3.1 Importance of the Study of Urban Heat Island (UHI)

Rapid urbanization, in the developing countries of world mostly located in the tropical/subtropical regions, in recent years, is bound to change the nature of the environment. This shows the relevance and priorities that studies of the climate of tropical cities must be given in coming decades.

Literatures on urban climatology in tropical environments have increased substantially in the present decade, it still represents a small fraction of the work done in mid-latitudes. This is mainly due to lack of infrastructure and limited availability of financial resources to provide for expensive and more sophisticated instrumentation required for urban climate research.

In the 1970's research on urban climate required simple instrumentation being mostly limited to description of fields of temperature, humidity, etc. of a city. During recent decades attention has been shifted to the causes underlying the observed effects and to the modelling of such processes.

The events on urban climatology periodically promoted by WMO, UNEP and other international agencies have been influential in promoting the development of this area of research in tropics (and of course elsewhere).

- Recommendations formulation in the WMO Technical conference of Urban Climatology in Mexico city led to the idea of TRUCE (The Tropical Urban Climate Experiment) with the aim of a better understanding of the atmosphere of the tropical cities.
- The urban climate, planning and building conference in Kyoto in 1989 registered the largest attendance ever with emphasis on building climatology and applications of climate information.
- TecTUC (Technical Conference on Tropical Urban Climate) Dhaka in 1993, gave importance to further development of plans and implementation of TRUCE.
- The ICUC (International Conference on Urban Climatology) at Essen in 1996 and at
Sydney in 1999 noticed the importance of urban climatology.

Every metropolitan area produces a significant UHI. The UHI is primarily a nocturnal phenomenon that results from the city’s substantial heat output and it’s greater efficiency at radiating heat upward from the ground. The UHI is most pronounced on calm, dry, clear nights, and is small or absent on windy, humid, overcast nights with precipitation, because under those conditions, the nocturnal inversion cannot form, and urban and rural heat budgets are similar.

The structure of the atmospheric boundary-layer in the region of an urban conurbation has many interesting features compared with that of rural areas. The most frequently observed effect is the increase in surface and air temperatures within the urban conurbation. The concrete buildings and water proof surfaces of cities that have high albedo and heat capacities which can store incoming solar radiation as sensible heat better than surrounding rural landscape. Hence, according to Oke, the surface temperature within an urban conurbation is usually warmer, peaking over the city center. However, these temperature anomalies are modified by local weather conditions.

The importance of the study of urban heat island at Delhi can thus clearly be understood in the backdrop of research on urban climatology undergoing in tropical areas.

3.2 Different Approaches to study the UHI effect

The influence of urbanization and land use/land cover (LULC) on several meteorological variables has been studied (e.g., Landsberg, 1981; Kukla et al., 1986; Karl et al., 1988; Changnon, 1992; Gallo et al., 1993, 1996, 1999). Gallo et al. (1996) observed that weather observation stations with a predominantly urban-related LULC usually displayed a lower diurnal temperature range (DTR = Maximum – minimum temperature) compared to those stations with predominantly rural LULC. He speculated that the transition of the LULC from predominantly rural to a more urban setting can have an impact on the trends in temperature (specifically a decrease in the diurnal temperature range) similar to that which would be expected under an enhanced greenhouse warming scenario. Thus, the LULC influences on temperatures, as opposed to greenhouse warming influences, need to be clearly defined.

Of the several areas that require additional observations, the urban outgoing long wave energy flux is of special significance for it determines the heat island intensity at any particular city. The intensity of the heat island in a tropical city is likely to be more dependent on the land-use characteristics. Urban development usually results in a dramatic alteration of the Earth's
surface, as natural vegetation is replaced by non-evaporating, non-transpiring surfaces (e.g., stone, metal, concrete, etc.). Under such alteration, the partitioning of incoming solar radiation into fluxes of sensible and latent heat is skewed in favour of increased sensible heat flux as evapotranspirative surfaces are reduced. From thermal infrared measurements (10.5-11.5 μm) acquired by the Advanced Very High Resolution Radiometer (AVHRR) aboard the NOAA series of polar orbiting satellites, Roth et al. (1989) derived surface temperature data and assessed its spatial distribution across several cities along the west coast of North America. The fundamental characteristics of most concern for a nocturnal heat island are its mixing height, intensity, temperature distribution, and heat-island-induced circulation. The mixing height is generally maximum near the city center and decreases with radial distance from the center.

There are different approaches to study the UHI phenomenon. They are briefly discussed in the following paragraphs:

Field experiments on urban boundary layers have been conducted in many cities, such as Montreal (Summers, 1965), New York City (Bornstein, 1968), Cincinnati (Clarke, 1969), Columbus (Clarke and McElroy, 1974), St. Luis (Ching et al. 1978; Shreffler, 1978, 1979a, b). Comprehensive reviews of early observational studies have been given by Oke (1974, 1979, 1987) and Landsberg (1981), and a more comprehensive listing has been given by Oke (1995). Field experiments are generally expensive and their findings are often limited to a particular city.

Logistical difficulties of field experiments have encouraged a number of physical simulations in the laboratory (Raman and Cermak, 1974, 1975; Torrance, 1979; Giovannoni, 1987; Noto, 1996), as well as theoretical and numerical models (Vukovich et al., 1976; Vukovich and Dunn, 1978; Byun and Arya, 1990). The advantages of a laboratory model are that the experimental parameters can be individually controlled to resolve the effects of each variable, and the turbulence need not be parameterised as in most mathematical models.

Although thermal infrared satellite measurements cannot directly measure the UHI effect, they can be coupled with satellite-derived measurements of vegetation density to substantially describe the contributing factors to the UHI effect. The role of vegetation in reducing amount of heat stored in the soil and surface structures due to transpiration, in contrast to relatively unvegetated urban areas, has been cited as a significant contributor to the UHI effect (Carlson et al., 1981; Goward, 1981). Vegetation indices computed from remotely sensed data have been demonstrated as useful estimators of the amount of leaf area and related variables associated
with agricultural crops as well as forests. Gallo et al. (1993) compared vegetation indices and radiant surface temperature acquired by the AVHRR with minimum air temperatures observed for urban and rural locations. The satellite-derived vegetation index data were linearly related to the difference in observed urban and rural temperatures. Given the relationship between the UHI effect, urban surface temperatures and the texture of land cover influenced by urban land use, this exercise employs a multi-sensor approach to assess the climatic implications of urbanization at the regional-scale.

Heat islands can be defined for different layers of the urban atmosphere (e.g. urban canopy layer, and urban boundary layer), and for various surfaces and the subsurfaces. Observation of atmospheric heat islands have been made using fixed sensors and traverse methodologies. Observation of surface urban heat islands (SUHI) have been made for specific surfaces with ground based instrumentation, using remote sensors mounted on satellites, or from model output. Remote observations of surface heat islands by satellite sensors provides advantages in the near simultaneous observation of spatial variations of surface temperature across the urban landscape.

Numerical models also have many limitations such as computer resources, grid resolution, turbulence closure, and initial and boundary conditions. They must be validated by comparison with laboratory or field experiments. Inspite of these limitations, the numerical models are very much useful in atmospheric modelling.

The study and analysis of the UHI at NCT of Delhi and to identify the mesoscale circulations that are responsible for transporting pollutants from higher pollution zones to lower pollution zones, meteorological data at each grid points (of 4 km x 4 km area) were generated from a meso-scale model named, ARPS with the use of upper air meteorological data and the soil and vegetation data of Delhi as model input. This approach is unique of its kind in India. The most commonly used technique to study the UHI phenomenon is to have differences in air temperatures of urban and rural stations (Landsberg, 1981; Jauregui et al., 1992).

3.3 Model Description

3.3.1 Introduction

A three dimensional nonhydrostatic model, the Advanced Regional Prediction System (ARPS) has been developed at the Center for Analysis and Prediction of Storms (CAPS) for the past several years (Droegemeier, et al., 1995; Xue, et al., 1995; CAPS, 1992). The prime
objective of CAPS is to demonstrate the feasibility of numerical weather prediction (NWP) on the storm scale. ARPS is appropriate for use on scales ranging from a few meters to hundreds of kilometres. It is based on compressible Navier-Stokes equations describing the atmospheric flow, and uses a generalized terrain-following coordinate system.

3.3.2 Model Equations

The governing equations of ARPS include the prognostic equations for Cartesian velocity components \((u, v, w)\), perturbation potential temperature \((\theta)\), perturbation pressure \((p')\), mixing ratios for water vapor \((q_v)\), cloud water \((q_c)\), rainwater \((q_r)\), cloud ice \((q_i)\), snow \((q_s)\) and hail \((q_h)\), and the subgrid scale turbulent kinetic energy \((E)\). The equation of state for moist air is also used. These equations are represented in a curvilinear coordinate system \((\xi, \eta, \zeta)\) that has a coordinate surface following the terrain at the model lower boundary. In addition to the above equations, ARPS also solves prognostic equations for the surface and deep soil temperature, surface and deep soil moisture, and the canopy moisture in a coupled two-layer soil model.

In ARPS, a special curvilinear coordinate being orthogonal in the horizontal directions is used. The transformation is defined as

\[
\xi = x, \eta = y, \zeta = (x, y, z)
\]  

(3.1)

where \(x, y\) and \(z\) are the independent variables in the Cartesian coordinate. As shown in Sharman et al. (1988), the contravariant velocities \(U^c, V^c, W^c\) can be expressed as the functions of Cartesian velocities \(u, v\) and \(w\), which in this special case are:

\[
U^c = uJ_3/\sqrt{G}, V^c = wJ_3/\sqrt{G}, W^c = (uJ_1 + vJ_2 + w)/\sqrt{G}
\]  

(3.2)

Here, the transformation Jacobians \(J_1, J_2, J_3\) and \(\sqrt{G}\) are defined as

\[
J_1 = -\frac{\partial z}{\partial \xi}, J_2 = -\frac{\partial z}{\partial \eta}, J_3 = -\frac{\partial z}{\partial \zeta}, \text{and} \quad \sqrt{G} = J_3
\]  

(3.3)

The transformation relations for spatial derivatives from Cartesian coordinate \((x,y,z)\) to the curvilinear coordinate \((\xi, \eta, \zeta)\) are

\[
\frac{\partial \phi}{\partial x} = \frac{1}{\sqrt{G}} \left[ \frac{\partial}{\partial \xi} (J_3 \phi) + \frac{\partial}{\partial \zeta} (J_1 \phi) \right], \quad \frac{\partial \phi}{\partial y} = \frac{1}{\sqrt{G}} \left[ \frac{\partial}{\partial \eta} (J_3 \phi) + \frac{\partial}{\partial \zeta} (J_2 \phi) \right], \quad \frac{\partial \phi}{\partial z} = \frac{1}{\sqrt{G}} \left[ \frac{\partial \phi}{\partial \zeta} \right]
\]  

(3.4)

For the convenience, we define

\[
\rho' = \sqrt{G} \rho, \quad u' = \rho' u, \quad v' = \rho' v, \quad w' = \rho' w, \quad W' = \rho' W^0
\]  

(3.5)
where, $\rho_1$ is the air density. In the curvilinear coordinate, the governing equations of ARPS for Cartesian momentum components ($u', v', w'$), perturbation pressure ($p'$), perturbation potential temperature ($\theta'$), mixing ratios ($q$) of water and ice species, and subgrid scale turbulent kinetic energy ($E$) become:

$$\frac{\partial u'}{\partial t} + \frac{\partial}{\partial \xi} \left( J_3 (\rho' - \alpha Diu^*) \right) + \frac{\partial}{\partial \eta} \left( j_1 (P' - \alpha Diu^*) \right) = -\text{ADV}(u) + \left[ \rho' f v - \rho' f w \right] \sqrt{GD_u} \quad (3.6)$$

$$\frac{\partial v'}{\partial t} + \frac{\partial}{\partial \eta} \left( J_3 (\rho' - \alpha Diu^*) \right) + \frac{\partial}{\partial \xi} \left( j_2 (P' - \alpha Diu^*) \right) = -\text{ADV}(v) - \rho' fu + \sqrt{GD_v} \quad (3.7)$$

$$\frac{\partial w'}{\partial t} + \frac{\partial}{\partial \zeta} \left( \rho' - \alpha Diu' \right) + \frac{\partial}{\partial \xi} \left( \frac{p' g}{\rho c_s^2} \right) = -\text{ADV}(w) + \rho' g \left[ \frac{\theta'}{\theta} + \frac{q'}{e - q_v} \right] \frac{q_v + q_l}{1 - q_v} + \rho' fu + \sqrt{GD_w} \quad (3.8)$$

$$\frac{\partial (J_3 P')}{\partial t} - J_3 \rho g w + \rho c_s^2 \left[ \frac{\partial}{\partial \xi} (J_3 u) + \frac{\partial}{\partial \eta} (j_3 v) + \frac{\partial}{\partial \zeta} (J_3 W_c) \right]$$

$$= - \left[ (J_3 u) \frac{\partial p'}{\partial \xi} + (J_3 v) \frac{\partial p'}{\partial \eta} + (J_3 W_c) \frac{\partial p'}{\partial \zeta} \right] + J_3 \rho c_s^2 \left[ \frac{1}{A} \frac{d\theta}{dt} - \frac{1}{A} \frac{dA}{dt} \right] \quad (3.9)$$

$$\frac{\partial (\rho' \theta')}{\partial t} + \left[ \rho' W \frac{\partial \theta}{\partial z} \right] = -\text{ADV}(\theta) - \sqrt{GD_\theta} \sqrt{GS_\theta} \quad (3.10)$$

$$\frac{\partial (\rho' q)}{\partial t} = -\text{ADV}(q) + \frac{\partial}{\partial \zeta} \left( \frac{\rho' V_q q}{\zeta} \right) + \sqrt{GD_q} + \sqrt{GS_q} \quad (3.11)$$

$$\frac{\partial (\rho' E)}{\partial t} = -\text{ADV}(E) + C \cdot \rho' (K_q D e f^2 - \frac{2}{3} E D i u) - \rho' \frac{C_2}{J} E^{3/2} + 2 \sqrt{GD_x} \quad (3.12)$$

where advection operator $\text{ADV}(\cdot)$ is defined as $\text{ADV}(\cdot) = u' \frac{\partial (\cdot)}{\partial \xi} + v' \frac{\partial (\cdot)}{\partial \eta} + W' \frac{\partial (\cdot)}{\partial \zeta}$.

In the above eqns., over-barred variables represent base-state atmosphere, and primed variables are deviations from the base state. In ARPS, the base state is defined as a function of the Cartesian coordinate $z$ only, it is, however, dependent on all three independent variables $\xi, \mu$ and $\zeta$ in the transformed coordinate. The base state satisfies the hydrostatic relation. In eqn.(3.9), $A = 1 + 0.61 q_r + q_k$, where $q_k$ is the total liquid and ice water.

Terms on the l.h.s. of horizontal momentum eqns. (3.6) and (3.7) are the local time tendency and perturbation pressure gradient force, and on the r.h.s. are, respectively, advection, Coriolis force, and turbulent and computational mixing terms. Here, all components of Coriolis force are included ($f = -2\Omega \sin\phi, f' = -2\Omega \cos\phi$, where $\Omega$ is the angular velocity of the earth and $\phi$ is the latitude). The contribution from vertical velocity can be as large as the horizontal
contribution for convective scale motion. Since the base-state pressure is assumed to be horizontally homogeneous, its horizontal gradient terms do not appear in the horizontal momentum equations. The absence of these terms reduces the computational error associated with terrain-following coordinates.

The vertical momentum eqn. (3.8) has, on the l.h.s., the local time tendency, pressure gradient force, and the buoyancy due to compression, and on the r.h.s., the advection, thermal and water loading buoyancy, Coriolis force and mixing terms. The total buoyancy, defined by \(-g \rho' \bar{p}\), can be derived from the eqn. of state for moist air

\[
\rho = \frac{P}{R_c T} (1 - \frac{q_v}{\varepsilon q_v})(1 \cdot q_v \cdot q_h)
\]

where \(g\) is the acceleration due to gravity, \(\varepsilon = R_c / R_T\) is the ratio of the gas constants for dry air \((R_c)\) and water vapor \((R_T)\). \(T\) is the air temperature. \(c_s = \sqrt{\gamma R_T T}\) in eqn. (3.8) is the acoustic wave speed, with \(\gamma = C_p / C_v\) being the ratio of the specific heat of air at constant pressure \((C_p)\) and constant volume \((C_v)\).

The terms involving \(\alpha \text{Div}'\) in eqns. (3.6) - (3.8) are artificial "divergence damping" terms designed to attenuate acoustic waves, where \(\text{Div}'\) is the density weighted divergence and \(\alpha\) the damping coefficient. Skamarock and Klemp (1993) showed that unstable acoustic modes due to mode-splitting time integration scheme can be effectively controlled by divergence damping. Since atmospheric flows are quasi-incompressible, the divergence damping has little adverse effect on meteorologically significant waves.

Pressure eqn. (3.9) is obtained by taking the material derivative of the eqn. of state and replacing the time derivative of density by velocity divergence using the continuity eqn. The terms in the pressure eqn. are, on the l.h.s., the base-state pressure advection and velocity divergence term, and on the r.h.s., the perturbation pressure advection and diabatic heating and moisture/water effect terms. The hydrostatic relation was used to substitute for the vertical gradient of \(\bar{p}\) in the \(\bar{p}\) advection term. In general, the divergence term is dominant for meteorological applications, and the diabatic and moisture/water effect terms are small and are, as is done in Klemp and Wilhelmson (1978), neglected in the ARPS.

In obtaining the above eqns., linearization approximations are made. The state variables that appear in the coefficients of certain terms are replaced by their base-state values. This is true for the density appearing in the pressure gradient force. These approximations are consistent

Eqn. (3.10) is a conservation equation for potential temperature. The terms in the eqns. are, from left to right, the local time tendency, vertical advection of $\theta$, $\theta$ advection, subgrid scale and computational mixing and diabatic source or sink terms. Moist and microphysical processes are included in the source/sink term and surface heat flux is handled by the subgrid scale mixing. The horizontal advection of $\bar{\theta}$ vanishes because $\bar{\theta}$ is horizontally homogeneous. The absence of this term reduces the noises associated with the calculation of horizontal gradients in transformed coordinates, similar to the treatment of gradient terms.

The conservation eqns. for the mixing ratios of water vapor $qv$, cloud water $qc$, rainwater $qr$, cloud ice $qi$, snow $qs$ and hail $qh$ are represented by eqn. (3.11) for a generic variable $q$. The r.h.s terms are, in order, advection, sedimentation, mixing and source terms. The source term $Sq$ includes all microphysical contributions. The sedimentation term represents the falling of hydrometeors ($qr$, $qs$ and $qh$) at their respective terminal speed ($V_q$). Cloud water and cloud ice are usually assumed to float with the air, therefore their flow-relative terminal velocity is zero.

The turbulent mixing process represented by $D$ in eqns. (3.6)-(3.11) has to be parameterized using a closure scheme. In ARPS, three closure options are available. They are, respectively, the first order Smagorinsky/Lilly scheme (Smagorinsky, 1963; Lilly, 1962), the 1.5 order TKE-based scheme (Klemp and Wilhelmson, 1978; Moeng, 1984) and Germano dynamic closure scheme (Germano et al., 1991; Wong, 1992; Wong and Lilly, 1994; Wong, 1994). In the present case, 1.5 order scheme is being used.

The 1.5 order scheme requires the solution of turbulent kinetic energy (TKE) eqn. (3.12). In eqn. (3.12), the terms on the right hand side are, respectively, the TKE advection, potential-kinetic energy conversion, shear production, dissipation and diffusion of turbulent kinetic energy. The potential-kinetic energy conversion term $C$ can be found in Xue et al. (1995). The coefficient in the dissipation term is related to the length scale $l$ and model grid scale $\Delta_l$ and is adjusted at the lowest grid level after Moeng (1984) and Deardorff (1980).

For the momentum eqns., $D$ is expressed in terms of Reynolds stress tensor $\tau_u$, which is in turn related to deformation tensor $D_u$:

$$D_{w l} = \frac{\partial x_{ij}}{\partial x_j} = \frac{\partial (\rho K^l_{ij} D_{ij})}{\partial x_j}$$

where index represents the Cartesian coordinate directions and the repeated lower indices denote
summation. $K^j_{af}$ is the turbulent mixing coefficient for momentum in $j$ direction.

The turbulent mixing for potential temperature and any water substance can be written in terms of their turbulent fluxes $H_j$, which is further related to their spatial gradients:

$$D_\phi = \frac{\partial H_j}{\partial x_j} = \frac{\partial}{\partial x_j} [\rho K^j_H \frac{\partial \phi}{\partial x_j}]$$  \hspace{1cm} (3.15)

where $K^j_H$ is the turbulent mixing coefficient for $\phi$ in $j$ direction. In general, $K_H$ is taken to be the same for heat, moisture or water/ice quantities and $K_H = K_{af}/P_r$ where $P_r$ is the turbulent Prandtl number.

In Smagorinsky scheme (Smagorinsky, 1963; Lilly, 1962), $K_{af}$ is given by

$$K^j_{af} = (k\Delta_j)^2 [\max(|Def|^2 - N^2/P_r.0)]^{1/2}$$  \hspace{1cm} (3.16)

where $k = 0.21$ after Deardorff (1972a, 1972b) and $|Def|$ is the magnitude of deformation. $\Delta_j$ is a measure of the grid scale. On a model grid with similar grid spacing in all directions, the turbulence is nearly isotropic, $\Delta_j = (\Delta_x \Delta_y \Delta_z)^{1/3}$ for all $j$. When the grid aspect ratio $(\Delta_x/\Delta_z)$ is large, $K^j_{af}$ can become too large for the vertical mixing, so the approximation $\Delta_j = \Delta_z = (\Delta_x \Delta_y)^{1/2}, \Delta_3 = \Delta z$ is used. This a case of anisotropic turbulence.

In the 1.5 order TKE scheme, the mixing coefficient is related to $E$ and length scale $l$:

$$K^j_{af} = 0.1 E l / 2 l_j$$  \hspace{1cm} (3.17)

For isotropic turbulence, $l$ is equal to $\Delta = (\Delta_x \Delta_y \Delta_z)^{1/3}$ in unstable or neutral case, but is further limited by length scale $l_j$ that is related to $E$ and static stability $N$ in stable case. The turbulent Prandtl number is given by $P_r = (1+2l/E\Delta)^{-1}$. For anisotropic turbulence, $J_1 = J_2 = (\Delta_x \Delta_y)^{1/2}, J_3 = \min(\Delta z l_j), \text{ and } P_r = (1+2l_j/\Delta z)^{-1}$

The Germano scheme and the detailed implementation can be found in Xue, et al. (1995). Apart from the turbulent mixing, terms $D$ in eqns. (3.6)-(3.11) also include optional second order and fourth order computational mixing terms. These terms act to damp computational noises of small scales.

### 3.3.3 Other Model Physics

Parameterization of cloud microphysics processes are important for storm scale prediction. In ARPS, the microphysics include the parameterized Kessler-type two-category liquid water
scheme and the three-category ice scheme after Lin et al. (1983). The implementation in ARPS follows that of Tao and Simpson (1993) and includes the ice-water saturation adjustment procedure of Tao et al. (1989). In addition to the explicit microphysics representations, Kuo cumulus Parameterization scheme is also available in ARPS for meso- $\alpha$ scale prediction.

A proper treatment of the surface processes including momentum, heat and moisture fluxes are equally important for the development and maintenance of storm scale weather and other types of local flow. A stability and roughness-length dependent surface flux model has been implemented in the ARPS using a modified Businger formulation (Businger, et al., 1971). An analytical procedure is used in the flux calculations for a much improved efficiency (Byun, 1990). Businger's formulation is further modified for more realistic results in highly stable or unstable environments (Deardorff, 1972). The model also distinguishes land and water surfaces for the roughness length calculations. The calculated surface heat fluxes are applied at the ground level in stable conditions, and are linearly distributed throughout the planetary boundary layer (PBL) in unstable conditions. This procedure is similar to the treatment in the PBL parameterization scheme of Blackadar (Zhang and Anthes, 1982). In ARPS, the PBL depth is predicted by a rate equation of Nieuwstadt and Tennekes (1981) for stable boundary layer and that of Gryning and Batchvarova (1990) for unstable boundary layer.

The surface flux model is coupled with a soil-vegetation model after Noilhan and Planton (1989) and Pleim and Xiu (1995). It is designed to simulate the essential processes involved in the surface-atmosphere interactions and requires data of soil and vegetation types at the land surface (Wong, et al., 1994). In the model, soil surface temperature changes as a result of the net surface energy balance among net radiation, surface sensible and latent heat fluxes, and the deep soil heat transfer. The time rate of change in volumetric soil moisture near the soil surface results from the residual of the precipitation reaching the ground, the evaporation from the ground, and the transfer of moisture between surface and deep soil layers. Separate equations describe the heat and moisture budget in deep soil and the amount of water in the canopy. A simple time-space varying radiation model is used to calculate the net surface radiation. The predicted surface soil temperature and moisture are used in the calculations of surface heat and moisture fluxes. The detailed equations of the soil-vegetation model, as well as the calculations of stability-dependent surface fluxes are described in Xue, et al. (1995).
3.3.4 Land-Surface Energy Budget and Soil-Vegetation Model:

3.3.4.1 Land-Surface Energy and Moisture Budgets:

This model is based on the soil-vegetation model developed by Noilhan and Planton (1989) and Pleim and Xiu (1995). It is designed to simulate the essential processes involved in surface-atmosphere interactions with the minimal amount of computation time and fewest parameters and complexities (Wong et al., 1994). It requires the horizontal distribution of soil texture at the land-surface. The model is based on five prognostic eqns.

\[
\frac{\partial T_s}{\partial t} = C_T (R_n - H - LE) - \frac{2\pi}{\tau} (T_s - T_2)
\]

\[
\frac{\partial T_2}{\partial t} = \frac{1}{\tau} (T_s - T_2)
\]

\[
\frac{\partial W_g}{\partial t} = \frac{C_1}{\rho_w d_1} (P_g - E_g) - \frac{C_2}{\tau} (W_g - W_{geq})
\]

\[
\frac{\partial W_r}{\partial t} = \frac{1}{\rho_w d_2} (P_g - E_g - E_r)
\]

where, eqn. (3.18) shows that the time rate of change in soil surface temperature is the residual of surface energy balance between net radiation $R_n$, surface sensible heat flux $H$, latent heat flux $LE$ and $T_s - T_2$. The soil heat transfer eqn. (3.20) shows that the time rate of change in volumetric soil moisture near the soil surface results from the residual of the precipitation rate at the ground,

\[
\frac{\partial W_r}{\partial t} = \text{veg} P - E_r
\]

and the evaporation rate from the ground, and the transfer of surface soil moisture with deep soil layer moisture. Eqns. (3.19) and (3.21) describe the heat and moisture budget in the deep soil. Eqns. (3.22) predicts the time rate of change of water $W_r$ in the canopy. The functional form of the various terms in the above set of equations are discussed as follows.
3.3.4.2 Radiation-Soil-vegetation Model

(a) Thermal coefficients

The thermal coefficients $C_T$ in eqn. (3.18) can be written as

$$C_T = \frac{1}{1 - \text{veg} \times \text{veg}} \frac{C_G}{C_v}$$

in which $\text{veg}$ is the fractional coverage of vegetation, and the thermal coefficient of vegetation is

$$C_v = 10^{-3} \text{ km}^2 \text{j}^{-1}$$

and the thermal coefficient of bare soil is

$$C_G = C_{G\text{sat}} \left( \frac{W_{\text{sat}}}{W_s} \right)^{b(2ln10)}$$

(b) Radiation flux

For surface heat balance, net radiative flux in (3.18) is given by (Wong et al., 1983)

$$R_n = R_{svw} (1 - \alpha) - \varepsilon_g \sigma T_s^4 - \varepsilon_a \sigma T_a^4$$

where $\varepsilon_g$ is the emissivity of the earth's surface, $\varepsilon_a = 0.725$ is the emissivity of the air, $\sigma = 5.67 \times 10^8$ Wm$^{-2}$ K$^{-4}$ is the Stefan-Boltzman constant, and $T_a$ is the air temperature at an atmospheric level. The total albedo $\alpha = \alpha_s + \alpha_z$, where $\alpha_s$ is the albedo at polar zenith and $\alpha_z$ the zenith angle adjustment to $\alpha$. The zenith angle adjustment is give by

$$\alpha_z = 0.01[\exp(0.003286Z^{1.5}) - 1]$$

where $Z$ is the solar zenith angle in radians and the minimum albedo with $Z = 0$ is

$$\alpha_s = 0.31 - 0.34 \frac{W_g}{W_{\text{sat}}}, \quad \frac{W_g}{W_{\text{sat}}} \leq 0.5$$

$$\alpha_s = 0.14 \quad \frac{W_g}{W_{\text{sat}}} > 0.5$$

The short-wave radiation is determined from

$$R_{sw} = \tau_{rg} \tau_{wv} S_o \left( \frac{a^2}{\gamma^2} \right) \cos Z$$
where the solar constant is \( S = 1353.0 \) Wm\(^2\), and Earth-Sun distance factor is from

\[
\frac{a^2}{r^2} = 1.000110 + 0.034221 \cos d_o + 0.001280 \sin d_o + 0.000719 \cos 2d_o + 0.000077 \sin 2d_o
\]

(3.28)

where \( d_o = 2\pi m/365 \) and \( m \) is the day number starting with 0 on Jan. 1 and ending 364 on Dec. 31. The solar zenith angle \( Z \) is defined by

\[
\cos Z = \sin \Phi \sin \delta + \cos \Phi \cos \delta \cos (h_r)
\]

(3.29)

where \( \Phi \) is the latitude is the solar declination:

\[
\delta = \frac{(23.5\pi)}{180} \cos \left[ 2\pi (J_d - 173)/365 \right]
\]

(3.30)

where \( J_d = m + 1 \) is Julian day.

The solar hour angle is defined by:

\[
h_r = \frac{15\pi}{180} (t_{GMT} - \lambda/15^\circ + E_r - 12)
\]

(3.31)

where \( E_r = 0.158 \sin \left[ \pi (J_d + 10)/91.25 \right] + 0.125 \sin \left[ \pi J_d/182.5 \right] \),

(3.32)

t_{GMT} is Greenwich Meridian Time, and \( \lambda \) is west longitude (in degrees).

To account for the attenuation by Rayleigh scattering and absorption by permanent gases for solar radiation, the transmission function in (3.27) has the form (Atwater and Ball, 1981)

\[
\tau_{rg} = 1.021 - 0.084 [m_{dirf}(949X10^{-8}P + 0.051)]^{1/2}
\]

(3.33)

where \( P \) is the surface pressure (kPa) \( m_{dirf} \) is a directional factor that is equivalent to air mass at pressure of 101.3 kPa and follows a formulation given by:

\[
m_{dirf} = \frac{35}{(1224 \cos^2 Z + 1)^{1/2}}
\]

(3.34)

The water vapour transmittance in eqn. (3.27) can be written as

\[
\tau_{wv} = 1 - \frac{2.9 \mu M_{dirf}}{(1 + 141.5 \mu M_{dirf})^{0.635} - 5.925 \mu M_{dirf}}
\]

(3.35)

where \( M_{dirf} = m_{dirf} \) above any clouds and \( M_{dirf} = 5/3 \) below and within cloud layers. The path length \( \mu \) at level \( P \) is computed from
\[
\mu = \frac{1}{g} |q_v \left( \frac{P}{101300} \right)^{273.16} T^{1/2} dp
\] (3.36)

(c) Sensible heat flux

The sensible heat flux
\[
H = \rho_a c_p C_{dh} V_a (T_s - T_a)
\] (3.37)
where \( c_p \) is the specific heat at constant pressure; \( \rho_a \) and \( V_a \) are, respectively, the air density and wind speed at an atmospheric level; \( C_{dh} \) is the exchange coefficient depending upon the thermal stability and roughness.

(d) Latent heat flux

The latent heat flux is the sum of the evaporation from the soil surface \( E_g \), transpiration \( E_{tr} \), and evaporation from wet parts of the canopy \( E_r \):
\[
LE = L(E_g + E_{tr} + E_r)
\] (3.38)
in which \( L \) is the latent heat of vaporization and
\[
E_g = (1 - \text{veg}) \rho_a C_{aq} V_a \left[ h_u q_{\text{vsat}}(T_s) - q_{va} \right]
\] (3.39)
where the relative humidity at the ground surface is
\[
h_u = 0.5[1 - \cos(\pi W_g/W_{fl})], \quad W_g < W_{fl}
\]
\[
= 1, \quad \text{otherwise}
\] (3.40)
with field capacity \( W_{fl} = 0.75 \ W_{sat} \). The saturation mixing ratio \( q_{\text{vsat}} \) is calculated using Teten's formula given in eqn. (3.36).

In eqns. (3.38) - (3.39),
\[
E_{tr} = \text{veg} \rho_a \frac{1 - F_w}{R_a + R_s} [q_{\text{vsat}}(T_s) - q_{va}]
\] (3.41)
\[
E_r = \text{veg} \rho_a \frac{F_w}{R_a} [q_{\text{vsat}}(T_s) - q_{va}]
\] (3.42)
in which the wet fraction of the canopy, \( F_w \), is defined as
\[
F_w = \left( \frac{W_r}{W_{\text{rmax}}} \right)^{2/3}
\] (3.43)
and \( W_{\text{rmax}} = 0.2 \ \text{veg LAI (mm)}. \) (3.44)
Here LAI is the leaf area index of vegetation and it depends on the vegetation type. The aerodynamic resistance is parameterized by

\[ R_a = \frac{1}{C_{dQ} V_a} \]  

(3.45)

The surface resistance for evapotranspiration is computed as

\[ R_s = \frac{R_{\text{sm}}}{LAI F_1 F_2 F_3 F_4} \]  

(3.46)

in which

\[ F_1 = \frac{R_{\text{sm}} / R_{\text{smax}}}{1 + f} \]  

(3.47)

with, \[ f = 0.55 \frac{R_G}{R_{GL}} \frac{2}{LAI} \]  

(3.48)

where \( R_{\text{smax}} = 5,000 \text{ s/m} \), \( R_G = R_{SW} \), and \( R_{GL} \) depends on the vegetation type.

\[ F_2 = \begin{cases} 1 & W_2 > W_{fl} \\ (W_2 - W_{wilt}) / (W_{fl} - W_{wilt}), & W_{wilt} \leq W_2 \leq W_{fl} \\ 0, & W_2 < W_{wilt} \end{cases} \]  

(3.49)

\[ F_3 = 1 - 0.06(q_{\text{vsa}}(T_a) - q_{va}), \quad q_{\text{vsa}}(T_a) - q_{va} \leq 12.5 \text{ g/kg} \]

\[ F_3 = 0.25, \quad \text{otherwise} \]  

(3.50)

\[ F_4 = 1 - 0.0016(298.0 - T_a)^2 \]  

(3.51)

e) Soil Surface moisture

In eqn. (3.20), the surface volumetric moisture \( W_{geq} \) when gravity balances the capillary force is computed according to

\[ W_{geq} = \alpha R \alpha^p (1 - x 8^p) \]  

(3.52)
\[
\begin{align*}
    x &= \frac{W_2}{W_{sat}} \\
\end{align*}
\]  

(3.53)

In (3.20), the coefficients are given by

\[
\begin{align*}
    C_1 &= C_{1, sat} \left( \frac{W_{sat}}{W_g} \right)^{b-1} \\
    C_2 &= C_{2, ref} \left[ \frac{W_2}{W_{sat} - W_2 + W_1} \right]
\end{align*}
\]  

(3.54)

(3.55)

where \( W_1 \) is a small numerical value that limits \( C_2 \) at saturation. The parameters \( C_{1, sat} \), \( C_{2, ref} \), \( b \), and \( p \) are soil-texture dependent.

3.3.5 Numerical Treatment

The continuous equations described earlier are solved using finite difference method on an Arakawa C-grid. The mode-splitting time integration technique of Klemp and Wilhelmson (1978) is employed. The large time-step integration uses leapfrog time differencing scheme. The advection terms use fourth-order-centered differencing while the other terms use second-order differencing. The small time step integration uses Crank-Nicolson scheme which solves the \( w \) and \( p \) equations implicitly in the vertical direction. Furthermore, ARPS has the option for evaluating the terms responsible for internal gravity waves in small time steps, which removes the limitation on the large time step size by static stability, thereby allowing bigger large time step size whenever possible. In eqns.(3.6)-(3.10), all terms on the left hand side are evaluated in the small time steps. Rigid wall boundary condition is used in this case.

3.4 Model Domain, Input data for simulation of ARPS

As stated earlier the study area is NCT of Delhi, sprawling over 1,438 sq. km. Between the latitude of 28° 24' and 28° 53' N and the longitude of 76° 50' and 77° 20' E. The model domain for simulation of ARPS is slightly larger one enclosing NCT of Delhi and rectangular in shape having 11x12 grid points, each grid size being 4x4 sq. km.

The model simulation requires two types of data. They are:

a. The sounding data for NCT of Delhi: The upper air meteorological data used for simulation are pressure (pa), temperature (K), relative humidity (non-dimensional), wind direction (degree), & wind speed (m/s).
b. Soil and Vegetation type data for NCT of Delhi: With the help of soil texture map of Delhi developed by NBSS & LUP and as per USDA soil type classification used by ARPS (Table 3.1, APPENDIX - 3A), the soil texture type for each grid of the domain is prepared. However, as per the above classification Delhi has got seven predominant soil classes: sand, sandy loam, sandy clay loam, loamy sand, silty clay loam, clay and loam. With the help of Survey of India topo-map for Delhi and its neighbourhood and the vegetation type classes used by ARPS (Table 3.2, APPENDIX - 3B), the entire domain of study is divided into different vegetation type classes. The leaf area index (LAI) and surface roughness at each grid point are also taken into account in the simulation.

3.5 Experiment

The model was run for 11-12th of January, April, August and October for the years 1991 to 1995. Since there was no major disturbance during these months over the years, these days were assumed to represent the Winter, Summer, Monsoon and Post-monsoon seasons respectively. In order to study the spatial variation of potential temperature, specific humidity, wind fields, etc., the soil texture and vegetation type, surface roughness and LAI values are incorporated at each grid point of the domain to the Land-Surface Energy Budget and Soil-Vegetation Model (included in the ARPS model). From the simulation results, the regions of warm pockets, cold pools, the humidity fields and the wind fields are plotted.

Since, the pattern of different fields are more or less the same for corresponding seasons over the years, the simulated results for 1995 only are discussed here.
3.6 Results and Discussion

Fig. 3.1 (a) - 3.4 (a) show the spatial variation of potential temperature fields indicating the warmer and cooler regions for different seasons of '95. From the figure 3.1 (a), it is seen that the regions of cold pockets appear towards the south-east part of the domain and the temperature increases gradually towards north. Towards north-east part of the domain, there appear warm pockets which correspond to urban center of Delhi. Maximum temperature appears in the north-western corner which being outside Delhi, is not discussed in the present study.

Magnitude wise, the simulated result for heat island intensity as well as the urban-rural temperature difference are low over the study region (magnitude of heat island intensity is ~ 1.0 K), whereas the observed heat island intensity is of the order of 2 to 3 K. This low value of the simulated result could be mainly due to two reasons: (a) the terrain feature of Delhi was not taken into consideration in the model simulation and (b) the rise of surface temperature due to pollution was also not taken into account. However, though magnitude wise, the observed and simulated heat island intensity differs, the regions of warm and cool pockets more or less are coinciding with the simulated positions, thus suggesting that the model is capable of simulating the heat island effect fairly over the region.

In all months discussed here, the location of warm and cool pools appear more or less in the same region, with slight shift northwards.

The urban-rural temperature anomaly is likely to generate a local circulation and this circulation if associated with the urban air pollution, act as sources and sinks of pollution. This local circulation may create stagnating conditions by not allowing the pollutants to leave the grid complex.

Fig. 3.1 (b) - 3.4 (b) show the spatial variation of humidity fields. From figure 3.1 (b) to 3.4 (b), it is seen that the humidity field has maximum intensity near water-body and in the vegetation area (in south-eastern and eastern part) and goes on decreasing towards the north (i.e. the city centre). Thus the spatial variation of humidity field show an inverse relationship with the potential temperature field.

Humidity islands play a key role in determining the residence times of pollutants, their rate of deposition and chemical action on the exposed objects.
Figs. 3.1 (c) & (d) to 3.4 (c) & (d) show the spatial variation of wind speed and horizontal wind field respectively. The wind field is influenced by topography, warm and cold pockets. The winds accelerated from south (cold pool) towards north (warm pocket) in January. Thus whichever grid is observed the wind direct towards warm pocket. In April the winds which are relatively weak converged towards the warm pocket in the north and diverged from the cold pool in the south. In the dense built up area winds further weakened (in the west). In the monsoon season, winds are strong and is generally westerly/easterly depending upon the location of monsoon trough. In the present case the strong westerly winds turned north-westerly/northerly probably the Coriolis force assisted in deviating the wind direction. In the post-monsoon season winds are strong south-westerly deviating towards east. Here too the winds originally blowing towards heat island in the north turned east most likely due to Coriolis effect.

Fig. 3.1 (e) - 3.4 (e) show the diurnal variation of model simulated potential temperature at warm and cold grid points. The diurnal variation shows an increase of potential temperature from 00 GMT reaching its maximum value at 11 GMT i.e. one-two hour after sunset, thereafter the potential temperature decreases slowly. The trend in potential temperature after midnight however is not in accordance with that of the observed one (discussed in the next figure). This could be the radiative characteristics of stratified pollution layers lying above the surface of the earth, the effect of which could not be accounted for in the model used.

Fig. 3.1 (f) - 3.4 (f) show the diurnal variation of observed potential temperature at a grid point in the vicinity of Safdarjung meteorological station. The variation of potential temperature from 00 GMT to 11 GMT show a constant increase attaining maximum value two hour after sunset and there after decreases rapidly.
Fig. 3.1 Numerical Simulation at 11 UTC of Jan. '95 for Spatial Variation of (a) pot. temp., (b) sp. humidity, (c) wind speed, (d) horizontal wind vector, & Diurnal Variation of pot.temp at(e) simulated(9,9),(9,3) & (f)observed(7,6) gd. pt.
Fig. 3.2  Numerical Simulation at 12 UTC of Apr. '95 for Spatial Variation of (a) pot. temp., (b) sp. humidity, (c) wind speed, (d) horizontal wind vector, & Diurnal Variation of pot.temp at(e) simulated(9,8),(9,5) & (f)observed(8,7)gd. pt.
Fig. 3.3  Numerical Simulation at 11 UTC of Aug. '95 for Spatial Variation of (a) pot. temp., (b) sp. humidity, (c) wind speed, (d) horizontal wind vector, & Diurnal Variation of pot.temp at(e) simulated(8,8),(5,3) & (f) observed(8,7) gd. pt.
Fig. 3.4 Numerical Simulation at 11 UTC of Oct. '95 for Spatial Variation of (a) pot. temp., (b) sp. humidity, (c) wind speed, (d) horizontal wind vector, & Diurnal Variation of pot.temp at (e) simulated(9,9),(9,5) & (f) observed(8,7) gd. pt.
3.7 Conclusion

i. For most part, the trends divide themselves nicely into two classes: significant warming of urban areas, and relatively less warming, or even slight cooling at the non-urban/rural sites.

ii. The UHI initially increases to a maximum after sunset before it first decreases rather fast, but from midnight until the morning with a lower rate.

iii. A more specific objective is to determine if the meteorological stations can be characterized as "urban" or "rural" based upon their climatologically significant surface characteristics. An objective method for determining whether a station is urban or rural would be beneficial for assessment of global climate change, as the influence of urban stations could be extracted for future analyses of temperature trends.

In the next chapter PBL layer, mixing height determination and model description to calculate mixing height, stability, Monin-Obukhov length, etc. are discussed.