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Numerical Simulation Of The Marine Atmospheric Boundary Layer Flow Over The Cape Canaveral Region And Gulf Stream Using AVHRR SST – An Atmospheric Seabreeze Model

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ABSTRACT

An atmospheric mesoscale numerical model has been developed and applied to the marine atmospheric boundary layer (MABL) in the vicinity of sharp sea-surface temperature (SST) gradients. The Gulf Stream offshore of East Central Florida near Cape Canaveral is the region of interest. The model equations which govern atmospheric behavior are based on the basic conservation laws of mass, momentum, and energy. A "primitive" equation formulation is used to produce preliminary predictions of horizontal velocity, potential temperature, and water vapor. The hydrostatic approximation is applied to the vertical momentum conservation equation and the anelastic form of the continuity equation is used to approximate mass conservation. Surface fluxes of momentum, heat, and moisture are estimated using high-resolution SST data obtained in near-real time from the AVHRR infrared instrument. Interaction of the boundary layer flow with the nearby Central Florida peninsula is simulated by inducing a diurnal seabreeze circulation across the coastline. It is found through a series of simulations that the distribution of sea-surface temperatures influences the boundary layer flow field - especially over the region of the Gulf Stream front and the low level convergence field, which may be an important factor for initiating convective precipitation over the land-water margin and Gulf Stream. The importance of the air-sea or air-land interfaces are thus fully recognized as being crucial to parameterize via remote sensing data in order to proceed with further model developments of the newly developed sea-breeze model. Further examination of methods for estimating surface temperatures of the water and land as well as inclusion of surface gravity wave forcing due to the diurnal seabreeze over the complex land-water margin in the Indian River Lagoon will need to be included in order to utilize the model for future application.

1. INTRODUCTION & MODEL BACKGROUND

The three-dimensional atmospheric model developed for this study is a "primitive equation" model. Primitive equation models describe fluid motions in terms of the basic physical laws of conservation of momentum, energy, and mass through prediction of the independent variables (u, v, w, θ, ρ), where u, v, and w, are the orthogonal components of the velocity field (ms-1), θ is the potential temperature (K), and ρ is the density (kgm-3). The model, through scaling assumptions, uses the hydrostatic approximation to the vertical momentum conservation (w) equation and the anelastic approximation to the mass conservation (ρ) equation. The evolution of the water vapor field is also examined through prediction of the specific humidity (q) field (gkg^{-1}). The atmosphere's water vapor content has an influence in the mass (density) field, and also contributes to the energy budget through latent heat release or absorption associated with phase changes. For this study, the water vapor field is assumed to have no sources or sinks and therefore latent energy is neglected. The water vapor's contribution is only to the mass field. With this assumption, the resulting dynamics suggests that latent energy releases are minimal in the atmospheric boundary layer below the lifting condensation level (LCL) which is the region under study. Written in vector form, the equation describing conservation of momentum on a flat rotating earth can be written as:

$$\frac{\partial \boldsymbol{V}}{\partial t} = -\boldsymbol{V}\cdot\nabla\boldsymbol{V} - f(\boldsymbol{k}\times\boldsymbol{V}) - \frac{1}{\rho}\nabla p - g\boldsymbol{k} + F_r \ . \qquad (1)$$

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Term I on the left hand side of equation (1) represents the local time rate of change velocity (momentum per unit mass), where the velocity vector V has the orthogonal components (u,v,w). The terms on the right hand side are the advection of the velocity field by the mean wind, the Coriolis force per unit mass, the pressure gradient force per unit mass, and the turbulent diffusion of the mean wind. The vertical (w) component of equation (1) may be expressed as:

$$\frac{\partial w}{\partial t} = -u\frac{\partial w}{\partial x} - v\frac{\partial w}{\partial y} - w\frac{\partial w}{\partial z} + \hat{f}u - g - \frac{1}{\rho}\frac{\partial p}{\partial z} + \nu\nabla^2 w\,, \tag{2}$$

where $f = 2\Omega \cos \phi$. Via scaling analysis of the individual terms in equation (2) (Holton 1992, Pielke 1981, McIlveen 1986), it may be seen that the magnitude of the gravity and pressure gradient force terms are much greater that the other terms, and equation (2) can be reduced to:

$$\frac{\partial p}{\partial z} = -\rho g\,. \tag{3}$$

Thus for the mesoscale model developed for purposes of this study, the atmosphere as assumed to be in hydrostatic equilibrium. The equation describing the conservation of mass in a fluid of density ρ can be written as:

$$\frac{\partial \rho}{\partial t} + \nabla \cdot p\boldsymbol{V} = 0\,, \tag{4}$$

or in Cartesian form is:

$$\frac{\partial \rho}{\partial t} + \frac{\partial \rho u}{\partial x} + \frac{\partial \rho v}{\partial y} + \frac{\partial \rho w}{\partial z} = 0\,. \tag{5}$$

Through scaling assumptions about the individual terms in equation (5), the expression for conservation of mass can be reduced to:

$$\rho\frac{\partial u}{\partial x} + \rho\frac{\partial v}{\partial y} = -\frac{\partial \rho w}{\partial z}\,. \tag{6}$$

The time rate of change of density term $\frac{\partial \rho}{\partial t}$ may be safely dropped for this model application. This is known as the *anelastic* or *soundproof* approximation and assures that acoustic waves will not be a solution in the continuity equation (Pielke 1981). For purely horizontal atmospheric flows, incompressibility is a reasonable approximation, that is, $\frac{D\rho}{Dt} = 0$, but when vertical motions are present the vertical variation of the density field must be taken into account and equation (6) holds. The equation for conservation of energy is known as the thermodynamic energy equation and is written in terms of a time rate of change of potential temperature, or:

$$\frac{\partial \theta}{\partial t} = -\boldsymbol{V} \cdot \nabla\theta + \frac{S}{c_p}\,. \tag{7}$$

Where θ is the potential temperature (K) that for a moist atmosphere can be expressed as:

$$\theta = T_v\left(\frac{p_0}{p}\right)^{R/c_p}\,. \tag{8}$$

In the above, p is the pressure of a parcel of air, p_0 is a reference pressure value conventionally defined as 1000 mb, R is the universal gas constant, and c_p is the specific heat at constant pressure. The virtual temperature T_v (K) is the temperature that an element of moist air has if it is as dense as an element of dry air at temperature T, or: $T_v = T(1 + 0.61q)$, where q is the specific humidity (gkg^{-1}). Thus the effect of moisture is to warm the air (increase buoyancy). The term S represents the local heating rate per unit mass due to radiative and latent sources of heat. The equation describing conservation of water vapor in the atmosphere for this model study is expressed in terms of a prediction of the specific humidity field q (gkg^{-1}), or:

$$\frac{\partial q}{\partial t} = -\mathbf{V} \cdot \nabla q + \frac{L}{c_p}S_q\,. \tag{9}$$

The source/sink term Sq will be assumed zero in this initial study since changes in the water vapor field due to condensation/evaporation will be assumed to be negligible in this simplified atmospheric boundary layer. Application of the Reynolds's (time) averaging process for a variable allows for definition of an instantaneous value as a sum of the mean and fluctuating components, or:

$$\phi = \overline{\phi} + \phi' . \tag{10}$$

Using the above and applying the result to the model predicted variables (u, v, w, q, θ) and replacing all occurrences of these values in equations (1), (7), and (9) allows the prognostic equations to be re-written as:

$$\frac{D\overline{u}}{Dt} = f\overline{v} - \frac{1}{\rho}\frac{\partial \overline{p}}{\partial x} + K\left(\frac{\partial \overline{u'u'}}{\partial x} + \frac{\partial \overline{u'v'}}{\partial x} + \frac{\partial \overline{u'w'}}{\partial x}\right), \tag{11}$$

$$\frac{D\overline{v}}{Dt} = -f\overline{u} - \frac{1}{\rho}\frac{\partial \overline{p}}{\partial y} + K\left(\frac{\partial \overline{v'u'}}{\partial x} + \frac{\partial \overline{v'v'}}{\partial x} + \frac{\partial \overline{v'w'}}{\partial x}\right), \tag{12}$$

$$\frac{D\overline{\theta}}{Dt} = K\left(\frac{\partial \overline{\theta'u'}}{\partial x} + \frac{\partial \overline{\theta'v'}}{\partial x} + \frac{\partial \overline{\theta'w'}}{\partial x}\right), \tag{13}$$

$$\frac{D\overline{q}}{Dt} = K\left(\frac{\partial \overline{q'u'}}{\partial x} + \frac{\partial \overline{q'v'}}{\partial x} + \frac{\partial \overline{q'w'}}{\partial x}\right), \tag{14}$$

where K represents a "zero order" general linearized eddy diffusion coefficient. We use this approach as an initial approach (although quite limiting) in order to complete the numerical modeling problem in a tractable manner as a first step, and then we utilize a stochastic approach described below. An additional equation must be added to the above to develop a set of governing equations in closed form. The equation of state for a two component gas (dry air and water vapor) which we assume will be expressed as:

$$\overline{p} = \rho R\overline{T_v} , \tag{15}$$

where the linearization has been made (Stull, 1988) such that the normalized products of the fluctuations $\rho' T_v' / \overline{\rho T_v'}$ is assumed much smaller than the other terms, so that the mean equation of state is valid. The fluctuation product terms or covariance's in equations 10-13 that result from the presence of the nonlinear advection terms are generally not known and are a function of the flow. Therefore parameterization of these fluxes must be made for mathematical and computational "closure" to the set of governing equations. For testing of the model we begin with a "flux-proportional-to-gradient" method (Libbey, 1994) where an eddy flux term (e.g. $\overline{u'w'}$) is assumed proportional to the local gradient through a transfer coefficient (the zero order model assumption) or:

$$\overline{u'w'} = -K\frac{\partial \overline{u}}{\partial z} \tag{16}$$

The parameterization for the eddy transfer coefficient K in the surface layer are based on the Monin-Obukov similarity theory and empirical values for these terms are shown by Businger (1972). Transfer coefficients for the transition layer are calculated using O'Brien's (1970) cubic polynomial. Uhlhorn's and Bostater's original atmospheric model, we now call *the Florida Tech **UTC-M** atmospheric seabreeze model* has been modified to allow refinement for realistic testing. For example a call to a subroutine which generates K_m, K_h, and K_q from the randomly generated fluctuating field terms u', v', w', θ', and q'. These values are generated from a random number generator subroutine originally described by (Wichmann and Hill) and improved upon by an embedded and repeating time seed by Bostater (1989). The random number generator uses a uniform distribution function when generating numbers between 0 and 1. The generated number r is then multiplied with the given range of the field term as follows: $m' = \sin(r * (2\pi)) * m_{range}$ where m is the field term (u, v, w, θ, or q) and m_{range} is the range of fluctuation of the field term as specified by the user. The ranges used to run model for preliminary testing "***the stochastic turbulence closure (STC) methodology***" are as follows: $u_{range} = \pm 0.1$ ms^{-1}, $v_{range} = \pm 0.1$ m^{s-1}, $w_{range} = \pm 0.05$ m^{s2}, $\theta_{range} = \pm 0.1^{o}$ K, and $q_{range} = \pm 0.0001$ gkg^{-1}. The K_m, K_h, and K_q diffusivity terms are then calculated as follows: $K_m = (u'w')/(\sqrt{\ }\ [(\delta u/\delta z)^2 + (\delta v/\delta z)^2]$, $K_h = (\theta' w')/(\delta\theta/\delta z)$, $K_q = (q'w')/(\delta q/\delta z)$. The atmospheric model was run using both the randomly generated flux and the method described above. In both runs, the same initial conditions are used. This includes the use of real sea surface temperature derived from AVHRR imagery.

2. METHODS

The above set of non-linear partial differential equations must be solved simultaneously using *numerical methods* in order to demonstrate the evolution of a time-dependent atmospheric circulation. The non-linearity of the coupled equations requires that numerical methods be applied in seeking solutions since no known analytical solutions exist. Finite difference, finite element, and spectral (wave number) methods are examples of common numerical treatment of partial differential equations. The numerical method used for purposes of this model is the finite difference method. The finite difference method utilizes a form of a Taylor series expansion about a point in space or time to approximate partial derivatives at that point (see e.g., Press et al. 1992, Roache 1973). In this section the various finite difference methods applied in seeking solutions to the model partial differential equations will be presented. The methods are shown on a term-by-term basis. The procedure for calculating individual terms in the conservation equations separately is called the splitting-up method and is detailed by Marchuk (1971). The reader is also referred to Long and Hicks (1975) and Messinger and Arakawa (1976). The horizontal advective terms in the model equations (e.g. $\partial u / \partial t = -u\partial u / \partial x - v\partial u / \partial y$) are computed using the two-step Lax-Wendroff scheme first proposed by Richtmyer (1963). The two-step Lax-Wendroff method is second-order accurate in time and avoids the computational/physical mode splitting problem of the staggered-leapfrog time differencing (Press et. al. 1992). Intermediate forecast values centered at half time-steps and half grid-points are used to predict a variable at the advanced time step. It can be shown that this particular method is conditionally stable based on the Courant-Friedrichs-Levy (CFL) criterion (Hess, 1959) that a wave speed $C = v\Delta x / \Delta t \le 1$ for velocity v, grid spacing Δx, and time-step Δt. Analogous expressions are obtained for the v-momentum, energy and water vapor conservation equations. In terms of the horizontal advection in the u-momentum equation, the continuum representation (eqn. 14) and finite difference analog (eqn.15) is:

$$\frac{\partial u}{\partial t} = -u\frac{\partial u}{\partial x} - v\frac{\partial u}{\partial y}, \tag{17}$$

$$\begin{aligned} u_{i,j}^{n+1} = u_{i,j}^{n} &- \alpha\left[\frac{1}{2}\left(u_{i+1,j}^{n} + u_{i,j}^{n}\right) - \frac{1}{2}\alpha\left(u_{i+1,j}^{n} - u_{i,j}^{n}\right) - \frac{1}{2}\left(u_{i,j}^{n} + u_{i-1,j}^{n}\right) + \frac{1}{2}\alpha\left(u_{i,j}^{n} - u_{i-1,j}^{n}\right)\right], \\ &- \beta\left[\frac{1}{2}\left(u_{i,j+1}^{n} + u_{i,j}^{n}\right) - \frac{1}{2}\beta\left(u_{i,j+1}^{n} - u_{i,j}^{n}\right) - \frac{1}{2}\left(u_{i,j}^{n} + u_{i,j-1}^{n}\right) + \frac{1}{2}\beta\left(u_{i,j}^{n} - u_{i,j-1}^{n}\right)\right] \end{aligned} \tag{18}$$

where in equation (18), $\alpha = u_{i,j}^{n} \Delta t / \Delta x$, $\beta = v_{i,j}^{n} \Delta t / \Delta x$, the i subscript denotes the i^{th} gridpoint in the x-direction, and similarly for j in the y-direction. Analogous expressions are obtained for the v-momentum, energy, and water vapor conservation equations. The vertical advection terms are calculated using the upwind finite difference method. Because the vertical grid mesh is not necessarily equally spaced, the space derivatives cannot be exactly centered about a gridpoint. A conditionally stable first-order truncation error method is the upwind method, where the differences are either forward or backward in space, depending on the local direction of the flow. The upwind method representation for the vertical advection term in the u-momentum conservation equation $\partial u / \partial t = -w \, \partial u / \partial z$ is:

$$u_k^{n+1} - u_k^n \begin{cases} = \dfrac{-w_k^n \Delta t}{z_k - z_{k-1}}\left(u_k^n - u_{k-1}^n\right), & w_k^n > 0 \\[2ex] = \dfrac{-w_k^n \Delta t}{z_{k+1} - z_k}\left(u_{k+1}^n - u_k^n\right), & w_k^n \le 0 \end{cases} \tag{19a,b}$$

where in equation (16) the k subscript denotes the k^{th} vertical gridpoint, w is the local vertical velocity, and z is the altitude. The horizontal turbulent diffusion equations are calculated using a semi-implicit finite-difference formulation known as the Alternating-Direction-Implicit (ADI) method (Roache 1972). The ADI method utilizes the concept of time splitting, in which each direction is treated in a time-implicit manner at alternating half time-steps (while the other direction is calculated explicitly at that half time-level). Solutions to implicit finite difference schemes must be determined through matrix manipulations, and the ADI method solves a matrix in tridiagonal form for which computations are highly efficient. Vertical turbulent diffusion terms in the model equations are calculated using the Dufort-Frankel finite difference scheme (Anderson, Tannehill, and Pletcher, 1984) which is an explicit method and can be shown to be unconditionally stable for any wave speed.

For *boundary and initial conditions*: at the lateral (horizontal) boundaries, an open-type boundary condition is utilized. The method is achieved by definition of a zero gradient of all variables at the boundaries, that is, $\partial\phi/\partial x = \partial\phi/\partial y = 0$ for a scalar ϕ. This particular method has been employed successfully by Berri and Nunez (1993). It should be noted that a better method exists for treatment of lateral open boundaries and is known as the radiation condition described by Orlanski (1976), applied to hyperbolic steady flows. The Sommerfeld boundary condition previously discussed in Kreiss (1966) and Pearson (1974) is extended by Orlanski to include an explicit wave speed (C) calculation at each boundary grid point, and from that the boundary condition $\partial\phi/\partial t + C\,\partial\phi/\partial x, y = 0$ is determined.

The top of the numerical model domain is placed just within the upper troposphere at an altitude of 10 km. A rigid top boundary is utilized, meaning that vertical velocities are zero ($w = 0$). (see e.g., Estoque 1961). The local coastal atmospheric (seabreeze) circulation to be modeled is assumed to be of much shorter time scale than the synoptic scale disturbance propagation. Therefore the flow at the top boundary is assumed to be in geostrophic balance according to the initial hydrostatic pressure field, or:

$$\overline{u} = \overline{u}_g,\ \ \overline{v} = \overline{v}_g \ \ @ \ \ z = z_{top}.$$

The temperature and moisture at the top boundary are the initial observed values, denoted as:

$$\overline{\theta} = \overline{\theta}_{obs},\ \ \overline{q} = \overline{q}_{obs} \ \ @ \ \ z = z_{top}.$$

The surface of the earth is the only well-defined physical interface in the atmosphere. For the flow field over the land surface, a no-slip boundary condition is used, that is, $\overline{u}(z_0) = \overline{v}(z_0) = \overline{w}(z_0) = 0$. The potential temperature at the surface of the earth is taken to be temporally invariant over the sea surface during the model integration period, and is spatially invariant but time dependent over the land surface, or:

$$\overline{\theta} = \begin{cases} \overline{\theta}(t) & \text{over land,} \\ \overline{\theta}(x, y, z_0) & \text{over sea.} \end{cases}$$

The above is used as an initial testing of the model, and is a necessary first step at this juncture in the model development for simplicity. The land surface water vapor field (q) is initialized from griddng the limited surface observations over the land surface, and the first grid level is assumed to be saturated over the sea surface. The pressure field at the surface is obtained from gridding the limited observations taken at various reporting stations across the Central Florida region. Upper-air (rawinsonde) soundings are deployed every twelve hours from the Cape Canaveral Air Force Station (XMR). Measurements of wind speed and direction, temperature, and dewpoint temperature are made at mandatory (surface, 1000, 925, 850 mb, etc.) and significant levels when conditions warrant (i.e., strong gradients). Each of these measurements are linearly interpolated to the vertical levels within the model vertical grid. The altitude is then determined through integration of the hydrostatic equation (Wallace and Hobbs, 1977) from the surface. Wind speed and direction are converted to u and v-wind components and the specific humidity q is computed from the dewpoint temperature T_d. The boundary conditions of the model are such that the surface temperature over the land varies with the diurnal solar cycle. The surface temperature over the land is assumed to be accurately represented by $\tilde{\theta} = \Theta + \theta(t)$, where $\tilde{\theta}$ is the instantaneous potential temperature in Kelvin's at the surface of the earth (z=0), Θ is the minimum value for the diurnal cycle, and $\theta(t)$ is the time dependent departure of the temperature from the minimum value. For this model study, the minimum value is defined to occur at 0600 local daylight time (10Z) which corresponds approximately to sunrise during the summer months along the Space Coast and is also the time the upper-air sounding is made from Cape Canaveral. The coupling of the thermodynamic and momentum fields is provided by the pressure $\tilde{p} = P + p$, where P is the initial hydrostatic pressure field and p is the perturbation of the hydrostatic pressure field which is related to the temperature fluctuation by (Dutton and Fichtl, 1969) as:

$$\frac{\partial p}{\partial z} = \frac{\alpha}{\alpha_0^2} g \cong \frac{g}{\alpha_0}\frac{\theta}{\Theta} \quad . \tag{20}$$

A case study of the model described above is performed using AVHRR SST data as initial conditions at the air-sea interface. The model initial conditions are presented for July 31, 1996. The location of the model domain with respect to the East Central Space Coast of Florida is shown in Figure 1. The sea surface temperature contour field is shown in Figure 2, along with the subjectively analyzed Gulf Stream center axis location. Figures 3, 4, and 5 show model output at selected intervals through out a model cycle of one day. In this case, the AVHRR SST is grided & thus defined across the model air-sea surface domain. The actual AVHRR data is then used for the case where the SST distribution is provided by remotely sensed satellite data. The results for the satellite based SST forcing is shown in Figures 4, through 1, with Figure 2 showing the utilized lower boundary SST (contours from grided data from NOAA's Coastwatch data) used as the boundary conditions.

3. RESULTS

As shown in Figure 1, the model domain covers the region from approximately 27^0 N to 19.16^0 N latitude and -79^0 W to -81.5^0 W latitude where each grid cell is on the order of 5x5 km. The model domain is limited to a 24 by 24 grid region due to computational considerations. The coastline is seen as a solid line with Cape Canaveral located at approximately 28.5^0 W and 80.5^0 W. Thus the model domain is from Daytona to just north of West Palm Beach, Florida. The dashed line represents the approximate axis location of the south to north flowing Gulf Stream current.

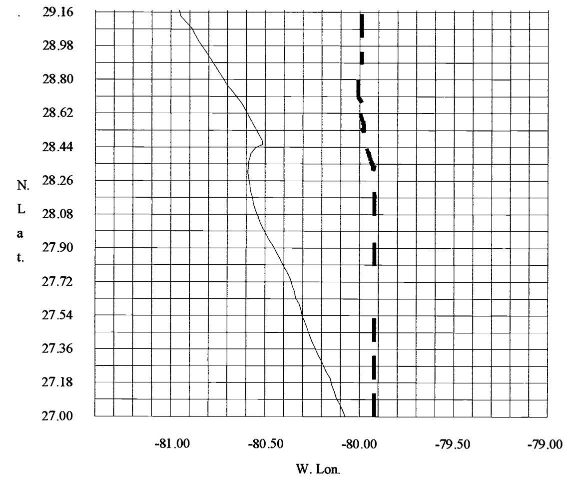


Figure 1 (right). **UTC-M** model grid showing the 25x25 Km grid cells for the model domain from approximately Daytona Beach, Fl. to West Palm Beach, Florida. This model domain covers the North-South region of the Indian River Lagoon. The Eastern boundary of the grid Lies within the eastern side of the 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.

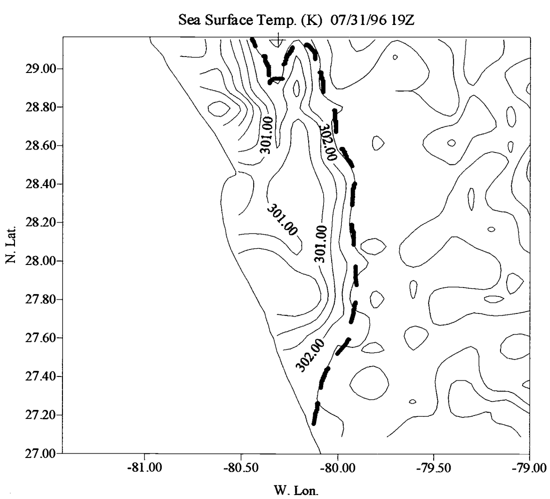


Figure 2. (above) Sea surface temperature derived from AVHRR imagery for 7-31-96 (19Z). The location of the Gulf Stream can be seen as the sharp temperature gradien running roughly parallel to the coastline. Cool waters exist just offshore north of Cape Canaveral. Dashed line shows approximate location of the west wall of the Gulf Stream. Data from this SST Image is used for the bottom boundary condition initial conditions.

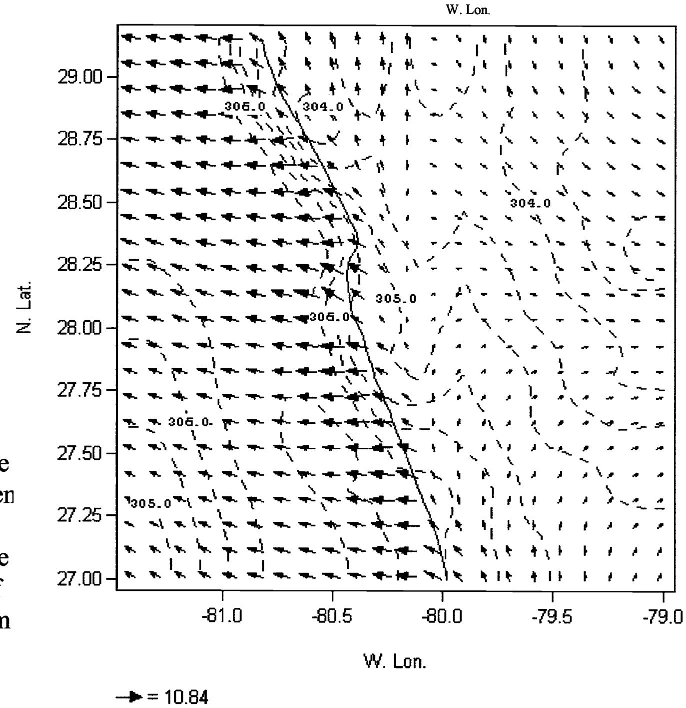


Figure 3. (left) Boundary layer (30 m) winds in ms^{-1} see arrow) and potential temperature (dashed lines) in K, 6hrs into simulation using grided SST data in figure 2.

Figure 4. (left) Boundary layer (30 m) winds in ms^{-1} see arrow) and potential temperature (dashed lines) in K, twenty four hours into a simulation (31 July, 1996, 1800 LDT) using the grided surface SST distribution showed in figure 2. Note strong shear in the wind can be seen over the Gulf Stream front at the northern boundary of the model domain. The seabreeze has penetrated to its farthest distance inland as solar heating is rapidly weakening. The pressure gradient across the coastline is weakening along with the resulting inland flowing seabreeze.

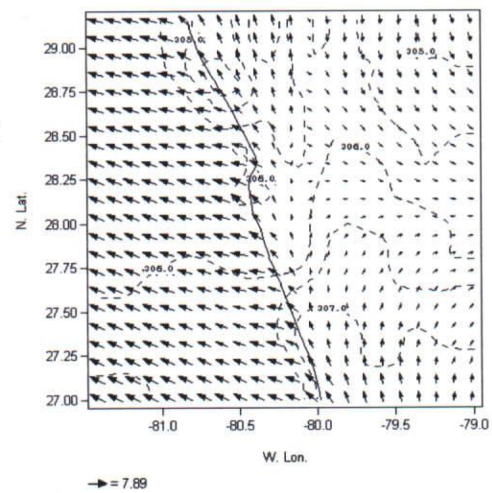
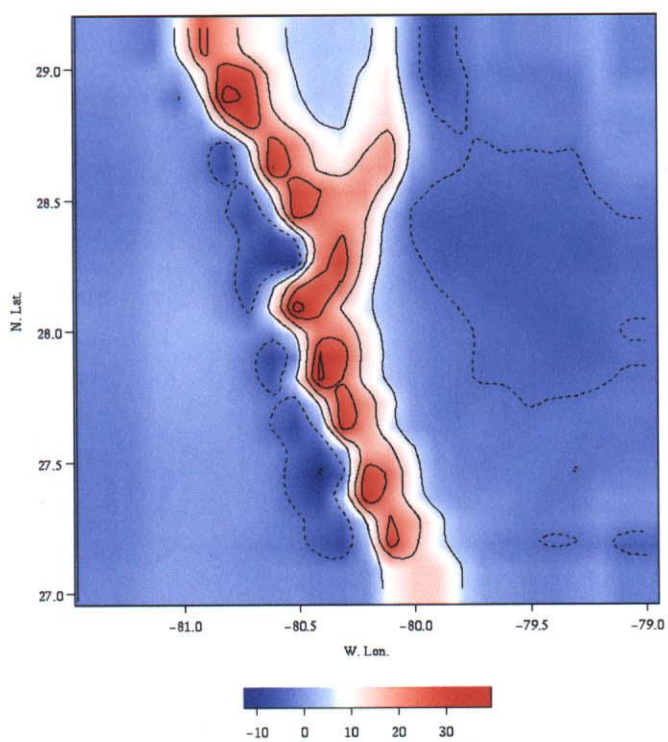
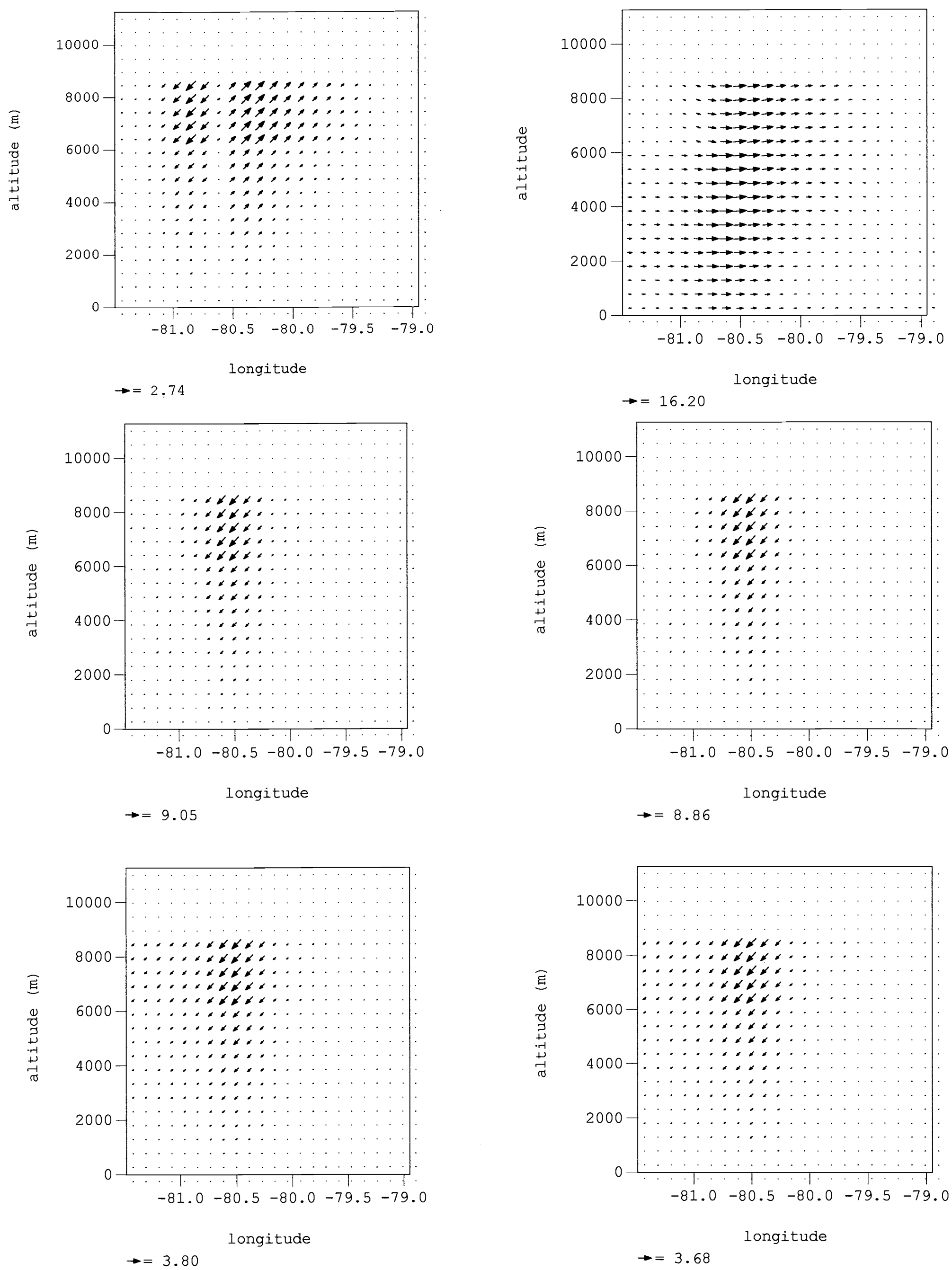


Figure 5 (below) shows the convergence field (m/s/deg) at the model 30 meter altitude after the model has run 12 hours (July 31, 1800 LDT). The dashed lines indicate the convergence filed and indicates low level convergence over the coastline. This suggests the vertical advection and increase in motion and cloud developments could be expected to occur in the intensified convergence areas long this land-water margin.



The Monte-carlo stochastic turbulence closure scheme was applied to test and compare the technique with the eddy terms proportional to the mean flow assumptions described above and referenced below. Figures 6-11 on the following page compares the resulting vertical velocities and indicated vertical motion as a function of altitude from a crossection taken in the Melbourne Florida area These results suggest the two techniques give similar results after the model spins up or after 6 hours of simulation time.



Figures 6-11. The figures on the right show the resulting **UTC-M** model velocities as a function of altitude (m) from a crossection in the model domain near Melbourne Florida. A model simulation using the stochastic Monte Carlo turbulence closure (**STC-M**) scheme with AVHRR sea surface temperature initialization. The figures on the right indicate the standard "eddy diffusivity" closure approach described. Model results are different for the first 6 hours of the simulation, however 14 hour and 22-hour simulations indicate quite similar results, showing the realism capable with the stochastic turbulence closure scheme.

4. DISCUSSION

A diurnal land/sea-breeze oscillation is typically known to exist along the eastern coastline of the Space Coast of Florida during the summertime or "rainy" season. The associated rapid land surface heating during this period of the year is sufficient to erode nocturnal thermal inversions, which are typically combined with moisture from the Atlantic Ocean and the complex estuarine and lagoon system in central Florida. The complex air-sea-land processes and local geography can produce numerous short-lived convective thunderstorms (Nichols, Pielke, and Cotton 1991; Blanchard and Lopez 1985; and others). At nighttime, with a lack of incoming solar radiation, the land surface cools to temperatures below that of the ocean. This allows for a reversed land-to-sea atmospheric flow (land-breeze) (Zhong and Takle 1992). As this cooler air is advected out to sea it may encounter an unstable marine layer overlying the Gulf Stream current. This instability is due to the cooler air from the land in contact with relatively warm waters. (Other more complex mechanisms such as rain-cooled thunderstorm outflow boundaries may also be responsible). Convective shower and thunderstorm activity may be initiated as a result of the unstable marine atmospheric boundary layer. The results in this paper has 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 marine boundary layer of the numerical model is initialized using sea-surface temperatures derived from high-resolution (~ 4 km) AVHRR infrared imagery. 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. The three-dimensional hydrostatic meso-g scale numerical model (**UTC-M**, developed at Florida Tech,) has been shown to generate the seabreeze circulation when initialized with AVHHR SST data along the Indian River Lagoon watershed region. The semi-implicit model which makes use of a k-e model for the parameterization of sub-grid scale turbulent fluxes of momentum, heat, and water vapor or a Monte carlo based "stochastic" turbulence closure methodology (**STC-M**) for the turbulent flux terms as shown in this paper will need further testing of the anelastic assumptions to produce realistic vertical advection association with gravity wave processes which are generated over the complex air-land-water geography in the Cape Canaveral region. The "dry" flow and thermodynamic fields in the atmospheric boundary layer below the lifting condensation level will also need to be examined and latent heating of the atmosphere due to condensation of water vapor will need to be developed in the model along with other physics such as cloud generation physics. Results for initializing the model using AVHRR derived sea-surface temperature data, land-surface hourly observations, and upper-air data from radiosonde sites across the state of Florida can be used to run the model. Future research may include the nesting of the model with NWS model output along with examination of the UTC-M model output at various time increments and the joint occurrence of model-computed convective instability regions along with NEXRAD data from the NWS and the intensive meteorological data available at NASA, Kennedy Space Center, as well as the inclusion of radiative transfer model similar to the mathematical and numerical techniques developed by Bostater, McNally and Ma[26] to the Princeton Ocean Model for large scale geophysical fluids.

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