperature algorithm to estimate the spectral, angular scene emissivity from land cover classification. Kernel model are (semi-) empirical models based on linear combinations of “kernels”.

\[ \rho = \int_{iso} + \int_{geo} k_{geo} + \int_{vol} k_{vol}, \quad (2.6) \]

where:

\[ \rho = \text{is the surface reflectance as a function of component reflectance } \left( f_{i} \right) \]

and the kernels \( k \) which are mathematical functions that depend on sun (or incident) and view (or observer) angles. The subscripts “geo” and “vol” refer to the physical bases for some kernels in which there is an identification of a “geometric” or hotspot factor and a “volume” or path length and scattering factor (Snyder and Wan, 1998).
2.3.1.3 EOS Data Gateway and MODIS Reprojection Tool

Albedo and emissivity data sets for October 23rd 2002 were downloaded from the EOS Data Gateway. The Earth Observing System Data Gateway allows users to search and order earth science data products from NASA and affiliated centers.

The projection of the MODIS hdf files downloaded from the gateway is the integerized sinusoidal (ISIN) projection (figure 2.4). The hdf files had to be reprojected with the MODIS Reprojection tool to a geographic (Lat and Lon) coordinate system. The Reprojection software and the user guide can be downloaded for free from the EROS Data Center (EDC) Distributed Active Archive Center (DAAC).

Figure 2.4 Integerized Sinusoidal Grid. Tiles are 10 degrees x 10 degrees. The tile coordinate system starts at (0,0) (horizontal tile number, vertical tile number) in the upper left corner and proceeds right (horizontal) and downward (vertical). The tile in the bottom left corner is (35,17). Dark blue tiles contain only water (no land).
2.3.2 Upwelling Shortwave Radiation - Albedo

The upwelling reflected short-wave radiation can be found using:

\[ K(\lambda)^\uparrow = -a(\lambda) K(\lambda)^\downarrow, \]  

(2.7)

where:

- \( K(\lambda)^\uparrow \) = the upwelling reflected shortwave (solar) radiation (W m\(^{-2}\)),
- \( K(\lambda)^\downarrow \) = the downwelling shortwave radiation transmitted through the air (W m\(^{-2}\)),
- \( a(\lambda) \) = the albedo or surface reflectance of the surface (dimensionless).

The ground albedo is a fraction of downwelling radiation at the surface that is reflected. It can vary from about 0.95 over fresh snow, 0.4 over light-colored dry soils to 0.05 over dark wet soils.

The 7 ground reflectance bands utilized for the calculation of upwelling radiation are used to interpolate albedo at every wavelength to allow wavelength dependency as shown in figure 2.4 and table B.1.
Figure 2.5 Albedo interpolations from 7 MODIS albedo bands to allow the wavelength dependency for the ground surface reflectance parameter. The MODIS 09 data set was collected on October 22nd 2002. This graph was created from the dataset of the upper left corner of the domain (-81.375W, 29.1622N).
2.3.3 Upwelling Longwave Radiation - Emissivity

The outward longwave infrared radiation is expressed as:

\[ I(\lambda)^\uparrow = \varepsilon(\lambda) \sigma_b T^4, \quad (2.8) \]

where:
\( I(\lambda)^\uparrow \) = the longwave infrared radiation emitted up (W m\(^{-2}\) nm\(^{-1}\)),
\( \varepsilon(\lambda) \) = the emissivity of the surface (dimensionless),
\( \sigma \) = the Stephan-Boltzmann constant (W m\(^{-2}\) K\(^{-4}\)),
\( T \) = the temperature of the surface (Kelvin).

Kirchhoff’s law states that matter will emit radiation at a rate that depends both on its absorptive properties and its absolute temperature \( T \). A planet sheds its energy solely by this emission process (radiating to space).

This negative feedback of the planetary temperature (through Kirchhoff’s law) acts like a giant thermostat, continuously adjusting the rate of cooling to compensate for excess warming (Xiu, 2001). The situation in which thermal emission is balanced locally at all wavelengths by heating due to solar radiation is called local radiative equilibrium.

The emissivity of an object is equal to the energy emitted by the object divided by the amount of energy the same object would emit at that temperature if it were a blackbody (Lyon 1997) as:
\[ \varepsilon = \frac{E}{E_b}, \quad (2.9) \]

where:

\( \varepsilon \) = the emissivity (dimensionless),

\( E \) = the total emissive power of a surface (\( \text{W m}^{-2} \)),

\( E_b \) = the total emissive power of an ideally radiating surface (blackbody) at the same temperature (\( \text{W m}^{-2} \)).

The infrared emissivity is in the range 0.9 to 0.99 for most surfaces (Stull, 1988).

**2.4 Downward Radiation Transfer Sub-Model Equations**

The radiation transfer Sub-model can be divided into 2 distinct parts. First, the model will calculate the solar declination, the equation of time, the true solar time, the hour angle and the local time of sunrise and sunset. Then, it computes the solar zenith angle, atmospheric path length, the earth-sun distance factor as given by Spencer (1971), and the atmospheric transmittance coefficient for both diffuse and direct radiation as a function of wavelength using Neckel and Labs (1981) solar constant data (see Table B2) and the atmospheric model of Gregg and Carder/Bird and Riordan (1986).
2.4.1 Solar Declination

Declination is the angular distance of the sun north or south of the earth's equator. The earth's equator is tilted 23.45 degrees with respect to the plane of the earth's orbit around the sun, so at various times during the year, as the earth orbits the sun, declination varies from 23.45 degrees north to 23.45 degrees south.

This gives rise to the seasons. Around December 21, the northern hemisphere of the earth is tilted 23.45 degrees away from the sun, which is the winter solstice for the northern hemisphere and the summer solstice for the southern hemisphere. Around June 21, the southern hemisphere is tilted 23.45 degrees away from the sun, which is the summer solstice for the northern hemisphere and winter solstice for the southern hemisphere. March 21 and September 21 are the fall and spring equinoxes when the sun is passing directly over the equator. Note that the tropics of Cancer and Capricorn mark the maximum declination of the sun in each hemisphere.

The declination of the sun in degrees is given by (Anonymous, 1978):

\[
\delta = 23.45 \sin \left[ 1.008 \left( n - 180 \right) \right], \quad (2.10)
\]

over the interval \( 1 \leq n \leq 80 \),

\[
\delta = 23.45 \sin \left[ 0.965 \left( n - 80 \right) \right], \quad (2.11)
\]

over the interval \( 81 \leq n \leq 266 \),

\[
\delta = 23.45 \sin \left[ 0.975 \left( n - 266 \right) \right], \quad (2.12)
\]

over the interval \( 267 \leq n \leq 365 \),

41
where \( n \) is the Julian day of the year.

### 2.4.2 The Equation of Time

Solar day is the length of time between one local noon (when the Sun is highest in the sky) to the next. As it turns out, the length of the solar day is not always 24 hrs (its average over the course of a year defines 24 hrs). The solar day would always be 24 hrs if the Sun 'moves' east against the background of fixed stars at a constant rate. The real Sun moves at a variable rate, however, because of the tilt of the Earth rotation axis relative to its orbit around the Sun (the obliquity), the same reason as for the changing length of daytime hours. Also, the Earth's orbit is elliptical and so it moves faster at perihelion (around Jan 2) than at aphelion (Jul 3). Both effects combine to create an offset in the time of local noon (and those of sunrise and sunset) by as much as +/- 16 min. The equation of time corrects for this anomaly (Anonymous, 1978).

\[
E_{qt} = -14.2 \sin \left[ \left( n - 7 \right) \times \frac{180}{111} \right],
\]

over the interval \( 1 \leq n \leq 106 \),
\[
E_{qt} = 4 \sin \left[ \left( n - 106 \right) \times \frac{1}{106} \right],
\]

over the interval \( 107 \leq n \leq 166 \),
\[
E_{qt} = -6.5 \sin \left[ \left( n - 166 \right) \times \frac{180}{80} \right],
\]
over the interval $167 \leq n \leq 246$,
\[
E_{qt} = 16.4 \sin \left[ \left( n - 247 \right) \times \left( \frac{180}{113} \right) \right],
\] (2.16)

over the interval $247 \leq n \leq 365$,
where $n$ is the Julian day of the year.

### 2.4.3 True Solar Time

True solar time differs from clock time. The difference is due to differences between the site longitude and the standard meridian (for example: Greenwich meridian), the Equation of Time, and summer time or daylight-saving conventions.

The true solar time, $H_s$ in hours is given by (Anonymous, 1978):

\[
H_s = H_{ls} + \frac{1}{60} E_{qt} + \frac{1}{15} \left( L_{sm} - L_{og} \right),
\] (2.17)

where:
- $H_{ls}$ = the local standard time (hours),
- $L_{sm}$ = the longitude of the standard meridian of the time zone (degrees),
- $L_{og}$ = the longitude of the grid point (degrees),
- $E_{qt}$ = the equation of time (hours).

### 2.4.4 Hour Angle

The hour angle of the sun is the angle along the arc traversed by the sun across the sky. The hour angle of the sun, $h$, is defined such that $h = 0$ at the local solar noon. (i.e., when the sun is at its maximum solar angle for the day, or
equivalently, when the sun crosses the local meridian of longitude). In the morning, the hour angle is negative and in the afternoon the hour angle is positive.

The hour angle ($h$ in degrees) about solar noon is given by (Anonymous, 1978):

\[
h = 15 \times (12 - H_s),
\]

(2.18)

where $H_s$ is the true solar time (hours).

### 2.4.5 Local Time of Sunset and Sunrise

The time of sunset and sunrise control the radiative transfer model and they are given by (Anonymous, 1978):

\[
H_{sr} = \frac{1}{15} \arccos (\tan \phi \tan \delta) - Q,
\]

(2.19)

where:

$H_{sr} =$ the time of sunrise after midnight (hours),
\[
\phi = \text{the latitude of the grid point (degrees),}
\]
\[
\delta = \text{the solar declination (degrees).}
\]

### 2.4.6 Solar Zenith Angle

The solar zenith angle of the sun is a function of the latitude of a grid point, the solar declination (equation 2.10 – 2.12) and the hour angle (equation 2.18). The zenith angle at each grid point is calculated by (Anonymous, 1978):
\[
\cos \theta = \cos \phi \cos \delta \cos(h) + \sin \phi \sin \delta ,
\]

(2.20)

where:
\( \theta \) = the zenith angle (degrees),
\( \delta \) = the solar declination (degrees).

### 2.4.7 Atmospheric Path Length

The non-dimensional path length \( M(\theta) \) through the atmosphere is required for atmospheric transmittance due to attenuation by all constituents. It can be determined using an equation from Kasten (1966) valid at all zenith angles:

\[
M(\theta) = \frac{1}{\cos \theta + 0.15 (93.885 - \theta)^{-1.253}} ,
\]

(2.21)

where:
\( \theta \) = the solar zenith angle (degrees).

A second path length corrected for nonstandard atmospheric pressure is also used and its calculation is given by Gregg and Carder (1990) as:

\[
M' = \frac{M(\theta)P}{P_o} ,
\]

(2.22)

where:
\( P_o \) = the standard atmospheric pressure (101.325 kPa),
\( P \) = the actual atmospheric pressure (kPa).
Ozone is mainly concentrated in the stratosphere thus it requires a slightly longer path length for accurate transmittance computations. Its calculation \( M_\infty \) is given by Paltridge and Platt (1976).

\[
M_\infty (\theta) = \frac{1.0035}{(\cos^2 \theta + 0.007)^{\frac{1}{2}}}, \quad \text{(2.23)}
\]

where:

\( \theta \) = the solar zenith angle (degrees).

### 2.4.8 Earth-Sun Distance Factor

The extraterrestrial irradiance corrected for earth-sun distance is given by Spencer (1971):

\[
D = 1.00011 + 0.034221 \cos \varphi + 0.00128 \sin \varphi + 0.000719 \cos 2\varphi + 0.000077 \sin 2\varphi, \quad \text{(2.24)}
\]

where:

\( \varphi = \frac{2\pi(n - 1)}{365} \) = the day angle in radians where \( n \) is the Julian day of the year.
2.4.9 Atmospheric Transmittance Calculations

2.4.9.1 Rayleigh Scattering

Rayleigh scattering refers to the scattering of light off of the molecules of the air, and can be extended to scattering from particles up to about a tenth of the wavelength of the light. It is Rayleigh scattering off the molecules of the air, which gives us the blue sky (Slater, 1980). The Rayleigh transmittance $T_r$ is given by Bird and Riordan (1986) as:

$$T_r = \exp \left\{ -\frac{M'}{\lambda^4 \left( 115.6406 - \frac{1.335}{\lambda^2} \right)} \right\},$$

where:
$
\lambda = \text{the wavelength (\text{\mu m})},$

$M' = \text{the corrected non-dimensional atmospheric path length}.$

2.4.9.2 Water Vapor Absorption

The transmittance after water vapor absorption is given by Leckner (1978) as:
\[ T_w = \exp \left[ - \frac{0.2385a_w WM}{(1 + 20.07a_w WM)^{0.45}} \right], \]  

(2.26)

where:

- \( W \) = the precipitable water vapor (cm) in a vertical path,
- \( a_w \) = the water vapor absorption coefficient as a function of wavelength, given by Bird and Riordan (1986) (see table B3),
- \( M \) = the non-dimensional atmospheric path length.

### 2.4.9.3 Ozone Absorption

The transmittance after ozone absorption is given by Leckner (1978) as:

\[ T_{oz} = \exp \left[ - a_{oz} H_{oz} M_{oz} \right], \]  

(2.27)

where:

- \( M_{oz} \) = the non-dimensional ozone atmospheric path length,
- \( a_{oz} \) = the ozone absorption coefficient (km\(^{-1}\)) given by Bird and Riordan (1986) (see table B4),
- \( H_{oz} = \frac{DU}{1000} \) = the ozone scale height and is a function of the ozone concentration in Dobson Units (\( DU \)). 1 Dobson Unit (DU) is defined to be 0.01 mm thickness at STP.

The ozone concentration (\( DU \)) used in the ozone scale height calculations is estimated from a MODIS 08 L3 dataset (figure 2.6). MODIS 08 L3 are statistical datasets derived from four Level-2 MODIS atmosphere products: aerosol, water vapor, cloud, and atmosphere profile. MODIS 08 L3 has different temporal scales:
MOD08_D3 (Daily), MOD08_E3 (8-Day), and MOD08_M3 (Monthly). The different parameters measured in MODIS 08 are atmospheric parameters related to atmospheric aerosol particle properties, total ozone burden, atmospheric water vapor, cloud optical and physical properties, and atmospheric stability indices. This product also provides standard deviations, quality assurance weighted means and other statistically derived quantities for each parameter (MODIS, 2003).

Figure 2.6 shows a column of air, 10 deg x 5 deg, over Labrador, Canada. The amount of ozone in this column (i.e. covering the 10 x 5 deg area) is conveniently measured in Dobson Units. If all the ozone in this column were to be compressed to standard temperature and pressure (STP) (0 deg C and 1 atmosphere pressure) and spread out evenly over the area, it would form a slab approximately 3mm thick. (The copyrighted graphic is based on a page developed by Owen Garrett for the Centre for Atmospheric Science at Cambridge University, UK. Dr. Glenn Carver has kindly given me the permission to reproduce it.)
In this research, the daily data set (MOD08_D3) is used for October 23rd 2003 and the parameters used are the ozone burden for the ozone absorption calculation (2.26) and the angstrom exponent for land and ocean for the aerosol scattering and absorption calculation (2.29).

Let’s assume that \( X_1, X_2, \ldots, X_n \) represent the retrieved pixel values of a Level 2 parameter over a \( 1^\circ \times 1^\circ \) grid box, \( W_i \) is the weighting factor (1 for the daily case), then the simple statistic is defined as:

\[
\overline{X}_d = \frac{\sum_i W_i - X_1}{\sum_i W_i},
\]  

(2.28)

\( \overline{X}_d \) = the regular mean for the ozone and angstrom parameters used in this research (MODIS, 2003).

The domain used in this research is a 2x2 degree grid. The ozone value was extracted from the daily data set MOD08_D3 (see figure C1). In appendix C, figure C1 shows a 1x1 degree grid average value of total ozone burden for the world. In order to obtain the ozone value for the research domain, the world ozone file was imported in ENVI where a spatial query was made to identify the ozone burden for our particular area (see figure C2). Figure 2.7 is a GIS representation of the domain of interest with the corresponding ozone burden data. The world file with the subset area can be seen in appendix C. The ozone burden value for the area is a simple average of the four 1 degree cells.
Figure 2.7 Ozone burden (DU) over the grid domain for October 23rd 2002. Data for the studied domain was extracted from MODIS08_D3 (figure C1). The ozone burden value for the area is a simple average of the four 1 degree cells.

2.4.9.4 Uniformly Mixed Gas Absorption

The transmittance after uniformly mixed gas absorption used by Bird and Riordan is after Leckner (1978) and is given by:
\[ T_a = \exp \left[ -\frac{1.41 a_u M'}{(1 + 118.93 a_u M')^{0.45}} \right], \]  

(2.29)

where:

\( M' \) = the corrected non-dimensional atmospheric path length,
\( a_u \) = the combination of an absorption coefficient and gaseous amount. The values used here are those of Bird and Riordan (1986) (see table B5).

### 2.4.9.5 Aerosol Scattering and Absorption

Aerosols may influence the atmosphere in two important ways, through direct and indirect effects. Direct effects refer to the scattering and absorption of radiation and their subsequent influence on planetary albedo and the climate system. Through their scattering and absorption of radiation as well as altering cloud optical properties (indirect effects), aerosols can also induce some other important environment effects. For example, aerosols are a major contributor to visibility problem in urban centers as well as rural areas (Slater, 1980). The aerosol concentration parameter \( \beta \) is assumed to be a function of the visibility in kilometers but is independent of wavelength (Gregg and Carder, 1990).

The transmittance after aerosol scattering and absorption is given by Gregg and Carter (1990) for maritime atmospheres as:

\[ T_a = \exp \left[ -\tau_a M \right], \]  

(2.30)

where:
\[ \tau_a = \text{the aerosol transmittance coefficient (dimensionless) given by:} \]
\[ \tau_a = \beta \lambda^{-\alpha}, \]  \hspace{0.5cm} (2.31)

where:

\[ \lambda = \text{the wavelength (\(\mu m\)),} \]
\[ \beta = \text{the turbidity non-dimensional coefficient from the aerosol concentration,} \]
\[ \alpha = \text{the Angstrom non-dimensional exponent.} \]

The Angstrom exponent both for land and ocean are estimated from the MODIS 08_D3 dataset described in section 2.4.9.3. The same technique used for the ozone burden was used to extract the angstrom value for both land and ocean. Figures C3 and C5 in Appendix C represent world files for the angstrom exponent for both land and ocean respectively. The angstrom exponent for land or ocean for the area is a simple average of the two 1 degree cells (see figures C4 and C6). Figure 2.8 is a GIS representation of the domain of interest with the corresponding land and ocean angstrom exponent data.

The wavelength independent parameter \( \beta \) can be computed with equation (2.29) by setting the wavelength to 0.55 \( \mu m \) and by computing the aerosol optical thickness \( \tau_a(550) \) with:

\[ \tau_a(550) = c_a(550) H_a, \]  \hspace{0.5cm} (2.32)

where:

\[ H_a = \text{the aerosol scale height taken to be 1 km by Gregg and Carder (1990),} \]
\[ c_a(550) = \text{the aerosol extinction coefficient at 550 nm (km}^{-1}). \]
where Gregg and Carder used Koschmieder formula from Fitzgerald (1975) to relate visibility to the aerosol total extinction coefficient at 550 nm:

\[
c_a(550) = \frac{3.91}{V},
\]

(2.33)

where:

- \( V \) is the visibility (km).

Figure 2.8 Land and ocean Ångström exponent for the domain. Data was extracted from MODIS08 (figures C3 & C5) and combined over the studied domain. The Ångström exponent for land or ocean for the area is a simple average of the two 1 degree cells.
2.4.10 Atmospheric Transmittance for Diffuse Radiation

2.4.10.1 Rayleigh Scattering Component

The Rayleigh scattering component on a horizontal surface is given by (Bird and Riordan, 1986) as:

\[ I_{r,k} = E_{dir,k}^d \cos (Z) T_{ao,k} T_{u,k} T_{w,k} T_{aak} \left( 1 - T_{r,k}^{0.95} \right)^{0.5}, \]  

(2.34)

where:

- \( T_{x,k} \) = the transmittance terms calculated using equations (2.26 for \( T_{ao,k} \), (2.27 for \( T_{u,k} \), (2.25 for \( T_{w,k} \), (2.28 for \( T_{a,k} \), and (2.24 for \( T_{r,k} \),
- \( T_{aak} \) = the aerosol absorption (dimensionless). It can be estimated as described by Bird and Riordan (1986) as:

\[ T_{aak} = \exp \left[ \left( 1 - \omega_{\lambda} \right) \tau_{a,k} M \right], \]  

(2.35)

where:

- \( \omega_{\lambda} \) = the aerosol single scattering albedo as a function of wavelength. It can be estimated as described by Bird and Riordan (1986) as:

\[ \omega_{\lambda} = \omega_{0.4} \exp \left\{ - \omega' \left[ \ln \left( \frac{\lambda}{0.4} \right) \right]^2 \right\}, \]  

(2.36)

where:

- \( \omega_{0.4} = 0.945 \) and is the single scattering albedo at 0.4 \( \mu m \) wavelength,
- \( \omega' = 0.095 \) and is the wavelength variation factor.
2.4.10.2 Aerosol Scattering Component

The Aerosol scattering component on a horizontal surface is given as (Bird and Riordan, 1986):

\[
I_{a\lambda} = E_{dir\lambda}^d \cos(Z) T_{a\lambda} T_{w\lambda} T_{e\lambda}^{1.5} (1 - T_{a\lambda}) F_s, \quad (2.37)
\]

where:

- \( T_{a\lambda} \) = the transmittance term for aerosol absorption (dimensionless) estimated with equation (2.33),
- \( T_{a\lambda} \) = the transmittance term for aerosol scattering (dimensionless). It can be estimated as described by Bird and Riordan (1986) as:

\[
T_{a\lambda} = \exp(-\omega_{\lambda} \tau_{a\lambda} M), \quad (2.38)
\]

where:

- \( \omega_{\lambda} \) = the aerosol single scattering albedo as a function of wavelength described in equation (2.34),
- \( F_s \) = a fraction of the aerosol scattered downward and is a function of the solar zenith angle. It can be estimated as described by Bird and Riordan (1986) as:

\[
F_s = 1 - 0.5 \exp\left( (AFS + BFS \cos Z) \cos Z \right), \quad (2.39)
\]

where:

- \( AFS = ALG [1.459 + ALG (0.1595 + ALG \ 0.4129)] \),
- \( BFS = ALG [0.0783 + ALG (-0.3824 - ALG 0.5874)] \),
- \( ALG = \ln(1 - (\cos \theta)) \).
2.4.10.3 Ground Air interactions

The ground air multiple interactions represent multiple reflection of irradiance between the ground and the air on a horizontal surface and are given as (Bird and Riordan, 1986):

\[
I_{g\lambda} = \left( E_{dir\lambda} \cos(Z) + I_{r\lambda} + I_{a\lambda} \right) r_{s\lambda} r_{g\lambda} / \left( 1 - r_{s\lambda} r_{g\lambda} \right),
\]

(2.40)

where:

\( r_{g\lambda} = \) the ground albedo as a function of wavelength (dimensionless),
\( r_{s\lambda} = \) the sky reflectivity (dimensionless),

where:

\[
r_{s\lambda} = T'_{o\lambda} T'_{w\lambda} T'_{a\lambda} \left[ 0.5 \left( 1 - T'_{r\lambda} \right) + \left( 1 - F_s' \right) \left( 1 - T'_{a\lambda} \right) \right],
\]

(2.41)

where:

\( T'_{x\lambda} = \) the primed transmittance terms. They are the regular atmospheric transmittance terms evaluated with the non-dimensional atmospheric path length \( M = 1.8 \),
\( F_s' = \) the primed fraction of the aerosol scattered downward and is a function of the solar zenith angle. It is evaluated with the non-dimensional atmospheric path length \( M = 1.8 \).
The ground albedo values used in equation (2.35) come from the same data set used for the calculation of upwelling shortwave radiation (see section 2.3.1).

Below is a table detailing the differences between the radiative transfer model modified by Bostater et al. (1995-2000) and the one described in this research.

Table 2.4 Differences in the calculation of diffuse radiation between the radiative transfer model used in the subroutine of this research and the one by Bostater and McNally (1996).

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Diffuse radiation</strong></td>
<td>$E_{\text{diff}}^d (\lambda) = \nu E_{\text{dir}}^d (\lambda)$</td>
<td>$E_{\text{diff}}^d (\lambda) = I_r(\lambda) I_a(\lambda) I_g(\lambda)$</td>
<td></td>
</tr>
<tr>
<td><strong>Angstrom exponent</strong></td>
<td>Taken from Gregg and Carder (1990).</td>
<td>Obtained from MODIS 08 data. Daily 1 x 1 degree grid average values of angstrom exponent for October 23rd 2002.</td>
<td></td>
</tr>
<tr>
<td><strong>Total ozone burden</strong></td>
<td>Taken from Gregg and Carder (1990).</td>
<td>Obtained from MODIS 08 data. Daily 1 x 1 degree grid average values of total ozone burden for October 23rd 2002.</td>
<td></td>
</tr>
</tbody>
</table>
CHAPTER 3

THERMAL MODEL PART II

THERMAL INERTIA SIMULATION

3.1 Definition

To understand thoroughly the concept and the importance of thermal inertia in mesoscale modeling, let’s first define thermal inertia.

The sun rises every morning and shines more or less equally on the earth surface depending on the time of the year, the time of the day and the relief of the region. Assuming clear conditions, the response in terms of surface temperatures varies enormously due to land surface differences. Then, at night, the sunsets and the regions that experience a large temperature rises during the day will experience large temperature drops at night. The most extreme difference are between bare rocks on the one hand and water on the other hand. At night, the sea is warmer than the land, but during the day, the land becomes warmer than the sea. The amplitude of the temperature variations of the land surface can be of several tens of degrees Celsius whereas the sea surface variations are only of the order of a few degrees Celsius. Therefore, the variation of temperatures over 24-h period can vary enormously for different types of surface. In qualitative terms, thermal inertia can
be defined as the resistance of a material to a change in temperature. And it is the most important thermal property, which governs the variations of surface temperature. Thermal Inertia is usually expressed by:

\[ P = \sqrt{\rho c K} , \]  

(3.1)

where:
- \( P \) is expressed in thermal inertia units (1 TIU = 1 W s\(^{1/2}\) m\(^{-2}\) K\(^{-1}\)),
- \( K \) = the thermal conductivity (J m\(^{-1}\) s\(^{-1}\) K\(^{-1}\)),
- \( \rho \) = the density (kg m\(^{-3}\)),
- \( c \) = the specific heat of the material (J kg\(^{-1}\) K\(^{-1}\)).

For a given heat transfer, high thermal inertia values lead to small changes in temperatures, while low thermal inertia values lead to large changes in temperatures (Pratt and Ellyett, 1979).

The determination of thermal inertia from in-situ measurements is difficult, particularly for urban areas or surfaces covered with vegetation. Remote sensing data provide a unique way for monitoring this parameter on a mesoscale. Watson (1970, 1973, and 1975) was the first to work on mapping thermal inertia with satellite data. Watson’s model is based on an original theory introduced by Jaeger (1953), who used a one-dimensional periodic heating model to simulate temperature variations of the lunar surface.

Xue and Cracknell (1995) created a simple and operational thermal inertia model to overcome all the drawbacks from previous authors. Sobrino and Kharraz
(1999) modified Cracknell and Xue’s model in order to be able to find thermal inertia values only from satellite data. Their model will be used in this project to estimates the thermal inertia for every pixel. Emissivity and ground reflectance values from MODIS were used in the radiative transfer model in chapter 2. In order to estimate thermal inertia, AVHRR data sets for 3 different satellites passes are used.

3.2 NOAA-AVHRR Characteristics

NOAA-AVHRR is at the present time the most adequate way for mapping thermal inertia. This is due to its spatial, temporal and spectral resolution. For example, the spatial resolution of 1 km² at nadir offers the accuracy needed for the mesoscale grid. AVHRR channels 1 and 2 provide spectral channel data for determination of albedo and AVHRR channels 3, 4 and 5 provide spectral data for determination of brightness temperatures (Table 3.1).

Table 3.1 Channel characteristics for NOAA AVHRR satellites.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Spectral Range (micrometers)</th>
<th>Detector type</th>
<th>Resolution (km)</th>
<th>Temperature Range (K)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ch. 1</td>
<td>0.58-0.68</td>
<td>Silicon</td>
<td>1.09</td>
<td>-</td>
</tr>
<tr>
<td>Ch. 2</td>
<td>0.725-1.0</td>
<td>Silicon</td>
<td>1.09</td>
<td>-</td>
</tr>
<tr>
<td>Ch. 3</td>
<td>3.55-3.93</td>
<td>InSb</td>
<td>1.09</td>
<td>180-335</td>
</tr>
<tr>
<td>Ch. 4</td>
<td>10.3-11.3</td>
<td>HgCdTe</td>
<td>1.09</td>
<td>180-335</td>
</tr>
<tr>
<td>Ch. 5</td>
<td>11.5-12.5</td>
<td>HgCdTe</td>
<td>1.09</td>
<td>180-335</td>
</tr>
</tbody>
</table>
NOAA-AVHRR 12, 14, 15, 16 and 17 allow the retrieval of different satellite time passes necessary for the “four temperature algorithm”. MODIS data cannot be used since the satellite Terra passes above the region once a day. Finally, NOAA 15 passes at approximately 1:00 pm and NOAA 17 passes at approximately at 3:20 pm and NOAA 16 passes at approximately at 7:25 pm (see table 3.2 for characteristic information). The actual images for the three passes can be seen in Appendix C. The data necessary for the estimation of thermal inertia values is retrieved from these three passes. Data sets were downloaded from the Satellite Active Archive (SAA). The SAA is part of the National Environmental Satellites, Data, and Information Service (NESDIS).

Table 3.2 Characteristic information on the satellite passes used.

<table>
<thead>
<tr>
<th>Start Time</th>
<th>End Time</th>
<th>Sat ID</th>
<th>Dataset Name</th>
</tr>
</thead>
</table>

AVHRR data have 3 different formats: GAC, HRPT, and LAC. GAC stands for Global Area Coverage and is intended to allow worldwide coverage with a manageable amount of data and its spatial resolution is 16 km. The only difference between Local Area Coverage (LAC) and High Resolution Picture Transmission (HRPT) reside in the way data is transferred from the satellite to the ground station.
LAC data are selectively recorded on board for subsequent playback while HRPT data are sent to Earth continuously in real-time (NOAA, 1990). The format used in this research is HRPT for its 1.1km resolution at nadir and its 0.7°C accuracy on sea surface temperature.

### 3.3 Fourier Series to Determine Thermal Inertia

#### 3.3.1 Fourier Series Definition

A Fourier series is a powerful tool used in solving differential equations. An infinite series in which the terms are constants \( A, B \) multiplied by sine or cosine functions of integer multiples \( n \) of the variable \( x \). The general Fourier series is given by:

\[
A_0 + \sum_{n=1}^{\infty} \left( A_n \cos(nx) + B_n \sin(nx) \right),
\]

Let \( f \) be continuous on \( I = [-\pi, \pi] \). Suppose series \( f \) converges uniformly to \( f \) for all \( x \in I \) then,

\[
A_n = \frac{1}{\pi} \int_{-\pi}^{\pi} f(t) \cos(nt) \, dt \quad n = 0, 1, 2, \ldots,
\]

\[
B_n = \frac{1}{\pi} \int_{-\pi}^{\pi} f(t) \sin(nt) \, dt \quad n = 1, 2, \ldots,
\]
The numbers $A_n$ and $B_n$ are called the **Fourier coefficients** of $f$. When $A_n$ and $B_n$ are given by (3.3, 3.4), the trigonometric series (3.2) is called the **Fourier series** of the function $f$. The function $f$ used to determine thermal inertia is the diffusion equation stated and solution derived by Carslaw and Jaeger (1959).

### 3.3.2 Thermal Inertia Algorithm

#### 3.3.2.1 Diffusion equation

Presently, all thermal models have assumed one-dimensional periodic heating of a uniform half-space (a region bounded by a plane on its upper side and extending downward to infinity) of constant thermal properties. The temperature is driven by a form of the diffusion equation:

$$k \frac{\partial^2 T(x,t)}{\partial x^2} = \rho c \frac{\partial T(x,t)}{\partial t},$$

where:
- $T(x,t) = \text{the temperature at depth } x \text{ below the surface (Kelvin)}$,
- $t = \text{time (seconds)}$,
- $k = \text{the thermal conductivity of the half space (J m}^{-1} \text{ s}^{-1} \text{ K}^{-1})$,
- $\rho = \text{the density (kg m}^{-3})$,
- $c = \text{the specific heat of the material (J kg}^{-1} \text{ K}^{-1})$.

This is called the one-dimensional Fourier heat conduction law in solid.
The solution to (3.5) for periodic heating of angular frequency \( \omega \) is given by Carslaw and Jaeger (1959) as:

\[
T(x, t) = \sum_{n=0}^{\infty} D_n \exp \left( -k \sqrt{n} x \right) \cos \left( n \omega t - \xi_n - k \sqrt{n} x \right)
\]  

(3.6)

where;

\( D_n \) and \( \xi_n \) are arbitrary coefficients and \( k \equiv \sqrt{\omega/2k} \) is the wavenumber of the first harmonic. The arbitrary coefficients are evaluated by expressing the surface boundary condition as an energy balance between incoming solar and sky radiation, outgoing ground radiation, and conduction into the ground.

\( T(x, t) \) will oscillate with a period of 24 hours plus various harmonics of that period. The amplitude of these oscillations will decay exponentially as we go deeper below the surface. In equation (3.6), heat transfer mechanisms such as atmospheric conduction and convection, convective heat transfer in the ground, and latent and sensible heat effects one must specify the appropriate boundary conditions in order to solve equation (3.5). The ones used in this research are (Watson, 1975):

\[
- K \frac{\partial T(x, t)}{\partial x} \bigg|_{x=0} = \left( 1 - a \right) S_0 C_i \cos Z' - [A_e + BT(0, t)], 
\]

(3.7)

and

\[
T(x, t) \text{ is finite as } x \to \infty, 
\]

(3.8)

where:

\( a \) = the surface albedo (dimensionless),
\( S_o \) = the solar constant (W m\(^2\)),
\( C_t \) = the atmospheric transmittance (dimensionless),
\( A_c \) and \( B \) = the linearization coefficients of the boundary condition which is from the dynamic energy balance equation (3.9) at the ground surface.

\[
-K \frac{\partial T(x,t)}{\partial x} \bigg|_{x=0} = (1 - A) S_o \ C_t \ \cos \theta - R_{earth} + R_{sky} - H - LE
\]

(3.9)

where:
\( R_{earth} \) is the Earth emitted radiation,
\( R_{sky} \) is the downward longwave sky radiation,
\( H \) is the sensible heat flux to the atmosphere,
\( LE \) is the latent heat flux to the atmosphere.

The solution of the diffusivity equation (3.5) subject to the boundary conditions in equation (3.6) and (3.7) for the ground surface temperature can be obtained (Xue and Cracknell, 1992):

\[
T(x,t) = -A \frac{x}{B} + (1 - A) S_o \ C_t \ \sum_{n=1}^{\infty} A_n \ \exp \left( -k_n \sqrt{n \omega} \right) \cos \left( n \omega t - k_n \sqrt{n \omega} \delta_n \right) \cos \left( \sqrt{\omega n^2 \beta^2 - \cos^2 \theta} \right), \quad (3.10)
\]

where:
\( P \) = the thermal inertia (TIU) = 4.1854*10\(^4\) cal cm\(^{-2}\) s\(^{-1}\) \(^\circ\)C\(^{-1}\),
\( k_n = \frac{P \sqrt{\omega}}{K \sqrt{2}} \),

\( \delta_n = \arctan \left( \frac{P \sqrt{\omega}}{2B + P \sqrt{\omega}} \right) \),

\( A_n = \frac{2}{\pi} \sin \delta \ \sin \alpha + \frac{1}{2\pi} \cos \delta \ \cos \alpha \ [\sin(2\psi) + 2\psi] \),

(3.11)
\[ A_n = \frac{2 \sin \delta \sin \lambda}{n \pi} \sin(n\psi) + \frac{2 \cos \delta \cos \lambda}{\pi(n^2 - 1)} \left[ n \sin(n\psi) \cos \psi - \cos(n\psi) \sin \psi \right], \quad (3.14) \]

with \( n = 2,3,\ldots, \)

\[ \psi = \arccos(tg \delta tg \alpha), \quad (3.15) \]

where:

\( \theta = \) the zenith angle (radians),

\( \delta = \) the solar declination (radians),

\( \lambda = \) the latitude of the grid point (radians),

\( \omega = 7.272 \times 10^{-5} \) and is the angular velocity of rotation of the Earth.

Dr. Xue explains in more details the solution of the heat diffusion equation with these specific boundary conditions in his Ph-D dissertation. Below is a summary of his derivation:

Let \( \phi(x,t) = T(x,t) - \frac{K}{B} \frac{\partial T(x,t)}{\partial x}, \quad (3.16) \)

where \( \phi(x,t) \) satisfies:

\[ D \frac{\partial^2 \phi(x,t)}{\partial x^2} = \frac{\partial \phi(x,t)}{\partial t}, \quad \text{with two boundary conditions:} \]

1) \( \phi(0,t) = -\frac{A}{B} + \frac{(1-A) S_c C \cos Z}{B}, \quad (3.17) \)

2) \( \phi(x,t) \) is finite as \( x \to \infty. \quad (3.18) \)

The initial condition Dr. Xue used was not specified in the appendix of his dissertation but appears to be:
\[ \phi(x, 0) = 0. \]

\( \phi(x, t) \) is also a periodic function and we have:

\[ \phi(0, t) = \sum_{n=0}^{\infty} A_n \cos(n \omega t - \epsilon_n) \] as seen in Carslaw and Jaeger (p81, 1959) where the surface is considered a plane \( x = 0 \).

where:

- \( A_n \) and \( \epsilon_n \) are amplitude and phase difference for each harmonic term in the Fourier expansion series of \( \cos Z \).

The temperature at depth \( x \) is:

\[ \phi(x, t) = \sum_{n=0}^{\infty} A_n e^{-kx \sqrt{n}} \cos \left( n \omega t - \epsilon_n - \frac{1}{2k n^2} \right), \quad (3.19) \]

where:

- \( k = \left( \frac{\omega}{2k} \right)^{\frac{1}{2}} \).

\( A_n \) tends to become zero quickly with the increment of \( n \). Price (1977) concluded that the first term in the series is dominant. A first-order approximation is obtained from (3.8) for thermal inertia \( P \) (Xue and Cracknell, 1992). However, the first order is only a good approximation for dry area where surface evaporation and vegetation are constant in the domain studied. In this research, surface moisture and vegetative cover vary from one pixel to another; therefore surface evaporation reduces the amplitude of soil flux in comparison with the amplitude in dry areas (Price, 1985). In this case, according to Xue and Cracknell (1995), the second order of the Fourier series should be used. Consequently a second-order approximation thermal inertia model can be obtained and is given by Xue and Cracknell (1995) as:
\[
P = \frac{(1-a)S_oC_i}{\Delta T \sqrt{\omega}} \left[ A_1 \left[ \cos(\omega t_1 - \delta_1) - \cos(\omega t_1 - \delta) \right] \right. \\
+ \left. A_2 \left[ \cos(\omega t_2 - \delta_2) - \cos(2\omega t_1 - \delta_2) \right] \right] \\
\sqrt{1 + \frac{1}{b} + \frac{1}{2b^2}}
\]

where:

\( P = \) the thermal inertia (TIU) = \(4.1854 \times 10^4\) cal cm\(^2\) s\(^{-1}\) °C\(^{-1}\),

\( a = \) the surface albedo or land surface reflectance (dimensionless),

\( S_o = \) the solar constant (Wm\(^{-2}\)nm\(^{-1}\)),

\( \omega = \) the angular velocity of rotation of the Earth (rad s\(^{-1}\)),

\( C_i = \) the atmospheric transmittance in the visible spectrum (dimensionless),

\( A_1 \) and \( A_2 = \) the first and second coefficients of Fourier series (dimensionless),

\( \Delta T = \) the day-night surface temperature difference, \( t_1 \) and \( t_2 \) are the time (hours) of diurnal and nocturnal satellite passes,

\( b = \frac{\tan(\delta)}{1 - \tan(\delta)} \) (dimensionless),

\( \delta_2 = \arctan \left( \frac{b\sqrt{2}}{1 + b\sqrt{2}} \right) \) (dimensionless),

\( \delta_1 = \arctan(\xi) + (2m+1)\pi, \quad m = 0,1,2,\ldots \) (dimensionless),

\( \xi = \frac{(T_j - T_k) \left[ \cos(\omega t) - \cos(\omega t_j) \right] - (T_i - T_j) \left[ \cos(\omega t_k) - \cos(\omega t_k) \right]}{(T_i - T_j) \left[ \sin(\omega t) - \sin(\omega t_j) \right] - (T_j - T_k) \left[ \sin(\omega t_k) - \sin(\omega t_k) \right]} \),

where:

\( t_i, t_j, \) and \( t_k = \) the actual local solar times (hours) of the three different satellite passes for which three surface temperatures correspond, respectively.

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The importance of equation (3.21) is that it provides the phase angle information of the diurnal temperature change $\delta_1$ as a function of the times of three satellite overpasses instead of the time of maximum diurnal surface temperature (Sobrino and Kharraz, 1999). All parameters in equation (3.20) have been defined except for the thermal difference term $\Delta T$ and $A_1$ and $A_2$, the first and second coefficients of Fourier series.

### 3.3.2.2 Thermal Differences – Split-Window Algorithm

The method widely used to estimate temperature differences is a multispectral method called the split-window method. It makes use of the observation that the transmission of a path through a moist atmosphere at one wavelength is closely correlated with the transmission through the same path at a second, nearby wavelength (Prata, 1993). The split-window method uses channel 4 and 5 from the AVHRR sensor, which can be used to calculate this differential absorption effect. Many authors have used such technique in recent years (Price, 1984; Sobrino et al., 1991, 1994; Prata, 1993, Becker and Li, 1990, 1995). The split-window equation to find the temperature, used by these authors, is shown below:

\[
T_s = A_o + P \frac{(T_4 + T_5)}{2} + M \frac{(T_4 - T_5)}{2},
\]  

(3.22)

where:
\( T_s \) = the surface temperature (K),
\( A_o, P \) and \( M \) = coefficients worked out by statistical analysis and given by:

\[
A_o = 1.274 \text{ (dimensionless)},
\]

\[
P = 1 + \left[ 0.15616 \left( \frac{1 - \varepsilon}{\varepsilon} \right) - 0.482 \left( \frac{\Delta \varepsilon}{\varepsilon^2} \right) \right] \text{ (dimensionless)},
\]

\[
M = 6.26 + \left[ 3.98 \left( \frac{1 - \varepsilon}{\varepsilon} \right) + 38.33 \left( \frac{\Delta \varepsilon}{\varepsilon^2} \right) \right] \text{ (dimensionless)},
\]

\[
\varepsilon = \frac{(\varepsilon_4 + \varepsilon_5)}{2}, \quad \Delta \varepsilon = (\varepsilon_4 - \varepsilon_5) \text{ (dimensionless)},
\]

\( T_4 \) and \( T_5 \) are brightness temperatures in channel 4 and 5 of AVHRR respectively.

The split-window explained above requires knowledge of the total atmospheric water vapor content and effective surface emissivity. The second order Fourier series (3.20) only uses surface temperature differences between the satellite passes, which could simplify the calculations. The time between 2 satellite passes does not exceed 12 hours; we can assume that emissivity and the total water vapor would remain constant. Thus, the temperature difference maybe written from (3.22) and for two satellite passes \( i, j \) (Sobrino et al., 1999) as:

\[
\Delta T = T_{si} - T_{sj} = \Delta T_{4ij} + \Delta T_{\text{atmos}}, \quad (3.23)
\]

with:

\[
\Delta T_{4ij} = (T_{4i} - T_{4j}), \quad (3.24)
\]

\[
\Delta T_{\text{atmos}} = 1.4 \left( T_{45i} - T_{45j} \right) + 0.32 \left( T_{45i}^2 - T_{45j}^2 \right), \quad (3.25)
\]

\[
T_{45i} = (T_4 - T_5), \quad (3.26)
\]

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where:

\( T_{4i} \) and \( T_{4j} \) = the brightness temperatures (K) from channel 4 for two satellite passes \( i \), \( j \).

\( T_{5i} \) and \( T_{5j} \) = the brightness temperatures (K) from channel 5 for two satellite passes \( i \), \( j \).

### 3.3.2.3 First and Second Fourier Coefficients

The first and second Fourier coefficients are essential to obtain thermal inertia (equation 3.20). They are given by:

\[
A_1 = \frac{2}{\pi} \sin \delta \sin \psi + \frac{1}{2\pi} \cos \delta \cos \lambda \left[ \sin(2\psi) + 2\psi \right], \tag{3.27}
\]

\[
A_2 = \frac{\sin \delta \sin \lambda}{\pi} \sin(2\psi) + \frac{2\cos \delta \cos \lambda}{\pi} \left[ 2\sin(2\psi) \cos \psi - \cos(2\psi) \sin \psi \right], \tag{3.28}
\]

where:

\( \delta \) = the solar declination in radians,

\( \lambda \) = the latitude of the region studied in radians,

\( \psi = \arccos(\tan \delta \tan \lambda) \).
3.4 Brightness temperatures extraction from AVHRR images

3.4.1 Radiometric Correction and Calibration

NOAA-AVHRR HRPT level 1b data have been widely used for environmental research and estimation of numerous variables such as radiance, reflectance, albedo, and surface temperature. The level 1b data contain raw AVHRR spectral data as well as calibration coefficients, solar zenith angles, earth location and other auxiliary data such as ground control points (GCPs) necessary to preprocess AVHRR level 1b data. Indeed, the spectral data contain both geometric and radiometric errors, which must be removed in order to quantitatively, analyze AVHRR data. Radiometric errors are usually due to change in scene illumination, atmospheric conditions, viewing geometry, and instrument-response characteristics (Lillesand and Kiefer, 1987).

The radiometric correction is due to the large scanning angle of the AVHRR (about ± 55.4 degrees) resulting in quite different amounts of solar radiation received by a ground object. Therefore, the solar zenith angle varies significantly along one scan line causing spectral errors in the visible and near infrared channels of AVHRR data. A cosine correction can remove the errors created by the solar zenith angle:
\[ DN_\theta = \frac{DN_\theta}{\cos \theta}, \quad (3.29) \]

where:
\( DN_\theta = \) a digital number with a solar-zenith angle of \( \theta \) degrees,
\( DN_0 = \) a digital number with an angle of 0 degrees.

The calibration of AVHRR data converts channels 1 and 2 to reflectance and channels 3, 4 and 5 to brightness temperatures. The equation to convert digital numbers (DNs) from the thermal infrared bands to their brightness temperatures is given by NOAA (1990) as:

\[ T_s(E) = C_2 \nu / \ln \left(1 + C_1 \nu^3 / E \right), \quad (3.30) \]

where:
\( T_s = \) the surface brightness temperature (K),
\( E = \) the energy value (irradiance at instrument aperture),
\( \nu = \) the central wave number of channel filter (cm\(^{-1}\)),
\( C_1 = 1.1910659 \times 10^{-5} \text{ mW/(m}^2\text{sr cm}^{-4}), \)
\( C_2 = 1.438833 \text{ (cm K)}. \)

The radiant energy can be obtained by converting the DNs and is given by NOAA (1990) as:

\[ E = c \ast TIR + d, \quad (3.31) \]

where:
\( TIR = \) a DN from AVHRR channel 3, 4 or 5,
\( c, d = \) constants appended from the AVHRR file.
Constants $c, d$ differ from AVHRR 15, 16 and 17. The software used in this research to geometrically and radiometrically correct the brightness temperatures data sets from each satellite, is ENVI 3.6. ENVI is written in Interactive data Language (IDL), a programming language that offers integrated image processing tools.

### 3.4.2 Geometric Correction

It is usually necessary to preprocess the remotely sensed data before analyzing it. Indeed, errors creep into the data acquisition process and can degrade the quality of the remote sensor data collected (Duggin and Robinove, 1990). These errors can be divided into 2 classes: 1) Those that can be corrected using data from platform ephemeris and knowledge of internal sensor distortion and 2) those that cannot be corrected with acceptable accuracy without a sufficient number of ground control points (Jensen, 1996). Therefore the purposes of geometric correction are to remove these path radiance errors but also to relate the digital remote sensing data to a map projection. The map projection used for the mesoscale is evidently a geographic projection to represent every grid point with a latitude and longitude coordinate. This operation involves relating the pixel coordinates (row and column) of ground control points (GCPs) with their corresponding map coordinates which would be the latitude and longitude in this
case. This will establish the nature of the geometric coordinate transformation that must be applied to rectify or relocate every pixel in the original input image \((x', y')\) to its proper position in the rectified output image \((x, y)\). This process is called spatial interpolation. The typical projection equations relating the map coordinates and image coordinate are polynomials such as:

\[
x' = a_0 + a_1 x + a_2 y, \quad (3.32)
\]

\[
y' = b_0 + b_1 x + b_2 y, \quad (3.33)
\]

where:

- \(x\) and \(y\) are positions in the output-rectified image, while \(x'\) and \(y'\) represent positions in the original input image. The coefficients are determined by regression analysis of GCPs. The order of the polynomial is calculated by both the magnitude of the distortion in the raw image and the number of available GCPs (Jensen, 1996). There are 51 possible ground reference values for each scan line. ENVI 3.6 offers radiometric and geometric correction IDL scripts as well as calibration scripts for a number of satellite data such as NOAA AVHRR. IDL scripts in ENVI 3.6 allow the user to specify the projection and the number of ground control points.

### 3.5 Cloud Removal Procedures

After channel 4 and 5-satellite imagery, have been calibrated and radiometrically corrected with ENVI, pixels need to be analyzed to ensure that no
clouds are affecting the data. In order to know if a pixel is a cloudy pixel; the reflectance obtained after calibration of channel 1 is used. First, a mask is created with ENVI from channel 1 where every pixel with reflectance above a 10% threshold is considered a cloudy pixel (see figures 3.1 and 3.2).

Figure 3.1 Central East Florida AVHRR satellite picture representing band 3, 4 and 5 (brightness temperature) on October 3rd 2001 at 12:48:22 UTC with clouds.
Figure 3.2 Central East Florida cloud mask created from reflectance values of AVHRR band 1 on October 3rd 2001 at 12:48:22 UTC.

This cloud mask is then applied on channel 4 and 5 images of the AVHRR data and allows removal all cloudy pixels (see figure 3.3).
Figure 3.3 Central East Florida AVHRR HRPT satellite picture composed of channels 3, 4 and 5 (brightness temperature) on October 3rd 2001 after applying the cloud mask. The grid resolution is 100x100.

Finally, an interpolation method (based upon IDW techniques) is applied to estimate the brightness temperatures of the cloud-covered pixels. In order to the IDW correctly, Geographic Information Systems (GIS) techniques were utilized to execute the interpolation such as the use of “barrier input line theme” where these
themes are created as boundaries in order to limit the influence of land and water landscape boundaries.

3.6 GIS Interpolation

3.6.1 Barrier Florida file

Each line in a barrier input line theme is used as a break that limits the search for input sample points. A line can represent a cliff, ridge, or some other interruption in a landscape.

Figure 3.4 Central East Florida water barrier file used during the IDW interpolation.
3.6.2 IDW interpolation technique

The AVHRR HRPT satellite data for all three passes is by now cleared of all pixel clouds. These pixels have a null value; therefore an interpolation needs to be applied to associate them with a brightness temperature.

3.6.2.1 Preparing the interpolation

The data transfer from one platform to another can be sometime cumbersome. The data was saved as an ASCII file and the header file had to be modified to suit the GIS platform used to interpolate the temperatures. The data was first displayed as a grid file where grid cell would have temperatures values and null values for previously removed cloud pixels. Then, the data had to be extracted into an x, y and z table to be reinserted as a point file so the interpolation could take place.

3.6.2.2 IDW method

IDW or inverse distance weighting is a technique in which interpolated estimates are based on values of nearby known values that are weighted by their distance from the interpolation location. This technique makes a basic assumption that nearby points ought to be more closely related than distant points to the value at the interpolation location. The neighborhood around the interpolated point is
identified and a weighted average is taken of the observation values within this neighborhood. The weighting function used is inverse power.

\[
W(d) = \frac{1}{dp},
\]

(3.34)

where:
\(W\) = the weight of a nearest neighbor,
\(d\) = the distance of nearest neighbor from point of interest,
\(p\) = the power (\(p > 0\), and \(< \infty\) in ArcView Spatial Analyst; however, values 8 or less are recommended).

We chose a power of 8 nearest neighbors to perform the interpolation. This combination assures that all nearest neighbors are within about 2 km of each interpolated point. This information is based on direction within ArcView™ Spatial Analyst 2.0.
Figure 3.5 Flowchart summarizing the different procedures to obtain cloudless brightness temperature grid.
3.7 Thermal inertia comparison between three consecutive days

The UTC-M mesoscale model was meant to simulate mesoscale circulation up to at least three days. The idea would be to assume thermal inertia constant for three consecutive days. In order to make this assumption, thermal inertia was calculated for three consecutive days (October 22\textsuperscript{nd} to 24\textsuperscript{th} 2002). To be able to compare three different thermal inertia datasets, the three days would need to be cloud free or almost cloud free days. Unfortunately, October 23\textsuperscript{rd} and 24\textsuperscript{th} are very cloudy days and the interpolation of brightness temperatures values is very important. The interpolation of brightness temperature has an important impact on the accuracy of thermal inertia. Additionally, the thermal inertia model relies on temperature difference calculations. Therefore, satellite time passes have to be close to the time of lowest and highest temperature. Satellite passes on the 23\textsuperscript{rd} and the 24\textsuperscript{th} of October are not at these ideal times increasing the error in the thermal inertia calculation.

The thermal inertia results obtained on October 23\textsuperscript{rd} and 24\textsuperscript{th} are not acceptable for all the reasons explained above. However, a cloudless subset was extracted from these three days to analyze the accuracy of the thermal inertia calculation.

Figure 3.6 shows the subset taken from the three thermal inertia calculation. The subset is a 10x10 grid cell size composed of both land and ocean pixels.
Figure 3.6 Thermal Inertia subset grid (10x10 cell grid), subset calculated for 3 consecutive days (22\textsuperscript{nd} – 24\textsuperscript{th} October 2002). The subset represented in this figure is for October 22\textsuperscript{nd} 2002.

Figure 3.7 represents three thermal inertia histograms for three consecutive days. Differences observed between these three histograms are the result of a
limited interpolation for the thermal inertia calculated for both October 23rd and 24th and the satellite passes for these 2 days are not at the times of maximum and minimum day temperatures which is necessary to estimate thermal inertia correctly.

![Thermal Inertia Histograms](image)

Figure 3.7 Thermal inertia histograms for October 22nd 2002 (upper left), October 23rd 2002 (upper right) and October 24th 2002 (lower left).

Most of the pixel count is around 500TIU which represent an average thermal inertia for land. The second peak is at 1500 TIU which is a familiar thermal
inertia for ocean water. A cumulative frequency curve shows however that results for the thermal inertia calculation are acceptable (figure 3.8).

Figure 3.8 Cumulative frequency curves respectively representing thermal inertia for three consecutive days, October 22\textsuperscript{nd} to October 24\textsuperscript{th} 2002.
CHAPTER 4

THERMAL MODEL PART III

TEMPERATURE TIME-DEPENDANT EQUATION

4.1 Conduction of heat in solids

From Carslaw and Jaeger (1959), the temperature in degrees in a semi-infinite solid subjected to flux $f(t)$ per unit time per unit area at $z=0$ and with the region $z>0$ initially at constant temperature is:

$$T = \frac{1}{\sqrt{\pi xt}} \int_{0}^{\infty} f_o(L) e^{-\frac{L^2}{4xt}} dL + \frac{1}{K} \sqrt{\frac{x}{\pi}} \int_{0}^{t} f(t-\gamma) \frac{d\gamma}{\sqrt{\gamma}},$$  \hspace{1cm} (4.1)

where:
- $T =$ the temperature in degrees ($^\circ$C),
- $f_o(L) =$ the initial vertical soil temperature profile ($^\circ$C),
- $f(t-\gamma) =$ the incremental flux acting over $d\gamma$ (cal cm$^{-2}$s$^{-1}$),
- $x =$ the diffusivity (cm$^2$ s$^{-1}$),
- $K =$ the conductivity (cal cm$^{-2}$s$^{-1}$$^\circ$C$^{-1}$),
- $t =$ the time (seconds),
- $x^{1/2}K^{-1} =$ the same as the thermal inertia ($TI$).
At the initial time (6:00 am), a near-perfect balance usually exists between downward atmospheric long-wave radiation and upward surface radiation. Under these conditions, the initial soil temperature profile \( f_0 \) can be considered constant. If the net flux is assumed to be constant at each time step, equation (4.1) can be written as (Gannon, 1978):

\[
T^{n+1} = T^n + \frac{2Q^{n+1}}{P} \left( \frac{\Delta t}{\pi} \right)^{1/2},
\]

where:

\( T^{n+1} \) = the temperature at the next time step \( n + 1 \) (Kelvin),
\( T^n \) = the initial temperature at the current time step (Kelvin),
\( Q^{n+1} \) = the net flux across the air-ground interface which is considered constant over \( \Delta t \) (cal cm\(^{-2}\) s\(^{-1}\)),
\( P \) = the thermal inertia (cal s\(^{1/2}\) cm\(^{-2}\) °C\(^{-1}\)),
\( \Delta t \) = the time step (seconds).

The contribution of the soil heat flux to the net flux is determined by the value of thermal inertia. Thermal Inertia can be defined as the resistance of a material to temperature change. In the case of the land surface, the variation in surface temperature depends on ground properties and characteristics.

The Florida Tech UTC-M Mesoscale Model is a complex model as it combines both land and sea surfaces. The case of the sea is somewhat easier than that of the land for several reasons. First, sea surfaces temperatures tend to fluctuate
less over a diurnal period than land temperatures and second, emissivity of the sea is virtually constant and its value is assumed to be known. On land, emissivity is highly variable both spatially and temporally. Discussions of the determination of sea surface temperatures from thermal infrared satellite data are given, for example, by Robinson et al. (1984) Robinson (1985), Stewart (1985) and Fuiza (1992).

4.2 Scaling Analysis of the temperature time dependant equation

The importance of this scale analysis is to provide insight into which processes or forces may be responsible for the increase or decrease in surface temperature. The scaling analysis will provide an estimate of the change of temperature for a time step of 2 hours or 3600 seconds. The UTM-C Mesoscale Model region of focus is the planetary boundary layer and this research will focus below the level of cloud formation where water phase changes of water begin to occur. Also, we will assume a cloudless atmosphere with complete transmission of shortwave energy and rapid absorption of longwave radiation. For testing purposes, a dry atmosphere will be assumed which permits us to neglect latent heat sources. To sum up, the largest source of longwave energy in the atmosphere is the radiation of longwave radiation by the solid surface of the earth and the rapid absorption of the longwave energy by the atmosphere (McIleveen 1986).
The conservation of heat relation cannot be part of a scale analysis because of the complex mathematical form of the source-sink term (Pielke, 1981).

The energy conservation equation can be written in this form (Uhlhorn, 1996):

\[
\frac{d\theta}{dt} = \frac{\partial \theta}{\partial t} + u \frac{\partial \theta}{\partial x} + v \frac{\partial \theta}{\partial y} + w \frac{\partial \theta}{\partial z} + S_R, \tag{4.3}
\]

Therefore, sources or sinks are due entirely to radiative absorption of longwave energy emissions by the earth’s surface.

\( S_R \) is the source and sink component of heat due to longwave radiative energy absorption.

<table>
<thead>
<tr>
<th>Table 4.1 Scale analysis of the energy conservation equation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Scale</td>
</tr>
<tr>
<td>-------</td>
</tr>
<tr>
<td>Dimension</td>
</tr>
<tr>
<td>Value</td>
</tr>
<tr>
<td>Reference</td>
</tr>
</tbody>
</table>

The temperature time dependent equation used in this model can be scale using these scaled term estimates:
\[ T^{n+1} = T^n + \frac{2Q^{n+1}}{P} \left( \frac{\Delta t}{\pi} \right)^{\frac{1}{2}} \]

Table 4.2 Scale analysis of the temperature time dependant equation.

<table>
<thead>
<tr>
<th>Scale</th>
<th>( T^n )</th>
<th>( Q^{n+1} )</th>
<th>( P )</th>
<th>( \Delta t )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Units</td>
<td>Kelvin</td>
<td>Cal cm(^{-2}) s(^{-1})</td>
<td>Cal s(^{-1/2}) cm(^{-2}) oC(^{-1})</td>
<td>Seconds</td>
</tr>
<tr>
<td>Value</td>
<td>300</td>
<td>5.10(^{-3}) and 2.5.10(^{-3})</td>
<td>0.04-0.6</td>
<td>3600</td>
</tr>
</tbody>
</table>

This scale analysis shows that after sunrise, a temperature gradient can be observed every hour depending on the thermal inertia and the amount of energy available at the surface. For example, a pixel with a thermal inertia of 0.2 Cal s\(^{-1/2}\) cm\(^{-2}\) oC\(^{-1}\) (forested area) and with a net radiation of 4.4 \( 10^{-3} \) Cal cm\(^{-2}\) s\(^{-1}\) (~ 8:00am) will experience an increase of 0.03 degrees at every time step (45 seconds) into the simulation.

4.3 Flow chart summary of the thermal model

The net surface heat flux \( Q^{n+1} \) is determined from part one of the thermal model (see figures 4.1 and 4.2). The radiative transfer model allows the estimation of the downwelling shortwave radiation at the surface (figure 4.1). The addition of the upwelling shortwave and the downwelling and upwelling longwave radiation
give the total energy flux at the surface (figure 4.2). The thermal inertia $P$ is
determined from part two of the thermal model (see figure 4.3). The temperature
time dependent equation relating the total net heat budget and the thermal inertia to
the change in temperature is described in figure 4.4. Finally, a detail diagram
describing the source of all satellite data can be seen in figure 4.5.
Figure 4.1 Flow chart of the thermal model part one. The radiative transfer is divided in two parts, the earth positioning and atmospheric transmission calculations. These 2 calculations allow the estimation of direct and diffuse radiation at the surface for downwelling shortwave radiation.
Figure 4.2 Flow chart of the thermal model part two. The surface heat budget is divided into four parts, the downwelling shortwave (see figure 4.1), the upwelling shortwave, and the downwelling and upwelling longwave. These 4 calculations allow the estimation of the total net energy flux at the surface for shortwave and longwave.
Thermal Inertia Model
Thermal-Model Flow Chart (Part II)

Figure 4.3 Flow chart of the thermal model part three. The thermal inertia model requires the knowledge of albedo, thermal differences and the first and second coefficient of Fourier series.
Figure 4.4 Flow chart of the thermal model part four. Temperature time dependant model showing the relation between all parts of the thermal model: the radiative transfer (part one and two) and the thermal inertia model (part three).

\[ T^{n+1} = T^n + \frac{2Q^{n+1}}{P} \left( \frac{\Delta t}{\pi} \right)^{\frac{1}{2}} \]
Figure 4.5 Diagram describing the contribution of satellite data from MODIS and AVHRR sensors to the thermal model.
4.4 Courant Friedrichs Lewy adjustment

The CFL condition is a particular effect of computational instability (error in approximate numerical computations which increases rapidly as the computation proceeds) associated with numerical weather forecasting. If the grid size is less than the distance traveled in the time-step interval by the fastest waves permitted by the equation, errors will grow and swamp the physical solution. Courant, Friedrichs, and Lewy made the remarkable discovery that the steps in time could not be chosen arbitrarily but had to be smaller than some constant times the steps in the space variable. For the wave equation that constant was the reciprocal of the speed of propagation. For other equations there may be many such speeds or one may have a different kind of propagation, but there is always a limitation of space step ($\Delta x$) and time step ($\Delta t$) (Hess, 1979).

\[
\frac{\Delta x}{\Delta t} \geq C \quad (4.4)
\]

where:

$\Delta x$ is the grid cell size (meters),

$\Delta t$ is the time step (seconds),

$C$ is the speed of the fastest wave permitted by the equations (m/s).
The increase in resolution in terms of the number of grid cells (25 to 200) created a computational instability. This error was corrected by reducing the time step from a 45-second to a 30-second time step to agree with the CFL condition.
CHAPTER 5
PARALLEL PROCESSING

5.1 What is a parallel computer?

A parallel computer is a set of processors that are able to work cooperatively to solve a computational problem (Pacheco, 1997). Given a problem to be solved, it is broken into a number of sub problems. These sub problems are solved simultaneously, each on a different processor. A workstation computer consists essentially of a single processing unit, or processor, that executes a single sequence of instructions on a single sequence of data, as shown in figure 5.1. The program tells the processor how to solve a certain problem. The control unit emits one instruction that operates on a set of data obtained from the memory unit. In a parallel computer, there are several processors (48 for Bluemarlin, supercomputer at Florida Tech) and they communicate with one another to exchange partial results and then the results are combined to produce an answer to the original problem as shown in figure 5.2.
Figure 5.1 Diagram describing a sequential (or serial) computer

Figure 5.2 Diagram describing a distributed-memory parallel computer.
It is now possible to construct very fast, low-cost processors. Currently, the speed of off-the-shelf microprocessors is within one order of magnitude of the speed of the fastest serial computers. And microprocessors cost many orders of magnitude less. Thus by connecting only few microprocessors together to form a parallel computer, it is possible to obtain raw computing power comparable to that of the fastest serial computers and the cost of such parallel computers is considerably less. The supercomputer used in this research is the Beowulf at Florida Tech. A Beowulf is a high-performance, massively parallel computer made up of a cluster of nodes connected by a high-speed network that perform intense computing tasks. The Beowulf at Florida Tech is a distributed memory supercomputer cluster that is in the MIMD paradigm and is named Bluemarlin (Figure 5.3). It contains 47 compute nodes and one head node running Red Hat Linux 6.2. The head node contains 2 PIII processors and the compute nodes each have a PIII processor and 512Mb of RAM. A cluster is a type of parallel or distributed system that consists of a collection of interconnected whole computers (one or more processors, adequate amount of memory, I/O facilities and an operating system) and is utilized as a single, unified computing resource (Pfister, 1995).
5.2 Why Parallel Computation?

Parallel computers are used primarily to speed up computations. A parallel algorithm can be significantly faster than the best possible sequential solution. There are a growing number of applications that involve processing huge amounts of data, or performing a large number of iterations or both. Parallel computation is the only way known today that would make these computations faster and feasible. Developments at the high end of computing have been motivated by numerical
simulations of complex systems such as weather, climate, mechanical devices, electronic circuits, manufacturing processes, and chemical reactions. In summary, the need for faster computers is driven by the increasing demand of the scientific and engineering world where applications become more data intensive and perform more sophisticated computations (Kumar et al., 1994).

5.2.1 Why parallel computation in this research?

In this research, several improvements to the UTC-M Mesoscale model call for the use of parallel processing techniques. First, the increase of resolution from a 25x25 grid cell size to a 200x200 grid cell size will require computing speeds in order to reduce the time of a 24-hour simulation. Second, the insertion of a thermal model that has a radiative transfer model will increase the computational time of runs.

5.2.2 Limitations of the UTC-M serial code

Problems are parallelizable to different degrees. For some problems, assigning partitions to other processors might be more time consuming than performing the task on one processor. In the case of the UTC-M model, the code was written in a serial mindset and therefore the output of one subroutine is used as inputs in the next subroutine making it difficult to use parallel techniques where
instructions are executed simultaneously. The thermal model developed in this research was written in such a way a simple parallelization could be applied to the code. In addition to the parallelization of the thermal model, a profiling work was done on the UTC-M mesoscale model to estimate the time of each subroutine and decide which subroutine should be parallelized.

5.3 Thermal Model and UTC-M Mesoscale Model profiles

The Pgprof Profiler was used to profile the code. The profiler allows users to discover which functions and lines were executed as well as how often they were executed and how much time of the total time they consumed. This information can be used to identify the portions of a program that will benefit the most from performance tuning.

5.3.1 Profile of the thermal model

The thermal model was profiled with Pgprof the results can be seen in Table 5.1. The atmos1 subroutine is taking 98% of the time and should therefore be parallelized in order to gain computational time. The “atmospheric” subroutine calculates the different transmittance coefficient for direct and diffuse radiation for every wavelength of the spectrum.
The advantage with the thermal model is that it was written so that it would be easily to parallelize. The schematic use to parallelize the thermal model can be viewed in section 5.4.

Table 5.1 Profiling chart for the thermal model. Calls for all subroutines and functions are represented and time is represented both in seconds and in percent of time.

<table>
<thead>
<tr>
<th>Function</th>
<th>Calls ..........</th>
<th>Time ..........</th>
</tr>
</thead>
<tbody>
<tr>
<td>atmos1</td>
<td>75625</td>
<td>352,777</td>
</tr>
<tr>
<td>upwelsht</td>
<td>75625</td>
<td>3,901,250</td>
</tr>
<tr>
<td>ti</td>
<td>75625</td>
<td>0.862064</td>
</tr>
<tr>
<td>radiatrans1</td>
<td>75625</td>
<td>0.401477</td>
</tr>
<tr>
<td>radiation</td>
<td>1</td>
<td>0.246781</td>
</tr>
<tr>
<td>tempdep</td>
<td>75625</td>
<td>0.124718</td>
</tr>
<tr>
<td>pdata</td>
<td>12</td>
<td>0.102715</td>
</tr>
<tr>
<td>upwellong</td>
<td>75625</td>
<td>0.061353</td>
</tr>
<tr>
<td>files</td>
<td>12</td>
<td>5.63e-05</td>
</tr>
</tbody>
</table>

5.3.2 Profile of the improved UTC-M model

The completed UTC-M mesoscale model was profiled with Pgprof and the results can be seen in Table 5.2. In this case, the horzadv subroutine is spending 65% of the time followed by the atmos1 subroutine with 17%. The “horizontal advection” subroutine performs a two-step “Lax-Wendroff” finite difference scheme. For the UTC-M Mesoscale model, the parallelization effort will be focused on the horizontal advection and can be viewed in section 5.4.
Table 5.2 Profiling chart for the improved UTC-M Mesoscale Model. Calls for all subroutines and functions are represented and time is represented both in seconds and in percent of time.

<table>
<thead>
<tr>
<th>Function</th>
<th>Calls</th>
<th>Time</th>
</tr>
</thead>
<tbody>
<tr>
<td>horzadv</td>
<td>9570032</td>
<td>65.29%</td>
</tr>
<tr>
<td>atmos1</td>
<td>486627</td>
<td>16.83%</td>
</tr>
<tr>
<td>dens</td>
<td>19585000</td>
<td>7.96%</td>
</tr>
<tr>
<td>horzdif</td>
<td>13916</td>
<td>1.81%</td>
</tr>
<tr>
<td>vertadv</td>
<td>1.13e+08</td>
<td>1.62%</td>
</tr>
<tr>
<td>tridiag</td>
<td>9916400</td>
<td>1.56%</td>
</tr>
<tr>
<td>mm2</td>
<td>1</td>
<td>1.13%</td>
</tr>
<tr>
<td>vertdif</td>
<td>6960</td>
<td>0.58%</td>
</tr>
<tr>
<td>upweelsht</td>
<td>1087500</td>
<td>0.41%</td>
</tr>
<tr>
<td>smooth</td>
<td>12182</td>
<td>0.40%</td>
</tr>
<tr>
<td>retpdpn</td>
<td>1740</td>
<td>0.26%</td>
</tr>
<tr>
<td>thetak</td>
<td>17420000</td>
<td>0.26%</td>
</tr>
<tr>
<td>dfvert</td>
<td>6956</td>
<td>0.23%</td>
</tr>
<tr>
<td>xisobr</td>
<td>1740</td>
<td>0.22%</td>
</tr>
<tr>
<td>bcadj</td>
<td>62628</td>
<td>0.19%</td>
</tr>
<tr>
<td>relhum</td>
<td>1740</td>
<td>0.17%</td>
</tr>
<tr>
<td>pgrad</td>
<td>6958</td>
<td>0.15%</td>
</tr>
<tr>
<td>tfilter</td>
<td>6956</td>
<td>0.14%</td>
</tr>
<tr>
<td>contint</td>
<td>1740</td>
<td>0.12%</td>
</tr>
<tr>
<td>mffconv</td>
<td>1740</td>
<td>0.12%</td>
</tr>
<tr>
<td>turbcs</td>
<td>1740</td>
<td>0.11%</td>
</tr>
<tr>
<td>ti</td>
<td>1087500</td>
<td>0.09%</td>
</tr>
<tr>
<td>shrcoefs</td>
<td>1740</td>
<td>0.06%</td>
</tr>
<tr>
<td>sfclyr</td>
<td>1740</td>
<td>0.05%</td>
</tr>
<tr>
<td>geostr</td>
<td>1740</td>
<td>0.05%</td>
</tr>
<tr>
<td>pprime</td>
<td>1740</td>
<td>0.04%</td>
</tr>
<tr>
<td>radiatransl</td>
<td>1087500</td>
<td>0.04%</td>
</tr>
<tr>
<td>retpdp</td>
<td>1740</td>
<td>0.03%</td>
</tr>
<tr>
<td>forcef</td>
<td>1741</td>
<td>0.02%</td>
</tr>
<tr>
<td>mixlnth</td>
<td>1740</td>
<td>0.01%</td>
</tr>
<tr>
<td>tempdep</td>
<td>1087500</td>
<td>0.01%</td>
</tr>
<tr>
<td>thermodata</td>
<td>174</td>
<td>0.01%</td>
</tr>
<tr>
<td>pldata</td>
<td>145</td>
<td>0.01%</td>
</tr>
<tr>
<td>horzdata</td>
<td>145</td>
<td>0.01%</td>
</tr>
<tr>
<td>upwellong</td>
<td>1087500</td>
<td>0.01%</td>
</tr>
<tr>
<td>specchu</td>
<td>10000</td>
<td>0.00%</td>
</tr>
<tr>
<td>hydpress</td>
<td>1</td>
<td>0.00%</td>
</tr>
<tr>
<td>files</td>
<td>464</td>
<td>0.00%</td>
</tr>
<tr>
<td>corparm</td>
<td>1</td>
<td>0.00%</td>
</tr>
</tbody>
</table>
5.4 Parallelization Methods

5.4.1 Parallelization scheme for the thermal model

The thermal model was written with the knowledge that it would be parallelized. Therefore all calls to functions and subroutines were written so that they could be executed simultaneously on different processors. In order to run a job simultaneously on different processors, calculations on each processor must be independent from one another. In other words, results obtained from node 1 are independent from results obtained on node 2 and etc. Every grid cell in the domain is independent from each other and therefore columns of cells can be associated to a processor as shown in figure 5.4.

Each processor is therefore working on a set of columns simultaneously reducing the computational time. First, all inputs files are read by all processors and the head node. This operation can be done in 2 different ways. The processors can read all input files from the Raid 5 box, which is the local hard drive accessible from all processors. The other way would consist in sending all input files to all local hard drive located on each of the nodes. The latter would reduce communication time from the processors to the Raid 5 box. After running several jobs, the size of the matrices being either 25x25 or 200x200, are not large enough to be a factor in communication time.
Figure 5.4 An 8x8 matrix is divided into 4 identical size matrices. Each processor is working simultaneously on a 2x8 matrix.

Each processor is now ready to work on a set of columns of the grid. In order to assure that each processor is working on a set of columns, the pseudo code below explains how it divides the domain grid to every processor. When all work is done on all processor, the head node will gather all columns and output the results to the Raid 5 box.

```plaintext
if (node=head node (or node 0)) then
  do m=1,xgrid
    do n=1,local_n
```
call subroutines and functions
end do
end do
else (if nodes = 1 to 25)

do m=1,xgrid
  do n=1,local_n
    call subroutines and functions
    matrix(m, node number*local_n+n)
  end do
end do
endif

call MPI_GATHER (local_matrix(m,n), xgrid*local_n, MPI_DOUBLE_PRECISION, final_matrix (m,n), xgrid*local_n, MPI_DOUBLE_PRECISION, 0, MPI_COMM_WORLD, ierr)

write output to the Raid 5 box.

where:

\[
\text{local}_n = \frac{\text{xgrid}}{\# \text{processors}} = \frac{25}{1 \text{ or } 5 \text{ or } 25} \text{ or } \frac{100}{1 \text{ or } 5 \text{ or } 10 \text{ or } 20 \text{ or } 25}
\]

The parallelized thermal model was run with 1 to 25 processors and the time for each run can be seen in figure 5.5. The size of the matrices is 100x100 and the run time decreases with an increase in processors. These times are for a 24-hour simulation. The time gained from a parallelized run compared to a serial run is around 7:30 hours. A profiling work of the parallelized code can be obtained with the new PGI 4.0 compilers but unfortunately they were not installed in time for this project. The profiling work can be very interesting since it gives a complete report
Runtime Comparison for Different Thermal Model Runs on Bluemarlin

Figure 5.5 Runtime comparison for different thermal model runs on the supercomputer Bluemarlin. The model was run on 1, 2, 5, 10, 20, and 25 processors. The size of the matrices is 100x100 and the runs are for a 24-hour simulation.
on time spent not only on computational time but also on communication time between all processors. The parallelization of the thermal model did not necessitate much communication time since all processors could work independently from each other. This is not the case with the parallelization work on the UTC-M mesoscale model.

5.4.2 Parallelization scheme for the UTC-M Mesoscale Model

The UTC-M Mesoscale Model was written in 1995-1996 (Uhlhorn, 1996). The code is a serial program and it was not developed at the time to accommodate a transfer to a parallelized code. The code follows a chronological mode where results obtained in subroutine 1 are used as inputs in subroutine 2 etc. In addition, the finite difference schemes used in this model, especially the staggered leapfrog scheme for the calculation of the horizontal advection, necessitate the dependency between 3 columns at one time. In other words, to estimate the wind velocity at (x,y,z) in the u direction, uwind (x-1,y,z) and uwind (x+1,y,z) are needed (see figure 5.6). In this case, grid columns cannot be sent to different processors since they are not interdependent. The UTC-M Mesoscale model is a 3-dimensional model where the vertical dimension is divided into 16 layers. From the profiling job of the improved UTC-M Mesoscale Model (see section 5.3.2), the horizontal advection subroutine is taking 65% of the complete runtime of a 24-hour
Figure 5.6 Numerical molecules for the staggered-leapfrog method. Solid dots represent known values and open dots represent predictions. Solid lines are space differences and dashed lines are time differences (modified from Uhlhorn, 1996).

simulation. The parallelization work should be focused on this subroutine and since the parallelization scheme used for the thermal model is not adequate for this subroutine, another scheme is used where the layers in the vertical are sent to different processors since they are independent from one another (Figure 5.7). Each processor is now ready to work on a set of matrices. When all work is done on all processor, the head node will gather all columns and output the results to the Raid 5 box.
Figure 5.7 A 3-dimensional matrix divided into 4 equal 3-dimensional matrixes. Each one of them is sent to a processor.

The main difference with the scheme used for the thermal model is that results obtained from the parallelized subroutine are used as inputs in the following subroutines. In addition, the inputs used in the parallelized subroutine needs to be sent from the head node and received by all processors. In order to assure that each processor is working on a set of matrices, the pseudo code below explains how it divides the domain grid to every processor.

```plaintext
if (node=head node (or node 0)) then
    MPI_SEND (to send the data to each processor)
    do m=1,xgrid
```

121
do n=1,local_n
    call subroutine horzadv
end do
end do
else (if nodes = 1 to 25)

MPI_RECV(to receive the data on each processor)

do m=1,xgrid
    do n=1,local_n
        call subroutines horzadv
        matrix(m, node number*local_n+n)
    end do
end do
endif

call MPI_GATHER (gather all local matrices (m,n) back to the head node)

write output to the Raid 5 box.

where:
local_n = \# of columns = \frac{zgrid}{\# processors} = \frac{16}{1, 2, 8, or 16}

The parallelized UTC-M Mesoscale model was run with 1 to 16 processors
and the time for each run can be seen in figure 5.8. The size of the matrices is
25x25 and the run time decreases with an increase in processors. These times are
for a 24-hour simulation. The time gained from a parallelized run compared to a
serial run is around 3:15 hours.
Figure 5.8 Runtime comparison for different UTC-M Mesoscale model runs on the supercomputer Bluemarlin. The model was run on 1, 2, 8, and 16 processors. The size of the matrices is 25x25 and the runs are for a 24-hour simulation.
Even though the profiler which comes with the new pg compilers 4.0 can profile MPI commands, it is possible to use time commands such as MPI W-time and Etime to get an understanding with the time spent on communications or computations. W-time measures both the time spent on computations and communications. The cputime from Etime measures the computational time. Communication time can be found by subtracting W-time with the cputime calculated by Etime. The speed-up time is the ratio between W-time with 1 processor over the W-times of multiples processors. The computational speed-up time is good with a speed-up of 20 between 1 processor and 16. However, due to the time spent in communication time and the system time used by the operating system, the overall speed-up time is not appreciable (see Table 5.3).

Table 5.3 Speed-ups times determined by 1 time step on 100x100x16 matrices. Times recorded are averages of 4 different runs.

<table>
<thead>
<tr>
<th>Number of processors</th>
<th>W-time (s)</th>
<th>Cputime from Etime (s)</th>
<th>Communication time (s)</th>
<th>Speed-up</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>75.40</td>
<td>73.44</td>
<td>______</td>
<td>______</td>
</tr>
<tr>
<td>2</td>
<td>63.24</td>
<td>36.47</td>
<td>26.77</td>
<td>1.19</td>
</tr>
<tr>
<td>4</td>
<td>44.65</td>
<td>17.45</td>
<td>27.2</td>
<td>1.69</td>
</tr>
<tr>
<td>8</td>
<td>35.69</td>
<td>8.53</td>
<td>27.16</td>
<td>2.11</td>
</tr>
<tr>
<td>16</td>
<td>30.79</td>
<td>3.63</td>
<td>27.16</td>
<td>2.44</td>
</tr>
</tbody>
</table>

Table 5.4 below show the difference in time spent between system time and user time. One can see that system time is about 80% of the total CPU time. This
abnormality is believed to be due to excessive data movement in the code within
the RAM (apparently not out of core movement to the RAID 5 box), which keeps
grabbing more memory frequently. I thank Dr. Richard Ford for his help with this
analysis. He has also suggested that possible solutions to this problem memory
swapping overload may be:

1) Optimizing the UTC-M model to be more sensitive to paging problems.

2) Or run the code on a 64 chip computer where more virtual memory is
   addressed for each job.

Table 5.4 CPU, User and System time for a one-time step run of the UTC-M
mesoscale model with 16 processors on the existing Beowulf cluster. Times
recorded are averages of 4 different runs.

<table>
<thead>
<tr>
<th>Processors</th>
<th>CPU time</th>
<th>User time</th>
<th>System Time</th>
<th>% of User</th>
<th>% of System</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>3.65</td>
<td>0.73</td>
<td>2.92</td>
<td>20</td>
<td>80</td>
</tr>
<tr>
<td>1</td>
<td>3.62</td>
<td>0.77</td>
<td>2.85</td>
<td>21</td>
<td>79</td>
</tr>
<tr>
<td>2</td>
<td>3.63</td>
<td>0.79</td>
<td>2.84</td>
<td>21</td>
<td>79</td>
</tr>
<tr>
<td>3</td>
<td>3.62</td>
<td>0.829</td>
<td>2.79</td>
<td>23</td>
<td>77</td>
</tr>
<tr>
<td>4</td>
<td>3.63</td>
<td>0.65</td>
<td>2.98</td>
<td>18</td>
<td>82</td>
</tr>
<tr>
<td>5</td>
<td>3.64</td>
<td>0.89</td>
<td>2.75</td>
<td>24</td>
<td>76</td>
</tr>
<tr>
<td>6</td>
<td>3.65</td>
<td>0.84</td>
<td>2.81</td>
<td>23</td>
<td>77</td>
</tr>
<tr>
<td>7</td>
<td>3.64</td>
<td>0.78</td>
<td>2.86</td>
<td>21</td>
<td>79</td>
</tr>
<tr>
<td>8</td>
<td>3.62</td>
<td>0.769</td>
<td>2.85</td>
<td>21</td>
<td>79</td>
</tr>
<tr>
<td>9</td>
<td>3.62</td>
<td>0.82</td>
<td>2.8</td>
<td>22</td>
<td>78</td>
</tr>
<tr>
<td>10</td>
<td>3.62</td>
<td>0.92</td>
<td>2.7</td>
<td>25</td>
<td>75</td>
</tr>
<tr>
<td>11</td>
<td>3.64</td>
<td>0.749</td>
<td>2.89</td>
<td>20</td>
<td>80</td>
</tr>
<tr>
<td>12</td>
<td>3.66</td>
<td>0.839</td>
<td>2.82</td>
<td>23</td>
<td>77</td>
</tr>
<tr>
<td>13</td>
<td>3.64</td>
<td>0.78</td>
<td>2.86</td>
<td>21</td>
<td>79</td>
</tr>
<tr>
<td>14</td>
<td>3.64</td>
<td>0.8</td>
<td>2.84</td>
<td>22</td>
<td>78</td>
</tr>
<tr>
<td>15</td>
<td>3.64</td>
<td>0.79</td>
<td>2.85</td>
<td>21</td>
<td>79</td>
</tr>
</tbody>
</table>
CHAPTER 6

SIMULATIONS AND RESULTS

6.1 Radiative Transfer Model Simulations

6.1.1 Total downwelling shortwave radiation

One of the unique features of this research is the wavelength dependency in calculations for the radiative transfer parameterization. The total downward shortwave radiation is divided into 2 main radiations, the direct and diffuse. The simulated total downwelling shortwave radiation and the solar radiation for the upper left pixel of the domain are graphed in figure 6.1. The graph clearly shows the different attenuations due to ozone, water and carbon dioxide absorption in the spectra. The attenuation window at 555nm is due to ozone attenuation. The one at 725nm is due to water and oxygen attenuation. The ones at 850, 900 and 1150 nm are due to water attenuation alone. Finally, the one at 1400nm is due to carbon dioxide.

The improvements applied to the UTC-M Mesoscale model should enable a better understanding of the importance of the different land use. Figure 6.2 and 6.3 represent the land use in the domain studied and the water bodies respectively. The
Figure 6.1 Solar, simulated direct and diffuse radiation between 300 to 3000nm for the upper left pixel of the domain (81.375W, 29.1552N) for October 23rd 2002.
Figure 6.2 Land use repartition over Central-East Florida. Land use information was downloaded from St Johns River Water Management District based on 1994-1995 color-infrared aerial photography and edited and projected in ArcView GIS.
Figure 6.3 Water bodies over Central-East Florida. Water bodies information was downloaded from St Johns water Management District and edited and projected in ArcView GIS.

land use represented in figure 6.2 was downloaded from the St Johns River Water Management District based on 1994 – 1995 color – infrared aerial photography. The data downloaded by segments had to be reprojected and combined in ArcView GIS. The GIS land use and the water bodies’ pictures will be used as background information in order to compare them with simulated thermal inertia and convergence areas. Three important urban areas can be depicted from figure 6.2. (The Fort Pierce urban area in the southern part of the domain, the Melbourne-Palm Bay area, south of Cape Canaveral and the Orlando area, in the northeast corner.) Three classes of water bodies are important in this research, the Atlantic Ocean, the
different rivers and lagoons, and the lakes. The three major lakes of importance are Lake Okeechobee, Lake Istokpoga and Lake Kissimmee (see figure 6.3).

One difference between models used by Bostater, McNally (1996) and McNally (1997) and the one used in this research is the use of a different angstrom exponent for the ocean and land, since we now use actual MODIS data, not literature values for this coefficient. The angstrom exponent affects the total downwelling shortwave radiation at the surface. Figure 6.4 and Figure 6.5 demonstrate the differences in the total downwelling shortwave radiation on the ocean and land surface. The shortwave downwelling radiation is uniform over the land and ocean in figure 6.4 where the angstrom exponent is constant over both surface. On the other hand, the radiation is less over the ocean than over the land in figure 6.5 where the angstrom exponent varies from the ocean (1.206) to land (0.568). This observation can be explained with the different aerosol concentrations applied over the two surfaces. Marine aerosols are composed mainly of sea salt particles (see table 6.1) and sulfate (Fitzgerald 1991).
Figure 6.4 Total downwelling shortwave radiation with an angstrom exponent constant over land and ocean of 0.2 (value used by Gregg and Carder, 1990 and Bostater et al. 1995-2000) for October 23\textsuperscript{rd} 2002 at 0845 EDT after sunrise.
Figure 6.5 Total downwelling shortwave radiation with an angstrom exponent of 0.568 over land and of 1.206 over the ocean (value obtained from MODIS08_D3) for October 23rd 2002 at 0845 EDT after sunrise.
Table 6.1 Global emission estimates for major aerosol types in the 1980s. (Reproduced from "Atmospheric Physics and Chemistry", Seinfeld and Pandis, 1998).

<table>
<thead>
<tr>
<th>Source</th>
<th>Estimated Flux (Tg yr(^{-1}))</th>
<th>Particle Size Category*</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Low</td>
<td>High</td>
</tr>
<tr>
<td><strong>NATURAL</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Primary</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Soil dust (mineral aerosol)</td>
<td>1000</td>
<td>3000</td>
</tr>
<tr>
<td>Sea salt</td>
<td>1000</td>
<td>10000</td>
</tr>
<tr>
<td>Volcanic dust</td>
<td>4</td>
<td>10000</td>
</tr>
<tr>
<td>Biological debris</td>
<td>26</td>
<td>80</td>
</tr>
<tr>
<td>Secondary</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Sulfates from biogenic gases</td>
<td>80</td>
<td>150</td>
</tr>
<tr>
<td>Sulfates from volcanic SO(_2)</td>
<td>5</td>
<td>60</td>
</tr>
<tr>
<td>Organic matter from biogenic VOC</td>
<td>40</td>
<td>200</td>
</tr>
<tr>
<td>Nitrates from NO(_x)</td>
<td>15</td>
<td>50</td>
</tr>
<tr>
<td>Total natural</td>
<td>2200</td>
<td>23500</td>
</tr>
<tr>
<td><strong>ANTHROPOGENIC</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Primary</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Industrial dust, etc. (except soot)</td>
<td>40</td>
<td>130</td>
</tr>
<tr>
<td>Soot</td>
<td>5</td>
<td>20</td>
</tr>
<tr>
<td>Secondary</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Sulfates from SO(_2)</td>
<td>170</td>
<td>250</td>
</tr>
<tr>
<td>Biomass burning</td>
<td>60</td>
<td>150</td>
</tr>
<tr>
<td>Nitrates from NO(_x)</td>
<td>25</td>
<td>65</td>
</tr>
<tr>
<td>Organics from anthropogenic VOC</td>
<td>5</td>
<td>25</td>
</tr>
<tr>
<td>Total anthropogenic</td>
<td>300</td>
<td>650</td>
</tr>
<tr>
<td>Total</td>
<td>2500</td>
<td>24000</td>
</tr>
</tbody>
</table>

*Coarse and fine size categories refer to mean particle diameter above and below 1 μm, respectively.

**Note:** Sulfates and nitrates are assumed to occur as ammonium salts. Flux unit: Tg yr\(^{-1}\) (dry mass).

**Source:** Kiehl and Rodhe (1995).

Ocean water and sea salt are transferred to the atmosphere through air bubbles at the sea surface. As the water evaporates, the salt is left suspended in the atmosphere. Haywood et al. (1999) suggest that naturally occurring sea salt is the
leading aerosol contributor to the global-mean clear-sky radiation balance over oceans. He added that all aerosol species reduce the total shortwave radiation at the surface (see table 6.2). The sea-salt aerosol is the one that diminishes the most the amount of solar irradiance at the surface.

Table 6.2 Optical properties (specific extinction $k_{e}$, single-scattering albedo $\omega$, and asymmetry factor $g$) of different aerosols at a wavelength of 0.55 $\mu$m, and their respective effects on the global-and-annual-mean, clear-sky net short-wave irradiances at the top-of-the-atmosphere (TOA) and surface in oceanic regions. (Reproduced from Haywood et al., 1999).

<table>
<thead>
<tr>
<th>Species</th>
<th>$k_{e}$ (m$^2$/g)</th>
<th>$\omega$</th>
<th>$g$</th>
<th>TOA (W/m$^2$)</th>
<th>Surface (W/m$^2$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Natural sulfate</td>
<td>9.74</td>
<td>1.00</td>
<td>0.75</td>
<td>-0.93</td>
<td>-0.86</td>
</tr>
<tr>
<td>Anthropogenic sulfate</td>
<td>9.74</td>
<td>1.00</td>
<td>0.75</td>
<td>-0.72</td>
<td>-0.68</td>
</tr>
<tr>
<td>Organic carbon</td>
<td>8.04</td>
<td>1.00</td>
<td>0.75</td>
<td>1.02</td>
<td>0.96</td>
</tr>
<tr>
<td>Black carbon</td>
<td>9.26</td>
<td>0.21</td>
<td>0.34</td>
<td>+0.17</td>
<td>-1.12</td>
</tr>
<tr>
<td>Natural dust</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>-0.58</td>
<td>-1.10</td>
</tr>
<tr>
<td>Anthropogenic dust</td>
<td>*</td>
<td>*</td>
<td>*</td>
<td>-0.54</td>
<td>-1.07</td>
</tr>
<tr>
<td>Low sea-salt</td>
<td>2.5</td>
<td>1.00</td>
<td>0.78</td>
<td>-1.51</td>
<td>-1.55</td>
</tr>
<tr>
<td>High sea-salt</td>
<td>2.5</td>
<td>1.00</td>
<td>0.78</td>
<td>-5.03</td>
<td>-5.17</td>
</tr>
</tbody>
</table>

*The optical parameters in the case of natural and anthropogenic dust are based on a segregation of the aerosol distributions in terms of eight different classes. The parameters are approximately those used in (15).

The total downward shortwave radiation spatial distribution depends on different factors such as the solar declination, the solar zenith angle and the time of the year and time of the day. Figure 6.5 and 6.6 show a plot of total downward shortwave radiation that illustrates these dependencies. Figure 6.5 is a plot of the total radiation at the surface at 8:45 am when the sun rise. In October, the sun rises in the South East and the plot shows the intensity decreasing from the South East towards the North West. Figure 6.6 is a plot of the total radiation at the surface at
5:00 pm when the sun sets. In October, the sun sets in the South West and the plot shows the intensity decreasing from the Southwest towards the North East. And since the sun is setting the intensity is less than in the morning.

6.1.2 Upwelling Radiation

The shortwave and longwave radiation were calculated using equation (2.6) and (2.7) respectively. The shortwave radiation is dependant on albedo or ground surface radiation. The longwave radiation is dependant on the emissivity of the surface. Figure 6.7 and figure 6.8 illustrate both respectively shortwave and longwave upwelling radiation. The albedo or ground reflectance values being very small between 0 and 0.3 in general, the upwelling shortwave is low. On the other hand, emissivity values range between 0.97 and 0.99, therefore the calculation of upwelling longwave radiation relies on the Stephan-Boltzmann constant and the initial temperature.
Figure 6.6 Total downwelling shortwave radiation with an angstrom exponent of 0.568 over land and of 1.206 over the ocean (value obtained from MODIS08_D3 October 23\textsuperscript{rd} 2002) for October 23\textsuperscript{rd} 2002 at 1700 EDT.
Figure 6.7 Total upwelling shortwave radiation simulation (2km x 2km) for October 23rd 2002 at 1400 EDT.
Figure 6.8 Total upwelling longwave radiation simulation (2km x 2km) for October 23rd 2002 at 1400 EDT.
6.1.3 Net radiation budget at the earth surface

The Net radiation at the surface is the difference between the downwelling radiation and the upwelling radiation at the surface. The difference is the net radiation, which is the driving factor for warming when the sun rises and cooling when the sun sets. Figure 6.9 and figure 6.10 show the net radiation in the morning close to sunrise and at the end of the afternoon close to sunset. The plot has positive values during the day and negative values as soon as the sun sets. The spatial variation on land surface is due to the combination of ground reflectance and emissivity values. Urban areas have a lower ground reflectance and emissivity values and therefore a lower net energy available at the surface than water areas such as the ocean or Lake Okeechobee. One can still see the pattern observed on the total downwelling shortwave radiation. The intensity in the morning is higher in the Southeast corner in October.
Figure 6.9 Net budget radiation simulation (2km x 2km) at the surface for October 23rd at 0845 EDT.
Figure 6.10 Net budget radiation simulation (2km x 2km) at the surface for October 24th at 1530 EDT.
6.2 Thermal Inertia Model Simulations

The thermal inertia simulations are of importance to see where the regions of low and high thermal inertia are. The low thermal inertia regions can be considered as urban areas because these areas absorb a lot of heat during the day and cools at night. On the other hand, regions of high thermal inertia such as water bodies or the ocean won’t be experiencing a high temperature change during a 24-hour simulation. Figure 6.11 and 6.12 represent respectively the thermal inertia for the ocean and water bodies and for land surface. One can observe that regions in blue are low thermal inertia areas and can be considered urban areas. The blue area in the Northwest corner represents the Orlando area. The area in the Southeast represents Fort Pierce and the blue area north of Fort Pierce represents the Melbourne – Palm Bay urban area. Figure 6.11 shows the land use repartition in the domain grid. The areas of low inertia correspond to the urban and built-up areas. In figure 6.12, thermal inertia for the ocean surface is constant except for a couple whit patchy areas where values are higher than 3500 TIU. For these areas, temperatures do not change much during the day.
Figure 6.11 Thermal Inertia simulation for land surface (2km x 2km) estimated for October 23rd 2002.
Figure 6.12 Thermal inertia simulation (1 km x 1 km) for the ocean for October 23rd 2002.
6.3 Time - Dependant Surface Temperature Simulations

The temperature simulations are obtained with the net budget radiation simulations at the surface and the thermal inertia simulations. The temperature model output is plotted at a six-hour time intervals during a simulation period of one day. Figure 6.13 is a temperature plot at 8:45 am, 3 hours into the simulation. Figure 6.14 is a temperature plot at 12:30 pm, 6 hours into the simulation, when the sun is at its zenith above Florida. Most land surfaces are warmed up especially the urban areas. Figure 6.15 is a temperature plot at 6:00 pm, 12 hours into the simulation. The sun has set for 30 minutes and land is cooling faster than ocean surface. Figure 6.16 is a plot of different air temperatures records (60cm, 2m, and 10m above the surface in degree C) of Fort Pierce (grid point (50, 88) on land (27° 25.565, 80° 24.118 ) on October 23rd 2002. The data obtained for Fort Pierce was downloaded from the FAWN (Florida Automated Weather Network). The data used to graph figure 6.16 can be found in Appendix K in table K1. Skin temperatures observations are not available or have not been archived for the area; therefore it is hard to compare the modeled skin temperatures with observations. However, simulated skin temperature can be compared to air temperatures. Figure 6.16 shows a comparison between the different air temperatures recorded
Figure 6.13 Land and ocean surface temperatures 3 hours into simulation. (October 23rd 2002; 0845 EDT). The urban areas such as Fort Pierce, Melbourne-Palm Bay and Orlando are heating at a faster rate than other land use surfaces.
Figure 6.14 Land and ocean surface temperatures 6 hours into simulation. (October 23rd 2002; 1230 EDT). Land and ocean surface have reached their highest temperature for the day.
Figure 6.15 Land and ocean surface temperatures 12 hours into simulation. (October 23rd 2002; 1800 EDT). After the sun sets, the land surface cools rapidly while the ocean waters remains proportionally warm.
Figure 6.16 Comparison between air temperatures and the modeled skin temperature at Fort Pierce on land (27° 25.565, 80° 24.118) for October 23rd 2002.
at Fort Pierce and the modeled skin temperatures from the thermal model at or near the surface. Skin temperatures are expected to vary more than air temperature during the day. Skin temperature is the actual temperature of the surface and its temperature is directly dependent on solar radiation. That is the reason why skin temperatures are cooler than air temperatures when there is no solar radiation. On the other hand, skin temperatures are warmer than air temperatures during the day when solar rays hit the surface. In order to do an accurate comparison between the simulated skin temperatures and observational skin temperatures, the latter would need to be recorded at different locations where land uses change. Then, analysis could demonstrate for which land types the simulated temperature match best the observational skin temperatures.

6.4 Improved UTC-M Mesoscale Model Simulations

The UTC-M Mesoscale atmospheric model was run with the newly inserted thermal model to simulate land and ocean surface temperatures. Additionally, the grid size of the domain was increased from a 25 x 25 grid to 200 x 200 grid which represents a change in resolution from 10 km to 1 km. The simulations are based on conditions observed on October 23rd 2002 (Julian Day 296). Model outputs are plotted at six-hour time intervals during a simulation period of one day (October 23rd 2002). Data plotted include u- and v- winds, vertical velocity, and convergence/divergence calculated as shown below using continuity (equation 6.1).
An altitude of 30 meters is chosen as an altitude representative of the boundary layer where low-level mass convergence is being studied. Fields plotted in this chapter include the horizontal winds in the form of vectors, the vertical component of velocity as contours. The vertical velocity is calculated through vertically integrating the continuity equation.

\[
\frac{\partial w}{\partial z} = \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}
\]  

(6.1)

Results of the simulations are graphed below. The first plot (figure 6.17) is 6 hours into simulation. The wind field over the ocean is almost inexistent and random. At 2:00 pm, the sea breeze circulation does not exist yet. However, we can see positive vertical velocities along the coast with the highest winds south of Cape Canaveral. Over land, the wind field is weak and random except along the coast where a southward wind blows. The increase in grid resolution allows a better view of the Cape. At the Cape, rising motion can be observed whereas sinking air can be observed above the Banana River, the Mosquito lagoon and the Indian River lagoon. Figure 6.18 depicts the wind field 12 hours into the simulation (8:00 pm). The end of the afternoon is usually the time when sea breeze showers happen. The wind field over the ocean has less noise than 6 hours into the simulation. The vertical wind speed is almost inexistent off shore but picks up as we approach the coast. The area of vertical motion over land seen in Figure 6.19 continues and the
Figure 6.17 Winds and vertical velocity contours (m/s) at 30 meters altitude. 6 hours into simulation. (23rd October 2002; 1400 EDT).
Figure 6.18 Winds and vertical velocity contours (m/s) at 30 meters altitude. 12 hours into simulation. (23rd October 2002; 2000 EDT).
Figure 6.19 Winds and vertical velocity contours (m/s) at 30 meters altitude. 18 hours into simulation. (24th October 2002; 0200 EDT).
Figure 6.20 Winds and vertical velocity contours (m/s) at 30 meters altitude. 24 hours into simulation. (24th October 2002; 0800 EDT).
sea breeze has penetrated to its farthest distance inland as solar radiation is
decreasing as the sunsets and therefore solar heating is rapidly weakening. The
wind vectors are oriented towards the land indicating the sea breeze influence over
land. A large rising air area can be seen from the Cape to the Orlando area. Over
the south, the effect of Lake Okeechobee generated an area of sinking motion. The
same effect can be seen over Lake Istokpoga in the South west and over Lake
Kissimmee in the east. Figure 6.19 depicts the wind field at 2:00 am, 18 hours into
the simulation. The winds are weak, the land has cooled, and the land breeze is
slowly starting to appear. Figure 6.20 shows better the land breeze circulation. It
depicts the wind field at 8:00 am, 24 hours into the simulation. Conditions are
similar to the ones at the beginning of the simulation. The surface winds are weak.
The sun has risen and starts to warm the land surface and rising and sinking air can
be seen at the land sea margin at the Cape.

In order to understand the impact of the sea breeze on the east coast of
Florida, a couple snapshots are added below. As seen above, figure 6.19 represents
the condition at 2:00 pm; the areas of rising and sinking air are not yet well defined.
Figure 6.21 represents the conditions at 6:00 pm in the afternoon. At this time of
the day, the rising and sinking air areas are well defined. They are mainly located at
the land-sea margin. The horizontal winds are at their highest velocities (<30m/s)
and there are directed in the southwest direction, pushing inland the rising air areas
Figure 6.21 Winds and vertical velocity contours (m/s) at 30 meters altitude. 10 hours into simulation. (23rd October 2002; 1800 EDT). Yellow lines represent the east coast and water-land boundaries.
Figure 6.22 Winds and vertical velocity contours (m/s) at 30 meters altitude. 14 hours into simulation. (23rd October 2002; 2200 EDT). Yellow lines represent the east coast and water-land boundaries.
located at the land sea margin. This activity can be seen throughout the evening.

Figure 6.22 shows the major rising air area up to 10 miles inland. The horizontal winds are starting to weaken (<12.5 m/s) and the areas of rising and sinking air are decreasing. These diagrams are essential to visually observe the influence of the seabreeze on the location of storm activity.

In addition to these graphs, hourly graphs of the 2 km and 1 km simulations for horizontal winds and vertical winds can be found in Appendix G (1 km horizontal winds), Appendix H (2 km horizontal winds), Appendix I (1 km convergence field) and J (2 km convergence field). General observations can be made from these graphs. First, both 1 km and 2 km horizontal wind fields simulations give the same wind pattern. Over land, winds start to be north westerlies in morning and rotate clockwise during a 24 hour simulation. Offshore, winds start their rotation from the east and rotate clockwise. With regards to the convergence field simulations, one can observe the same patterns on the 1 km and 2 km simulations. During the morning leading to the afternoon, one can observe a built-up of rising air at the coast and this built-up is pushed inland by the onshore winds due to the sea breeze.

To assess if the model predicts the sea breeze accurately, a comparison between an AVHRR picture showing cloud formation and a simulated convergence field are shown in Appendix K. In figure K1, the cloud formation along the coast is
the proof of the presence of rising air. The NOAA AVHRR satellite passed at 8:00 pm on October 23rd 2002. In figure K2, the simulated convergence field shows the effect of the sea breeze where the rising air which occurred at the coast around 4:00 pm is now inland. Figure K3 shows the overlapping of cloud formation from the AVHRR image (see figure K1) on the simulated vertical winds diagram for 7:00 pm on October 23rd 2002. The patch of clouds offshore Port Canaveral is overlapping a patch of rising air. On one hand, cloud formation overlaps also rising air regions over land and on the other hand, sinking air regions are not overlapped by clouds.
CHAPTER 7

MODEL IMPROVEMENTS AND CONCLUSIONS

An atmospheric mesoscale numerical model has been improved in different areas. First, the grid cell resolution was increased from a 10x10 km to a 1x1 km. This increase in resolution necessitated a better land and sea surface temperatures parameterization. A thermal model was incorporated to the existing model. This thermal model is comprised of a radiative transfer and a thermal inertia model which allowed a parameterization of the temperature change on both land and sea surface. Due to the incorporation of a thermal model and the increase in cell size resolution, the improved model was parallelized in order to run it on Bluemarlin, the supercomputer at Florida Tech. The insertion of a thermal model to better estimate land and sea surface temperature allowed a better prediction of rising motion and therefore forecast of cloud formation. The increase in grid resolution allowed a better understanding of the effect of the different river breezes such as the Indian River breeze but also the Mosquito lagoon and Banana river breezes.

Many improvements can still be accomplished on the UTC-M Mesoscale model such as a better parameterization of latent and sensible heat in the net budget equation for calculation of land and ocean surface temperatures. The sensible and latent heat fluxes from the surface are a major part of the atmospheric forcing
Sensible heat transfer between the ground and air is a function of the magnitude of the ground-air temperature difference, wind speed and atmospheric stability (Pratt, 1979). Sensible heat equation is given by Pratt (1979) as:

\[
H = d \left( \frac{T_a - T_g}{T_{aH}} \right) - \frac{U k_v^2 \rho c}{\left( \frac{\ln z}{z_0} - \Delta \psi_H \right) \left( \frac{\ln z}{z_0} - \Delta \psi_M \right)},
\]

where:
\[
d = - \frac{\overline{U} k_v^2 \rho c}{\left( \frac{\ln z}{z_0} \right)^2},
\]
\[
\overline{U} = \text{average wind speed (m/s)},
\]
\[
k_v = \text{Von Karman’s constant (0.4) (dimensionless)},
\]
\[
\rho = \text{density of dry air (kg/m}^3),
\]
\[
c = \text{specific heat of air at constant pressure},
\]
\[
z_o = \text{surface roughness},
\]
\[
z = \text{height of wind and air temperature sensors}.
\]

The latent heat flux can also be considered in the same manner as the sensible heat flux and the latent heat equation is given by Pratt (1979) as:

\[
E = e \left( \overline{h_a} - h_g \right)
\]

where:
\[
e = - \frac{\overline{U} k_v^2 \rho L}{\left( \frac{\ln z}{z_0} \right)^2},
\]
\[
L = \text{latent heat of vaporization (J/kg)},
\]
\[
\overline{h_a} = \text{average humidity of the air (g/m}^3),
\]
\[
h_g = \text{average humidity at the ground surface (g/m}^3).\]
A good parameterization of cloud physics should be added especially as the resolution of the grid domain increases. Indeed, an accurate representation of clouds would have a tremendous impact on both the temperature and the mesoscale circulation at the surface and the boundary layer.

Finally, a parallelization of the UTC-M Mesoscale model where more than 16 processors would be used can be implemented. The large system time could be reduced by optimizing the algorithm to be more sensitive to memory swapping.

Uhlhorn’s conclusions also suggested the importance of an accurate representation of the dynamic state of the sea-surface. And a coupled ocean-atmosphere model would allow a three-dimension structure of the temperature and salinity fields and the sea surface elevation. The improved UTC-M Mesoscale model could be nested inside a larger lower resolution model grid to capture influences of synoptic scale energy. Finally, a better treatment of lateral boundary conditions and a higher order turbulence closure method is required to get a better turbulence parameterization.
References


Layer Atmospheric Model – Sea Breeze Predictions along Central Florida Using a Thermal Sub-Model.” SPIE vol.


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Appendix A

Features and limitations of Various Mesoscale Models
Table A-1. Features and limitations of various mesoscale models.

<table>
<thead>
<tr>
<th>MESOSCALE MODELS</th>
<th>ADVANTAGES</th>
<th>LIMITATIONS</th>
</tr>
</thead>
</table>
| MM5 (The Fifth-Generation NCAR / Penn State Mesoscale Model) Anthes and Warner (1978) | • Used for theoretical and real-time studies, including applications of both predictive simulation and four-dimensional data assimilation to monsoons, hurricanes, and cyclones.  
• Used for studies involving mesoscale convective systems, fronts, land-sea breezes, mountain-valley circulations, and urban heat islands. | • Does not account for shadowing effects due to terrain.  
• Underestimates the spatial and temporal variability observed in surface and upper air data at ranges.  
• The vertical resolution in MM5 does not allow proper treatment of thin cloud layers. |
| ETA Model (NCEP) Mesinger (1984) | • Removes the large errors which are known to occur when computing the horizontal pressure gradient force, as well as the advection and horizontal diffusion, along a steeply sloped coordinate surface, such as the sigma surfaces in the NGM model.  
• Because the eta coordinate is pressure based and normalized (i.e. quasi-horizontal), it leads to a much simpler solution of the equations of motion in areas such as the pressure gradient force, horizontal advection, and diffusion.  
• Has a resolution of 29 kilometers. In other words, the grid box is much smaller than in the NGM.  
• ETA coordinate system (mathematical coordinate system that takes into account topographical features such as mountains).  
• Has a much more accurate picture of the terrain across the USA. | • The step nature of the eta coordinate makes it difficult to retain detailed vertical structure in the boundary layer over the entire model domain, particularly over elevated terrain.  
• Eta models do not accurately depict gradually sloping terrain. Since all terrain is represented in discrete steps, gradual slopes that extend over large distances can be concentrated within as few as one step. This unrealistic compression of the slope into a small area can be compensated, in part, by increasing the vertical and/or horizontal resolution.  
• Eta models have difficulty predicting extreme down slope wind events. |
### Table A-1. Continue

<table>
<thead>
<tr>
<th><strong>MESOSCALE MODELS</strong></th>
<th><strong>ADVANTAGES</strong></th>
<th><strong>LIMITATIONS</strong></th>
</tr>
</thead>
</table>
| Nested Grid Model (NGM)       | • The terrain following system simplifies the treatment of processes at the bottom of the model atmosphere.  
   • The same vertical structure of 16 layers is carried throughout the analysis, initialization, and forecast components of the NGM to eliminate inconsistencies that may arise through vertical interpolation. | • The NGM does not include cloud physics.  
   • Nested Grid Model (NGM), ignored vegetation type in its simple land surface parameterization. |
| NORAPS6 (Navy Operational Regional Atmospheric Prediction System Version 6) | • Large-scale and convective precipitation modules were included in the model. Evaporation of precipitation is allowed to occur.  
   • Cloud temperatures and mixing ratios follow a moist adiabat, thereby representing a nonentraining cloud (Hodur, 1987). | • NORAPS has no land-use model.  
   • It does not have options for a nonhydrostatic equation set. A nonhydrostatic model is critical in the simulation of a variety of buoyancy-driven atmospheric features. |
| Regional Atmospheric Modeling System (RAMS) | • Large-scale and convective precipitation modules were included in the model, and large-scale processes are treated explicitly.  
   • A range of option exists when specifying how the microphysics is modeled, and mixed-phase microphysics is allowed. | • May not correctly portray weather events in lee of mountains. |
Table A-1. Continue

<table>
<thead>
<tr>
<th>MESOSCALE MODELS</th>
<th>ADVANTAGES</th>
<th>LIMITATIONS</th>
</tr>
</thead>
<tbody>
<tr>
<td>Relocatable Window Model (RWM)</td>
<td>• It employs a single, unstaggered horizontal grid</td>
<td>• It does not account for land surface processes, diagnosed cloud fraction,</td>
</tr>
<tr>
<td></td>
<td>and uses a terrain-following vertical coordinate system.</td>
<td>and solar/terrestrial radiative processes.</td>
</tr>
<tr>
<td></td>
<td>• It uses a basic boundary layer physics package in which bulk transfers of</td>
<td>• It does not have the ability to add a nested grid.</td>
</tr>
<tr>
<td></td>
<td>sensible and latent heat are allowed over the ocean when the sea surface is</td>
<td></td>
</tr>
<tr>
<td></td>
<td>greater than the air temperature, and surface friction is modeled with a</td>
<td></td>
</tr>
<tr>
<td></td>
<td>terrain-dependent drag coefficient (Kopp et al. 1994).</td>
<td></td>
</tr>
</tbody>
</table>
Appendix B

2 km and 1 km grid domain
Figure B1 UTC-M model grid showing the 2 x 2 km grid cells used in this research domain from approximately Daytona Beach, Fl. to West Palm Beach, Florida. The dashed line shows the long. axis location of the longitudinal axis of the Gulf Stream current for October 23rd 2002.
Figure B2 UTC-M model grid showing the 1 x 1 km grid cells used in this research domain from approximately Daytona Beach, Fl. to West Palm Beach, Florida. The dashed line shows the longitudinal axis location of the Gulf Stream current for October 23rd 2002.
Appendix C

MODIS Ozone and Angstrom exponent
MOD08_D3 Mean Total Ozone, Mean Angstrom Exponent for land and for ocean
Figure C1 Mean Total Ozone in Dobson Units. It shows daily 1 x 1 degree grid average values of total ozone burden for October 23rd 2002.
Figure C2: Zoom of Florida for extraction of ozone burden for the domain. The red square represents the research domain. The ozone burden for the domain research is the average of four 1 degree pixels.
Figure C3 Mean Angstrom Exponent calculated for land. It shows daily 1 x 1 degree grid average values of the Angstrom exponent for land for October 23rd 2002.
Figure C4 Zoom of Florida for extraction of angstrom exponent for land. The red square represents the research domain. The angstrom exponent for land for the domain research is the average of two 1 degree pixels.
Figure C5 Mean Angstrom Exponent calculated for ocean. It shows daily 1 x 1 degree grid average values of Angstrom exponent for ocean for October 23rd 2002.
Figure C6 Zoom of Florida for extraction of angstrom exponent for ocean. The red square represents the research domain. The angstrom exponent for ocean for the domain research is the average of two 1 degree pixels.
Appendix D

AVHRR data input files:
AVHRR data files for the three passes on
October 22nd 2002.
Figure D1 AVHRR image from data NOAA AVHRR 15 around 1:00 pm on October 22nd 2002. In this image, band 1, 2 and 3 are represented.
Figure D2 AVHRR image from data NOAA AVHRR 17 around 3:20 pm on October 22\textsuperscript{nd} 2002. In this image, band 1, 2 and 3 are represented.
Figure D3 AVHRR image from data NOAA AVHRR 16 around 7:25 pm on October 22nd 2002. In this image, band 1, 2 and 3 are represented.
Figure D4 AVHRR NOAA 15 Band 4 brightness temperature (K) for the 1:00 pm October 22\textsuperscript{nd} 2002 pass (1km x 1km grid). This image is the final result after elimination and interpolation of the cloudy pixels. Input file name: 20010221257ch4.txt. (See Appendix K for all listings of input files).
Figure D5 AVHRR NOAA 15 Band 5 brightness temperature (K) for the 1:00 pm October 22\textsuperscript{nd} 2002 pass (1km x 1km grid). This image is the final result after elimination and interpolation of the cloudy pixels. Input file name: 20010221257ch5.txt. (See Appendix K for all listings of input files).
Figure D6 AVHRR NOAA 17 Band 4 brightness temperature (K) for the 3:20 pm October 22\textsuperscript{nd} 2002 pass (1km x 1km grid). This image is the final result after elimination and interpolation of the cloudy pixels. Input file name: 20010221518ch4.txt. (See Appendix K for all listings of input files).
Figure D7 AVHRR NOAA 17 Band 5 brightness temperature (K) for the 3:20 pm October 22\textsuperscript{nd} 2002 pass (1km x 1km grid). This image is the final result after elimination and interpolation of the cloudy pixels. Input file name: 20010221518ch5.txt. (See Appendix K for all listings of input files).
Figure D8 AVHRR NOAA 16 Band 4 brightness temperature (K) for the 7:20 pm October 22\textsuperscript{nd} 2002 pass (1km x 1km grid). This image is the final result after elimination and interpolation of the cloudy pixels. Input file name: 20010221926ch4.txt. (See Appendix K for all listings of input files).
Figure D9 AVHRR NOAA 16 Band 5 brightness temperature (K) for the 7:20 pm October 22\textsuperscript{nd} 2002 pass (1km x 1km grid). This image is the final result after elimination and interpolation of the cloudy pixels. Input file name: 20010221926ch5.txt. (See Appendix K for all listings of input files).
Appendix E

MODIS Input files:
MOD09GHK Albedo, MOD11A1 Emissivity
Figure E1 MOD09GHK surface reflectance band 1 data for October 23\textsuperscript{rd} 2002 (1km x 1km grid). Input file name: 200galbedobd1.txt. (See Appendix K for all listings of input files).
Figure E2 MOD09GHK surface reflectance band 2 data for October 23rd 2002 (1km x 1km grid). Input file name: 200galbedobd2.txt. (See Appendix K for all listings of input files).
Figure E3 MOD09GHK surface reflectance band 3 data for October 23rd 2002 (1km x 1km grid). Input file name: 200galbedobd3.txt. (See Appendix K for all listings of input files).
Figure E4 MOD09GHK surface reflectance band 4 data for October 23\textsuperscript{rd} 2002 (1km x 1km grid). Input file name: 200galbedobd4.txt. (See Appendix K for all listings of input files).
Figure E5 MOD09GHK surface reflectance band 5 data for October 23rd 2002 (1km x 1km grid). Input file name: 200galbedobd5.txt. (See Appendix K for all listings of input files).
Figure E6 MOD09GHK surface reflectance band 6 data for October 23rd 2002 (1km x 1km grid). Input file name: 200galbedobd6.txt. (See Appendix K for all listings of input files).
Figure E7 MOD09GHK surface reflectance band 7 data for October 23\textsuperscript{rd} 2002 (1km x 1km grid). Input file name: 200galbedobd7.txt. (See Appendix K for all listings of input files).
Figure E8 MOD11A1 emissivity band 11 data for October 23rd 2002 (1km x 1km grid). Input file name: 200emiss31.txt. (See Appendix K for all listings of input files).
Appendix F

Radiative Transfer input files:
Albedo, Solar Constant, Transmittance Coefficients
(water vapor, ozone and uniformly mixed gas)
Table F.1 Albedo values interpolated from 7 MODIS albedo bands to allow the wavelength dependency for the ground surface reflectance parameter. The MODIS 09 data set was collected on October 23rd 2002. This graph was created from the dataset of the upper left corner of the domain (-81.375W, 29.1622N).

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Appendix G

Boundary layer 1km grid simulation
Hourly u and v winds (m/s) simulations at 30 meters altitude for October 23\textsuperscript{rd} 2002.
Figure G1 Horizontal (u and v) winds (m/s) at 30 meters altitude. 1 hour into simulation (October 23rd 2002, 0900 EDT). Winds are weak with no general circulation over the land and very weak onshore winds over the ocean. At the coast, winds are blowing from the North.
Figure G2 Horizontal \( (u \text{ and } v) \) winds (m/s) at 30 meters altitude. 2 hours into simulation (October 23\textsuperscript{rd} 2002, 1000 EDT). Winds are weak with no general circulation over the land and onshore winds ceased over the ocean. At the coast, winds are blowing from the North.
Figure G3 Horizontal (u and v) winds (m/s) at 30 meters altitude. 3 hours into simulation (October 23rd 2002, 1100 EDT). Winds are picking up with no general circulation over the land and no activity over the ocean. At the coast, winds are blowing from the North.
Figure G4 Horizontal (u and v) winds (m/s) at 30 meters altitude. 4 hours into simulation (October 23\textsuperscript{rd} 2002, 1200 EDT). Winds are picking up with no general circulation over the land and no activity over the ocean. At the coast, winds are blowing from the North.
Figure G5 Horizontal (u and v) winds (m/s) at 30 meters altitude. 5 hours into simulation (October 23rd, 2002, 1300 EDT). Winds are picking up with no general circulation over the land and no activity over the ocean. At the coast, winds are blowing from the North.
Figure G6 Horizontal (u and v) winds (m/s) at 30 meters altitude. 6 hours into simulation (October 23rd 2002, 1400 EDT). Winds are at their highest speed of the day with no general circulation over neither land nor ocean. Northerly winds are still blowing at the coast.
Figure G7 Horizontal (u and v) winds (m/s) at 30 meters altitude. 7 hours into simulation (October 23rd 2002, 1500 EDT). Winds are slowly decreasing with no general circulation over neither land nor ocean. Northerly winds are still blowing at the coast.
Figure G8 Horizontal (u and v) winds (m/s) at 30 meters altitude. 8 hours into simulation (October 23rd 2002, 1600 EDT). Winds are slowly decreasing with no general circulation over neither land nor ocean. Northerly winds are still blowing at the coast.
Figure G9 Horizontal ($u$ and $v$) winds (m/s) at 30 meters altitude. 9 hours into simulation (October 23rd 2002, 1700 EDT). Winds are slowly decreasing with no general circulation over neither land nor ocean. Northerly winds are still blowing at the coast.
Figure G10 Horizontal (u and v) winds (m/s) at 30 meters altitude. 10 hours into simulation (October 23rd 2002, 1800 EDT). Winds are slowly decreasing with no general circulation over neither land nor ocean. Winds at the coast are slowly changing from northerly winds to northeasterly winds. The sea breeze circulation is slowly starting.
Figure G11 Horizontal (u and v) winds (m/s) at 30 meters altitude. 11 hours into simulation (October 23rd 2002, 1900 EDT). Winds are slowly decreasing with no general circulation over neither land nor ocean. Winds at the coast are slowly changing from northerly winds to northeasterly winds. The sea breeze circulation is slowly starting.
Figure G12 Horizontal (u and v) winds (m/s) at 30 meters altitude. 12 hours into simulation (October 23rd 2002, 2000 EDT). Winds are slowly decreasing with no general circulation over neither land nor ocean. Winds at the coast are slowly changing from northerly winds to northeasterly winds. The sea breeze circulation is slowly starting.
Figure G13: Horizontal (u and v) winds (m/s) at 30 meters altitude. 13 hours into simulation (October 23rd, 2002, 2100 EDT). Winds on land are mainly from the east. Winds at the coast are from the northeast. The sea breeze circulation is slowly starting.
Figure G14 Horizontal (u and v) winds (m/s) at 30 meters altitude. 14 hours into simulation (October 23rd 2002, 2200 EDT). Winds on land are mainly from the east. Winds at the coast are changing from north easterlies to easterlies. The sea breeze circulation is in full effect. Winds offshore are becoming westerly.
Figure G15 Horizontal (u and v) winds (m/s) at 30 meters altitude. 15 hours into simulation (October 23rd 2002, 2300 EDT). Winds on land are mainly from the east. In the south, winds are slowly changing from south easterlies to southerlies. Winds at the coast are easterlies. The sea breeze circulation is in full effect. Winds offshore are becoming westerly.
Figure G16 Horizontal (u and v) winds (m/s) at 30 meters altitude. 16 hours into simulation (October 23rd 2002, 2400 EDT). Winds on land are mainly from the east while offshore winds are from the west. In the south, winds are slowly changing from south easterlies to southerlies. Winds at the coast are easterlies. The sea breeze circulation is in full effect.
Figure G17 Horizontal (u and v) winds (m/s) at 30 meters altitude. 17 hours into simulation (October 23rd 2002, 0100 EDT). Winds on land are changing from easterlies to south easterlies while offshore winds are still from the west. In the south, winds are from the south. Winds at the coast are easterlies. Winds in the northwest corner are slowly becoming southerlies.
Figure G18 Horizontal (u and v) winds (m/s) at 30 meters altitude. 18 hours into simulation (October 24th, 2002, 0200 EDT). Winds on land are mainly from the south east while offshore winds are from the west. In the south, winds are slowly changing from south easterlies to southerlies. Winds at the coast are south easterlies and weakening. The sea breeze circulation is decreasing in intensity. Winds in the northwest corner are slowly becoming southerlies.
Figure G19 Horizontal (u and v) winds (m/s) at 30 meters altitude. 19 hours into simulation (October 24th 2002, 0300 EDT). Winds on land are mainly from the south east changing to southerlies while offshore winds are from the west. Winds at the coast are south easterlies and weakening. The sea breeze circulation is decreasing in intensity. Winds in the northwest corner are slowly becoming southerlies.
Figure G20 Horizontal (u and v) winds (m/s) at 30 meters altitude. 20 hours into simulation (October 24th, 2002, 0400 EDT). Winds on land are mainly from the south while offshore winds are from the southwest and northwest. Winds at the coast are easterlies and weakening. The sea breeze circulation is almost over.
Figure G21 Horizontal (u and v) winds (m/s) at 30 meters altitude. 21 hours into simulation (October 24th, 2002, 0500 EDT). Winds on land are mainly from the south while offshore winds are from the southwest and northwest. Winds at the coast are south easterlies and weakening. The sea breeze circulation is over and a land breeze will start shortly after.
Figure G22 Horizontal (u and v) winds (m/s) at 30 meters altitude. 22 hours into simulation (October 24th 2002, 0600 EDT). Winds on land are mainly from the south slowly becoming south westerlies while offshore winds are from the southwest and northwest. Winds at the coast are south easterlies and weakening. The sea breeze circulation is over and a land breeze will start shortly after.
Figure G23 Horizontal (u and v) winds (m/s) at 30 meters altitude. 23 hours into simulation (October 24th 2002, 0700 EDT). Winds on land are mainly from the south west while offshore winds are from the southwest and northeast rotating and becoming north - northwest. Winds at the coast are from the south and weakening. The land breeze circulation is increasing in intensity.
Figure G24 Horizontal (u and v) winds (m/s) at 30 meters altitude. 24 hours into simulation (October 24th 2002, 0800 EDT). Winds on land are mainly from the southwest while offshore winds are from the southwest and northeast rotating and becoming north - northwest. Winds at the coast are from the south and changing to south westerlies and increasing. The land breeze circulation is increasing in intensity.
Appendix H

Boundary layer 2km grid simulation
u and v winds (m/s) at 30 meters altitude. 23rd October 2002 simulation
Figure H1 Horizontal (u and v) winds (m/s) at 30 meters altitude. 1 hour into simulation (October 23rd 2002, 0900 EDT). Winds are weak with no general circulation over the land and very weak onshore winds over the ocean. At the coast, winds are blowing from the North.
Figure H2 Horizontal (u and v) winds (m/s) at 30 meters altitude. 2 hours into simulation (October 23\textsuperscript{rd} 2002, 1000 EDT). Winds are weak with no general circulation over the land and onshore winds ceased over the ocean. At the coast, winds are blowing from the North.
Figure H3 Horizontal (u and v) winds (m/s) at 30 meters altitude. 3 hours into simulation (October 23rd 2002, 1100 EDT). Winds are picking up with no general circulation over the land and no activity over the ocean. At the coast, winds are blowing from the North.
Figure H4 Horizontal (u and v) winds (m/s) at 30 meters altitude. 4 hours into simulation (October 23rd 2002, 1200 EDT). Winds are picking up with no general circulation over the land and no activity over the ocean. In the south west corner, winds are blowing to the west. At the coast, winds are blowing from the North.
Figure H5 Horizontal (u and v) winds (m/s) at 30 meters altitude. 5 hours into simulation (October 23rd 2002, 1300 EDT). Winds are picking up with no general circulation over the land and no activity over the ocean. In the south west corner, winds are blowing to the west. At the coast, winds are blowing from the North.
Figure H6 Horizontal \((u\text{ and } v)\) winds (m/s) at 30 meters altitude. 6 hours into simulation (October 23rd 2002, 1400 EDT). Winds are at their highest speed of the day with no general circulation over neither land nor ocean. In the south west corner, winds are blowing to the west. Northerly winds are still blowing at the coast.
Figure H7 Horizontal (u and v) winds (m/s) at 30 meters altitude. 7 hours into simulation (October 23rd 2002, 1500 EDT). Winds are slowly decreasing with no general circulation over neither land nor ocean. In the south west corner, winds are blowing to the west. Northerly winds are still blowing at the coast.
Figure H8 Horizontal (u and v) winds (m/s) at 30 meters altitude. 8 hours into simulation (October 23rd 2002, 1600 EDT). In the south west corner, winds are blowing to the west. Northerly winds are still blowing at the coast. A sea breeze pattern is emerging with winds at the coast slowly changing to onshore winds.
Figure H9 Horizontal (u and v) winds (m/s) at 30 meters altitude. 9 hours into simulation (October 23rd 2002, 1700 EDT). Northerly winds are still blowing at the coast. A sea breeze pattern is emerging with winds at the coast slowly changing to onshore winds.
Figure H10 Horizontal (u and v) winds (m/s) at 30 meters altitude. 10 hours into simulation (October 23rd 2002, 1800 EDT). Winds at the coast are slowly changing from northerly winds to northeasterly winds. The sea breeze circulation is in full effect.
Figure H11 Horizontal (u and v) winds (m/s) at 30 meters altitude. 11 hours into simulation (October 23rd 2002, 1900 EDT). Winds at the coast are northeasterly winds. The sea breeze circulation is in full effect.
Figure H12 Horizontal (u and v) winds (m/s) at 30 meters altitude. 12 hours into simulation (October 23rd, 2002, 2000 EDT). Winds at the coast are northeasterly becoming easterly winds. The sea breeze circulation is in full effect.
Figure H13 Horizontal (u and v) winds (m/s) at 30 meters altitude. 13 hours into simulation (October 23rd, 2002, 2100 EDT). Winds on land and at the coast are mainly from the east. The sea breeze circulation is in full effect.
Figure H14 Horizontal (u and v) winds (m/s) at 30 meters altitude. 14 hours into simulation (October 23rd 2002, 2200 EDT). Winds on land and at the coast are mainly from the east. The sea breeze circulation is in full effect.
Figure H15 Horizontal (u and v) winds (m/s) at 30 meters altitude. 15 hours into simulation (October 23rd 2002, 2300 EDT). Winds on land are mainly from the east. In the south, winds are slowly changing from south easterlies to southerlies. Winds at the coast are easterlies. The sea breeze circulation is in full effect. Winds offshore are becoming westerly.
Figure H16 Horizontal (u and v) winds (m/s) at 30 meters altitude. 16 hours into simulation (October 23rd 2002, 2400 EDT). Winds on land are mainly from the east while offshore winds are from the west. In the south, winds are slowly changing from south easterlies to southerlies. Winds at the coast are easterlies. The sea breeze circulation is in full effect.
Figure H17 Horizontal (u and v) winds (m/s) at 30 meters altitude. 17 hours into simulation (October 23rd 2002, 0100 EDT). Winds on land are changing from easterlies to south easterlies. In the south, winds are from the south. Winds at the coast are south easterlies. The sea breeze is weakening.
Figure H18 Horizontal (u and v) winds (m/s) at 30 meters altitude. 18 hours into simulation (October 24th, 2002, 0200 EDT). Winds on land are mainly from the south east while. Winds at the coast are south easterlies and weakening. The sea breeze circulation is over. Winds in the northwest corner are south easterlies.
Figure H19 Horizontal (u and v) winds (m/s) at 30 meters altitude. 19 hours into simulation (October 24th 2002, 0300 EDT). Winds on land are mainly from the south east changing to southerlies. Winds at the coast are south easterlies and weakening. The sea breeze circulation is over. Winds in the northwest corner are southerlies.
Figure H20 Horizontal (u and v) winds (m/s) at 30 meters altitude. 20 hours into simulation (October 24th, 2002, 0400 EDT). Winds on land are mainly from the south. Winds at the coast are from the south and weakening.
Figure H21 Horizontal (u and v) winds (m/s) at 30 meters altitude. 21 hours into simulation (October 24th 2002, 0500 EDT). Winds on land are mainly from the south. Winds at the coast are southerlies and weakening. The land breeze will start shortly after.
Figure H22 Horizontal (u and v) winds (m/s) at 30 meters altitude. 22 hours into simulation (October 24th 2002, 0600 EDT). Winds on land are mainly from the south slowly becoming south westerlies. Winds at the coast are south westerlies and weakening. The land breeze will start shortly after.
Figure H23 Horizontal (u and v) winds (m/s) at 30 meters altitude. 23 hours into simulation (October 24th, 2002, 0700 EDT). Winds on land are from the south west. Winds at the coast are from the south and weakening. The land breeze circulation is increasing in intensity.
Figure H24 Horizontal (u and v) winds (m/s) at 30 meters altitude. 24 hours into simulation (October 24th 2002, 0800 EDT). Winds on land are mainly from the south west. Winds at the coast are from the south and changing to south westerlies and increasing. The land breeze circulation is increasing in intensity.
Appendix I

**Boundary layer 1 km grid simulation**
Convergence field (m/s) at 30 meters altitude. 23rd October 2002 simulation
Figure II Vertical velocity contours (m/s) at 30 meters altitude. 1 hour into simulation (October 23rd 2002, 0900 EDT). The heating of the land surface has just started. The vertical winds are weak and no general pattern is observed. The noise contour field may be attributed to the weak winds.
Figure I2 Vertical velocity contours (m/s) at 30 meters altitude. 2 hours into simulation (October 23rd 2002, 1000 EDT). The heating of the land surface has just started. The vertical winds are increasing and no general pattern is observed. The noise contour field may be attributed to the weak winds.
Figure I3 Vertical velocity contours (m/s) at 30 meters altitude. 3 hours into simulation (October 23rd 2002, 1100 EDT). The vertical winds are increasing and no general pattern is observed. The noise contour field may be attributed to the weak winds.
Figure I4 Vertical velocity contours (m/s) at 30 meters altitude. 4 hours into simulation (October 23rd 2002, 1200 EDT). The vertical winds are increasing and no general pattern is observed. The noise contour field may be attributed to the weak winds.
Figure I5 Vertical velocity contours (m/s) at 30 meters altitude. 5 hours into simulation (October 23\textsuperscript{rd} 2002, 1300 EDT). The vertical winds have reached their maximum velocity. The noise contour field may be attributed to the weak winds. An area of positive vertical motion has begun to appear along the Banana River and Indian River Lagoon.
Figure I6 Vertical velocity contours (m/s) at 30 meters altitude. 6 hours into simulation (October 23rd 2002, 1400 EDT). An area of positive vertical motion has begun to appear along the Banana River and Indian River Lagoon. An area of negative vertical motion has begun to appear along the coast.
Figure 17 Vertical velocity contours (m/s) at 30 meters altitude. 7 hours into simulation (October 23rd, 2002, 1500 EDT). The winds are slowly weakening. An area of positive vertical motion has begun to appear along the Banana River and Indian River Lagoon. Over land areas of rising and sinking air motion are starting to be more defined. An area of negative vertical motion has begun to appear along the coast.
Figure I8 Vertical velocity contours (m/s) at 30 meters altitude. 8 hours into simulation (October 23rd 2002, 1600 EDT). The winds are slowly weakening. An area of positive vertical motion has begun to appear along the Banana River and Indian River Lagoon. Over land areas of rising and sinking air motion are starting to be more defined. An area of negative vertical motion has begun to appear along the coast.
Figure I9 Vertical velocity contours (m/s) at 30 meters altitude. 9 hours into simulation (October 23rd 2002, 1700 EDT). The winds are slowly weakening. The area of positive vertical motion is well defined along the Banana River and Indian River Lagoon. Over land areas of rising and sinking air motion are starting to be more defined. An area of negative vertical motion has begun to appear along the coast. A well defined area of sinking air motion can be observed above the St Johns River Lagoon and Lake Okeechobee.
Figure I10 Vertical velocity contours (m/s) at 30 meters altitude. 10 hours into simulation (October 23\textsuperscript{rd} 2002, 1800 EDT). The winds are slowly weakening. The area of positive vertical motion is well defined along the Banana River and Indian River Lagoon Over land areas of rising and sinking air motion are starting to be more defined. An area of negative vertical motion has begun to appear along the coast. A well defined area of sinking air motion can be observed above the St Johns River Lagoon and Lake Okeechobee.
Figure I11 Vertical velocity contours (m/s) at 30 meters altitude. 11 hours into simulation (October 23rd 2002, 1900 EDT). The winds are slowly weakening. The area of positive vertical motion is well defined along the Banana River and Indian River Lagoon Over land areas of rising and sinking air motion are starting to be more defined. An area of negative vertical motion has begun to appear along the coast. A well defined area of sinking air motion can be observed above the St Johns River Lagoon and Lake Okeechobee.
Figure 112 Vertical velocity contours (m/s) at 30 meters altitude. 12 hours into simulation (October 23rd 2002, 2000 EDT). The winds are slowly weakening. The area of positive vertical motion is slowly starting to shift to the left due to the onshore winds of the sea breeze.
Figure I13 Vertical velocity contours (m/s) at 30 meters altitude. 13 hours into simulation (October 23rd 2002, 2100 EDT). The winds are slowly weakening. The area of positive vertical motion is slowly starting to shift to the left due to the onshore winds of the sea breeze.
Figure I14 Vertical velocity contours (m/s) at 30 meters altitude. 14 hours into simulation (October 23rd, 2002, 2200 EDT). The winds are slowly weakening. The area of positive vertical motion is slowly starting to shift to the left due to the onshore winds of the sea breeze.
Figure 115 Vertical velocity contours (m/s) at 30 meters altitude. 15 hours into simulation (October 23rd 2002, 2300 EDT). The winds are slowly weakening. The area of positive vertical motion is slowly starting to shift to the left due to the onshore winds of the sea breeze.
Figure 116 Vertical velocity contours (m/s) at 30 meters altitude. 16 hours into simulation (October 23rd 2002, 2400 EDT). The winds are slowly weakening. The area of positive and negative vertical motion is getting less defined due to the weakening of the vertical winds.
Figure I17 Vertical velocity contours (m/s) at 30 meters altitude. 17 hours into simulation (October 24th 2002, 0100 EDT). The winds are slowly weakening. The area of positive and negative vertical motion is getting less defined due to the weakening of the vertical winds.
Figure I18 Vertical velocity contours (m/s) at 30 meters altitude. 18 hours into simulation (October 24th 2002, 0200 EDT). The winds are slowly weakening. The area of positive and negative vertical motion is getting less defined due to the weakening of the vertical winds.
Figure I19 Vertical velocity contours (m/s) at 30 meters altitude. 19 hours into simulation (October 24th 2002, 0300 EDT). The winds are slowly weakening. The area of positive and negative vertical motion is getting less defined due to the weakening of the vertical winds.
Figure I20 Vertical velocity contours (m/s) at 30 meters altitude. 20 hours into simulation (October 24th 2002, 0400 EDT). The winds are very weak. The areas of positive and negative vertical motion are not defined anymore due to the weakening of the vertical winds.
Figure I21 Vertical velocity contours (m/s) at 30 meters altitude. 21 hours into simulation (October 24th 2002, 0500 EDT). The winds are very weak. The areas of positive and negative vertical motion are not defined anymore due to the weakening of the vertical winds.
Figure I22 Vertical velocity contours (m/s) at 30 meters altitude. 22 hours into simulation (October 24th 2002, 0600 EDT). The winds are very weak but increasing again. An area of negative vertical motion has begun to appear at the coast.
Figure I23 Vertical velocity contours (m/s) at 30 meters altitude. 23 hours into simulation (October 24th 2002, 0700 EDT). The winds are very weak but increasing again. An area of negative vertical motion has begun to appear at the coast.
Figure I24 Vertical velocity contours (m/s) at 30 meters altitude. 24 hours into simulation (October 24th 2002, 0700 EDT). The winds are very weak but increasing again. An area of negative vertical motion has begun to appear at the coast.
Appendix J

Boundary layer 2 km grid simulation
Convergence field (m/s) at 30 meters altitude. 23rd October 2002 simulation
Figure J1 Vertical velocity contours (m/s) at 30 meters altitude. 1 hour into simulation (October 23rd 2002, 0900 EDT). The heating of the land surface has just started. The vertical winds are weak and no general pattern is observed. The noise contour field may be attributed to the weak winds.
Figure J2 Vertical velocity contours (m/s) at 30 meters altitude. 2 hours into simulation (October 23rd 2002, 1000 EDT). The heating of the land surface has just started. The vertical winds are increasing and no general pattern is observed. The noise contour field is slowly dissipating.
Figure J3 Vertical velocity contours (m/s) at 30 meters altitude. 3 hours into simulation (October 23rd 2002, 1100 EDT). The vertical winds are increasing and areas of rising air are observed at the coast.
Figure J4 Vertical velocity contours (m/s) at 30 meters altitude. 4 hours into simulation (October 23rd 2002, 1200 EDT). The vertical winds are increasing and areas of rising air are observed at the coast.
Figure J5 Vertical velocity contours (m/s) at 30 meters altitude. 5 hours into simulation (October 23\textsuperscript{rd} 2002, 1300 EDT). The vertical winds are increasing and areas of rising air are observed at the coast.
Figure J6 Vertical velocity contours (m/s) at 30 meters altitude. 6 hours into simulation (October 23rd 2002, 1400 EDT). The vertical winds are starting to decrease and areas of rising air are observed at the coast as well as defined areas of sinking motion over the ocean near the coast.
Figure J7 Vertical velocity contours (m/s) at 30 meters altitude. 7 hours into simulation (October 23\textsuperscript{rd} 2002, 1500 EDT). The winds are slowly weakening. The vertical winds are starting to decrease and areas of rising air are observed at the coast as well as defined areas of sinking motion over the ocean near the coast.
Figure J8 Vertical velocity contours (m/s) at 30 meters altitude. 8 hours into simulation (October 23rd 2002, 1600 EDT). The winds are slowly weakening. The vertical winds are starting to decrease and areas of rising air are observed at the coast as well as defined areas of sinking motion over the ocean near the coast.
Figure J9 Vertical velocity contours (m/s) at 30 meters altitude. 9 hours into simulation (October 23rd 2002, 1700 EDT). The winds are slowly weakening. The area of positive vertical motion is well defined along the coast. An area of negative vertical motion has begun to appear along the coast.
Figure J10 Vertical velocity contours (m/s) at 30 meters altitude. 10 hours into simulation (October 23rd 2002, 1800 EDT). The winds are slowly weakening. The area of positive vertical motion is well defined along the coast and the maximum positive winds can be observed in the south near the coast. An area of negative vertical motion has begun to appear along the coast.
Figure J11 Vertical velocity contours (m/s) at 30 meters altitude. 11 hours into simulation (October 23rd 2002, 1900 EDT). The winds are slowly weakening. The area of positive vertical motion is well defined along the coast and the maximum positive winds can be observed in the south near the coast. An area of negative vertical motion has begun to appear along the coast.
Figure J12 Vertical velocity contours (m/s) at 30 meters altitude. 12 hours into simulation (October 23\textsuperscript{rd} 2002, 2000 EDT). The winds are slowly weakening. The area of positive vertical motion is well defined along the coast and the maximum positive winds can be observed in the south near the coast. An area of negative vertical motion has begun to appear along the coast.
Figure J13 Vertical velocity contours (m/s) at 30 meters altitude. 13 hours into simulation (October 23rd 2002, 2100 EDT). The winds are slowly weakening. The area of positive vertical motion is slowly starting to shift to the left due to the onshore winds of the sea breeze.
Figure J14 Vertical velocity contours (m/s) at 30 meters altitude. 14 hours into simulation (October 23\textsuperscript{rd} 2002, 2200 EDT). The winds are slowly weakening. The area of positive vertical motion is slowly starting to shift to the left due to the onshore winds of the sea breeze.
Figure J15 Vertical velocity contours (m/s) at 30 meters altitude. 15 hours into simulation (October 23rd 2002, 2300 EDT). The winds are slowly weakening. The area of positive vertical motion is slowly starting to shift to the left due to the onshore winds of the sea breeze.
Figure J16 Vertical velocity contours (m/s) at 30 meters altitude. 16 hours into simulation (October 23rd 2002, 2400 EDT). The winds are slowly weakening. The area of positive and negative vertical motion is getting less defined due to the weakening of the vertical winds.
Figure J17 Vertical velocity contours (m/s) at 30 meters altitude. 17 hours into simulation (October 24th, 2002, 0100 EDT). The winds are slowly weakening. The area of positive and negative vertical motion is getting less defined due to the weakening of the vertical winds.
Figure J18 Vertical velocity contours (m/s) at 30 meters altitude. 18 hours into simulation (October 24th 2002, 0200 EDT). The winds are slowly weakening. The area of positive and negative vertical motion is getting less defined due to the weakening of the vertical winds.
Figure J19 Vertical velocity contours (m/s) at 30 meters altitude. 19 hours into simulation (October 24th 2002, 0300 EDT). The winds are slowly weakening. The area of positive and negative vertical motion is getting less defined due to the weakening of the vertical winds. An area of negative vertical motion has begun to appear at the coast.
Figure J20 Vertical velocity contours (m/s) at 30 meters altitude. 20 hours into simulation (October 24th 2002, 0400 EDT). The winds are very weak. The areas of positive and negative vertical motion are not defined anymore due to the weakening of the vertical winds. An area of negative vertical motion has begun to appear at the coast.
Figure J21 Vertical velocity contours (m/s) at 30 meters altitude. 21 hours into simulation (October 24th 2002, 0500 EDT). The winds are very weak. The areas of positive and negative vertical motion are not defined anymore due to the weakening of the vertical winds. An area of negative vertical motion has begun to appear at the coast.
Figure J22 Vertical velocity contours (m/s) at 30 meters altitude. 22 hours into simulation (October 24\textsuperscript{th} 2002, 0600 EDT). The winds are very weak but increasing again. An area of negative vertical motion has begun to appear at the coast.
Figure J23 Vertical velocity contours (m/s) at 30 meters altitude. 23 hours into simulation (October 24th, 2002, 0700 EDT). The winds are very weak but increasing again. An area of negative vertical motion has begun to appear at the coast.
Figure J24 Vertical velocity contours (m/s) at 30 meters altitude. 24 hours into simulation (October 24th 2002, 0700 EDT). The winds are very weak but increasing again. An area of negative vertical motion has begun to appear at the coast.
Appendix K

FAWN Data and AVHRR Cloud Mask
Comparison between skin temperature collected by FAWN and simulated temperatures.
Comparison between simulated convergence and actual cloud formation from AVHRR image.
Table K1 Comparison between air temperatures collected by FAWN at 60cm, 2 m and 10 m for Fort Pierce and simulated skin temperature.

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Figure K1 NOAA AVHRR image at 1 km resolution (R = band 3, G = band 2, B = band 1) taken at 7:20 pm EDT on October 23rd 2003.
Figure K2 Vertical velocity contours (m/s) at 30 meters altitude. 11 hours into simulation (October 23rd 2002, 1900 EDT).
Figure K3 Cloud mask (figure K1) overlapping the simulated convergence field (figure K2).
Appendix L

Computer user information
Input files listing:

The set of input files for serial and parallel runs are the same. The set of input files for the 2 km grid runs are resampled input files from the 1 km grid input files. A nearest neighbor method was used to obtain the 2km grid input files.

AVHRR brightness temperatures input files:

These files were created from images in Appendix D. Methods are detailed in Chapter 3 section 3.4.

20010221257ch4.txt: channel 4 AVHRR 200 x 200 matrix (1 km grid) for October 22\textsuperscript{nd} at 12:57 pm.
20010221257ch5.txt: channel 5 AVHRR 200 x 200 matrix (1 km grid) for October 22\textsuperscript{nd} at 12:57 pm.
20010221518ch4.txt: channel 4 AVHRR 200 x 200 matrix (1 km grid) for October 22\textsuperscript{nd} at 15:18 pm.
20010221518ch5.txt: channel 5 AVHRR 200 x 200 matrix (1 km grid) for October 22\textsuperscript{nd} at 15:18 pm.
20010221926ch4.txt: channel 4 AVHRR 200 x 200 matrix (1 km grid) for October 22\textsuperscript{nd} at 19:26 pm.
20010221926ch5.txt: channel 5 AVHRR 200 x 200 matrix (1 km grid) for October 22\textsuperscript{nd} at 19:26 pm.

MODIS albedo input files:

These files were created from images in Appendix E. Methods are detailed in Chapter 2 section 2.3.2.
200galbedobd1.txt: channel 1 MODIS 200 x 200 matrix for October 22\textsuperscript{nd} 2002.

200galbedobd2.txt: channel 1 MODIS 200 x 200 matrix for October 22\textsuperscript{nd} 2002.

200galbedobd3.txt: channel 1 MODIS 200 x 200 matrix for October 22\textsuperscript{nd} 2002.

200galbedobd4.txt: channel 1 MODIS 200 x 200 matrix for October 22\textsuperscript{nd} 2002.

200galbedobd5.txt: channel 1 MODIS 200 x 200 matrix for October 22\textsuperscript{nd} 2002.

200galbedobd6.txt: channel 1 MODIS 200 x 200 matrix for October 22\textsuperscript{nd} 2002.

200galbedobd7.txt: channel 1 MODIS 200 x 200 matrix for October 22\textsuperscript{nd} 2002.

Other MODIS input files:

200angstr.txt: MODIS angstrom 200 x 200 matrix for October 22\textsuperscript{nd} 2002.

200emiss31.txt: MODIS emissivity 200 x 200 matrix for October 22\textsuperscript{nd} 2002.

Other input files:

ozoneabs.txt: Ozone absorption coefficient.
solarcst.txt: Solar constant.
unigasabs.txt: Uniformly mixed gas coefficient.
waterabs.txt: Water absorption coefficient.
startup.txt: start up file indicating x and y-direction spacing and time step.

**Output files listing:**

The output files are filed in 3 different folders:
Hordir: folder with u, v, and z wind velocities
Pbldir: folder with u, v, and z wind velocities for vertical cross-sections.
Thermodir: folder with all thermo model results (net, upwelling (short and long), downwelling temperature, and thermal inertia).

**Checkpointing files listing:**

Checkpoint.txt: see section below for more explanations.
Neworold.txt: see section below for more explanations.
The other files saved in the “check” folder are necessary to restart the run at a latter time.

**User information for compiling and running the UTC-M Mesoscale model on the Beowulf supercomputer:**

The code named MM200.f90 needs to be compiled with mpif90, which allows the MPI commands to be compiled. A PBS file needs to be created to send your job to a queue.

**Example of a PBS file:**

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In the PBS file, you specify the number of nodes you want to use, the time you expect your code to run and the queue you are assigned to.

To run the program, make sure all inputs files and outputs folders are located under the same directory, and then use these couple lines to compile and create an object file.

```
mpif90  –c   MM200check.f90
mpif90     -o   MM200check    MM200check.o
qsub     MM200check.pbs
```

During a run on Bluemarlin, you can use several commands associated with the portable patch system (PBS):

```
qstat: checks the status of your program.
qstat  -n: gives a more detail status of your run (number of processors, time...).
qstat  -f: gives details about the amount of memory used.
qdel:"id#”   terminates your program.
```
There are 2 ways one can run the model. Either, one can run it as a new simulation or one can run it starting from a previous simulation.

The checkpointing capabilities of the UTC-M Mesoscale model are outlined below.

2 files used for checkpointing capabilities can be found under this directory:

/data/users/kingj/MM200check/check/checkpoint.txt
/data/users/kingj/MM200check/check/neworold.txt

cCheckpoint.txt:
1 ! "0" if you want your code to keep going
! "1" if you want to stop the code at the next loop

The checkpoint.txt file allows one to stop the code by changing the value from “0” to “1” during a run. At every time step, the program reads this file and if the value is changed to “1” then the program will save all files needed to restart the simulation later on at the point where it was stopped. When one resubmits the job, one would have to change the value back to “0”.

Files saved when one end a simulation are stored under the “check” folder and will be used as input files for the next simulation if one wants to continue the simulation.

neworold.txt:
1 ! "0" if brand new start
! "1" if continuous run
The neworold.txt file defines if one wants the program to read the initial conditions files or files saved at the end of the last simulation. At the beginning of each runs, this file is read and if the value is set to “1”, then the program will read files saved at the end of the last simulation. If it is set to “0”, then the program will read initial files from the input folder.

Listing of files saved after modifying the chexpoint.txt file:

- hydpcheck.txt
- hydptzcheck.txt
- qcheck.txt
- qtzcheck.txt
- rhocheck.txt
- temptzcheck.txt
- thetacheck.txt
- thetatzcheck.txt
- time.txt
- ucheck.txt
- uggcheck.txt
- vcheck.txt
- vggcheck.txt
- wccheck.txt