Physical Parameterizations in Canadian Operational Models

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The main objective of this presentation is to describe and discuss the main parameterizations of physical processes in the Canadian meteorological models. In addition of reviewing the main ideas behind these physical representations, we will examine their main weaknesses and reveal the short and long-term plans for their improvement.

Starting from data acquisition all over the world, the road leading to the analysis of the large variety of numerical weather prediction (NWP) products by expert forecasters is long and complicated (although it has to proceed rapidly). As shown by the simple diagram in Fig. 1, integrations of data assimilation schemes and numerical models (global and regional) are of course key components of the NWP cycle. Plenty of effort is given to improve these two aspects. In this presentation, we will concentrate on the numerical modelling aspect of NWP, and in particular on the representation of physical processes (e.g., precipitation, boundary layer, ...).

At the Canadian Meteorological Center (CMC) and the Recherche en prévision numérique (RPN) division, a few atmospheric models are available. The Global Environmental Multi-scale (GEM) model is currently used for operational regional forecasts and is planned to replace very soon the Spectral Finite-Element (SEF) model for the global operational runs. The Mesoscale Compressible Community (MC2) model is often used for research applications that require high-resolution on a limited area. Another model, finally, is the old regional operational model, that is, the Regional Finite-Element (RFE) model, which is use much less frequently since the advent of the GEM model.

As an example, the dynamical equations integrated in GEM are shown in Fig. 2. Although these equations may not be easy to recognize in their η-coordinate format (η is the vertical coordinate used in GEM), they are, respectively, the momentum, continuity, thermodynamic, humidity, and hydrostatic equations. Our purpose here is not to describe each of these equations (this has been discussed in the introduction to NWP part of the course), but rather to point out the ***F*** terms on the right hand side, which are forcing terms for the physical processes.

As is evidenced in these equations, the role of the various physical mechanisms in the atmosphere are treated in GEM (as in all the other models mentioned above) as *forcing terms separated from the dynamical processes* (all on the left hand side). This *process-splitting*, described schematically in Fig. 3, allow the model developers at RPN to treat the computer code for the dynamical and physical processes as *separate entities*. The role of the physical library is to evaluate the tendencies on temperature, humidity, and horizontal winds, and transmit them, through an elaborate interface, to the dynamical core. This interface is common to all the models at CMC/RPN, so that the same physical package can be used with any model (see Fig. 4). The physics, thus, must be able to represent a wide range of processes, and must be flexible enough so that it can be useful in models in which the horizontal resolution can vary from 100 km (global models) to 1 km and even less (MC2).

The rest of the presentation is structured as follows:

- Overview of physical processes;
- Atmospheric radiation;
- Surface processes;
- Turbulence and vertical diffusion;
- Clouds and precipitation

A) Overview

Figure 5, in addition of summarizing the main physical processes occurring in the atmosphere, gives a measure of the diversity and complexity of the processes that have to be represented (or parameterized) in the atmospheric models.

First, the *solar and infrared radiation* (section B). The sun is the main source of energy for all the atmospheric dynamics over the globe, and it is of course essential to consider in an appropriate manner the effects of the incoming solar radiation in order to represent adequately the diurnal cycle. The temperature tendencies resulting from the absorption, reflection, and scattering of solar and infrared radiation will also be discussed, as well as the role of clouds.

Second, the *surface processes* (section C). The exchanges of heat, moisture, and momentum between the surface and the atmosphere constitute an important contribution on the characteristics and structure of the low-level atmosphere (basically the planetary boundary layer – PBL). They depend on the energy budget at the surface, or, in other words, on the partition of the available energy at the surface (incoming minus reflected solar and infrared radiation) into sensible heat flux, latent heat flux, and ground heat flux. As will be discussed, the treatment of soil water (evaporation, drainage, and runoff) is critical to represent adequately the surface energy budget.

Third, the *boundary-layer turbulence* (section D). Atmospheric turbulence is generated (mainly in the layer near the surface) mechanically because of the intense wind shear due to surface friction and thermally because of the sensible heat flux. The main role of this low-level turbulence is to mix the atmospheric properties (including chemical components) in the PBL.

Fourth, *condensation processes* (section E). Whether or not it will rain (snow) or not is one the most important question we all have to answer. In atmospheric models, the mechanisms leading to deep convective activity, mainly responsible for severe weather events organized on the mesoscale (i.e., 20-2000 km) are parameterized using different strategies which will be described briefly. The condensation that occurs on the grid-scale (i.e., resolved by the model) are also represented. At the current horizontal resolution of the operational models, the condensation processes are always parameterized using a *coupled* set of physical schemes: one for the unresolved deep convective clouds (implicit), and one for the resolved more stratiform type clouds (explicit).

(Another physical aspect, of lesser importance, is the impact of unresolved gravity waves generated by subgrid-scale topography variance. The drag on the grid-scale circulations resulting from these unresolved gravity waves are considered in the models using what is appropriately called a *gravity wave drag* scheme. In order to spend more time on the other aspects of the RPN physics package – which are thought to be of greater importance, this drag won't be discussed in this presentation. For more information on how this effect is included in the models, the reader is referred to the documentation of the RPN physics package.)

B) Atmospheric radiation

As an attempt to review briefly the radiative processes occurring in the atmosphere, we tried to include in Fig. 6 the main concepts necessary to understand the physical parameterizations for radiation.

One should first note from this figure that two different "types" of radiation are represented, i.e., the "solar" (from the sun) and "infrared" (from the Earth and its atmosphere) radiation. Because the temperature of these two emitters are so far apart (T_{Earth} ~ 255 K, whereas T_{sun} ~ 5780 K), the wavelengths at which the intensity (wave energy per spectral unit) of the emitted radiation is maximum are quite different. Based on Wien's displacement law for blackbody radiation, the "warmer" (sun) body will have its maximum intensity at a shorter wavelength than the "colder" (Earth and atmosphere) body. Thus, the solar and infrared radiation depicted in Fig. 6 are not really different in nature (this would have been something !) but only have their maximum intensities at different wavelengths. This fact is showed explicitly in Fig. 7, in which the normalized intensity (i.e., the intensity multiplied by the wavelength) versus the wavelength is plotted. The curves for solar and infrared radiation are quite distinct, with almost no overlap. Because the interactions of radiation with the atmosphere (absorption, reflection, scattering) depend so much on the frequency of the electromagnetic (EOM) wave, distinct physical schemes are used for solar and infrared radiation.

Obviously, one major effect of radiation is, especially for the shortwaves, to warm the Earth surface. This surface warming is responsible for sensible heat fluxes at the surface, which lead to thermal destabilization of the low levels of the atmosphere that in turn results in the generation of boundary-layer turbulence (see the next two sections). But, before reaching the ground (for the downward waves), or exit the Earth's atmosphere (for the upward waves), the EOM waves have to travel over large distances in the atmosphere. The signal is thus attenuated due to absorption and scattering from gases such as water vapor (H_20), ozone (O_3), and carbonic gas (CO_2). Interactions with clouds and aerosols are also important.

Very simply, the heating/cooling tendencies H_{rad} associated with radiation in the atmosphere are given by the flux convergence:

$$H_{rad} = \left(\frac{\partial T}{\partial t} \right)_{rad} = \frac{g}{c_p} \frac{\partial F(z)}{\partial z} \tag{1}$$

where

$$F(z) = F \downarrow (z) - F \uparrow (z) \tag{2}$$

is the net radiative flux, including the effect of both solar and infrared radiation. The task of the radiation schemes is thus simply to evaluate at each model level the solar and infrared radiative fluxes.

It is worthwhile to give, without of course going into the fine details of the schemes, the main characteristics of the solar and infrared radiation schemes currently used in the operational models.

For the ***solar radiation,*** the scheme is based on the work of Fouquart and Bonnel. Its main characteristics are:

- Faster version of only one spectral interval
- Effects of water vapor, carbonic gas, ozone, and clouds are evaluated
- Rayleigh diffusion and multiple scattering considered
- Absorption by cloud liquid water in clouds
- Aerosols could be taken into account, but are not in the current version

For the ***infrared radiation***, the scheme was developed by Garand, one of us here at RPN. Its main characteristics are:

- Effects of water vapor, carbonic gas, ozone, and clouds are considered
- Broad-band model
- 4 spectral bands in effect (one for ozone, two for carbonic gas, and one for the rest, i.e., water vapor)
- Continuum effect for water vapor (225 bands at 10 cm^{-1} resolution)
- Strong line approximation for the broad bands (absorber mass is band-independent)
- Climatological values for ozone
- Carbonic gas concentrations are constant throughout the atmosphere

In order to reduce the computing cost associated with radiation calculations (which are pretty long – for both solar and infrared), the radiation packages are not called at every time steps, but once every 2 hours or so (in the operational model at least). For the infrared scheme, the temperature tendencies are kept constant in the interval between two successive calls of the scheme, whereas for the solar scheme, the fluxes and heating rates are modulated by the cosine of the sun's angle.

As the readers may have already guess, one of the main difficulty for adequately representing radiation effects in the atmosphere is more related to the application of computationnally viable physical numerical schemes. So-called high-resolution line-bands models (which include interactions of very-fine spectral bands with each of the main atmospheric absorbents) do exist and produce good results. But unfortunately these models are not applicable in atmospheric models of the type used here, due to their enormous computational cost.

Another difficulty is to have good knowledge of the absorbent's concentrations in each of the model layers. Right now, these concentrations are crudely determined (from tables for ozone, constant for CO_2) and could probably be improved in the future. This work is important, since there is not much hope of significantly improving the treatment of radiation processes (scattering, absorption, etc.) in our models if the concentrations of the main absorbents in the atmosphere are not known more accurately. Note that the same problem exists for clouds, for which the fraction coverage and effective cloud liquid water content are difficult to predict.

In summary, here's a list of probable future developments concerning the radiation schemes in the RPN physics package:

- More spectral bands in the visible (Barker scheme)
- Better interaction of both infrared and visible with clouds
- Better knowledge of absorbents concentrations (e.g., O_3, CO_2)

C) Surface processes

It may appear odd to emphasize the importance of well representing the surface processes in atmospheric models. Indeed, our main concern in weather forecasting is to predict the evolution of the atmosphere (rain, winds, temperature...). But, as we will see in the following, the exchanges between the surface (which includes the ground, water, vegetation, snow, and ice) and the atmosphere can have, in some cases, a significant impact on the state of the low-level atmosphere (PBL). It may thus be important to represent the physical processes at the surface as well as possible, since after all most human activity is confined to this low-level layer of the atmosphere.

The main role of the surface in atmospheric models is to provide the lower boundary conditions. For the models used at CMC/RPN, this translates into furnishing the lower conditions for the turbulent vertical diffusion (see section D). Thus the only way the models feel the effect of the surface (except for the gravity wave drag which won't be discussed here) is through turbulent exchanges of heat, moisture, and momentum at low levels, which are mostly determined by the surface fluxes of these quantities. The surface fluxes can have a crucial impact on the evolution of certain weather systems. Consider for example the case of the vertical destabilization and moistening of a summertime daytime PBL (due to intense surface fluxes of sensible heat and water vapor) ahead of an advancing cold front: this clearly helps the propagation and triggering of deep convective activity (see Fig. 8).

The role of the surface will surely become more prominent in the next few years, as "environmental" predictions involving the coupling of atmospheric and hydrological models will be attempted.

In the current versions of the operational models, the surface characteristics evolve according to the simple force-restore scheme. In the original (and more typical) versions of this scheme, the temperature and soil water content of two soil reservoirs (one superficial – 10 cm – and one including a deeper soil layer – 1-2 m) changes as a result of forcing (e.g., positive net solar radiation, precipitation, ...) and restoring (relaxing terms towards deep soil or superficial values) terms. In our version of the force-restore scheme, however, only the temperature of the superficial soil layer evolves:

$$\frac{\partial T_{surf}}{\partial t} = \frac{1}{C_{surf}}\left(\frac{4\pi}{k_{surf}\tau}\right)^{1/2}\left[R_n - H - LE\right] - \frac{2\pi}{\tau}\left(T_{surf} - T_{deep}\right) \tag{3}$$

while the deep temperature T_{deep} and the soil water content of the superficial and deep reservoirs are both kept constant. In the above equation, R_n is the net incident radiation on the surface:

$$R_n = (1-\alpha)F_{SS}^{-} + \varepsilon\left(F_{IS}^{-} - \sigma_{SB}T_{surf}^{4}\right) \tag{4}$$

T_{surf} is the surface temperature, C_{surf} is the heat capacity of the surface, k_{surf} is the heat conductivity of the surface, τ is a time constant of one day, H and LE are the sensible and latent heat fluxes at the surface, F_{SS}^{-} and F_{IS}^{-} are the downward solar and infrared radiation fluxes at the surface.

Obviously, this surface scheme is not very sophisticated (it is an overly simplified version of an already uncomplicated scheme). The soil water contents do not evolve with time (it remains constant even during periods of precipitation, drying, melting, etc.). There is no interception of rainfall by the vegetation, which is treated in a very simple manner in this scheme. There is no runoff, no drainage, which makes it impossible to connect with an hydrological model. Without a doubt, the treatment of surface processes is one of the weakest parts of the RPN physics package.

For this reason, it is proposed to improve this aspect of the physics package by replacing, probably in the next two years, the force-restore scheme by a more physically sound treatment of the surface. Two other surface schemes are available at RPN: the Canadian LAnd Surface Scheme (CLASS) developed by Diana Verseghy of the climatological branch of MRB, and the Interactions Soil-Biosphere-Atmosphere (ISBA) scheme developed by Noilhan and Planton of Meteo-France.

CLASS is certainly the most sophisticated of the two schemes, with three soil layers, layered vegetation canopy, 4 types of vegetation, interaction of radiation and vegetation canopy, vertical diffusion of heat and moisture between the soil layers, complex treatment of snow, inclusion of infiltration, runoff, and drainage.

ISBA, on the other hand, is a more advanced version of a force-restore type scheme (but this time including all the equations of the force-restore) including more physically-correct treatment of vegetation and snow. The temperature and soil water content of two soil layers (one superficial and a deeper one) evolve according to the FR equations. There is interception of rainfall by the vegetation, surface runoff, infiltration, and deep soil drainage. Finally, the treatment of snow in ISBA is a simplification of that coded in CLASS.

It is not clear yet which one of these two schemes is more suitable to be included in future versions of the operational models. While CLASS is unequivocally the more sophisticated of the two, it is not certain, first, that this scheme produces better results than ISBA, and, second, that the additional computing cost resulting from the use of CLASS is worth the improvement it brings. Other factors have to be considered, like for example the set-up of an initialization scheme for the soil water content (a crucial initial variable that determines the magnitude of the surface fluxes). It is possible that the construction of such an initialization scheme is easier with one of the two schemes.

After having calculated the surface characteristics using the force-restore scheme, the kinematic fluxes of heat, moisture, and momentum, used as lower boundary conditions for the implicit vertical diffusion, are:

$$\left(\overline{w'\theta'}\right)_{surf} = -C_T\, u_*\left(\theta_a - \theta_{surf}\right) \tag{5}$$

$$\left(\overline{w'q'}\right)_{surf} = -C_T\, u_*\left(q_a - q_{surf}\right) \tag{6}$$

$$\left(\overline{w'V'}\right)_{surf} = -C_M\, u_* V_a \tag{7}$$

where the "a" indices are for the anemometer level (first model level), u_* is the friction velocity, C_T and C_M are integrated transfer coefficients for the surface layer that depend on the vertical stability and wind shear (Richardson's number). The fluxes thus depends on both the differences between the surface and the first model level, and on the transfer coefficient in the surface layer.

Finally, the fluxes over water are calculated in the same manner, except that the surface temperature is kept constant. The surface roughness is recalculated every time step using the Charnock's relation.

D) Boundary-layer turbulence

The impact of atmospheric turbulence is mainly felt in the lowest levels in response to surface fluxes of heat (thermally induced) and momentum (mechanically induced). During the day, the main role of turbulence is to "mix" the lowest levels; as a result, the conservative (in dry conditions) variables like potential temperature (θ), specific humidity (q), and wind (V) are usually almost constant throughout the whole PBL (see Fig. 9). The effect of turbulence is less important during the night (it is very weak), when the low atmosphere is very stable because of radiational cooling near the surface.

Well representing the vertical mixing evolution of the boundary layer due to turbulence is valuable not only for weather forecasting, but also for the coupling of atmospheric models with chemical transport models. The concentrations of the chemical materials are greatly influenced by the vertical diffusion caused by turbulence.

In the RPN physics package, the atmospheric turbulence is associated to a vertical diffusion mechanism. The diffusive tendencies on the conservative variables are thus written as follows:

$$\frac{\partial \theta}{\partial t} = \frac{1}{\rho}\frac{\partial}{\partial z}\left[\rho K_T \left(\frac{\partial \theta}{\partial z}\right)\right] \qquad (8)$$

$$\frac{\partial q}{\partial t} = \frac{1}{\rho}\frac{\partial}{\partial z}\left[\rho K_T \left(\frac{\partial q}{\partial z}\right)\right] \qquad (9)$$

$$\frac{\partial V}{\partial t} = \frac{1}{\rho}\frac{\partial}{\partial z}\left[\rho K_M \left(\frac{\partial V}{\partial z}\right)\right] \qquad (10)$$

The problem of parameterizing turbulence processes is then reduced to specify correctly the vertical diffusion coefficients K_T and K_M. The magnitude of the mixing from turbulence depends on these coefficients, as well as on the vertical gradients. Note that all the turbulent fluxes are in the gradient direction (no countergradient fluxes in the current version of the scheme).

The diffusion coefficients are evaluated from a turbulent mixing length scale (λ) and from the turbulent kinetic energy (*E*) following:

$$K_M = \Pr K_T = c\lambda \, E^{1/2} \tag{11}$$

in which *Pr* is the Prandtl number and *c* is a parameter.

The mixing length λ is very simply determined as the ratio of a mixing length scale in neutral conditions (λ_n) over a stability function (ϕ_n) depending on the Richardson's number:

$$\lambda = \frac{\lambda_n}{\phi_n} \tag{12}$$

It should be noted that the calculations of this mixing length is *local*, i.e., it depends only on model variables and their gradients at the same level.

In our scheme, the turbulent kinetic energy $E = 1/2\left(\overline{u'^2} + \overline{v'^2} + \overline{w'^2}\right)$ is a prognostic variable evolving according to:

$$\frac{dE}{dt} = \underbrace{B\,E^{1/2}}_{\substack{\textit{Generation}\\ (\textit{Thermal and}\\ \textit{Mechanical})}} - \underbrace{C\,E^{3/2}}_{\textit{Dissipation}} + \underbrace{\frac{\partial}{\partial z}\left(K_M \frac{\partial E}{\partial z}\right)}_{\text{Redistribution}} \tag{13}$$

This way, the physical mechanisms for the generation (both thermal and mechanical), dissipation, and redistribution of turbulent kinetic energy are allowed to influence [as the square root – see Eq. (11)] the diffusivity coefficients.

Maybe the main weaknesses of this scheme concern its "local" nature which is not appropriate for simulating the rapid mixing and evolution of daytime boundary layer, and its "diffused" variables (i.e., θ and *q*), which are not conserved when condensation occurs. Work is currently in progress to improve these two aspects.

E) Clouds and precipitation processes

Predicting the onset, location, type, and intensity of precipitation is one of the most difficult problem in weather forecasting. From a fundamental point of view, there is the problem of the predictability of precipitation systems, which depends on the quality of the initial conditions (low-level humidity tongues, mesoscale convergenge zones, ...), and whether the precipitation systems are developing from downscale (cold front organizing deep convective activity) or upscale (random convective cells organizing themselves into a mesoscale convective complex – summertime case) cascades. More practically, the problem of triggering and simulating physical processes associated with deep convective activity that are still not well understood represents a great challenge. Finally, scales interactions between meso-γ-scale (i.e., 2-20 km), and meso-β-scale circulations obscure even more the situation.

Before describing with some details the different cloud schemes used in the Canadian operational models, it is worthwhile explaining the essential distinctions between subgrid-scale (i.e., unresolved, parameterized) and grid-scale (resolved) precipitation processes. Figure 10 and figure 11 attempt to clarify this aspect. In Fig. 10, the precipitation region associated with a synoptic-scale weather system is sketched. Deep convective cells (~ 1-2 km) are triggered along the advancing cold front forming a convective line (~ 10 km). Clearly, this convective activity can not be resolved by mesoscale models (with horizontal resolution of about 20 km) and thus needs to be parameterized. In this case, it is not required that the model grid boxes attain saturation before producing precipitation. The issue here is thus to represent this cloud activity (unresolved) implicitly based on the model grid-scale variables. Adding to the difficulty of the problem, this parameterization is found to depend on the resolution of the model (some parameterizations scheme developed for large-scale models may not be adequate for mesoscale models).

For the rear part of the system, the precipitation is more organized on a larger-scale (few hundreds of km, see Fig. 10), with less intense vertical motion. More “stratiform” type of precipitation is generated in this region. This precipitation region is certainly resolved by the model, and thus the condensation processes taking place there should only occur when the model grid boxes are saturated. This type of simulated precipitation is called "explicit" (or resolved, grid-scale) and the required parameterizations are not based on model columns convective processes such as updrafts/downdrafts, mass convergence, etc, but rather on model grid boxes in which cloud microphysics (condensation, evaporation, sublimation, etc.) are represented.

Figure 11 shows another type of situation, a squall line summertime system, with the same conclusions. In this case, the leading line of deep convective activity is not resolved by the model (grid boxes are shown), while the trailing stratiform precipitation region at the rear of the system is resolved by the model. Ideally, the leading part of the system should only be dealt with by the convective parameterization (implicit scheme), whereas the trailing stratiform portion of the system should only be represented by the grid-scale explicit scheme.

Convective schemes

For many years now, the Kuo type of convective scheme is used in both the regional and global operational models. According to this method, cumulus convection exists only in the presence of deep layers with convective instability and is driven by large-scale convergence of moisture and surface evaporation. More precisely, the convective scheme vertically redistributes water (available from large-scale convergence of humidity) and release latent heat due to condensation based on a partition of the available water into humidifying and precipitation parts. (A schematic representation of the Kuo scheme is depicted in Fig. 12.)

A convective layer is defined as a conditionally unstable layer in which there is a net moisture convergence. The bottom of the layer (generally the surface) is the level at which the cloudy parcel originates. During its ascent, the parcel cools dry adiabatically while being lifted to its lifting condensation level (LCL); above the LCL, the ascent follows a pseudo-adiabat slightly modified by entrainment. The top of the cloud layer is the level of non-buoyancy.

The net moisture convergence, or accession, available to convective precipitation processes, is defined as:

$$Q_{AC} = -\frac{1}{g}\int_{p_t}^{p_l} \nabla \bullet (Vq)\, dp + E_0 \tag{14}$$

in which the first term represents the large-scale moisture convergence and E_0 is the contribution from surface evaporation (here, p_l is the pressure of the lifting level, p_t is the pressure at cloud top). One of the main hypothesis of the Kuo scheme, is that a fraction *b* of that available water returns to humidify the environment, whereas the remaining part *(1-b)* falls as convective precipitation, therefore contributing to warming the environment. This partition parameter *b* varies with the mean saturation deficit in the cloud layer:

$$b = \left[\frac{1 - \frac{1}{p_b - p_t}\int_{p_t}^{p_b} RH\, dp}{1 - RH_c} \right]^n \tag{15}$$

in which p_b is the pressure at the bottom of the cloud, *RH* is the relative humidity, RH_c is a critical relative humidity parameter (set to 0.37), and *n* is an exponent parameter (set to 3).

According to the Kuo scheme, the humidification, Δq , is distributed vertically as a function of the local environmental saturation deficit, in relation to the mean deficit in the convectively active layer:

$$\Delta q = \frac{b\, Q_{AC}\left[q_{sat}(T_e) - q_e\right]}{\frac{1}{p_b - p_t}\int_{p_t}^{p_b}\left[q_{sat}(T_e) - q_e\right]dp} \tag{16}$$

and the heating from condensation processes (not including the vertical advection of heat) is:

$$\Delta T = \frac{L}{c_p}\frac{(1-b)\, Q_{AC}\left[T_{vc} - T_{ve}\right]}{\frac{1}{p_b - p_t}\int_{p_t}^{p_b}\left[T_{vc} - T_{ve}\right]dp} \tag{17}$$

in which the indices “e” and “c” refer to “environmental” and “cloud” values (i.e., grid-scale), L is the latent heating of vaporization, and c_p is the specific heat of air at constant pressure. It can be seen from the above equations that the total effects of convective clouds is to release latent heating according to the *(1-b)* fraction, and to redistribute vertically the non-precipitation fraction (b) of the moisture accession.

The convective precipitation rate R is given by:

$$R = \frac{1}{g} \frac{\int_{p_b}^{p_t} \frac{c_p}{L} \Delta T \, dp}{2\,\Delta T} \tag{18}$$

Originally developed for larger-scale models (~ 100 km and more), the use of the Kuo scheme has been pushed to horizontal resolutions for which its closure assumption (based on large-scale moisture convergence) may not be appropriate and would probably lead to incorrect stabilization of atmospheric model columns. Mesoscale models with resolution on the order of 10-30 km gives the possibility to represent convection in a more accurate fashion, as will be discussed below. The Kuo scheme, with its very simple treatment of the convective updrafts and of microphysics, and with its lack of downdrafts and entrainment/detrainment, is not equipped to produce the best possible convective “adjustments” for such models.

Work is thus currently done to replace in the near future the Kuo scheme by a more appropriate one (at least in the regional model). The best candidate at this time is certainly the Fritsch-Chappell scheme, based on removal of convective available potential energy (CAPE). In this scheme, the upward and downward mass fluxes (updrafts and downdrafts) are calculated at each level, and entrainment/detrainment play an important role (see Fig. 13). Additional features, such as more sophisticated microphysics and precipitation efficiency functions, make this scheme attractive.

In the medium to longer term, it is envisioned to use an even more sophisticated convective scheme for the regional model. This scheme, the Kain-Fritsch scheme, is an advanced version of the Fritsch-Chappell scheme, in which a more realistic one-dimensional entraining/detraining plume model was included (see Fig. 14), together with more sophisticated microphysics. Right now, this scheme is too costly to be used operationally, but it is planned to improve soon its numerical performance.

Finally, it is not excluded that the Kuo scheme could be replaced in the global model. Currently, a relaxed version of the Arakawa-Schubert convective scheme is available at RPN. Based on a cloud work function in which the consumption of convective energy is equal to its generation by the large-scale, and using a cloud population, this scheme is certainly appropriate for large-scale models. Unfortunately, this scheme was not tested much in our models environment.

Grid-scale condensation schemes

To go with the Kuo scheme, the grid-scale condensation scheme based on the work of Sundqvist is used in both the regional and global operational models. In this scheme, the equations describing the thermodynamics of the moist atmosphere are substantially simplified: the precipitation is supposed to fall instantaneously to the ground, condensation occurs when a critical relative humidity is reached, and a cloud cover fraction is supposed. Thus, this scheme is not exactly an *explicit* scheme, in the sense that it does not require the grid-scale to be saturated before generating some condensate.

The starting equations describing the thermodynamics of the moist atmosphere are:

$$\frac{\partial T}{\partial t} = A_T + \frac{L}{c_p}\left(P_{con} - P_{re}\right) - P_{fm} - F_{cT} \tag{19}$$

$$\frac{\partial q_v}{\partial t} = A_{qv} - \left(P_{con} - P_{re}\right) + F_{cq} \tag{20}$$

$$\frac{\partial q_c}{\partial t} = A_{qc} + P_{con} - \left(P_{ra} + P_{rc}\right) \tag{21}$$

$$\frac{\partial q_r}{\partial t} = A_{qr} + \left(P_{ra} + P_{rc}\right) - P_{re} + P_{rf} \tag{22}$$

where q_v, q_c, and q_r are the mixing ratios of water vapor, cloud water/ice, and rainwater/snow; A_T, A_{qv}, A_{qc}, and A_{qr} are the dynamic tendencies for temperature, water vapor, cloud water/ice, and rainwater/snow; P_{ra} is the accretion rate of cloud droplets by raindrops; P_{rc} is the autoconversion rate of cloud droplets to raindrops; P_{re} is the evaporation rate of rainwater/snow; P_{con} is the condensation or evaporation rate of cloud droplets; P_{fm} is the heating term due to melting freezing of particles; P_{rf} if the fallout of rainwater/snow; and F_{ct}, F_{cq} are tendency terms for temperature and water vapor from convective processes.

Without a doubt, the above equations represent a large range of microphysical processes occurring in the atmosphere. But one must be conscious *that only a simplification of this set*, proposed by Sundqvist, *is used in this scheme*. First, the scheme is based on the idea that condensation processes on a grid with horizontal resolution on the order of 50 km or more are still subgrid-scale processes, so that cloud formation actually starts before the grid-resolved humidity reaches the saturation value, and that, equivalently, a cloud cover fraction is defined. Also, with the typical space and time resolutions of meso-α-scale models, one may assume that precipitation falls instantaneously and, therefore, the predictive equation for rainwater/snow [Eq. (22)] can be neglected. Using this approach, only one predictive equation describing the cloud water content is added to the usual thermodynamic equations.

Assuming that only a portion (fraction *f*) of the model grid area is cloudy, the above equations can be simplified using:

$$\left(P_{ra} + P_{rc}\right) = P = f\,\hat{P} \tag{23}$$

$$P_{con} = f\,\hat{Q} - \left(1 - f\right)\tilde{E}_c = Q - E_c \tag{24}$$

$$P_{re} = \left(1 - f\right)\tilde{E}_r = E_r \tag{25}$$

in which the caret denotes average values in the cloudy part, whereas the tilde are average values in the clear part of the grid. Also, Q is the condensation rate, E_c is the evaporation rate for cloud water, and E_r is the evaporation rate for rain.

The basic set of equations can then be written:

$$\frac{\partial T}{\partial t} = A_T + \frac{L}{c_p} Q_H + F_{ct} \tag{26}$$

$$\frac{\partial q_v}{\partial t} = A_{qv} - Q_H + F_{cq} \tag{27}$$

$$\frac{\partial q_c}{\partial t} = A_{qc} + Q_H - \left(P - E_r\right) \qquad (28)$$

in which

$$Q_H = P_{con} - P_{re} = Q - E_c - E_r \qquad (29)$$

is the net heating term including subgrid-scale condensation and evaporation of cloud water and rainwater. In this predictive cloud water scheme, condensation may occur in columns which have already been treated with the convective scheme, in order to account for the effect of overlapping convective and stratiform clouds with a particular grid cell. The advection of cloud water is done using the semi-Lagrangian technique with three-dimensional winds.

The fractional cloud cover f is related to the relative humidity following:

$$f = 1 - \left(\frac{RH_s - RH}{RH_s - RH_{00}}\right)^{1/2} \qquad (30)$$

where RH_s =1 (saturation) and RH_{00} = 0.8 is a treshold value.

We believe it is not necessary to give here all the details of the microphysical equations (for precipitation release, coalescence process, Bergeron-Findeisen mechanism, and evaporation of precipitation) for the reader to understand the basic philosophy of the Sundqvist scheme.

Again, as was the case for the Kuo scheme, the Sundqvist scheme may not be appropriate for higher-resolution models (10-30 km) due to its basic closure assumptions. For instance, condensation for such models should occur only with the grid cell is saturated. Also, the precipitation fallout (predictive equation for rainwater) should be considered one way or another. Other schemes are currently tested for possible improvements in the next versions of the regional operational model. Maybe the best candidate at this time is the scheme developed at MRB by Tremblay. One should note, however, that the improvements resulting from changing the explicit condensation scheme are not as dramatic as those obtained from changing the convective scheme.

F) Conclusions

To close this presentation, we express our hope that we have successfully planted in the reader's mind the importance of each of the physical processes represented in the RPN's physics package, and that the interactions between all of these processes have been highlighted sufficiently. To keep the presentation short, some secondary aspects of the physics have not been discussed, such as the gravity wave drag and the shallow convective scheme. In our view, however, the most important aspects of the physics for NWP have been described here.

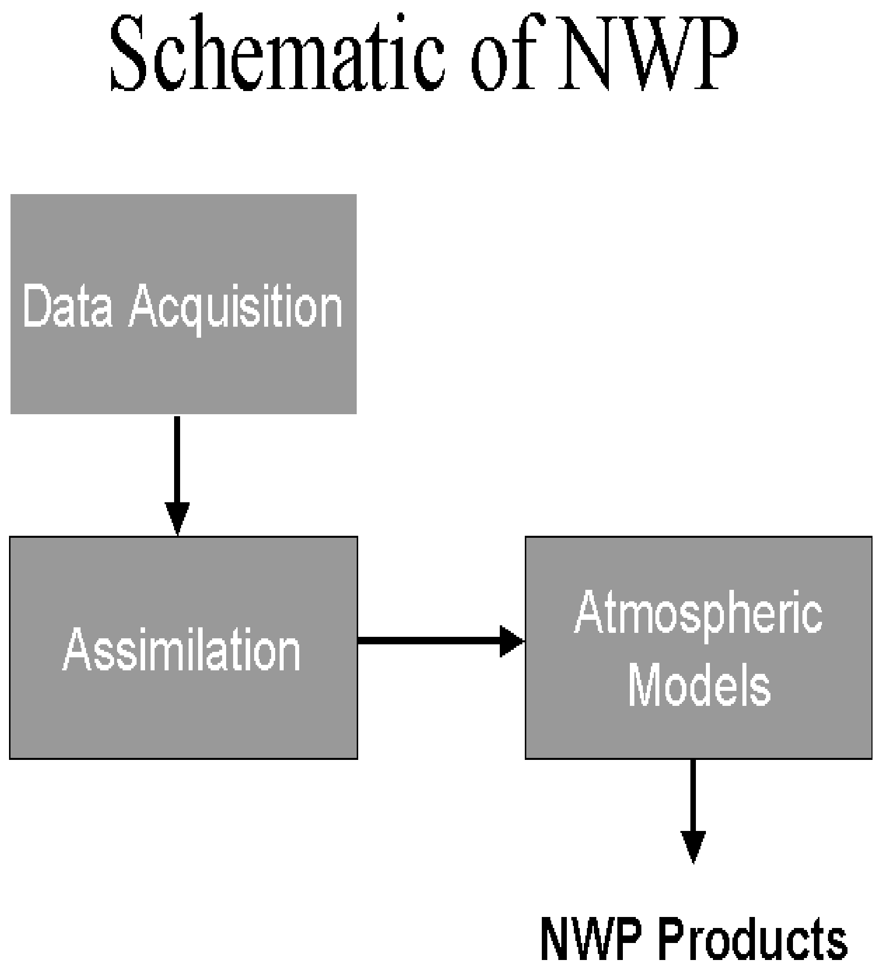
Schematic of NWP
Data Acquisition
Assimilation
Atmospheric Models
NWP Products

Figure 1

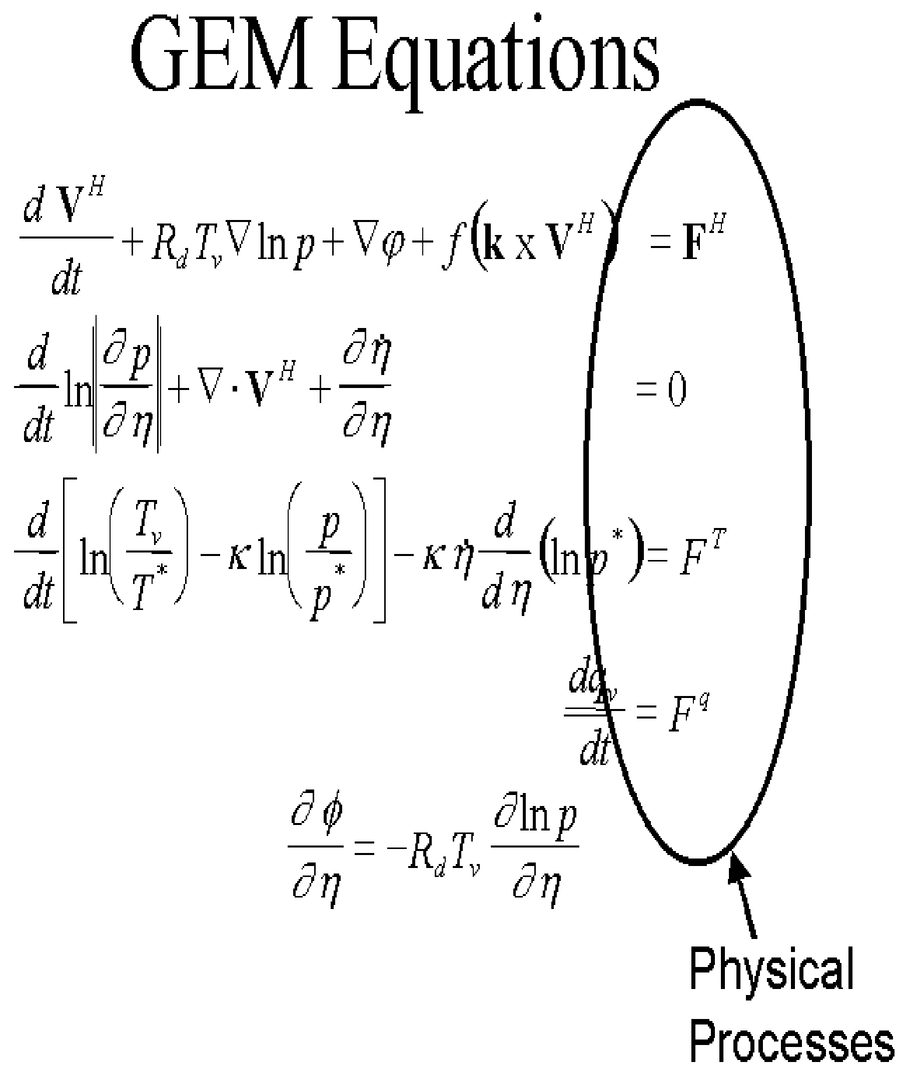
GEM Equations
$\frac{d\mathbf{V}^H}{dt} + R_d T_v \nabla \ln p + \nabla\varphi + f\left(\mathbf{k} \times \mathbf{V}^H\right) = \mathbf{F}^H$
$\frac{d}{dt}\ln\left|\frac{\partial p}{\partial\eta}\right| + \nabla\cdot\mathbf{V}^H + \frac{\partial\dot{\eta}}{\partial\eta} = 0$
$\frac{d}{dt}\left[\ln\left(\frac{T_v}{T^*}\right) - \kappa\ln\left(\frac{p}{p^*}\right)\right] - \kappa\dot{\eta}\frac{d}{d\eta}\left(\ln p^*\right) = F^T$
$\frac{dq_v}{dt} = F^q$
$\frac{\partial\phi}{\partial\eta} = -R_d T_v \frac{\partial \ln p}{\partial\eta}$
Physical
Processes

Figure 2

Splitting of Dynamical and Physical Processes

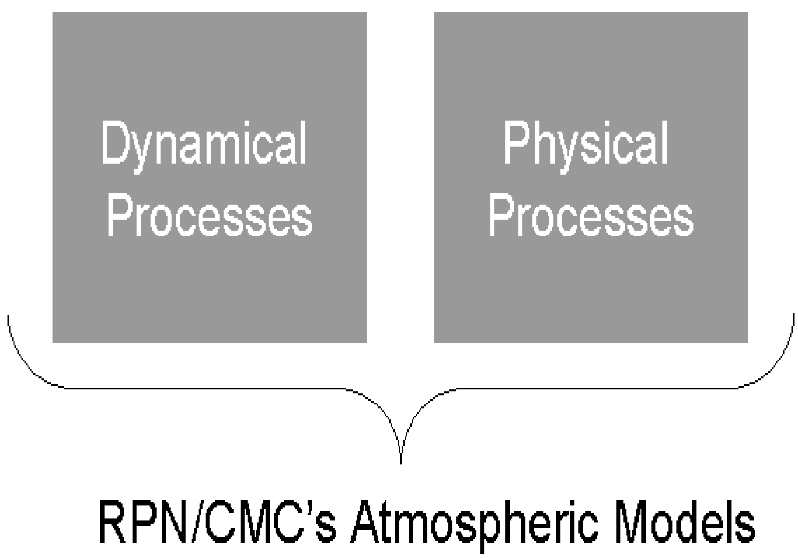

Figure 3

What Do All These Models Have in Common ?

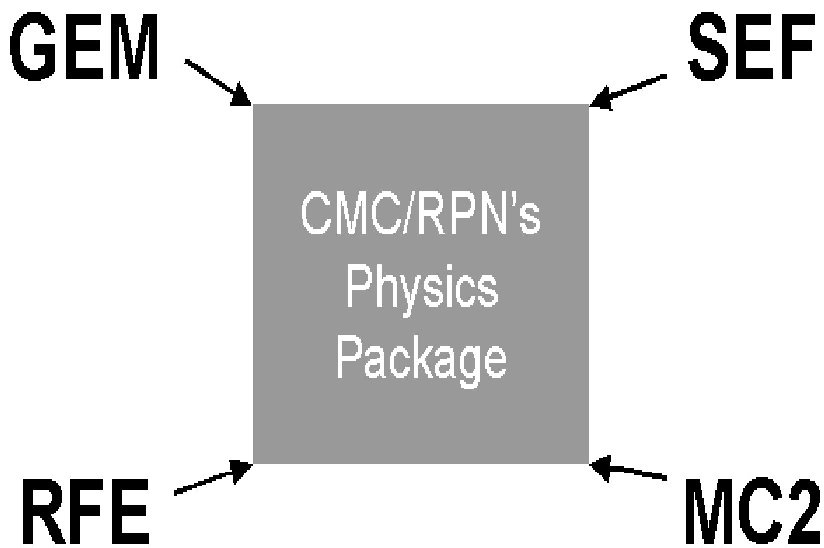

Figure 4

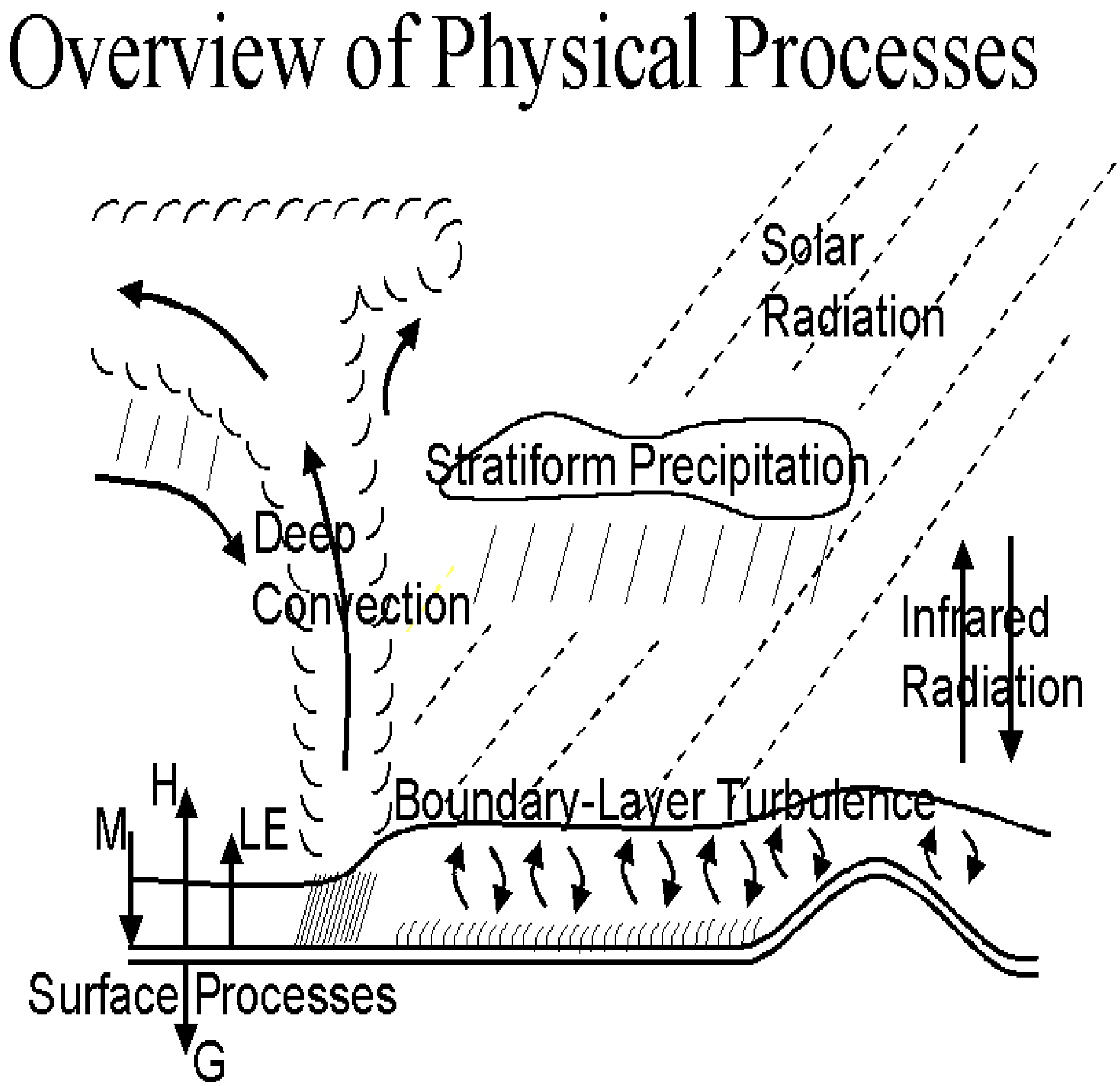
Overview of Physical Processes
Solar Radiation
Stratiform Precipitation
Deep Convection
Infrared Radiation
Boundary-Layer Turbulence
M
H
LE
Surface Processes
G

Figure 5

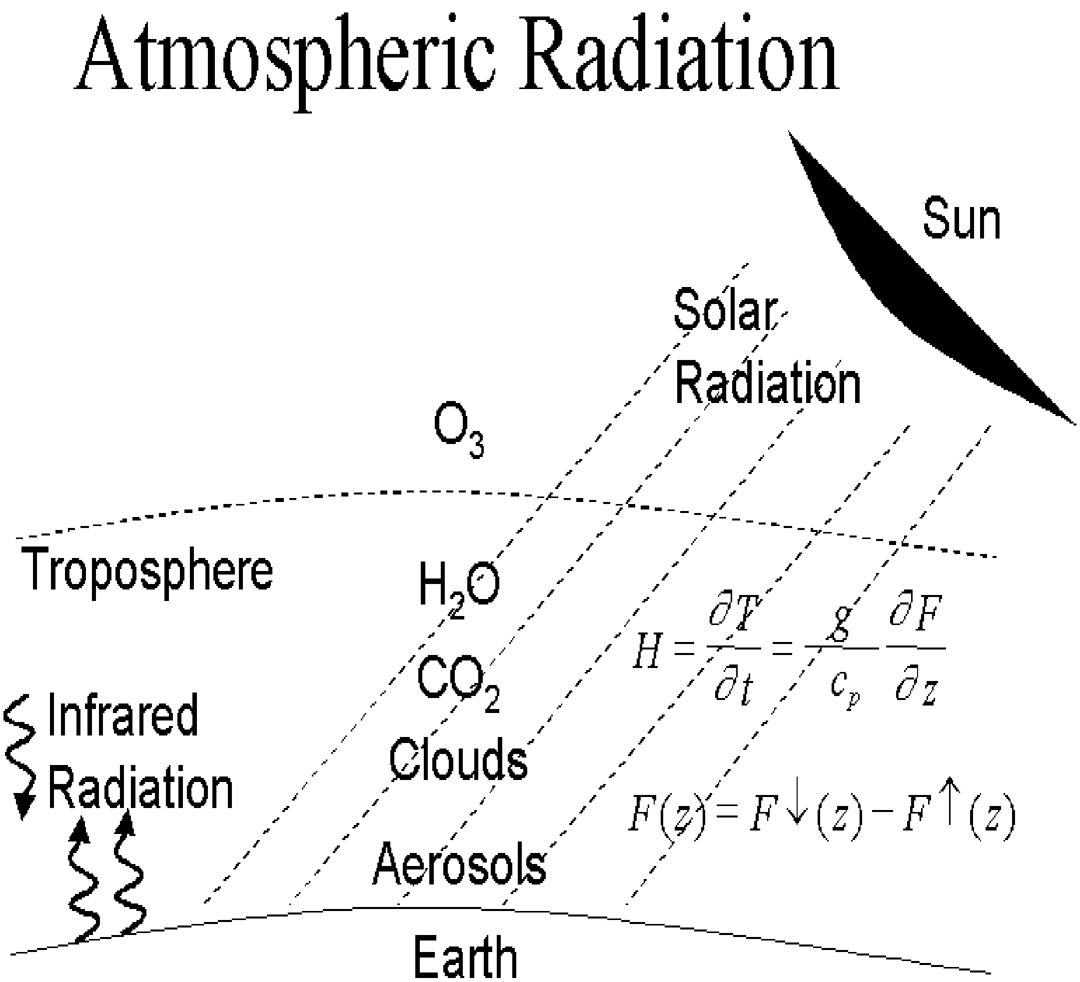
Atmospheric Radiation
Sun
Solar
Radiation
O_3
Troposphere
H_2O
CO_2
Clouds
Aerosols
Infrared
Radiation
Earth
$H = \frac{\partial T}{\partial t} = \frac{g}{c_p} \frac{\partial F}{\partial z}$
$F(z) = F\downarrow(z) - F\uparrow(z)$

Figure 6

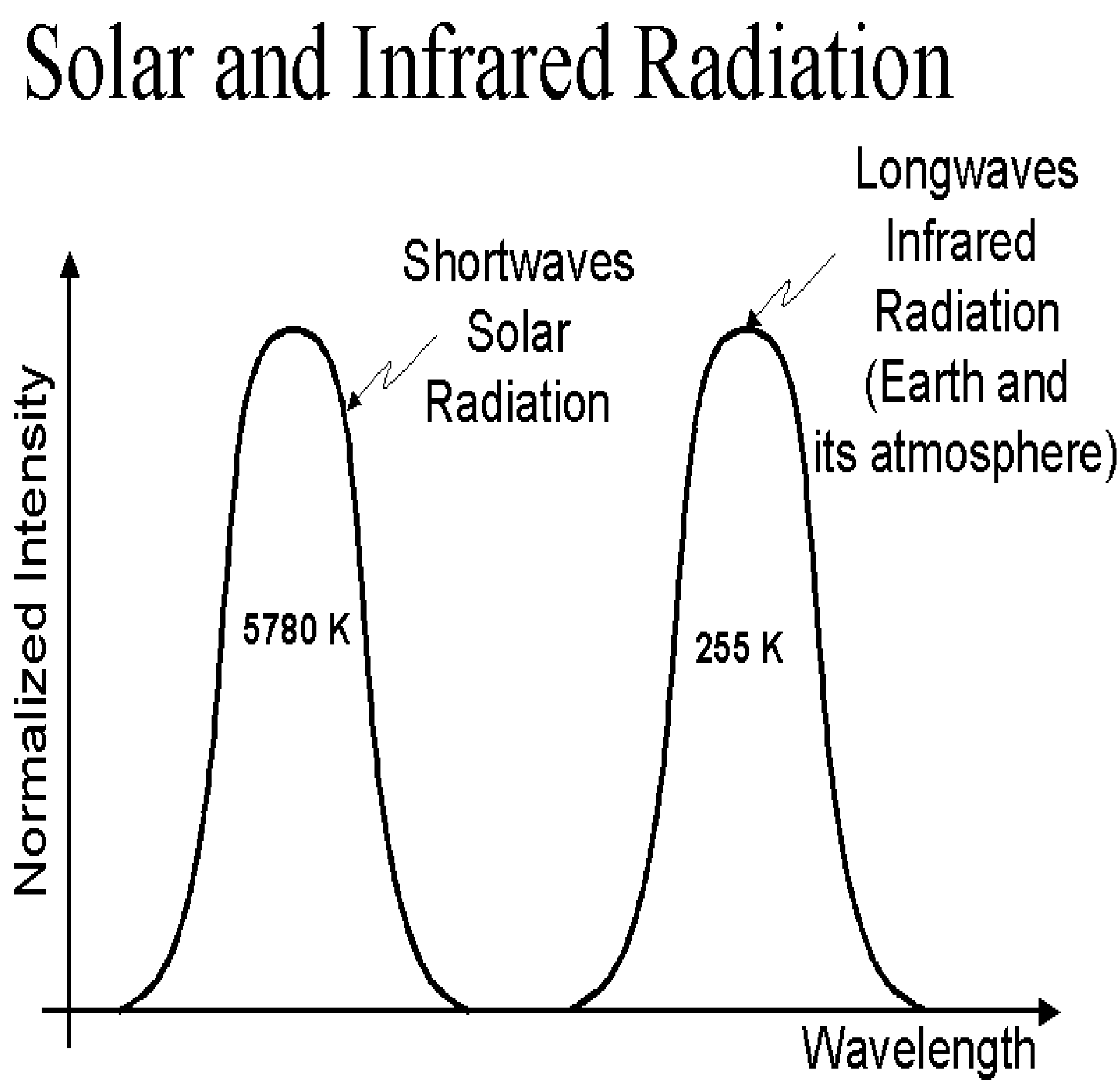
Solar and Infrared Radiation
Shortwaves
Solar
Radiation
Longwaves
Infrared
Radiation
(Earth and
its atmosphere)
Normalized Intensity
5780 K
255 K
Wavelength

Figure 7

Example: Surface Fluxes Influence

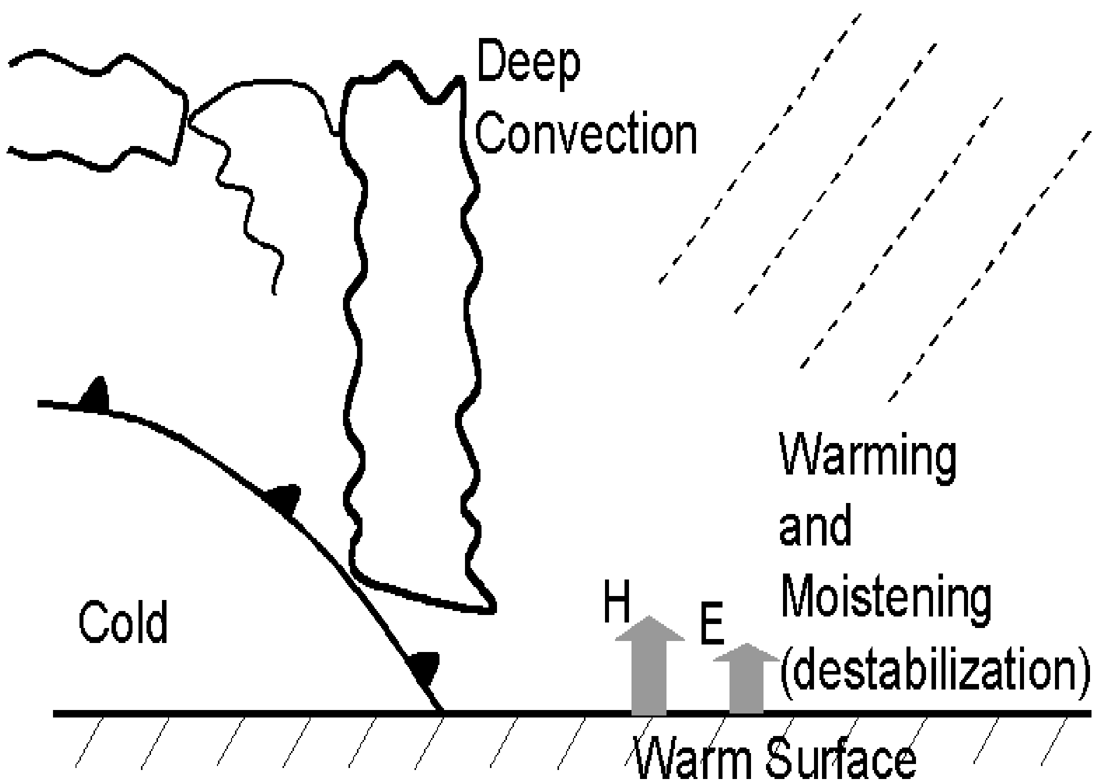

Figure 8

Daytime Boundary Layer

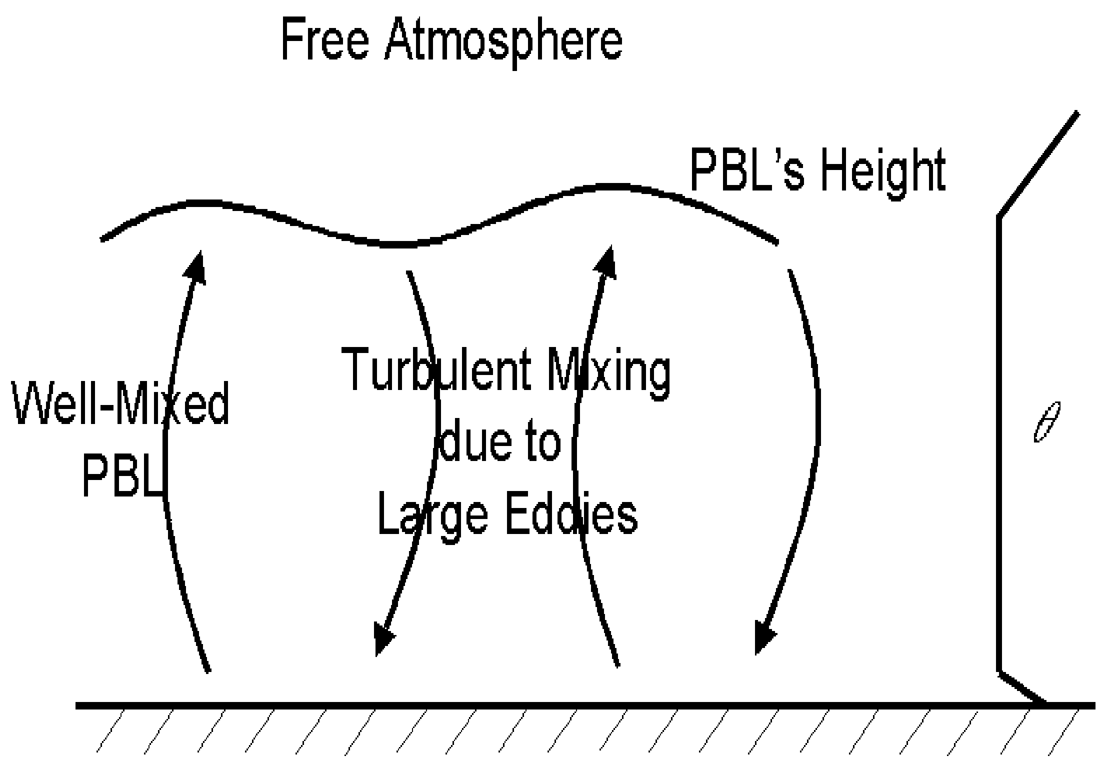

Figure 9

Example 1: Synoptic-Scale Weather System

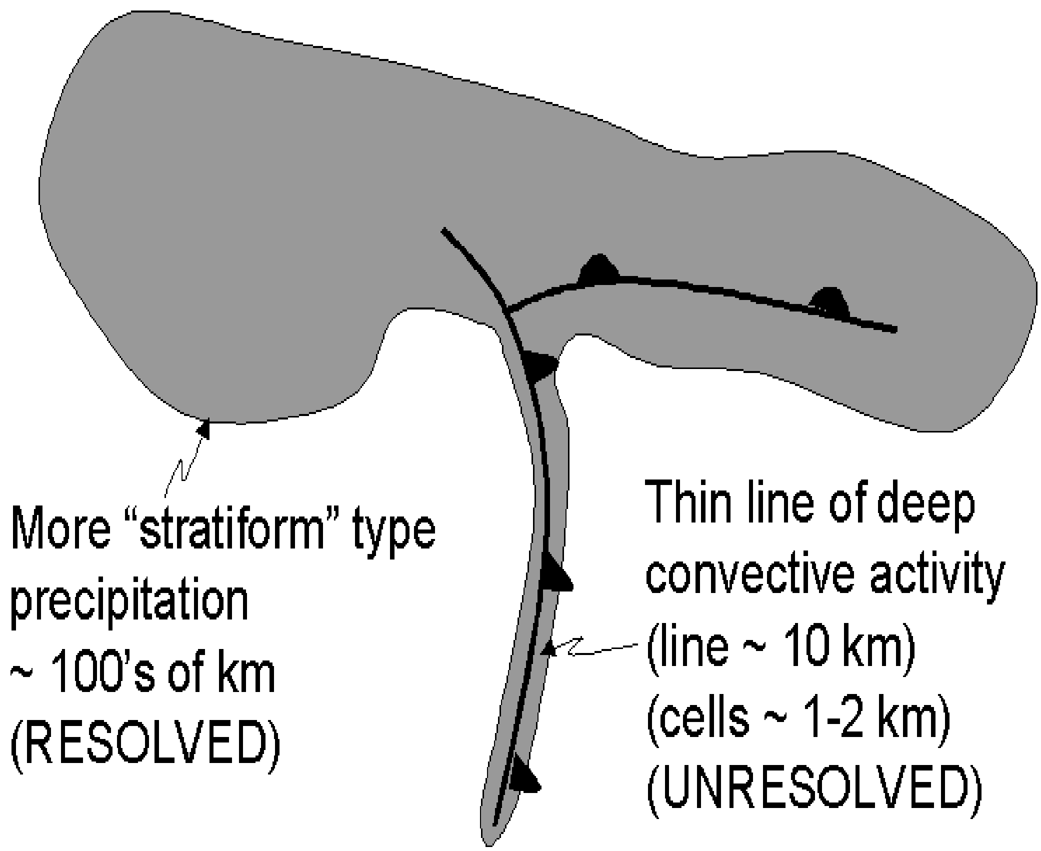

Figure 10

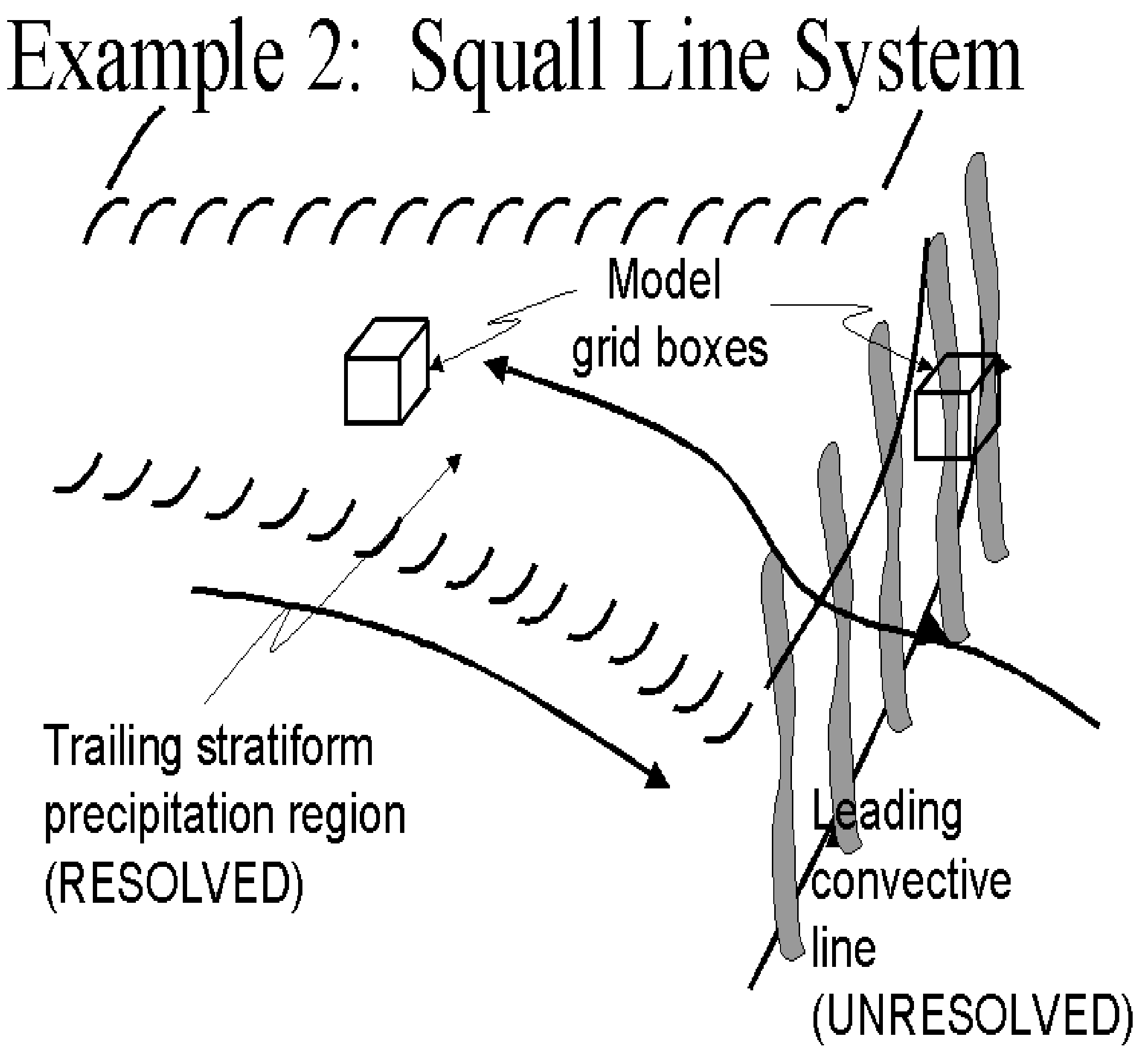
Example 2: Squall Line System
Model
grid boxes
Trailing stratiform
precipitation region
(RESOLVED)
Leading
convective
line
(UNRESOLVED)

Figure 11

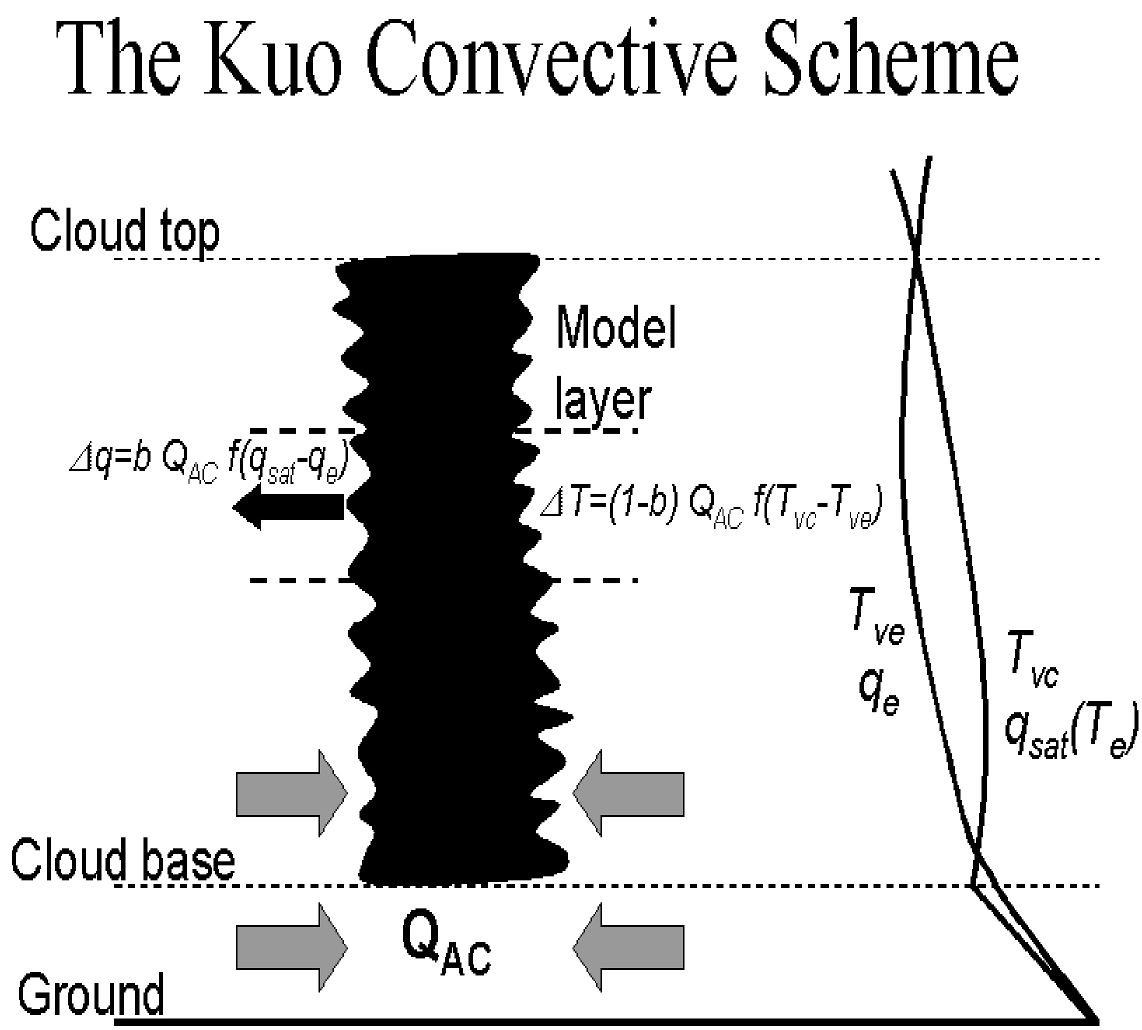
The Kuo Convective Scheme
Cloud top
Model
layer
$\Delta q = b\ Q_{AC}\ f(q_{sat} - q_e)$
$\Delta T = (1-b)\ Q_{AC}\ f(T_{vc} - T_{ve})$
T_{ve}
q_e
T_{vc}
$q_{sat}(T_e)$
Cloud base
Q_{AC}
Ground

Figure 12

Fritsch-Chappell Cloud Model

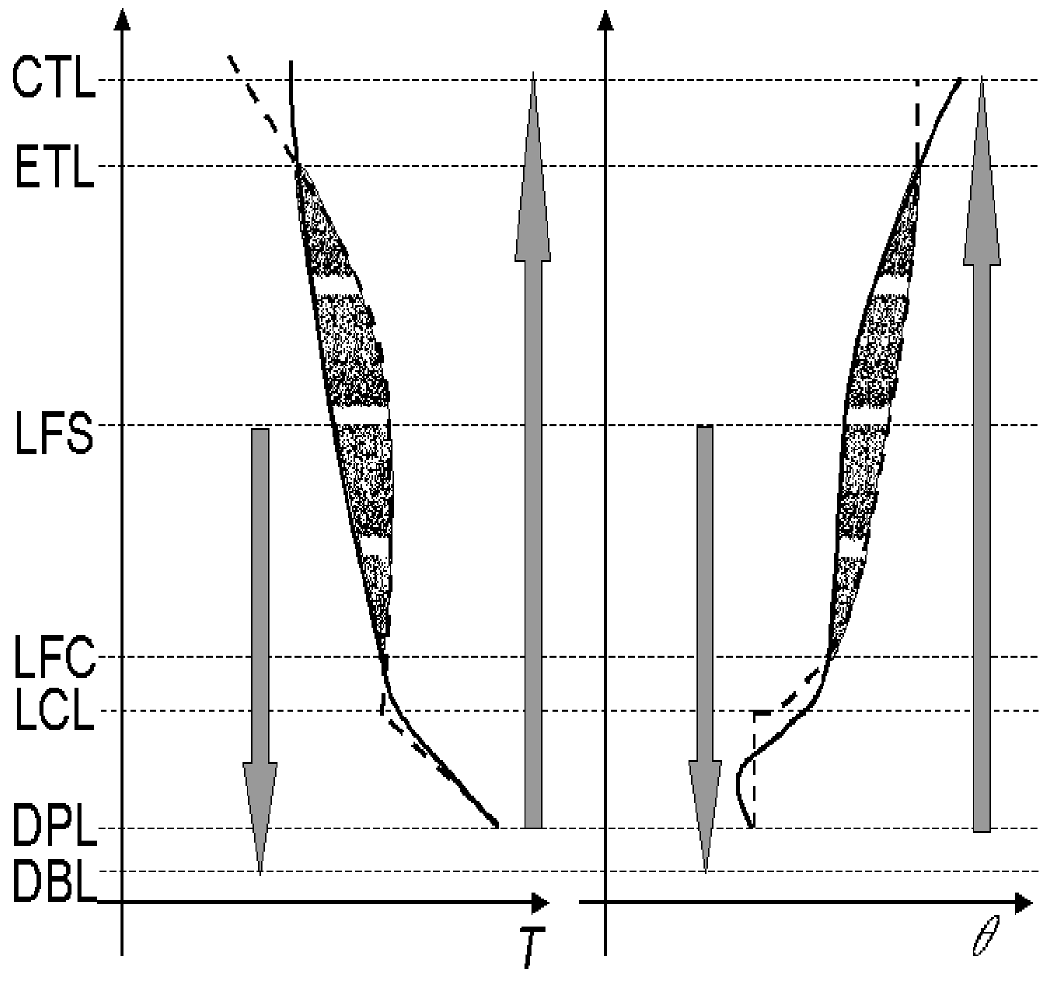

Figure 13

One-Dimensional Entraining Detraining Plume Model

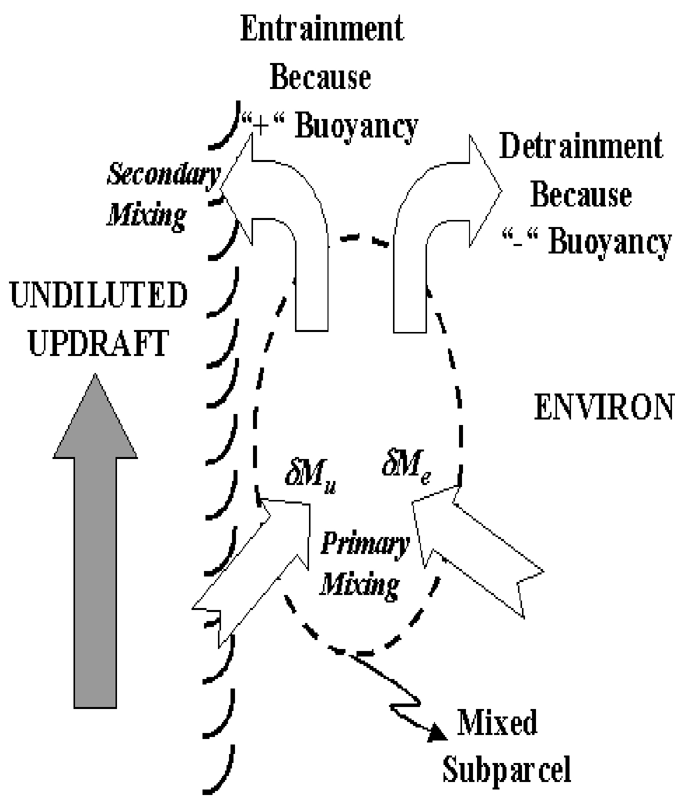

Figure 14