THESIS

OBSERVATIONS AND SIMULATIONS OF THE PLANETARY BOUNDARY LAYER AT A TALL TOWER IN NORTHERN WISCONSIN

Submitted by

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In partial fulfillment of the requirements

for the degree of Master of Science

Colorado State University

Fort Collins, Colorado

Summer 2002

COLORADO STATE UNIVERSITY

May 17, 2002

WE HEREBY RECOMMEND THAT THE THESIS PREPARED UNDER OUR SUPERVISION BY NI ZHANG ENTITLED OBSERVATIONS AND SIM-ULATIONS OF THE PLANETARY BOUNDARY LAYER AT A TALL TOWER IN NORTHERN WISCONSIN BE ACCEPTED AS FULFILLING IN PART RE-QUIREMENTS FOR THE DEGREE OF MASTER OF SCIENCE.

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ABSTRACT OF THESIS

OBSERVATIONS AND SIMULATIONS OF THE PLANETARY BOUNDARY LAYER AT A TALL TOWER IN NORTHERN WISCONSIN

The planetary boundary layer (PBL) is the lowest portion of the atmosphere where motion is strongly influenced by surface characteristics, and is highly interactive with Earth's biosphere. It is often turbulent, with a relatively fast response to surface forcing. A lot of weather activities happen in this portion of the atmosphere and most of human activities are also intimately tied to it. Carbon Dioxide (CO₂) is the most important greenhouse gas, which has significant impact for the global climate, and the concentration of CO₂ at Earth's surface is greatly influenced by the boundary layer mixing. A careful analysis of the PBL depth in the model development is essential for an accurate prediction of the global CO₂ budget and the future climate change.

Long-term continuous observation of the PBL depth (Zi) was achieved at an observational site in northern Wisconsin. The nighttime stable layer depth was estimated by detailed analysis of CO_2 mixing ratios measured at 6 levels on the tower. Daytime mixing layer depth was obtained with radar measurements and combined with CO_2 analysis in the early morning when mixing layer is shallower than 400 m.

The CSU single column GCM (SCM), forced by the Rapid Update Cycle (RUC) data and the tower measurements, was used to simulate the PBL depth and compared with the observational data. In general, the model captures most features of the diurnal variability of the PBL depth observed at the tower site. Exceptions tend to be associated with very calm conditions, which appears to reflect inadequate shear forcing of turbulence in the model. Simulated PBL depth tends to reach the maximum later than the observed and tends to remain high in the late afternoon. The simple bulk PBL model cannot capture the discontinuity in the late afternoon and sometimes in the early morning as well because it lacks a separate representation between the stable layer and the mixing layer.

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ACKNOWLEDGEMENTS

I would like to thank Dr. A. Scott Denning, my advisor, for his patience, understanding and support during the past few years. I would also like to thank my committee members Dr. David A. Randall and Dr. Paul W. Mielke Jr. for taking the time to review this thesis.

Many thanks to all the members in Denning's Group. Lara Prihodko, my longtime officemate, who has always been there with valuable advises and help through ups and downs in my life. Help from John Kleist and Ian Baker for computer problems and programming language was very important for me to get through this research. Connie Uliasz, Neil Suits, Elicia Inazawa, Lixin Lu, Kevin Schaefer and all other members of Dr. Denning's Group have contributed to make the past few years an enriching and exciting experience. Also many thanks to Mark Branson from Dr. Randall's group, who provided a lot of help with the model; Peter Bakwin at NOAA/CMDL, Ken Davis and Chuixiang Yi at Penn State University for their extraordinary field work and providing valuable observational data.

Special thanks to my fiancé Jean-Christophe Golaz for his tender love and solid support, which made the last few could-be-stressful months a very pleasant experience.

This research was funded by the U.S. Department of Energy (contract numbeer 1948-CSU-USDOE-3008) and through the National Institute for Global Environmental Change (Cooperative Agreement No. DE-FC03-90ER61010). Any opinions, findings, and conclusions or recommendations expressed in this publication are those of the authors and do not necessarily reflect the views of the DOE.

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Chapter 1

Introduction

1.1) Motivation

The planetary boundary layer (PBL) is the turbulent portion of the atmosphere directly affected by the Earth's surface, with a relatively quick response to surface forcing. The PBL is highly interactive with Earth's biosphere. A lot of weather activities happen in this portion of the atmosphere, and most of human activities are also intimately tied to it. Surface fluxes of heat, moisture, and momentum influence atmospheric general circulation, exerts a significant effect on the Earth's climate. Since the PBL plays such an important role in the Earth's climate, an accurate representation of the PBL in a general circulation model is essential for the future climate prediction.

1.2) Carbon Dioxide and Climate Change

Several natural factors can change the balance between the energy absorbed by the Earth and that emitted by it in the form of long wave infrared radiation. These factors produce a radiative forcing on climate. One of the most important factors is the greenhouse effect. Short-wave solar radiation can pass through the atmosphere easily but long-wave terrestrial radiation emitted by the warm surface of the Earth is partially absorbed and then re-emitted by a number of trace gases in the cooler atmosphere above. Because of this effect, both the lower atmosphere and the surface of the Earth are warmer than they would be without the greenhouse gases, since the overall out going long-wave radiation has to be balanced by the incoming solar radiation. Naturally occurring greenhouse gases keep the Earth warm enough to be habitable. However, increasing the concentrations of the greenhouse gases or adding new green house gases into the atmosphere will likely raise the global-average annual-mean surface-air temperature, which may result some unexpected climate changes (IPCC, 1990).

Carbon Dioxide (CO_2) is the most important greenhouse gas directly influenced by human activity. Many experiments and measurements have been conducted, and the records show a clear evidence of a seasonal cycle caused by photosynthesis and respiration of vegetation at temperate and subtropical latitudes as well as a strong longterm upward trend (Figure 1.1).



Figure 1.1: Global CO₂ distributions from 1991 through 2000. Data collected by NOAA/CMDL, Carbon Cycle Group which operates a network of a world wide observational program including more than 40 sampling site (Conway et al., 1988, 1994)

It is now a well-documented fact that the CO_2 mixing ratio has increased by about 25 percent within the past few hundred years (Watson et al., 1990), consistent with industrialization and large-scale land use changes. The increased mixing ratio of atmospheric CO_2 , along with other green house gases that absorb infrared radiation, will change the radiative balance of the atmosphere and the surface of the earth (Watson et al., 1990). Therefore, it has a significant impact on Earth's environment and consequently on human well-being.

In the past few decades, scientists have paid a lot of attention to the global carbon budget but the understanding of this subject is still not very clear, with the greatest uncertainty focused on the size and the location of the "missing CO₂ sink. Spatial and temporal variations of atmospheric CO₂ concentrations contain information about surface sources and sinks, which can be quantitatively interpreted through tracer transport inversion. Inverse modeling interprets variations of atmospheric CO₂ by estimating the strengths of response functions to unit surface fluxes carried with atmospheric tracer models. Early inversions used two-dimensional models to calculate the latitudinal distribution of fluxes (e.g. Tans et al , 1989; Enting and Mansbridge, 1989; Ciais et al., 1995). More recent inversions have used three-dimensional models to estimate the longitudinal distribution of fluxes (e.g. Fan et al., 1998; Bousquet et al., 1999a, Bousquet et al., 1999b, Kamisnki et al, 1999; Bruhwiler et al , 1999; Law, 1999; Bousquet et al., 2000; Baker, 2001).

The Tracer Transport Inversion Intercomparison Project (TransCom) is an international effort to quantify the sources of error and uncertainty in CO_2 inversions has been going on for nearly 10 years (Law et al, 1996; Denning et al, 1999; Gurney et al, 2002a,b). One of the most significant sources of uncertainties is due to covariance between surface CO_2 flux and vertical transport in the atmosphere, which has been termed the "rectifier effect" (Denning et al, 1995, 1996; 1999, Gurney et al, 2002a,b).

1.3) Rectifier Effect

The atmospheric rectifier effect is usually defined as a temporal co-variation between a surface flux and the atmospheric mixing or transport that produces a timemean spatial concentration gradient in the atmosphere. The term "rectifier" is analogous to the one in physics of electricity. A rectifier in the physics of electricity involves a filter circuit, used for removing alternating components of the current. The aim of the filter is to provide a low series resistance in one direction and a very high shunt resistance in another direction for the alternating current. Basically it is a device through which current can flow only in one direction, often used alone or in sets to convert AC current into DC pulsating current. Two common varieties of rectifier outputs are the full- and half-wave unidirectional voltages (non-sinusoidal periodic functions). Figure 1.2 shows the input signal and the half-wave output signal of a rectifier device.



Figure 1.2: Input and output signals of a rectifier device. Upper panel: the sinusoidal input signal. Bottom panel: the ideal half-wave rectifier output signal (non-sinusoidal).

In atmospheric science, the mechanism of the "rectifier effect" is not quite the same as the one in electricity, but the output is very similar to the half-wave rectifier output. Basically, it's a simple way to visualize the evolution of CO₂ concentration in the atmosphere. Figure 1.3 shows a conceptual rectifier model (Denning, personal communication) with two boxes: a PBL box and a "free troposphere" box. On the right side of the figure are the transport equations, where F is the surface flux; τ is the "mixing time scale"; C_1 is the PBL concentration and C_2 is the troposphere concentration. Figure 1.3B is the result of this conceptual rectifier model. The upper panel of figure 1.3B shows the input signal: the diurnal cycle of CO₂ flux due to photosynthesis and respiration and the diurnal variation of PBL depth, with a phase shift of 3 hours. The bottom panel is the output signal.



Figure 1.3A: Conceptual Rectifier Model with a varying PBL depth



Figure 1.3B: Results from the conceptual rectifier model (Denning, personal communication).

Biological exchange of CO_2 at the Earth's surface is accomplished by photosynthesis and respiration. When respiration exceeds photosynthesis, the surface is a source of CO_2 to the atmosphere, and vice versa. Photosynthesis requires solar radiation, so the surface tends to be a source at night and in winter, and a sink during the day and during the summer. Over much of the northern hemisphere, surface uptake of CO_2 is therefore associated with deep turbulent mixing in the planetary boundary layer (PBL) and by cumulus convection, since these phenomena are also associated with stronger solar radiation. Thus the influence of the terrestrial sink is "felt" through a deeper layer of the atmosphere than the influence of the source, even in the absence of a time-mean net flux. The global flask sampling network for CO_2 only "sees" surface air, so the covariance of source and sinks with vertical transport leads to elevated concentrations over much of the northern hemisphere where large temperate land regions experience this effect (Denning et al, 1995, 1996).

The rectifier effect leads to an apparent positive anomaly of CO_2 at the measuring stations that is not associated with a net source or sink. An atmospheric model which (underestimates/overestimates) the strength of the rectifier effect will therefore compensate for these elevated values by (overestimating/underestimating) the net land sink over the northern temperate zone. Gurney et al (2002b) found that the strength of the simulated rectifier effect among 16 transport models and model variants accounted for most of the variance in their estimation of northern terrestrial sinks (Figure 1.4). The stronger the simulated rectifier effect, the larger the estimated northern hemispheric sink.



Figure 1.4: TransCom3 Experiment. Gurney et al. 2002

1.4) Objectives of This Study

How the PBL depth (Zi) is modeled in the GCM significantly impacts CO_2 inversions. To better understand the "rectifier effect" and to improve the model prediction, I analyzed the data from measurements of CO_2 concentration, flux and PBL depth at a tower site in Northern Wisconsin. The result provides a good support of the "rectifier effect" in the model study.

A single column GCM (general circulation model) from CSU atmospheric science department is used to thoroughly study the atmospheric boundary layer depth. The model is forced by observational data at the tower site in Northern Wisconsin so we can compare the model result with observed PBL depth. Since PBL depth is a very important element in studying the atmospheric "rectifier effect", series of parameters are tested in the single column model to improve its PBL prediction, to provide a building block for the full GCM, which can be used in "inverse" modeling techniques to study the problem of the CO_2 "missing sink" in the Earth system.

Chapter 2

Methods

2.1) Sampling Site

The study area of this project is located in Chequamegon-Nicolet National Forest in northern Wisconsin (45.95N, 90.28W, elevation is 472m). It encompasses an area approximately 325,000 ha. The land surface of this site is covered by heavy forest of low relief and the dominant forest types are mixed upland pine, northern hardwoods, aspen, and lowlands and wetlands conifers (Figure 2.1). Much of the area was logged during the 1860-1920 period, mainly for pine, and has since re-grown. Human population density in the area is very low. The climate is cool continental, with average precipitation about 80 mm and mean annual temperature about 4.1°C, with a fluctuation of about 32°C from winter to summer.



Figure 2.1: Landscape of the study area, Northern Wisconsin.

The motivation of designing this observational site includes direct assessments of the exchange of carbon dioxide between ecosystem and atmosphere, trying to understand the role of vegetation in regulating microclimate, factors which contributing to the carbon dioxide mixing ratio, also to understand atmospheric boundary layer dynamics and the feedback between boundary layer dynamics and the vegetation.

The sampling platform for this study is a 447 m tall tower – the Wisconsin TV transmission tower (letter code WLEF), which hosts many instruments at various levels (Figure 2.2). Since 1994, continuous CO₂ mixing ratio measurements have been performed at 11 m, 30 m, 76 m, 122 m, 244 m and 396 m by two high precision Li-COR CO₂ analyzers, one measures air from 396 m continuously while the other cycles through all 6 levels (Bakwin et al., 1998). Micrometeorological data and eddy covariance flux are measured at three levels, 30 m, 122 m and 396 m. Three-axis sonic anemometers are used at these three levels to measure turbulent winds and virtual potential temperature. The air from these three levels is pumped down through a tube to three Li-COR analyzers on the ground to determine the fluctuations of CO₂ and water vapor mixing ratio for eddy covariance flux calculations. Sensible heat, latent heat, CO₂ vertical profile and CO₂ flux data are obtained from those measurements. Other observations from the tower, including net radiation, photosynthetically active radiation and rainfall, provide supporting meteorological data.



Figure 2.2: WLEF TV transmission tower and instruments at the different level.

2.2) Observational Data

2.2.1) Radar Measurements

Long-term, continuous observation of the PBL structure was impossible until the recent development of boundary-layer profiling radar and radio-acoustic sounding systems (RASS). A RASS was installed near the tall tower site (about 8 km east of the tower) and was operated continuously during the period between mid-March and the beginning of November, in 1998 and 1999 (Angevine et al., 1997; Yi et al., 2001). The profiler is a sensitive 915 mHz Doppler radar (Figure 2.3), which is designed to respond to fluctuations of the refractive index in clear air. From the reflectivity of the radar signal, the top of the PBL depth can be determined fairly accurately under good weather conditions. Also the residual mixing layer after sunset can sometimes be found (Figure 2.4, top panel). However, the profiler is very sensitive to large cloud droplets and rain drops so weather plays a very important role in this measurement. Figure 2.4 (bottom panel) shows a rainy day radar signal and we can see that PBL depth cannot be defined correctly. Because of the structure of the profiler, features shallower than 400 m cannot be captured by the radar signal. Therefore, the nocturnal boundary layer has to be estimated by other means, such as the vertical profile of CO₂ mixing ratio.



Figure 2.3: Radar profiler near WLEF tower site.



Figure 2.4: Response from radar profiler, for a clear day (June 17, 1999, top panel), and for a rainy day (June 23, 1999, bottom panel).

2.2.2) Carbon Dioxide (CO₂) Measurements

Continuous monitoring of the vertical profiles of CO₂ and other trace gases on existing tall communications towers was designed by NOAA/CMDL Carbon Cycle Group and was first operated on a 610 m tall WITN TV transmitter tower in eastern North Carolina in 1992. With the success of WITN tower, a second, 447 m tall WLEF TV transmitter tower in northern Wisconsin was put online in 1994 (Bakwin, et al., 1998). CO₂ mixing ratios are measured at 6 levels on WLEF tower by two high precision Li-COR CO₂ analyzers (infrared gas analyzer) as described before. Tubes with 1 cm inner diameter were mounted on the tower with inlets at 11, 30, 76, 122, 244 and 396 m above the ground. One of the analyzer measures air from 396 m continuously while the other analyzer cycles through all 6 levels, at a 2 minutes interval for each valve switch to obtain a steady reading. The air is pressurized and dried before entering the analyzer, so the water vapor interference and dilution effect can be minimized. A full CO₂ mixing ratio profile is produced every 12 minutes and a PC-based data acquisition and control system (Zhao et al., 1997) transfers raw data automatically every day from the tower site to the CMDL Lab. Data used in this study from 1998 to 1999 give us a very good view of the development of the nocturnal boundary layer during nighttime and the evolution of the mixed layers in the morning.

The atmospheric boundary layer is defined as the turbulent lower boundary of the atmosphere (approximately within 1 km from earth's surface) where motion is strongly influenced by surface characteristics, predominantly frictional drag and surface heating. A typical boundary layer diurnal cycle over land consists three major regimes: during daytime, a very turbulent mixed layer (convective mixing layer); and during nighttime, a nocturnal stable boundary layer with sporadic turbulence as well as a less-turbulent residual layer consisting of previously mixed-layer air above the stable layer (Figure 2.5).



Figure 2.5: A typical boundary layer over land (Stull, 1988)

During daytime, the radar profiler can measure the convective PBL depth accurately, and at night, the remaining residual layer can also be detected under good conditions. The stable boundary layer height can be estimated from the vertical profile of CO_2 mixing ratio because respiration causes CO_2 to build up near the ground. The top of the strong gradient of CO_2 mixing ratio is a good indicator for the top of the nocturnal boundary layer. In many cases, we can actually see the growth of the mixing layer in the morning hours and the formation of a stable layer in the evening by analyzing these CO_2 data, very much like what is depicted in Figure 2.5 by Stull. I will show some examples of these observations in Chapter 3 (Figure 3.1 to 3.2)

2.2.3) Boundary Layer (PBL) Depth (Zi)

We defined the nocturnal boundary layer depth as the top of the stable layer, which is indicated by a sharp change of CO₂ value. Using the vertical gradient of the CO₂ mixing ratio to estimate nocturnal boundary layer depth, we first evaluated the differences of CO₂ mixing ratio between each two adjacent levels [Δ CO₂ = CO₂ (h+1) – CO₂ (h)] and tried to find the maximum value for all levels. If the maximum value is below 3 ppm, then the stable boundary layer depth is not defined, or the PBL is well mixed. We then calculated ΔCO_2 from the top down to find where exactly CO₂ mixing ratio increased sharply (the first level from top of the tower where ΔCO_2 was much greater than 3ppm), and we defined this as the top of the stable layer. In the early morning hours, when mixing begins, the PBL depth is usually less than 400 m, so the lower level CO₂ is already well mixed but some top levels are still stable. We can also find this CO₂ jump at the top of the mixing layer by calculating ΔCO_2 from bottom up. Thus, we obtained both nighttime stable boundary layer depth and the early morning mixing layer depth from the CO₂ profiles. However, irregularities occurred quite often so we have to double check our results with the daily plots of CO₂ vertical profiles and time series by eye to determine the PBL depth as accurately as possible.

Also, as described above, daytime mixing layer depth was measured by radar under fair weather conditions. The PBL depth was estimated manually from daily plots of radar reflectivity by Chuixiang Yi at University of Minnesota at St. Paul (currently in Penn State University). Together with our estimated Zi from CO_2 mixing ratio, we produce the full boundary layer depth diurnal cycle as shown in Chapter 3 (Figure 3.6).

2.2.4) CO₂ Flux, Latent Heat and Sensible Heat Flux Data

Measurements of turbulent fluxes of CO₂, latent heat and sensible heat on the WLEF tower are located at three levels: 30 m, 122 m and 396 m, using the eddycovariance method as described by Baldocchi et al. (1987), Verma (1990), Wofsy et al. (1993), etc. The eddy-covariance method provides a relatively direct means of measuring fluxes. In this method the vertical flux of a transported variable at a point is obtained by correlating the fluctuations in the concentration of that variable with the fluctuations in the vertical wind speed. Over a horizontally homogeneous surface, fluxes of sensible heat (*H*), latent heat (λE) and CO₂ (*F_c*) are obtained by equation:

$$H = -\rho C_{p} \overline{w'T'}$$
$$\lambda E = -\lambda \overline{w'\rho_{v}'}$$
$$Fc = -\overline{w'\rho_{c}'}$$

where w is the vertical velocity, T is the potential air temperature, ρ_v is the absolute humidity, ρ_c is the carbon dioxide concentration, ρ is the air density, C_p is the specific heat of air at constant pressure, and λ is the latent heat of vaporization. The over-bars indicate time averages and the primes indicate deviations from the mean.

The flux measurement system at each level, in addition to the CO₂ "profiler" system, includes a sonic anemometer, which measures the three dimensional wind – u, v, w and the virtual temperature – T_v ; a sample inlet for both CO₂ flux and profile measurements; a temperature/water vapor probe for temperature and relative humidity measurements (Berger et al., 2000). These systems provided us with accurate values of the net ecosystem exchange (NEE) of CO₂, over a footprint for the larger area of the forest.

The surface or ecosystem CO₂ flux NEE is computed as the rate of change of CO₂ storage (FCst) plus turbulence flux of CO₂ (FCtb) from observations on the WLEF tower (NEE = FCst + FCtb). For a better understanding of the CO₂ rectifier effect in the atmosphere, we need to compare CO₂ flux (NEE) with the PBL depth and diurnal mean of CO₂ mixing ratio, which will be discussed more in chapter 3 (Figure 3.7A to D).

2.2.5) Monthly mean diurnal cycles

A monthly mean diurnal cycle of PBL depth (Zi) was produced by averaging the value of Zi at each level for the same hour of each day, then averaging over each month (Figure 3.7A, B, C, D). Monthly mean diurnal cycles of CO₂ mixing ratio and CO₂ flux

were also calculated by the same method (Figure 3.4, 3.5, 3.7A-D). Results of these diurnal cycles are compared with each other and also with the CSU single column GCM.

2.3) The Rapid Update Cycle (RUC) Data

We used analyzed weather information from NCEP-RUC to provide lateral boundary forcing for PBL model calculations. The Rapid Update Cycle (RUC), a highfrequency mesoscale analysis and forecast model system, operated by the National Center for Environmental Prediction (NCEP), is designed to provide frequently updated, accurate numerical forecast data for weather-sensitive users, such as aviation forecasts, for the nearest 12-hour period (Benjamin et al., 1999). The RUC assimilates recent observations aloft, such as aircraft data or profilers, and at the surface, such as synoptic data over the United States and surrounding areas, to provide high frequency updates of current conditions and short-range forecasts using a sophisticated mesoscale model. The forcing data we are using to drive the CSU single column GCM (SCM) is obtained from RUC-2, a new version of RUC available since April 1998. RUC-2 produces threedimensional analyses with 1-hour assimilation frequency and 40 km horizontal grid spacing. It has 40 vertical levels in a hybrid isentropic-sigma coordinate for most variables and 37 levels in pressure coordinate (25mb apart between two levels) for temperature, wind, humidity, etc. Table 2.1 lists all the variables extracted from RUC data that are needed for the SCM.

Because each RUC data point is on the 40 km x 40 km grid, I interpolated the data to the WLEF site (Figure 2.6). The data were interpolated with the four nearest grid points weighted by distance. I also use the nearest 16 points to calculate advective tendencies for temperature and moisture, as well as wind divergence.

For an advected variable *T*, the advective tendencies were calculated from:

$$\frac{\partial T}{\partial t}\Big|_{adv} = -\mathbf{V}\Box \nabla T - \omega \frac{\partial T}{\partial p} = -u \frac{\partial T_x}{\partial x} - v \frac{\partial T_y}{\partial y} - \omega \frac{\partial T_{\omega}}{\partial p}$$
(2.1)

where u and v are horizontal winds in m/s, and ω is the pressure vertical velocity in Pa/s at WLEF site. They are all interpolated from the nearest 4 RUC grid points. Horizontal differences of the variable T were calculated with a first order approximation over one grid area and vertical differences were calculated using a centered scheme. Horizontal wind divergence is calculated using a first order solution with 4 grid points and then averaged over 9 grid areas with 16 nearest grid points centered by WLEF site.

Tag	# of Levels	Level	Description[units]
HGT	37	1000mb - 100mb (25mb/lev)	Geopotential height [gpm]
TMP	37	1000mb - 100mb (25mb/lev)	Temperature. [K]
RH	37	1000mb - 100mb (25mb/lev)	Relative humidity [%]
UGRD	37	1000mb - 100mb (25mb/lev)	u wind [m/s]
VGRD	37	1000mb - 100mb (25mb/lev)	v wind [m/s]
VVEL	37	1000mb - 100mb (25mb/lev)	Pressure vertical velocity [Pa/s]
MSLSA	1	sfc	Mean sea level pressure (Std Atm) [Pa]
PTEND	1	sfc	Pressure tendency [Pa/s]
POT	1	2 m above gnd	Potential temperature [K]
UGRD	1	10 m above gnd	u wind [m/s]
VGRD	1	10 m above gnd	v wind [m/s]
TMP	1	2 m above gnd	Temperature [K]
SPFH	1	2 m above gnd	Specific humidity [kg/kg]
PRES	1	tropopause	Pressure [Pa]
POT	1	tropopause	Potential temperature [K]
PRATE	1	sfc	Precipitation rate [kg/m*2/s]
PRES	1	sfc	Pressure [Pa]
HGT	1	sfc	Geopotential height [gpm]
RH	1	2 m above gnd	Relative humidity [%]
PRES	40	hybrid level 1 - 40	Pressure [Pa]
HGT	40	hybrid level 1 - 40	Geopotential height [gpm]
VPTMP	40	hybrid level 1 - 40	Virtual potential temperature [K]
CLWMR	40	hybrid level 1 - 40	Cloud water [kg/kg]

Table 2.1: Variables used from RUC as forcing data for SCM



Figure 2.6: RUC data grid points used to interpolate for the WLEF site. Red dot is the tower location.

To make all data consistent with the pressure coordinate, I also interpolated all the 40-level hybrid data to the 37 pressure levels. The surface pressure at WLEF site is usually around 950 mb, so the first level of the data used starts at 950 mb and then decreases upward with 25 mb interval until 100 mb for a total of 35 levels.

All missing data points, if less than 6 hours consecutively, were filled by linear interpolation, or otherwise filled by calculations from other variables or from the surrounding area. Combining WLEF tower data and RUC data, we were able to produce several months of driving data for the simulations with the single column model (SCM). They are spread throughout July, part of August, September and October 1999.

Grid points near WLEF site

2.4) Model Description

2.4.1) General information

The model used for this study is a single-column model (SCM). It is one grid column of a full climate model – the CSU General Circulation Model (GCM), containing full GCM "physics", but with prescribed forcing by advection. Unlike the global climate model, in which the neighboring grid columns provide information that is needed to determine what will happen with in the grid column studied, the SCM has no neighboring grid columns. Therefore all information needed from neighboring grid columns, such as horizontal advection and divergence of mass, etc, is obtained from observations. This approach allows us to isolate problems with the parameterization from many other components of a global climate model and is an inexpensive way of testing GCM parameterizations.

Although the model includes a land-surface parameterization, I decided to use the WLEF tower data to prescribe the surface latent and sensible heat fluxes. The radiation parameterization of the model was developed by Harshvardhan et al. (1987). The cumulus cloud parameterization is based on the cumulus parameterization of Arakawa and Schubert (1974) and also Lord (1982), revised with the prognostic convective closure and multiple cloud-base levels described by Randall and Pan (1993), and Ding and Randall (1998).

As in the GCM, the SCM uses a stretched vertical coordinate, σ -coordinate (Figure 2.7) (Suarez et al., 1983). The lowest level of this coordinate follows the earth's topography (σ =2), and the top the atmospheric boundary layer (PBL) is the second level (σ =1). So the lowest layer (between σ =2 and σ =1) in the model represents the PBL. The PBL depth is a prognostic quantity.



Figure 2.7: The σ-coordinate in CSU GCM and SCM. There are 17 total layers from the bottom to the top of the model. The PBL is the lowest layer of the model.

The vertical coordinate - σ is defined as:

~

$$\sigma = \begin{cases} \frac{p - p_{I}}{p_{I} - p_{T}} , & (p_{I} \ge p \ge p_{T}) \\ \frac{p - p_{I}}{p_{B} - p_{I}} , & (p_{B} \ge p \ge p_{I}) \\ 1 + \left(\frac{p - p_{B}}{p_{S} - p_{B}}\right) , & (p_{S} \ge p \ge p_{B}) \end{cases}$$
(2.2)

where p_T , p_I , p_B , and p_S represent the pressure at the top of the model, the tropopause, the top of PBL, and the surface, respectively, as shown in Figure 2.7. The PBL depth is then calculated using:

$$\frac{\partial}{\partial t}(\delta p_M) + \nabla \bullet (\int_1^2 \delta p_M \mathbf{v} d\sigma) = g(E - M_B)$$
(2.3)

where $\delta p_M \equiv p_S - p_B$ is the pressure thickness of the PBL *E* is the turbulent entrainment rate at the PBL top M_B is the mass flux into the base of cumulus clouds

The turbulent entrainment rate is calculated from a prognostic equation for the turbulent kinetic energy (TKE) (Randall et al., 1989):

$$g^{-1}\delta p_M \frac{\partial e_M}{\partial t} + E e_M = B + S - D$$
(2.4)

where e_M is the turbulent kinetic energy (TKE)

B is the TKE production by buoyancy fluxesS is the TKE production by shearD represents dissipation of TKE

The production rates are given by

$$S = \int_{1}^{2} \mathbf{F}_{\mathbf{v}} \bullet \partial \mathbf{v} / \partial p \delta p_{M} d\sigma$$
(2.5)

$$B = \kappa \int_{1}^{2} F_{sv} / p \delta p_{M} d\sigma$$
(2.6)

where F_v is the momentum flux vector, v is the wind vector, κ is Poisson's constant and F_{sv} is the turbulence flux of virtual dry static energy.

It can be shown that both *S* and *B* are linear functions of the entrainment rate *E* (Suarez et al., 1983). *S* and *B* are divided into positive and negative contributions to the production rates, *P* and -N, both *P* and -N can also expressed in terms of *E*. Therefore

the vertically integrated conservation law for the turbulence kinetic energy of the PBL (Equation 2.4) can be rewritten as:

$$g^{-1}\delta p_M \frac{\partial e_M}{\partial t} + E e_M = P(E) - N(E) - D$$
(2.7)

The vertically integrated dissipation rate is given as:

$$D = \rho_M \sigma^3 \tag{2.8}$$

Here σ is a dissipation velocity scale and ρ_M is the vertically averaged PBL density.

Two assumptions are used in the conventional entrainment theories:

$$e_M = a_1 \sigma^2 \tag{2.9a}$$

and

$$\rho_M \sigma^3 = a_2 P(E) = D \tag{2.9b}$$

where $a_1 \cong 0.163$ and $a_2 \cong 0.96$, both are dimensionless constant (Randall, 1984). From (2.8), (2.9a) we obtain:

$$D = \rho_M (e_M / a_1)^{3/2}$$
(2.10)

Assuming that the local time-derivative term of (2.7) is negligible, then combining (2.7) and (2.10) we get:

$$e_M E + N = \rho_M \left(\frac{1 - a_2}{a_2}\right) \left(\frac{e_M}{a_1}\right)^{3/2} = \left(\frac{1 - a_2}{a_2}\right) D$$
 (2.11)

Here $e_M E$ is the "storage" term and N is the consumption term.

Use (2.10) and (2.11) in (2.7) we then obtain:

$$g^{-1}\delta p_M \frac{\partial e_M}{\partial t} = P - \frac{\rho_M}{a_2} \left(\frac{e_M}{a_1}\right)^{3/2}$$
(2.12)

Equation (2.12) is used in the present model to predict the TKE (the bulk PBL model). Different expressions of P and N in terms of entrainment rate E, give result in different schemes. Here we will discuss a solution (Randall, et al., 1989) only for the cloud free convective well-mixed layers because of the limitations of the radar measurements for the PBL depth.

Equation (2.11) can be rewritten as:

$$e_{\mu}E + N = \overline{D} \tag{2.13}$$

where
$$\overline{D} \equiv \rho_M \left(\frac{1-a_2}{a_2}\right) \left(\frac{e_M}{a_1}\right)^{3/2} = \left(\frac{1-a_2}{a_2}\right) D$$

For cloud-free mixed layer, the virtual dry static energy flux $F_{sv}(p)$ is assumed to be linear. The consumption rate N can be written as:

$$N = \frac{1}{2} \kappa \frac{\delta p_M}{p_B} \left[\frac{(E\Delta S_v)^2}{(F_{sv})_s + E\Delta S_v} \right]$$
(2.14)

where ΔS_v is the "jump" in virtual dry static energy at the mixed-layer top, assumed to be nonnegative. When $\Delta S_v = 0$, N = 0, then $E = \frac{D}{P} / e_M = \rho_M \left(\frac{1 - a_2}{a_2}\right) \sqrt{\frac{e_M}{a_1^3}}$. Suppose that $F_{sv}(p) < 0$ throughout the PBL, then the consumption rate is given by:

$$N = \frac{1}{2} \kappa \frac{\delta p_M}{p_B} \left[\frac{E \Delta S_v}{P_B} - \frac{(F_{sv})_s}{P_s} \right]$$
(2.15)

where $(F_{sv})s$ is the virtual dry static energy flux at surface. Then the entrainment rate can be expressed by:

$$E = (\overline{D} - N) / e_{M} = \frac{\overline{D} + \frac{1}{2}\kappa \frac{\delta p_{M}}{p_{s}} (F_{sv})_{s}}{e_{M} + \frac{1}{2}\kappa \frac{\delta p_{M}}{p_{B}} \Delta S_{v}}$$
(2.16)

For
$$(F_{sv})_s = 0$$
, $E = \frac{D}{2} / \left(e_M + \frac{1}{2} \kappa \frac{\delta p_M}{p_B} \Delta S_v \right)$, which is positive. So for $E < 0$, the

necessary condition is that $(F_{sv})_s < \frac{-2\overline{D}p_s}{\kappa\delta p_M} < 0$. This means that E < 0 is forced by a sufficiently negative surface buoyancy flux. Because $\overline{D} \sim (e_M)^{3/2}$, so the negative flux $(F_{sv})_s$ increases as TKE increases, and decreases as δp_M increases. So in this bulk PBL

 $(F_{sv})_s$ increases as TKE increases, and decreases as op_M increases. So in this bulk PBL model, it is easier for a deep PBL to undergo a rapid decrease during the evening transition than a shallow PBL originally.

This simple bulk PBL model described above was developed by Randall et al. in 1989 and we now refer this entrainment scheme as R89 scheme.

Another scheme we tested is referred to as K93, and was developed by Krasner, R.D. in 1993, which uses a different method to calculate E. Here I'd like to give a brief discussion about this scheme. For a positive entrainment rate without clouds, E is parameterized by assuming to be proportional to the square root of the TKE (Krasner, R.D., 1993):

$$E = \rho_B \sqrt{e_m} \frac{b_1}{(1 + b_2 R_i)}$$
(2.17)

where $\rho_B = \rho_{M_i}$ the vertically averaged PBL density, and R_i is the Richardson number. For a strong inversion $R_i >> 1$, so we get

$$E = \rho_B \sqrt{e_m} \frac{b_1}{b_2 R_i} \tag{2.18}$$

and for no inversion, Ri = 0, then

$$E = \rho_B \sqrt{e_m} b_1 \tag{2.19}$$

Krasner found that $b_1 = 0.624$ and $b_2 = 0.102$ gave the best fit to observations (Krasner, R.D., 1993).

Negative entrainment rates were parameterized by assuming that E and e_m are small compared to their values during rapid PBL growth. A tunable parameter was introduced to partition the tendency in the PBL integrated TKE into a contribution by the local rate of change of TKE and a weighted contribution by the loss of mass of the PBL. The TKE and E are determined using equation:

$$g^{-1}\Delta p_m \frac{\partial e_m}{\partial t} = weight(B_0 + S_0 - D)$$
(2.20)

and

$$E = \frac{(1 - weight)(B_0 + S_0 - D)}{e_m}$$
(2.21)

where B_0 and S_0 are the surface contribution to the buoyancy and shear.

Both schemes were run with the same forcing data and the results are analyzed in Chapter 3.

2.4.2) Data preparation for the model

The data requirements for the SCM are very challenging. Driving data for the vertical profile are obtained through data assimilation such as RUC data described in the last section. From the RUC data set, we not only can get temperature and pressure at the surface, the time varying vertical profiles of temperature, wind, pressure and pressure tendency directly, but also can produce moisture profile, temperature and moisture tendencies due to horizontal advection for the study site.

All forcing data from RUC contains 35 layers in pressure coordinate and the SCM reads in those layers and then translates them to the 17 layers in the model as the σ -coordinate.

2.4.3) Different forcing methods for prescribing advective tendencies

Consider an arbitrary scalar variable q. We can write the conservation equation in an advective form and the corresponding continuity equation:

$$\frac{\partial q}{\partial t} = -\left(\mathbf{V} \bullet \nabla q + \omega \frac{\partial q}{\partial p}\right) + P \tag{2.22}$$

$$\nabla \bullet \mathbf{V} + \frac{\partial \omega}{\partial p} = 0 \tag{2.23}$$

where P represents the physical processes (sinks and sources) that affect q.

Because the SCM cannot predict the horizontally domain-averaged divergence $\nabla \bullet \mathbf{V}$ and the advective tendency $-\nabla \bullet (\mathbf{V}q)$, we have to prescribe them. There are three different methods for prescribing advective tendencies in the CSU SCM: revealed forcing; horizontal advective forcing; and relaxation forcing (Randall, et. al., 1999). Because relaxation forcing adds an artificial "relaxation" term $(q_{obs}-q)/\tau$ to the right side

of equation 2.1 (where τ represents a specified relaxation timescale) to prevent the predicted value of q from drifting too far away from the observed value q_{obs} , it does not represent any real physical processes. Revealed forcing simply prescribes $-(\mathbf{V} \bullet \nabla q + \omega \partial q/\partial p)$ directly from observations $[\partial q/\partial t = -(\mathbf{V} \bullet \nabla q + \omega \partial q/\partial p)_{obs} + P]$. It is simple, but because temperature changes due to vertical motion are computed directly from observations, it cannot respond to the dry adiabatic vertical motions. Horizontal advective forcing uses the flux form of the conservation equation to obtain the following equations:

$$\frac{\partial q}{\partial t} = -\left[\left(\mathbf{V} \bullet \nabla q \right)_{obs} + q \left(\nabla \bullet \mathbf{V} \right)_{obs} + \frac{\partial}{\partial p} \left(\omega_{obs} q \right) \right] + P$$
(2.24)

$$\omega_{obs}(p) = -\int_0^p (\nabla \bullet \mathbf{V})_{obs} dp$$
(2.25)

For the purpose of this study, I chose to use the horizontal advective forcing. I used observed data from RUC to prescribe $\mathbf{V} \cdot \nabla q$ and $\nabla \cdot \mathbf{V}$, then calculated the advective tendencies of moisture and temperature from above equations.
Chapter 3

Results

3.1) **Observations**

3.1.1) CO₂ Mixing Ratio

Data from the measurements of CO_2 mixing ratio at the tower site have been analyzed carefully for both 1998 and 1999. Results from these analyses give a good picture of the formation and dissipation of the PBL mixing process, and show that the vertical profile of CO_2 mixing ratio is a good indicator of the nocturnal PBL depth.

Figure 3.1 left panel shows the progressive formation of a mixing layer (vertical line) on a typical morning, beginning around 6:00 AM LST (yellow line) and continuously growing deeper until it reaches the top of the tower at about 10:00 AM. We can see from the yellow line that mixing ratio of CO_2 started to drop dramatically from 6:00 AM and by 9:00 AM (brown line), the vertical part of the line had already reached the second highest level of the tower. The right panel of figure 3.1 shows the growth of a stable boundary layer indicated by the sharp change in CO_2 mixing ratio at the lowest levels, which grew higher and higher with each hour. Above the stable layer, CO_2 was still well mixed, indicating the residual mixing layer. Again at 6:00 AM (dark green line), mixing starts at the surface below the stable layer, which still exists above. From these characteristics of the CO_2 mixing ratio profiles, we calculated the height where the CO_2 jump occurs to estimate the height of the stable boundary layer as described in Chapter 2.



Figure 3.1: Examples of vertical profile of CO₂ mixing ratio during two "perfect" days. Left panel: early to mid morning of April 30th, 1998. Right panel: late afternoon through early morning of July 2nd, 1998. Numbers indicated in the small box of each plot are local standard time (LST).

However, "perfect" days like the ones shown above are relatively rare. Looking into daily CO_2 vertical profiles for both 1998 and 1999, we can find some trends but the lines like those in the Figure 3.1 are usually messy and cannot be determined clearly. However, if we analyze the mean value of CO_2 concentration at each level for the same hour of each day, averaged over each month, we can see how the "rectifier effect" takes place. Figure 3.2 and 3.3 depict the diurnal mean CO_2 vertical profiles for January and July of 1998, respectively.





Figure 3.2: Diurnal mean CO₂ concentration vs. height at each local hour of the day for January 1998. Numbers in the small box of each panel indicate local standard time (LST).

We can see from Figure 3.2 that, there is very little diurnal cycle in the wintertime due to reduced photosynthetic activity and weak mixing in the PBL. Soil respiration produces the higher CO_2 concentration observed near ground throughout the day and weak mixing occurs in the afternoon but stays below the top level of the measurement. The range of mixing ratio values between the top and the bottom levels are within 3 ppm. In July, however, we can find a much clearer evolution of the PBL activity (Figure 3.3). During the night, CO_2 concentration builds up continuously by the respiration from plants

and soils and diffuses upward as time extends until it reaches a maximum at 4:00 am. As the sun rises, the ground starts to be heated, and weak vertical mixing process in the early morning brings some lower level CO₂ rich air upward and the CO₂ mixing ratio at the surface starts to drop by mixing with the upper lower CO₂ air (upper right panel of Figure 3.3). Then photosynthetic process in the plants begins and strong vertical mixing process in the PBL starts as well. At 8:00 am we can see that the CO₂ mixing ratio at lowest levels (11 m and 30 m) becomes lower than that above. The CO₂ depleted air at the surface is being brought upward continuously until the mixing process dies down. The bottom left panel of Figure 3.3 was plotted with a different time span, from 8:00 AM to 7:00 PM so we can see a very nice progress of the development of this mixing phenomena with CO_2 depleted air at each hour. As a result, the bottom level CO_2 mixing ratio does not drop as sharply, as it would by photosynthesis if vertical mixing was absent. By 7:00 pm, CO₂ starts to build up again and the same story starts again. The range of the mixing ratio values is from about 351 ppm to 420 ppm, much larger than that in the wintertime, is caused by the photosynthesis and respiration. The 1999 data exhibit very similar behavior (not shown).





Figure 3.3: Diurnal mean CO₂ concentration vs. height at each local hour of the day for July 1998. Numbers in the small box of each panel indicate local standard time. The bottom left panel shows a time range from 8:00 to 19:00, demonstrate a clearer picture of the formation and dissipation of the PBL mixing process.

An overview of the CO_2 mean diurnal cycle for each month throughout the year is presented in Figures 3.4 and 3.5. CO_2 mixing ratio diurnal means are calculated as described in Chapter 2 for each month, then plotted from January to December with local hours marked within each month. From May until October, in both years, 1998 and 1999, we can see that during each afternoon, CO_2 in the boundary layer air is strongly depleted, with the minimum concentration occurring in July. This is caused by both processes - the diminishing CO_2 production due to photosynthesis in the summer days and the deep mixing of PBL air in the same time frame. It is also noticeable that the monthly mean vertical profile is always characterized by a maximum CO_2 mixing ratio at the surface, decreasing upward. The overall surface maximum is strongest during the growing season when the surface is a net sink.



Figure 3.4: CO₂ diurnal cycle at each height for 1998, numbers shown between the months are local hours of the day starting at midnight.



Figure 3.5: CO₂ diurnal cycle at each height for 1999, numbers shown between the months are local hours of the day starting at midnight.

3.1.2) The PBL Depth (Zi)

As described in previous chapter, we can use the characteristics of the CO_2 mixing ratio to estimate nighttime nocturnal boundary layer depth, and the daytime mixing layer depth can be measured by radar under fair weather conditions. Conbining

these data, evolution of a boundary layer depth through the whole diurnal cycle can be produced for many days as shown below.



Aug. 19-20, 1999

Figure 3.6A: Boundary Layer Depth at the WLEF tower site during the summer of 1999.

As we can see from Figure 3.6B, the stable PBL starts to form around 7:00 pm on the 19th of Aug 1999, gradually deepening until 8:00 am the next morning. This is a very calm night, so the depth of the stable PBL rises smoothly. The mixing starts at 7:00 am in the morning of the 20th of Aug 1999, rises quickly to above 2500 m in the early afternoon. The radar signal is not reliable in the late afternoon [Yi, et al., 2000]. When turbulence decays, no clear boundary can be seen at the top of the mixing layer from radar signals (see Figure 2.4). The stable layer does not form until 6 pm, so we have a gap between 4-5 pm, which is quite usual for all Zi values derived from observed data. Figure 3.6B shows two different days in July and September respectively. Sometimes we can see a decrease of the PBL depth and the stable layer forms before the top of the mixing layer disappears, like in the bottom panel of Figure 3.6B, which represents the existence of a residual mixing layer.



Figure 3.6B: Boundary Layer Depth at the WLEF tower site during the summer of 1999. Upper panel: a typical day in July. Lower panel: a typical day at the end of September.

3.1.3) CO₂ Flux

To better understand the CO_2 rectifier effect in the atmosphere, we need to compare CO_2 flux with the PBL depth.

Figure 3.7 top panel shows the CO_2 flux NEE (Net Ecosystem Exchange, calculated from eddy covariance measurements as described in Chapter). Middle panel shows the PBL depth, separated for stable PBL and mixing PBL. Error bars are the standard deviation from the mean diurnal cycle. Obviously daytime mixing layer depth varies significantly from day to day. Bottom panel is the time series of the CO_2 mixing ratios at 6 heights of the WLEF tower site.

By comparing the diurnal mean of CO_2 flux NEE with the diurnal mean of the PBL depth (Figure 3.7, top and middle), we can find that NEE is generally out of phase with Zi, with about 2 or 3 hours phase shift between the minimum flux and the maximum PBL height, for all four months shown here (Figure3.7 A through D, from July to October). Daytime NEE is about two times stronger (negative) than it is at night. Without a strong mixing during the daytime, surface CO_2 mixing ratio would have been very low as the result of photosynthesis. Instead, the plot of CO_2 time series (lower panel) shows a classic "half-wave rectifier" effect as described in Chapter 1, Figure 1.2, especially for the lower levels (11 m to 76 m). Daytime CO_2 vertical gradient is very weak, clearly, a result of the mixing process. Overall, it gives us a time mean vertical profile with the highest CO_2 concentration near surface, despite the fact that there is a large sink at the surface. This effect is strongest in July and weakest in October, as the amplitudes of the NEE and the PBL depth become much smaller.



Figure 3.7A: Diurnal mean of the CO2 flux NEE, PBL depth and CO2 mixing ratio at WLEF tower site. A. July; B. August; C. September; D. October, 1999



Figure 3.7B



Figure 3.7C



Figure 3.7D

3.2) Model Results and Comparisons With Observations

3.2.1) Different Schemes

As we have discussed earlier, the diurnal cycle of PBL depth due to summer daytime mixing is a key element of the "rectifier effect" in the study of CO_2 budget by inverse modeling (Gurney et al., 2002b). Different PBL schemes used in the full GCM give very different results for the PBL depth; therefore the choice of PBL parameterization will directly affect the results of CO_2 inversions. Figure 3.8 shows results from the CSU GCM run with two different parameterizations. As we have described in Chapter 2, K93 is a scheme that computes the PBL top entrainment rate (E) diagnostically and prognoses the PBL turbulent kinetic energy (TKE), and R89 is the classic scheme developed by Randall et al. (1989).



Figure 3.8: Mean Diurnal Cycle of the PBL Depth in July 1999. Blue lines are the observed data at WLEF site for both stable and mixing PBL depth, with error bars indicating the standard deviation.

As we can see from Figure 3.8, the K93 scheme is quite accurate for the prediction of the daytime mixing layer depth, but the nighttime stable layer is very unrealistic. The R89 scheme is pretty good for the nighttime prediction but it is too shallow during the day. Both cases may lead significant errors in predicting other variables that are closely associated with the PBL depth.

Entrainment is the mechanism that brings unmixed free-atmosphere air into the top of the PBL. The entrainment rate is positive if free-atmosphere air is being brought into the top of the mixed layer causing the mixed layer to grow. It is zero if no air is transported at the PBL top and it is negative if there is air being removed from the top of the mixed layer, and then the mixed layer is decaying. The K93 scheme works relatively well during the daytime in full GCM since the entrainment rate is predicted by the turbulent kinetic energy but it appears problematic at night, probably caused by a problem with the negative entrainment parameterization at night when the buoyancy forcing is negative. In the single column GCM (SCM) though, when the surface heating is prescribed from observational data, the problem is more than just at night. Figure 3.9 shows the results from both R89 and K93 schemes. We can see that the PBL depth (Zi) in the K93 scheme sometimes fails to collapse for days while the R89 scheme shows fairly realistic diurnal variation of Zi. For the purpose of CO₂ study, we should avoid using a scheme like the K93 in the full GCM. In this thesis, I will focus only on the R89 scheme, for a detailed analysis of the PBL depth with the single column GCM.



Figure 3.9: Overview of the SCM result with two different schemes

3.2.2) Minimum and Maximum Limits of Zi

The default minimum limit of the PBL depth (dpmmin) in the SCM is set at 10mb. The maximum allowable depth of the PBL (psblim) in the model, expressed as a fraction (σ) of the total atmosphere below 100mb, can effectively limit the growth of the PBL depth and the default is set at 0.2. To investigate the realism of these limits in the SCM, I performed 10 sensitivity experiments with the dpmmin set as 2, 4, 6, 8, 10 mb and σ set as 0.20, 0.25, 0.30, 0.35, 0.40 respectively. The minimum Zi and maximum Zi of each day are then picked out from both the SCM result and the observed data. The ratio of minimum and maximum Zi between SCM and observed data (SCM/OBS) are calculated for each day and then the time mean these ratios is calculated for each value of dpmmin and psblim. Results are plotted in Figure 3.10. The left panel of Figure 3.10 shows that the ratio is closest to 1 when dpmmin is 4 mb. The right panel shows that the " σ " should be 0.25 when the ratio approaches to 1. This means that the SCM results

compare to observed data best when we set the minimum limit of Zi to 4 mb and σ to 0.25 for the maximum limit of Zi.



Figure 3.10: Ratio of the PBL depth between SCM and observed. The left panel is the sensitivity to the minimum PBL thickness and the right panel is the sensitivity to the maximum PBL thickness.

From Figure 3.11 we can see that the maximum limit (σ) for Zi is quite important for an accurate prediction of mixing layer depth during the daytime. When σ is too big, Zi tends to grow too deep and when σ is too small, it will be cut off with a "flat top". This happens quite often during July when the PBL depth is at deepest through the whole year. Obviously, $\sigma = 0.25$ gives the best overall average maximum Zi.



Figure 3.11: Zi from observed data (dark blue) and SCM results with σ set to 0.2, 0.25 and 0.3.

3.2.3) Data Analysis and Comparisons

Figure 3.12 A and B are daily plots of Zi from both observed and SCM results. The SCM results are calculated using the R89 scheme with the minimum and maximum limit set as 4 and 0.25 respectively. Although I have run the simulation for four months, it is very difficult to find good days for the comparison. Many missing data in the forcing data (RUC) need to be filled and the SCM results from those periods should not be compared with the observations. Among those days we have good forcing data, many do not have valid observed Zi data due to bad weather or instrument problems. I was able to obtain some good comparisons in July, August and September but unfortunately, there is not a single good day in October.

Figure 3.12A shows four typical days in July 1999. These four days exhibited similar weather patterns. They were clear days without precipitation. Surface pressures are about 950mb and air temperatures at same hour of the day vary within 1 or 2°C. CO₂ measurements are typical and smooth, with mixing in the morning and a stable layer forming in the evening (see figure 3.13A for July 17, 1999). The observed PBL depth displays a discontinuity in the morning and evening. There are two different definitions of the PBL depth in the observations, one is the mixing layer depth at daytime and another is the stable layer depth at nighttime. The observed data demonstrate very well these two different depths, but the SCM cannot simulate these discontinuities. The model simulates a single layer where grows and shallows continuously through entrainment. When buoyancy forcing becomes negative in late afternoon, negative entrainment reduces Zi smoothly. By contrast the observations show the sudden appearance of a shallow stable layer at this time. The simulated stable layer forms from the "top down" as a result of negative entrainment. The real stable layer is formed from the "bottom up" as negative buoyancy flux undercuts the mixed layer, leaving a deep residual layer above. The simulated stable layer is slightly shallower than observed and does not grow as much

throughout the night. Wind speed on both July 2^{nd} and 6^{th} were higher than other days and it seems that the SCM tends to over-predict the PBL depth during windy or gusting days (see Figure 3.14, some measurements at certain level are missing due to instrument problems). When wind shear increased quickly in the late afternoon, the SCM predicted Zi remained deep (July 2^{nd} and 6^{th}) and maximum was truncated by the limit of maximum Zi that we set in the model. On the nights of July 16^{th} and 17^{th} , there were some gusty winds at night and sharp changes of wind direction. We can see from figure 3.13A that it was difficult to determine the stable layer depth from CO₂ data and it jumps up and down throughout the night until morning mixing starts. July 18^{th} was a very nice calm day and the wind speed was very low throughout the day. The predicted Zi was lower than observed. These results are pretty typical in July. We can say that the SCM is perhaps more sensitive in response to wind (shear) than to sensible heat flux (buoyancy). But we also have to keep in mind that observed mixing layer depth maybe higher on calm days (see Figure 3.15A and discussions).



Figure 3.12A: Daily plots of Zi from both observed data and SCM results for July 1999



Figure 3.12B: Daily plots of Zi from both observed data and SCM results for August and September 1999

Similar results can also be seen in August and September. Figure 3.12B shows two days in August and two days in September. In general daytime Zi becomes much lower as daytime mixing weakens when radiative heating and the resulting buoyancy forcing of TKE starts to drop in the fall. The SCM results tend to be lower than observed for most days in August and September when the weather was calm, except September 25th, a windy day in the fall (see figure 3.14).



Figure 3.13A: CO₂ mixing ratio (ppm) profiles for 17 Jul. 1999



Figure 3.13B: CO2 mixing ratio (ppm) profiles for 26 Aug. 1999



Figure 3.14: Daily plots of wind speed (m/s)

Figure 3.13B is the profile of CO₂ mixing ratio on Aug. 26th 1999. Most of the clear days in the Fall are similar to this with a stable layer remaining still at night and clear mixing in the morning. On a relatively windy night though, like Sep. 25th 1999, the nighttime stable layer cannot be determined clearly because some mixing occurred by gusty winds at night. The higher stable layer depth at night obtained from observed CO₂ data shown on the right bottom panel of Figure 3.12B may not be very accurate. Obviously, SCM cannot address this issue either and produces a very nice smooth stable layer. This may be caused by the forcing data, which are large-scale averages over an hour period. The mixing by the gusts is episodic and it may not be captured by the analyzed wind speed data that were used to drive the SCM. During the daytime of Sep. 25th, simulated Zi is higher and closer with observed data, compared to the days before which are less windy.

In general, PBL depths from the SCM results represent the real world fairly nicely. However, a few problems are noticeable. One of the problems shown in all the daily plots is that there is almost always a slight delay for simulated Zi to reach the maximum compared with the observations. In July, when the buoyancy forcing is usually large, if the wind is also high then the simulated Zi tends to grow much higher but matches the observations remarkably well in the morning mixing period. The maximum simulated Zi for these specific situations is limited by σ as we set in the model so it also appears realistic. However, the simulated Zi tends to reach the maximum 2 or 3 hours later than the observations during calm days or in the fall when the buoyancy forcing is relatively low. Another problem we noticed is that the single layer PBL model is not capable to capture the sudden formation of the stable layer from the bottom up. In fact, it often appears that the SCM predicted Zi stays high until late afternoon before it collapses in the evening. This looks more like a residual mixing layer in the late afternoon. The observed data shows that the top of the residual mixing layer gradually becomes nondistinguishable and occasionally shows a decrease of the residual mixing layer before it dissipates and stable layer may form before the mixing layer decays. This is a little different than what we previously believed, that the PBL collapses after the turbulence stops.







Figure 3.15A: An example of two very calm days. Top: Tower data; Middle: RUC data; Bottom: The simulated and observed PBL depths.

Figure 3.15A shows two extremely calm days in August 1999. Both days displayed relatively large discrepancy during daytime between observed and simulated data. We can see that part of the reason for this discrepancy is that observed Zi are actually very high (compare to Figure 3.12B). It is possible that under very calm conditions, buoyancy forcing becomes more important in the absence of shear, and radar signal reflects a higher top of the mixing layer. In the model, however, under the same buoyancy condition but without enough shear forcing, PBL may not be able to "grow" high enough.

Some significant errors do occur in the model predictions. Figure 3.15B shows one example, from Sep.17th, 1999. All forcing data were good and the instruments worked normally. Observed Zi behaved like other clear days, but the SCM predicted Zi remained deep all night. There was some precipitation simulated by the SCM from 6:00pm until the next morning but none was observed (Figure 3.15c). The SCM usually predicts lower Zi when precipitation is present but sometimes it gives a result like the one in Figure 3.15B. These kinds of results may significantly affect the result of averaged PBL depth diurnal means and monthly means, and may also affect the modeling of the CO_2 rectifier effect. Hopefully in the future we will have more observation sites of this kind with improved measurements, so we can test the SCM with a global distribution, in the end to improve the prediction ability of the full GCM.



Figure 3.15B: An example from the SCM for a perfect day as observed, clear and calm with good forcing data.



Figure 3.15C: Precipitations in the SCM for Sep. 17th, 1999, which corresponds to the hours in the SCM plots between 55-78.

3.2.4) Monthly Mean Diurnal Cycle and Monthly Means of the PBL depth

The mean diurnal cycles of the PBL depth for July, August and September are shown in Figure 3.16. Because observed data are only available for clear days, we left out those points in SCM simulated with precipitation, as well as few obviously bad data (like we see in Figure 3.16) before doing the average. The red line is the raw SCM result and the light green is the selected SCM result. As we can see, in July, the observed data and the simulated data match each other very well. It's not surprising to see that selected SCM simulation is slightly higher than the raw SCM result, because for most raining periods in the SCM, Zi is very low. The time of maximum simulated Zi is about two or three hours later than observed, which we have already noticed in the daily plots. Also the SCM result stays high into the late hours of the afternoon. August and September show the same trend except that they are much lower and the maximum are further behind. There may be different reasons for this but it is certainly interesting too see those wind speeds from the SCM results. Figure 3.17A, B and C show the PBL wind speeds from the SCM for July (1st to 31st), August (16th to 27th) and September (15th to 28th), with the time means as 8.22 m/s, 5.11 m/s and 6.94 m/s respectively. We can see that August has the lowest wind speed, as well as the lowest Zi from SCM.



Figure 3.16: Diurnal Cycle of PBL depth from both observed and simulated data for July, August and September 1999

One thing we might notice is that selected and non-selected SCM data have a larger discrepancy at night in September. This is because there were several bad days like the one in Figure 3.15B, when nighttime Zi does not come down at all. Since the CO_2 rectifier effect is so closely associated with the PBL depth, we should pay a special attention for those nights and hopefully we can eliminate this kind of problems in the future.



PBL WIND SPEED Grid Point (18,33) - Lat: 45.8, Lon: -90.5

Figure 3.17A: PBL wind speed of July 1999 from SCM





PBL WIND SPEED

Grid Point (18,33) - Lat: 45.8, Lon: -90.5 Time mean = 6.94 25.00 22.50 20.00 17.50 15.00 S/M 12.50 10.00 7.50 5.00 2.50 ol O 135 162 189 Elapsed Time (hours) 108 270 297 27 54 81 216 243 335



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Monthly mean Zi from both observed and simulated data are shown in Figure 3.18. Except in July, simulated Zi are lower than observed in both August and September. From the daily plots we can see that in July, simulated Zi tend to over shoot observed Zi on relatively windy days. That's why we have a much higher monthly mean from the simulated data. In both August and September however, simulated and observed Zi are just becoming closer with each other than those calm days, so the monthly averaged SCM result in these two months are much lower than that observed.



Figure 3.18: Monthly mean of the PBL depth

It is unfortunate that we only have three months data that are good enough for the comparison. So the seasonal cycle of the PBL depth is not very conclusive. There is certainly a lot more work that needs to be done. This thesis is a very small part of the study in CO_2 rectifier effect but it's also a very important part since the CO_2 rectifier effect is built purely upon the daily mixing activity in the boundary layer during summer growing seasons.

Chapter 4

Summary and Conclusions

The purpose of this research was to develop a method to evaluate the model prediction of the PBL depth as a building block for the study of the CO_2 rectifier effect. Observational data at a tower site in Northern Wisconsin were used to produce a data set of the PBL depth in the Summer and Fall of 1999. A method was developed to analyze the Rapid Update Cycle (RUC) data for using it as forcing data for the model. The CSU single column GCM was used to simulate the PBL depth and then compared with the observational data.

Conclusions regarding the observations and the performance of the model are as follows:

The use of CO_2 vertical profiling in conjunction with the radar data allowed us to make a complete characterization of diurnal variations in the PBL depth during the study period. CO_2 mixing ratios measured at six levels from 11 to 396 meter can be used not only as a good indicator for the nighttime stable layer depth, but also the mixing layer depth in the early morning when the PBL depth is lower than 400 meters, before the radar can pick up the signal.

Typical diurnal evolution of the PBL depth consists of rapid PBL growth by entrainment in the morning, followed by relatively deep mixing layer in the early afternoon. The evening transition involved the appearance of a stable layer near the ground with decoupling of the deep mixing layer to form a residual layer. It is apparent that the mixing layer depth and the stable layer depth evolves separately.

The SCM results captured many features of the observed variability. Overall results from the SCM were in good agreement with the results from the observations. On average, simulated Zi compared very well to observations in July. In August and September, simulated Zi was lower than observations.

The single bulk PBL in the model cannot capture the discontinuity in the late afternoon because it lacks a separate stable layer. Simulated Zi tends to stay high in the late afternoon and looks more like the residual mixing layer that can sometimes be seen in the observational data. The real mixing layer is usually not clear anymore in the late afternoon, while the stable layer starts to form near the surface. Model results were not able to capture this phenomena

The model appears overly sensitive to shear production of TKE, with overestimation of the PBL depth under windy conditions in July and underestimation of the PBL depth on calm days. Simulated PBL sometimes failed to collapse at night. These events were associated with precipitation in the SCM but were not present in the observations. Large cumulus mass flux exists at the PBL top in the model. The simulated Zi tend to reach the maximum height later than the observations, about 3 hours in July and almost 4 hours in August and September. This may be an important consideration when simulating the CO_2 rectifier effect.

Some thoughts about the future work:

The Rapid Update Cycle (RUC) is a highly useful tool for our model analysis. Unfortunately the current archived data contains a lot of errors and missing data. Improvement of the archived RUC data is highly needed. RUC now is available in 20 km resolution, which will result in considerable improvement in the effects of topography
and land-surface variations on winds and precipitation. It will certainly be worthwhile to adapt this new 20 km version of the RUC and to improve our archived RUC database. Data analysis is very time consuming and significant amount of my time was spent on data analysis. Hopefully in the future we'll have some ready to use RUC data.

More collaborative work between the modelers and the experimentalists should always be considered in the future work. We need more and reliable observed data. Out of four months in 1999, we were only able to obtain about 25 days of useful observed data. Weather conditions and instrument problems are the major obstacle. Collaboration is the only way for us to understand what data are really needed for our model study and what are the limitations of the instruments. With improved and guided observations, we could have better statistics with the daily comparisons between the model and the "real world".

Perhaps we can separate the two different concepts in the model in the future, a stable PBL, and a mixing PBL. Additional work should be done especially for the mixing PBL, to resolve the problem of delayed the growth of the PBL from the model and prolonged high Zi in the late afternoon. One layer for the PBL in the model may be too simple for the most complicated layer in the atmosphere.

The CO_2 concentrations and fluxes should be analyzed along with the full examination of the PBL depth. We can run the full GCM with adjusted limits for the minimum and maximum Zi. It might be also useful to run the GCM with a prescribed Zi diurnal cycle according to observed data and compare the results of CO_2 concentrations and fluxes between the models with prescribed Zi and simulated Zi.

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