THE IMPACT OF A NONLOCAL TURBULENCE SCHEME ON MODELING THE CONVECTIVE BOUNDARY LAYER OBSERVED DURING LAKE-ICE

by

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Abstract

The convective boundary layer is characterized by eddies of varying sizes. Large eddies are able to transfer heat and moisture throughout the depth of the boundary layer. In numerical weather prediction models a turbulence scheme based on standard K-theory can instantaneously transfer quantities only between adjacent grid boxes, along the local gradient of the quantity transferred. Standard K-theory does not account for the behavior and properties of the large eddies in the convective boundary layer. This is especially evident in the case of the convective boundary layer observed in lake effect snow events. Attempts at modeling the convective boundary layer of lake effect snow events has shown that standard K-theory requires a deep superadiabatic layer to develop in order for sub-grid scale turbulence to transfer heat and moisture through the depth of the convective boundary layer. In reality, a superadiabatic layer can only exist in a relatively shallow layer above the surface before convective eddies mix out the superadiabatic layer. By adding a dry nonlocal correction term to standard K-theory formulation, it is demonstrated that the UW-NMS is able to transfer quantities to the top of the convective boundary layer while maintaining a steep lapse rate.

Simulations with and without nonlocal vertical transport are presented for cases where the eddies are completely parameterized and for cases where the eddies are partially resolved. All the simulations with unresolved clouds were quite successful in using nonlocal transport. It is shown that a double accounting problem similar to those encountered with cumulus parameterization are encountered when cloud structures can be partially resolved.
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This work is dedicated to the memory of my Grandmother,
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1. Introduction
The atmospheric planetary boundary layer is defined as the lower part of the troposphere that feels the direct effects of surface fluxes. The atmospheric boundary layer (ABL) is usually capped by an inversion and is thus often viewed as separate from the free atmosphere above. While the ABL may be somewhat separated from the free atmosphere, the continual evolution and structure of the ABL plays an important role in many atmospheric phenomena. The localized structure of the ABL can make the difference in where and if thunderstorms will break out. Some atmospheric phenomena, such as land and sea breezes, are most intense in the ABL. The boundary layer structure is highly dependent on both the underlying surface, as well as the mechanical and convective mixing that occur within.

Numerical weather and climate prediction models must take into account subgrid scale turbulence within the atmospheric boundary layer. The quest for an accurate and efficient representation of the effects of turbulence has been an objective of atmospheric research for some time (Cuxart et al. 1994). Depending on the vertical resolution, a numerical weather prediction model may have relatively few grid points within the atmospheric boundary layer. This makes explicit resolution of three-dimensional boundary layer structures infeasible. Reynolds averaging procedures have lead to the formulation of turbulence "closure" models to represent the sub-gridscale effects. Horizontal resolution is considerably more coarse and generally incapable of resolving atmospheric thermal structure for
mesoscale application. The formulations lead to unknowns in the set of equations for turbulent flow statistics that must be accounted for. Efforts to do so lead to many new and difficult to verify assumptions. The task of representing the turbulent boundary layer accurately becomes a daunting one. The prognostic equations for the zero-order mean variables of wind, temperature, and humidity involve properties of sub-gridscale statistical correlations. First order closure can only transport along the local gradient of a quantity. As computers have become better over the years, higher order closures have been implemented. At the same time, better observations, and large eddy simulations have shown that within the dry convective boundary layer a counter-gradient transport exists. The presence of a counter-gradient transport is not represented by local closure techniques, even at higher orders.

The University of Wisconsin-Madison Nonhydrostatic Modeling System (UW-NMS) has traditionally relied on a local closure scheme for boundary layer diffusion. Both operational and research simulations of various weather phenomena have shown that the UW-NMS has a tendency to allow the growth of superadiabatic lapse rates through deep layers. In a convective boundary layer upward transport by the local closure can be obtained only with the presence of a superadiabatic lapse rate. However, convection does increase the value of eddy diffusivities, which can modify the extent of how superadiabatic the layer is. While the UW-NMS has been able to show some skill in predicting the cloud and precipitation structure in
convective boundary layer events like lake effect snow (Hoggatt 1996), this is achieved through unrealistic lapse rates.

This thesis will show improvement to simulations of a convective boundary layer observed during a lake effect snow event associated with the Lake-ICE field project. The improvements to the simulations (cloud resolving and coarser) were obtained through the implementation of a nonlocal diffusion scheme in the UW-NMS. Chapter Two provides a thorough background to nonlocal turbulence transport. Chapter Three describes the Lake-ICE project motivation and the synoptic setting of 10 January 1998. A description of the UW-NMS that emphasizes the formulation of the local turbulence closure, the nonlocal turbulence closure, and the boundary layer depth prediction is the focus of Chapter Four. The results of simulations performed are found in Chapter Five. Conclusions and direction of future work compose Chapter Six.
2. Background

a. *Nonlocal Turbulent Transport*

Turbulent mixing plays a critical role in the evolution and structure of the planetary boundary layer. The scale on which the turbulent mixing takes place always includes eddies too small to be resolved and thus must be parameterized. Local K-theory is a method for parameterizing the effects of turbulent mixing based on how small eddies will mix quantities along a local gradient of the transported quantity. This method requires a superadiabatic lapse rate in order for potential temperature to be transported upward. Field projects in the 1950’s (Bunker 1956; Webb 1958) found that there was still an upward heat flow within a positive potential temperature gradient. This indicated that large eddies were able to mix heat in a manner that was counter to the local gradient.

Deardorff (1966) was one of the first to look at this apparent counter-gradient heat flux. Deardorff proposed a physical explanation for the counter-gradient transport in a dry convective boundary layer. The proposal was that the counter-gradient heat flux consisted of eroding parcels or plumes of relatively warm air that form in the underlying superadiabatic layer and penetrate a slightly stable region above (Deardorff 1966). The counter-gradient heat flux was deemed necessary in a region called appropriately the "counter-gradient region". In order to account for this counter gradient flux, a new eddy coefficient was defined, and a positive parameter ($\gamma$) was introduced (1.1).
An estimate of $6.5 \times 10^{-4}\text{Km}^{-1}$ for $\gamma$ was suggested. Deardorff believed that while more observations were needed for an accurate estimate of $\gamma$, the estimate he proposed was enough to correct previously calculated potential temperature profiles so that they agreed with observations. The addition of the nonlocal term results in cooling the surface and warming the upper part of the boundary layer. While the term, $\gamma$, was meant for the counter-gradient region, Deardorff suggested that it remained small enough relative to $d\theta/dz$ in superadiabatic layers that it could also be applied there.

Deardorff’s work on the counter-gradient heat flux continued and in 1972 resulted in a “Theoretical Expression for the Countergradient Vertical Heat Flux”. Deardorff contended that in a boundary layer 3km deep, the addition of the counter-gradient term could predict values for $\theta$ at the top of the planetary boundary layer which were as much as 2K warmer than predictions that neglected the counter-gradient term (Deardorff 1972). A slightly different value was proposed for $\gamma$ of $0.7 \times 10^{-5}\text{K cm}^{-1}$. This new value for $\gamma$ was based on a theoretical derivation (1.2), rather than an arbitrary value.

$$\gamma_{c} = \frac{g \left\langle \theta^2 \right\rangle}{\theta' \left\langle w'^2 \right\rangle}$$ (1.2)
Deardorff related the countergradient term to the buoyancy production term. The objective of that paper was not only to provide a physical derivation of the countergradient term but also to provide greater incentive for use of the countergradient vertical heat-flux formulation within numerical models (Deardorff 1972). However, Deardorff did not explicitly describe how to obtain $\left\langle \theta^2 \right\rangle$ and $\left\langle w^2 \right\rangle$ in equation (1.2) making implementation of the counter-gradient transport, from his paper alone, rather difficult.

In the 1980’s there was an increase in the number of atmospheric scientists who believed that standard K-theory was no longer adequate for modeling the

![Figure 1: Depiction of the convective plume (right) and the environment in which it exists (left) by Zhang and Anthes (1982). This method is based on the Blackadar (1978) method. The area denoted by P represents a positive area on a thermodynamic diagram, where N represents a negative area.](image-url)
convective boundary layer. Zhang and Anthes (1982) implemented a convective
plume model under unstable conditions (Figure 1). The convective plume model is
based on Blackadar’s moist PBL model (1978). The convective plume model relies
on the idea that buoyant plumes rising from the superadiabatic layer near the
surface will exchange heat, moisture, and momentum as they mix upward. These
plumes were believed responsible for mixed layer growth and cooling of the
inversion. When visualizing this vertical exchange (Figure 1), it is from the surface
to each level in the boundary layer, rather than between adjacent layers as in K-
theory. The prognostic equations for entropy and moisture are given by:

\[
\frac{\partial C}{\partial t} = \overline{m} (C_a - C_i), \quad C = \theta, q_v, \text{or } q_c
\]

(1.3)

where the mixing coefficient (\(\overline{m}\)) is given by:

\[
\overline{m} = H_1 \left[ \rho_a c_p m (1 - \varepsilon) \int_{z_1}^{h} \theta_v \theta (z') dz' \right]^{-1}
\]

(1.4)

\(H_1\) is the heat flux at the top of the surface layer computed by the Priestly (1956)
equation, \(\varepsilon\) is the entrainment coefficient, and \(z_1\) is the depth of the surface layer.

The variable predicted in the surface layer (\(C_a\)) is predicted by:

\[
C_{a}^{r+1} = C_{a}^{r-1} + \left( \frac{F_s z_1}{mh^2} - \frac{F_s}{mh} + \frac{F_i}{mh} \right) \times \left[ \exp \left( -\frac{\overline{m}h \Delta t}{z_1} \right) - 1 \right] + \frac{F_s \Delta t}{h}
\]

(1.5)

Where \(F_s\) is the surface flux, \(F_i\) is the flux at the top of the surface layer, \(h\) is the
height of the PBL, and \(\Delta t\) is the time-step.
Zhang and Anthes found the buoyancy-driven mixed layer to be highly dependent on surface characteristics. This is not surprising, given that the unstable surface layer is driving plumes that mix throughout the depth of the boundary layer. Forms of Blackadar’s PBL model and Zhang and Anthes PBL model are used today in the MRF, AVN, and MM5.

Transient turbulence was introduced by Roland Stull (1984) as a different theory for explaining mixing by various sized eddies. The word transient was chosen from a Latin word meaning “leap across”, implying that mixing can occur between rudiments separated in space. In terms of a numerical model, this method allows transfer from one grid point to a non-adjacent grid point (Figure 2a). This method keeps track of the total amount of any passive tracer being transported so that the sum of the amount of tracer in all of the boxes is conserved. The final concentration in any one box ($S'_i$), after a discrete time period for mixing ($\Delta t$), is dependent on the sum of the concentration in each of the other boxes multiplied by a transient coefficient ($c_{ij}$):

$$
[S'_i] = [c_{ij}(\Delta t)][S_j]
$$

(1.6)

where $S_j$ represents the original concentration at grid point $j$ and the brackets indicate that matrix operations are appropriate. The transient coefficient is used to represent the amount of the total concentration from each box that is transferred into
each of the other boxes. The transilient coefficients are dependent on the time interval because the larger the time interval the more mixing that can occur. The transilient coefficients make up a matrix that is solved. This method is computationally more expensive than K-theory. However, it is more physical than K-theory because it allows transfer in the cases of a zero gradient or a counter-gradient. Stull’s transilient turbulence, like Deardorff’s counter-gradient method, attempted to represent mixing done by both large and small eddies. Measurements of the transilient matrix for a convective boundary layer were made by Ebert et al. (1989), by running a large eddy simulation model with injected tracers to track the
sources and destinations of air within the model. This work showed the amount of knowledge within a transient matrix, and was supportive of the need for implementation of a transient turbulence scheme in weather predictions models.

While the work of Stull and his colleagues continued to show the importance of nonlocal parameterization through transient turbulence, other groups in the boundary layer community were also exploring the need for something beyond K-theory. The failings of K-theory to capture the planetary boundary layer were explored by Wyngaard (1984 & 1985). Wyngaard used large eddy simulations, observations and results from laboratory experiments in combination with current theory to question where PBL modeling should be headed. Wyngaard questioned if higher resolution models and higher order closures are truly the answer to better boundary layer structure prediction. Wyngaard seemed to be putting out a "call to arms" for development of better boundary layer parameterizations, "One wonders: Is it worth spending 40% of the computation time on a model component with inherent limitations? In some applications the answer is clearly no." (Wyngaard 1985) The LES (large eddy simulations) community understands that parameterizations are for fluxes by eddies on a smaller scale than the grid box. Therefore, algorithms depend on grid scale and parameterized fluxes decrease as resolution increases.

A landmark paper on nonlocal turbulence is that of Troen and Mahrt (1986). Troen and Mahrt present a simple boundary layer formulation intended for use in
large-scale weather and climate prediction models. This paper remains one of the most widely referenced works when dealing with implementation of nonlocal turbulence into an existing model. There are several strong points to Troen and Mahrt's approach. The method used for determining the boundary layer depth does not require a resolution of a capping inversion and allows for continuous transition between the stable and unstable boundary layer. The top of the boundary layer is found iteratively through use of the Richardson number and includes influences of mixing by both shear and surface heating. The counter-gradient correction term is generalized to be consistent with surface layer similarity theory, and uses a velocity scale \( w_s \) that differs from the free convection velocity scale \( w^* \) suggested previously.

The work of Troen and Mahrt paved the way for easy implementation of nonlocal turbulence closure schemes. The ease in implementation as well as small computational cost, made the formulation of Troen and Mahrt more feasible than the transilient turbulence of Stull. Holtslag et al (1990) used Troen and Mahrt's work as a catalyst for designing a high resolution air mass transformation model. The air mass transformation model was used in short-range weather forecasting of boundary layer temperature and humidity profiles as well as boundary layer height and amount of boundary layer clouds. The simple boundary layer model, which included nonlocal transport, was used to characterize the air mass that was advected along trajectories determined by the large-scale flow.
The nonlocal correction term was reevaluated by Holtslag and Moeng (1991) for a physical interpretation that was consistent with similarity theory. Using the heat flux equation, Holtslag and Moeng derived a counter-gradient term that results from the third moment transport effect. Using the results from large eddy simulations, a new eddy diffusivity for heat was obtained in combination with the nonlocal term.

\[ \gamma(z) = a \frac{w_s(w'\theta')_{\text{SPC}}}{w'^2 z_i} \]  

(1.7)

Holtslag and Moeng also address the issue of top-down and bottom-up diffusion. Bottom-up diffusion refers to diffusion driven by surface flux with a zero entrainment flux of the scalar being diffused. Similarly, top-down diffusion is driven by the entrainment flux, with a zero surface flux. While they derive an expression for the nonlocal term in both top-down and bottom-up diffusion, top-down flux budgets allowed the top-down nonlocal term to be empirically set to zero. Therefore, the nonlocal correction term is applicable only in bottom-up diffusion. Holtslag and Moeng suggest that their formulation may "serve as simple, as well as realistic alternatives to the transilient turbulence concept", and note that their goal in this formulation is to model the mean quantities of the boundary layer. Considering that the nonlocal term is meant to parameterize the transport done by large eddies, many of the first scientists to notice the nonlocal transport were those from the large
eddy and boundary layer communities. Thus, it is understandable that attempts to validate the value for the nonlocal term would be done with large eddy simulations.

In the last decade, work with nonlocal turbulence transport has shifted from identifying the magnitude of the nonlocal transfer and attempting to mathematically represent the transport, to testing how well nonlocal parameterizations work compared to local diffusion schemes. Raymond and Stull (1990) implemented a transilient turbulence scheme into the NCAR/Penn State Mesoscale Model (an early version of the MM5). Raymond and Stull believed that numerical weather forecasts could be improved when "one unified turbulence-closure approximation is utilized for the whole model domain at all times, instead of a collection of specialized empirical approximations" (1990). They separated the numerical smoothing done for stability, from the turbulence parameterization. The transilient turbulence scheme which was implemented mandated that the horizontal and vertical K-theory diffusion and the convective adjustments (both moist and dry) be removed from the model. The results of the implementation of transilient turbulence were not very promising. While the scheme was very stable numerically, and improved some of the fields in the boundary layer, it also produced excessive rainfall, an over-exaggerated diurnal variation in precipitation rate and omega fields, an underestimate of surface evaporation, little difference in the forecast at 500mb and above, and was computationally expensive (Raymond and Stull 1990). The authors remained optimistic that computational time could be decreased by using a non-
staggered grid, and that transient turbulence could still be useful in numerical weather prediction.

Holtslag and Boville (1993) used both local and nonlocal boundary layer diffusion schemes in the NCAR Community Climate Model, Version 2 (CCM2). Simulations using each of the schemes were validated against radiosonde observations. The nonlocal diffusion scheme was found to transport the moisture more rapidly away from the surface, and deposited the moisture at higher levels. The local scheme's inability to transport the moisture away from an ocean surface resulted in a tendency for unrealistic low-level saturation and thus fictitious low-level clouds.

Holtslag and Boville were the first among many to begin testing how nonlocal diffusion schemes perform under various conditions. Cuxart et al (1994) compared a transient turbulence model to an exchange coefficient model for four convective days of the HAPEX-MOBILHY experiment. Both models included nonlocal transport, and no conclusion could be reached as to which model performed better. Cuxart et al did comment on the extra computation time required of the transient turbulence model. A local closure, a profile closure, and two nonlocal closures were tested during simulations of an Arctic cold air outbreak by Lupkes and Schlunzen (1996). The nonlocal closures implemented by Lupkes and Schlunzen handled the convective boundary layer better than the local closure and profile closure. Hong and Pan (1996) implemented a nonlocal vertical diffusion
scheme, based on the work of Troen and Mahrt, in the NCEP Medium Range Forecast. The work of Hong and Pan demonstrated that the nonlocal diffusion showed an improvement in predicted boundary layer structure over the local vertical diffusion used operationally. Interestingly, Hong and Pan found that there was little sensitivity to the magnitude of the counter-gradient mixing and the Richardson number. The nonlocal scheme was most dependent on the surface flux, implying that the best nonlocal scheme would be one that works in conjunction with a realistic surface layer model.

The inclusion of nonlocal aspects were tested in a TKE based boundary layer scheme by Belair et al (1999) for the Canadian Meteorological Centre’s CM2. Previous literature on nonlocal transport had discussed implementations in boundary layer schemes based primarily on similarity theory. Belair et al found little improvement to the PBL simulations due to nonlocal inclusion. Several different formulations of the nonlocal term (Deardorff 1972; Troen and Mahrt 1986; Mailhot and Benoit 1982; Holtslag and Moeng 1991) were tested with little variation in the end result. Belair et al found better improvements through alternative formulations to turbulent length scales. This was the first documented effort found to use the similarity theory based nonlocal term and apply it to a system that used eddy diffusivities based on TKE.

Recently, nonlocal boundary layer closure schemes have been tested with regard to very specific weather phenomena rather than the general guise of a
convective boundary layer. Basu et al. (2002) and Bright and Mullen (2002) both tested the impact of nonlocal turbulence closure in monsoon simulations. Basu et al. looked at the impact on the Indian monsoon, while Bright and Mullen's simulations were for the southwest U.S. monsoon. Both of these papers illustrate the important role that the boundary layer can play in weather phenomena. The prediction of the deep convective storms of the southwest U.S. monsoon is highly dependent on accurate predictions of convective available potential energy (CAPE), downdraft convective available potential energy (DCAPE), and convective inhibition (CIN). CAPE, DCAPE, and CIN can all be significantly changed by small changes to the structure of the boundary layer. Bright and Mullen found that the nonlocal diffusion schemes of the MM5 better predicted the boundary layer structure and consequentially the amount of CAPE and DCAPE. Basu et al. found that the use of a nonlocal scheme in the NCMRWF resulted in little change to the predicted flow pattern, but did make a noticeable improvement to the skill when predicting the precipitation maximum.

Over the last few decades, scientists have gained a better understanding of the convective boundary layer. One of the unique characteristics of the convective boundary layer is the behavior of large eddies. Large eddies possess the ability to transport heat and moisture throughout the depth of the boundary layer, at times in a direction counter to the local gradient. This counter-gradient transport has lead to the need to reformulate boundary layer turbulence closure schemes. There is a need
to represent this counter-gradient transport, while remaining physically consistent with the rest of the turbulence, and still maintain computational efficiency. Several different approaches have been put forward to the modeling community; transilient turbulence, convective plume models, and nonlocal transport terms. The various approaches have varying degrees of ease in implementation, as well as varying computational time. The ability of nonlocal schemes to accurately predict boundary layer structure has been demonstrated in multiple instances. The testing of nonlocal schemes over local schemes has been done at operational resolutions. The effect of an accurate boundary layer structure on weather phenomena in the free atmosphere (such as the deep convection of the southwest monsoon) has only recently been explored. While the formulation of the nonlocal term has been validated by large eddy simulations (Holtslag and Moeng 1991; Ebert et al. 1989), and tested for accuracy over local diffusion schemes (Basu et al. 2002; Bright and Mullen 2002; Cuxart et al 1994, etc), nonlocal turbulence parameterization has yet to be tested against explicit cloud resolving simulations. The testing of nonlocal parameterizations thus far have been from two points of view; the small scale large eddy simulators and the those who use the parameterization in global and operational scale weather prediction models. Examination of the nonlocal transport by large eddies from the viewpoint of a cloud resolving model has not yet been done.
b. *Lake Effect Snow*

Lake effect snow events have been greatly studied over the years. These events were studied not only because of the obvious human impact that these events have, but also because they give scientists an idealized opportunity to study the convective boundary layer. In winter, stable arctic air masses approach the Great Lakes, and are greatly modified as they traverse the lakes. Lake effect events can manifest themselves in various forms depending on the synoptic situation and prevailing winds. The lake effect clouds may become arranged in multiple wind/shear parallel bands, one large shore parallel band, or a mesovortex (Hjelmfelt 1990). Regardless of which cloud structure it arranges into, lake effect snow events occur in a convectively driven boundary layer. The size of the lakes over which the lake effect snow occurs are small enough to allow the entire convective boundary layer to fit easily in the domain of a numerical model.

Attempts at modeling lake effect snow has generally fallen into two categories; 1.) modeling efforts aimed at improving forecasting of lake effect snow amounts and locations 2.) modeling efforts aimed at creating a better understanding of the theory behind what creates the convective roll structure of the clouds. The work of Lavoie (1972) and Hjelmfelt (1990) showed the importance of prevailing wind flow (speed and direction), temperature lapse rates, and height of the capping inversion on the morphology of lake effect snow events. These findings were applicable knowledge for weather forecasters in the Great Lakes region.
Kelly (1984) used observations taken during the University of Chicago Lake Snow Project, in conjunction with output from numerical weather prediction models to study a wind parallel lake effect snow event. The observations were closely examined and related to current theories on roll convection. Kelly found a positive correlation between the Rayleigh number and wavelength of the convective rolls, and only a weak correlation between the Richardson number and wavelength. This led to the conclusion that shear instability was the primary determinant of wavelength, with modification imposed through thermal instability (Kelly 1984).

The role of shear versus thermal instability remains at the crux of understanding the convection organization in the boundary layer during lake effect snow events. Cooper et al. (2000) further explored the importance of shear in the lake effect environment, where strong surface heating occurs. Many theories for roll convection have been proposed over the years; inflection point instability (Brown 1972), wind shear curvature (Kuettner 1971), gravity wave interaction (Clark et al. 1986), etc.. Cooper et al. felt that many of the past studies used to support the various theories had been conducted in a slightly to moderately unstable boundary layer, and "do not appear to hold true for the lake-effect environment when strong surface heating is present." While the work of Cooper et al. furthered the knowledge of convective roll environments, it did not answer all of the questions surrounding clouds in the lake effect convectively driven boundary layer.
In recent work, Tripoli (in preparation) studied, through numerical simulations, the cloud structures of the convective boundary layer observed during a Lake-ICE intensive observation period. Tripoli performed a series of cloud resolving simulations of the evolution of moist convective rolls in a lake effect snow boundary layer. An idealized environment was designed by Tripoli in order to test the sensitivities of the banded cloud structure associated with roll convection. The cloud resolving simulations performed by Tripoli provided an excellent opportunity to test a diffusion scheme employing nonlocal turbulence against a cloud resolving model.
1. Lake-ICE
   a. Motivation for Field Project

   The Lake Induced Convective Experiment (Lake-ICE) was conducted during the 1997/1998 winter season lasting from December 4th, 1997 to January 19th, 1998. Lake-ICE took place in order to study the unstable boundary layer over Lake Michigan during wind parallel lake effect convection. The boundary layer in these events is characterized by roll convection (Figure 3). The location selected for the Lake-ICE experiment was central Lake Michigan. This location was selected for several reasons. This location was believed to represent an "ideal laboratory" for studying the boundary layer convection because arctic air masses approach the

Figure 3: (taken from Stull, 1988) Lake effect snow events over Lake Michigan are often characterized by roll convection, which aligns the convective clouds in cloud streets.
western shore of central Lake Michigan without having passed over any large lakes that may have modified the boundary layer structure. The fetch of Lake Michigan is large enough to support lake effect snow development (80 km fetch is typically needed to form lake effect snow), yet was still an observable domain. The entire transition from a stable arctic air mass to an unstable boundary layer convectively driven by the warm waters of the lake is observed over the relatively short distance of Lake Michigan. Observations were taken on both the windward and leeward shores of Lake Michigan by various instruments, and over Lake Michigan by aircraft during Intensive Observation Periods (IOPs).

b. 10 January 1998

On 9 January 1998 a cold front associated with a cyclone in southern Canada descended into the Great Lakes region of the United States. Extremely cold air followed the frontal passage, and by 00z 10 January 1998 surface temperatures on the western shore of Lake Michigan were a chilly 0°C. By 12z 10 January 1998 the cold front was just to the east of Michigan (Figure 4) and arctic air continued to funnel into the Great Lakes Region. Temperatures on the western shore of Lake Michigan were around -18°C. Upstream temperatures in northern Minnesota were as cold as -26°C. With Lake Michigan water temperatures as warm as 5°C, a convectively unstable environment developed over Lake Michigan. Wind speeds near the surface of around 5ms⁻¹ helped to increase the heat flux off of Lake Michigan. This environment was conducive to lake effect snow, and the Lake-ICE
Figure 4: By 12z, 10 January 1998, the cold front had already passed over Lake Michigan. Behind the cold front, arctic air was being brought over the warm waters of Lake Michigan.

project conducted one of its IOPs. On 10 January 1998 both the King Air and Electra flew over Lake Michigan. The UW-Lidar and the Penn State Cloud Observing System (PSU-COS) were in operation. Radiosondes were launched on both the windward shore (Sheboygan, WI) and the leeward shore (Montague, MI and Greenville, MI) of Lake Michigan. Super-rapid scan satellite images captured the banded cloud structure of the wind parallel event at 1km resolution in 1-minute increments (Figure 5). While this lake effect snow event was not particularly strong in terms of the amount of snow that fell, the influence of El Nino on the winter
season made this event one of the first good events for the Lake-ICE project. The lake effect snow bands persisted for several hours.

Several interesting features are noticeable on the satellite image from 10 January 1998. The clouds are orientated in a band structure parallel to the wind. These bands have a horizontal spacing of about 6km, and develop in a boundary layer with a maximum depth of 1.3km. The bands start over the warm Lake Michigan waters, but are most distinctive over the land of Michigan. The mechanism(s) behind the wind parallel bands remains a point of contention and is being actively studied. Perpendicular to the 6km bands are what appears to be
gravity waves. While noticeable on Figure 5, these waves are more clearly distinguishable on satellite animations. Unlike the wind parallel bands, the shore parallel gravity waves are most distinctive near the windward shore and become harder to distinguish farther downwind. The cloud structure as a whole starts to be visible a few kilometers from the windward shore (Figure 5). Cloud resolving simulations with the UW-NMS have found it difficult to reproduce the cloud structure so close to the windward shore (Tripoli, in preparation). This is because even in a convectively unstable boundary layer, like the one over Lake Michigan on 10 January 1998, the UW-NMS relies on local gradients to transport heat and moisture. The large heat and moisture fluxes off of Lake Michigan can diffuse their way up in the convective boundary layer only through a growing absolutely unstable layer with a superadiabatic lapse rate. The inability of the UW-NMS cloud resolving simulations to reproduce the cloud field close to the windward shore, led to the idea of a nonlocal term addition to the turbulent diffusion scheme. The goal of implementing a nonlocal term was to account for the large eddies that transport the large heat and moisture fluxes through the depth of the boundary layer.
4. University of Wisconsin-Madison Nonhydrostatic Modeling System (UW-NMS)

The University of Wisconsin-Madison Nonhydrostatic Modeling System (UW-NMS) is a quasi-compressible, nonhydrostatic model. The UW-NMS is fully scalable (with proper adjustments to sub-gridscale parameterizations) and has been used for simulations ranging from operational forecasts over the continental United States to cloud resolving simulations to large eddy simulations with resolutions smaller than 50m. Dynamic properties of flow such as vorticity, kinetic energy, and enstrophy are conserved by the advection scheme in the UW-NMS. A summary of the dynamics and numerics used in the UW-NMS is summarized in Table 1. The local diffusion scheme, the nonlocal diffusion scheme, and the formulation for

<table>
<thead>
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<th>Table 1: Summary of UW-NMS dynamics and numerics.</th>
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<td>• Variably Stepped Topography (see Appendix I.)</td>
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<td>• Integrates using a hybrid 2nd order leapfrog (dynamics) and Crowley (scalars) advection scheme employing a time-split compressible dynamics scheme.</td>
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<tr>
<td>• Thermodynamics based on prediction of moist ice-liquid variable, ( \tilde{\theta}_i ) (Tripoli and Cotton, 1981)</td>
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determining the boundary layer depth are described in detail in the following sections.

a. Local Diffusion Scheme

The UW-NMS employs a local diffusion scheme based on K-theory. This approach parameterizes subgrid-scale fluxes by making them proportional to the local gradient of the transported quantity. The name K-theory comes from the use of an eddy diffusivity, which is notated as $K$. The eddy diffusivity ($K$) is multiplied by the vertical gradient of the transported quantity ($\partial c/\partial z$) to estimate the vertical flux ($w'c'$).

$$\frac{\partial C}{\partial t} = \frac{\partial}{\partial z}[w'C'] = \frac{\partial}{\partial z}[\frac{K_c \partial C}{\partial z}]$$  \hspace{1cm} (2.1)

The eddy diffusivity, $K_c$, is based on predicted turbulent kinetic energy (TKE) in the UW-NMS. This differs from many mesoscale models which use a set profile for $K$ within the boundary layer. The set profile is often based on similarity theory. Basing the eddy diffusivities on the turbulent kinetic energy makes the assumption that eddy diffusivities are similar to molecular diffusivities. The UW-NMS local diffusion is a 1.5 level closure. This level of closure retains both prognostic equations for zero-order statistics such as temperature and humidity and equations for the variance of those variables. The vertical and horizontal eddy diffusivities are
set to use the same scale length based on the vertical grid spacing. This effectively produces a minimal sub-grid scale turbulence mixing compared to what many LES and cloud models employ, but is sufficient for this model because the eddy mixing is not being used to combat excessive numerical enstrophy cascade present in other numerical models finite differenced in the momentum form. The eddy diffusivity for momentum \( (K_m) \) is diagnosed from predicted turbulent kinetic energy and given by:

\[
K_m = c_1 l e^{1/2}
\]  

(2.2)

where \( c_1 = 0.21 \), \( l \) is the scale length set to the vertical grid scale, and \( e \) is the turbulent kinetic energy. The eddy diffusivity for heat is proportional to the eddy diffusivity for momentum.

\[
K_h = 3K_m
\]  

(2.3)

The turbulent kinetic energy \( (e) \) is predicted in a manner consistent with the standard form for a predictive scalar specific mass variable (Tripoli), with the source term given by the following set of equations:

\[
S_e = S_m + S_b + S_d + S_c
\]  

(2.4)

\[
S_m = c_1 l D^2 e^{1/2}
\]  

(2.5)

\[
S_b = -c_1 l \frac{K_h}{K_m} N^2 e^{1/2}
\]  

(2.6)
The total source term for TKE is based on mechanical ($s_m$), buoyancy ($s_b$), divergence ($s_d$) and dissipation ($s_e$) source terms. For these source terms (equations 2.5-2.8), $D$ is the three-dimensional deformation computed from the predicted velocity field, $D_v$ is the three-dimensional velocity divergence, and $c_e = 0.7$ is the dissipation constant. The Brunt-Vaisala frequency ($N^2$) used in computing the TKE source terms is rather complex. The Brunt-Vaisala frequency used allows the TKE prediction to account for moist subgrid scale processes. This is accomplished by using a moist water loaded virtual temperature ($\theta_v$) which takes into account the specific humidities of vapor ($q_v$), liquid ($q_l$) and ice ($q_i$) precipitation. The Brunt-Vaisala frequency used also has a vertical temperature adjustment ($\alpha$) that accounts for the effects of condensation and evaporation to the potential temperature gradient. The multifaceted Brunt-Vaisala frequency ($N$) employed is given by:

\[
N^2 = \frac{g}{\theta_v} \left( \theta_{vy_2} - \theta_{vy_1} \right) - \alpha \Delta z
\]  

(2.9)

\[
\theta_{vy} = \theta \frac{(1 + 0.61q_v)}{(1 - q_l - q_i)}
\]  

(2.10)
\[ \alpha = -\frac{L_{ab} \Delta q_v}{c_p T} \] (2.11)

where \( L_{ab} \) is the latent heat of a phase change for a process going from phase a to phase b, \( T \) is temperature, \( \Delta q_v \) is the change in mixing ratio for a rising parcel and sinking parcel and \( c_p \) is the specific heat at a constant pressure.

K-Theory/Local Diffusion

Nonlocal Diffusion

Figure 6: K-theory can mix only along the local gradient (darker colors represent warmer temperatures), whereas nonlocal diffusion can transport entropy, despite the local gradient (nonlocal transport in dashed lines). This allows a warm surface that would produce deep thermals to transport entropy and moisture thru regions whose local gradient opposes the transport.

This local method for handling diffusion (2.1) is also known as a gradient scheme because the transport relies on the local gradient of the scalar being
transported. Figure 6 illustrates that in local diffusion/K-theory transport is done
only between adjacent boxes, and only along the gradient of the transported
quantity. When the local gradient is zero, no transport can occur. In the real
atmosphere, convective thermals are known to penetrate into stable regions. K-
theory leaves no room for this type of phenomena.

\textit{b. Nonlocal Diffusion Scheme}

The convective boundary layer is often dominated by large eddies. These large eddies are capable of transporting quantities in a way that can oppose
the local gradient (Figure 6). This so-called "counter-gradient" transport cannot be
represented in standard K-theory. Nonlocal diffusion schemes proposed by
Deardorff (1973), Troen and Mahrt (1986), Holtslag et al. (1990), and Holtslag and
Boville (1993) add a correction term to the turbulence diffusion equation for
prognostic variables (2.1), resulting in:

\[
\frac{\partial C}{\partial t} = \frac{\partial}{\partial z} \left[ w'C' \right] = \frac{\partial}{\partial z} \left[ -K_c \left( \frac{\partial C}{\partial z} - \gamma_c \right) \right] \tag{2.12}
\]

where $\gamma_c$ is the correction to the local gradient. This correction term is an attempt to
represent the contribution of large eddies to the total flux. It should be noted that
while this term is often referred to as the "counter-gradient" term, it only transports
entropy "counter-gradient" in regions where the boundary layer is stable. In the unstable regions of the boundary layer the nonlocal term transports in the same direction as the local gradient. For the transport of moisture the term "counter-gradient" is even more ambiguous, as the moisture gradient in the boundary layer is almost always directed upward throughout the depth of the boundary layer. The large eddies in a convective boundary layer still transport moisture and need to be accounted for, but they are also transporting the moisture in the same direction as the local moisture gradient \( \left( \partial C / \partial z \right) \). The physical limits of the nonlocal term when applied to moisture will be discussed more later on. The nonlocal diffusion scheme is only applicable in the boundary layer for convective conditions where the surface flux of potential temperature is positive. In the free atmosphere above the boundary layer, standard local diffusion is applied.

Several formulations for the nonlocal correction \( \gamma_c \) have been proposed over the years. Deardorff first proposed a value \( \gamma \) based on aircraft observations:

\[
\gamma = 0.7 \times 10^{-3} \ (Km^{-1})
\]  

(2.13)

Troen and Mahrt (1986) propose the formulation:

\[
\gamma = a \frac{\left( w' \theta \right)_{SFC}}{w \cdot h}
\]

(2.14)
Where \( a \) is a proportionality constant, \( h \) is the height of the top of the boundary layer, and \( w_s \) is a velocity scale. The formulation by Troen and Mahrt (1986) is based on similarity theory.

The nonlocal diffusion term used in the UW-NMS follows that of Holtslag and Moeng (1991):

\[
\gamma = a \frac{w_s \left( \frac{w' \theta'}{\theta_{sfc}} \right)_{sfc}}{w'^2} \frac{1}{z_i} \tag{2.15}
\]

\[
w_s = \left( \frac{g}{\theta_i (k1)} z_i \frac{w' \theta'}{\theta_{sfc}} \right)^{1/3} \tag{2.16}
\]

where \( k1 \) represents values at the lowest grid box. A value of \( a=2 \) was used. Holtslag and Moeng (1991) used a velocity variance based on LES simulations, which had the following form:

\[
\left( \frac{w'^2}{u'^2} \right)^{3/2} = \left[ 1.6 u_i^2 \left( 1 - \frac{z}{z_i} \right) \right]^{3/2} + 1.2 w_i^3 \left( \frac{z}{z_i} \right) \left( 1 - 0.9 \frac{z}{z_i} \right)^{3/2} \tag{2.17}
\]

The first term is meant to describe mechanical shear turbulence induced by the surface. The second term expresses the effect of convection. In UW-NMS simulations using equation (2.17), there was an improvement to the structure of the convective boundary layer over Lake Michigan. However, the boundary layer still remained dominated by negative Richardson numbers and a superadiabatic lapse rate. The velocity variance profile (2.18), given by Stull (1988), based on similarity
theory for a convective well-mixed layer was applied to the UW-NMS with much greater success:

\[
\frac{\overline{w'^2}}{w'^2} = 1.8 \left( \frac{z}{z_i} \right)^{2/3} \left( 1 - 0.8 \frac{z}{z_i} \right)^2
\]  

(2.18)

The nonlocal term is applied to the scalars of potential temperature and moisture. The nonlocal term remains dependent on the thermal structure, regardless of the variable being transported. The difference in the nonlocal term for each variable is dependent only on the differences in surface flux. Several restrictions are put on the nonlocal term, in order to make sure it represents the real world. In a convectively stable atmosphere \((\overline{w'\theta_{sfc}} \leq 0)\), the convective velocity scale \((w_c)\) is not applicable, and the nonlocal term is not applied \((\gamma = 0)\). The nonlocal flux term is not allowed to be larger than the surface flux. This assumes that all of the large eddies are transporting moisture and entropy that originates from the surface. The nonlocal term is also not allowed to exceed 0.005 Km\(^{-1}\). This bound is placed on the nonlocal term mostly as a check to keep errors from growing.

c. Depth of the Boundary Layer

The nonlocal diffusion term \((\gamma_c)\) is both dependent on the boundary layer depth \((h)\), and applied throughout the depth of the boundary layer. In order for the nonlocal flux term to be applied in a manner that makes physical sense, a good estimate of the boundary layer height is needed. If the depth is
underestimated, then the large eddies the counter-gradient term represents would not be mixing through the depth of the boundary layer. If the height of the boundary layer is over-estimated, the nonlocal term may transport moisture and entropy into the capping inversion layer, resulting in an unrealistic moisture profile and a strengthening inversion. The depth of the boundary layer is determined using the following equation from Troen and Mahrt (1986):

\[
h = Ri_c \left( \frac{\theta_s(k) |U(h)|^2}{g(\theta_s(h) - \theta_s)} \right)
\]

where \( Ri_c \) is the critical Richardson Number with a value of 0.25. The boundary layer height is represented with \( h \). Values evaluated at \( k1 \) are taken at the lowest grid box. The surface potential temperature (\( \theta_s \)) is meant to represent the surface air temperature.

Several different methods were attempted to represent \( \theta_s \). Using the skin temperature of the ground resulted in an overestimate of the boundary layer height. Using the potential temperature value at the lowest grid point underestimated the boundary layer depth. This is because potential temperature value is in the middle of the grid box, and thus the potential temperature at the lowest grid box may represent a temperature that is 50m or more off of the ground. The shelter temperature, which is used in the UW-NMS for analysis, represents the temperature a few meters off of the ground. The shelter temperature also failed to adequately
represent the $\theta_s$ needed to correctly predict the depth of the boundary layer. The formulation for the surface potential temperature $(\theta_s)$ that worked best was that based on Troen and Mahrt (1986).

\[ \theta_s = \theta_s (k1) + \theta_r \]  
\[ \theta_r = C \frac{\left( w' \theta_s \right)_{SFC}}{w_s} \]

The term $\theta_r$ is the scaled virtual temperature excess near the ground. The coefficient of proportionality ($C$) uses a value of 8.5 after Holtslag et al. (1993). The velocity scale $w_s$ corresponds to the thermal turnover time (Troen and Mahrt 1986). The velocity scale is based on the surface frictional velocity scale, $u_s$, and the dimensionless wind shear, $\Phi_m$.

\[ w_s = u_s \Phi_m^{-1} \]
\[ \Phi_m = \left( 1 - 15 \left( \frac{0.1 z_s}{L} \right) \right)^{-1/3} \]
\[ L = \frac{-u_s^3 \theta_{skin}}{k g w' \theta} \]

Where $L$ is the Monin-Obukhov length and $k$ is the Von Karman constant ($k = 0.4$). The value for $\theta_r$ is limited so that it may not exceed a maximum value of 3K. In the case where the surface flux is negative, $\theta_s$ is set equal to $\theta_r (k1)$. This method for determining the boundary layer depth requires multiple iterations, as the solution
for $h$ is dependent upon $h$. An initial guess of 3km for the boundary layer height is used. If the predicted height is less that 3km the next lowest grid box is guessed.

The model continues to guess lower until the predicted boundary layer depth is greater than the last used depth. The opposite occurs if the first predicted height is higher than the first guess of 3km. In that case the model guesses consecutively higher grid boxes, until the predicted height is lower than the last used depth. However, in the case of the convective boundary layer observed during Lake-ICE, and simulated for this paper, the boundary layer depth was always under 3km in height.

Figure 7: The method employed in The UW-NMS to diagnosis boundary layer depth (after Troen and Mahrt) predicts the boundary layer height to be 1.05km at the grid point this sounding is taken from. The arrow identifies where 1.05km is located. This sounding is the observation taken from Sheboygan, WI on 10 January 1998 during Lake-ICE.
With this method there is no need to explicitly resolve any capping inversion on the boundary layer. Also, this method can work for boundary layers that contain both stable and unstable temperature profiles within them. Boundary layer depths predicted by this method agree well with the depth determined subjectively from forecasted soundings (Figure 7). The boundary layer depth is carried as a new variable in the UW-NMS, which allows for further applications in other areas where the height of the boundary layer is important, such as air pollution applications.
5. Simulations

a. Experiment Design

The set of experiments conducted were designed to test the use of the nonlocal diffusion scheme against simulations with explicitly resolved convection. Nonlocal diffusion schemes have been tested at operational resolutions in models and validated by large eddy simulations (Holtslag and Moeng 1991; Ebert et al. 1989; Basu et al. 2002; Bright and Mullen 2002; Cuxart et al. 1994). The set of experiments presented in this chapter were designed to test how well nonlocal diffusion on the coarse resolution can produce the same effect as explicitly resolved convection.

An idealized cloud resolving simulation (\(\Delta x = \Delta y = 400m\)), of the lake-effect snow event of 10 January 1998 performed by Tripoli (in preparation), produced the banded cloud structure similar to what was observed by satellite. The cloud resolving simulation of Tripoli (here after referred to as the CR-simulation) was able to resolve the 6 km wind parallel cloud bands and the 12km shore parallel gravity waves. In order to use the CR-simulation of Tripoli as a “truth” to the convective structure of 10 January 1998, simulations at coarser resolution (\(\Delta x = \Delta y = 2000m\)) were set up in the same manner.

The grid was placed over the central part of Lake Michigan (Figure 8a). The grid used includes the distance across the lake as well as enough land on either side (determined by observations) to capture the transitions between the land and water.
On the eastern side of the lake there is enough land in the grid box to examine if the banded cloud structure remains over the land as observed by satellite (Figure 5). A cyclic boundary condition was used in the $y$-direction in order to give the lake a seemingly endless length in the north-south direction. This was necessary for the CR-simulation in order to make the lake longer without having to worry about meridional variations in atmospheric soundings. This also kept the number of grid points in the CR-simulation down, which shortened computational time. While the cyclic boundary could have been neglected with the coarse resolution, in favor of more grid points in the $y$-direction, the cyclic boundary condition was kept for consistency between

![Figure 8: Grids used for Lake-ICE simulations. Grid A was used for control and most of other simulations. Grid B was used for the simulation with more land. Dashed arrow shows location of east-west cross sections.](image)
simulations. In the vertical direction a resolution of 100m was used for the first 1.2km and then slowly stretched to a 750m spacing (Tripoli, in prep.). An absorbing layer was applied to the western edge of the domain. This layer was applied to be consistent with the work of Tripoli (in preparation). The absorbing layer was applied to keep the gravity waves parallel to the shore from growing, and to keep the flow approaching the shore laminar. The flow approaching the lake was kept laminar to prevent localized influences on turbulence over the land from playing a role in the convective boundary layer.

A sounding (Figure 7), taken by a radiosonde launched during the 10 January 1998 IOP of Lake-ICE, was used to initialize the simulations. The radiosonde was launched on the windward shore of Lake Michigan, near Sheboygan, WI. Before the sounding was used for initialization of the model it was run in a 1-d dry simulation for 72 hours in order to determine the gradient wind that matched that sounding (Tripoli, in preparation). The empirically derived gradient wind was used with the sounding to initialize the UW-NMS with a horizontally homogeneous environment. Initializing with the observed sounding means that initially the environment across the lake is unrealistic, as the lake greatly modifies the air as it traverses the lake. The simulations were run for 6 hours, or about twice the time scale it takes flow at the top of the boundary layer to traverse the entire domain. This was to give enough time to create a realistic relationship between position and history over the lake. The
lake temperature was varied in a manner meant to be consistent with satellite derived sea surface temperatures obtained during Lake-ICE.

The cumulus parameterization was turned off for these simulations. Once again, this was to remain consistent with the experimental design of Tripoli, where clouds were being explicitly resolved and a parameterization of them was unnecessary. Explicit microphysics were used to predict the formation of pristine ice crystals, aggregate crystals, and rimed crystals. Rain and graupel were ignored for these cold cloud cases. The differences between the coarse and fine resolution simulations can be found in the Table 2.

<table>
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<th>Table 2: Summary of Fine and Coarse Resolution Simulations</th>
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<td><strong>Fine Resolution</strong></td>
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<tr>
<td>Horizontal Resolution</td>
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<tr>
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<tr>
<td>Number of Points in Y-direction</td>
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<tr>
<td>Timestep</td>
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<tr>
<td>Number of Points in Absorbing Layer on Western Edge of Domain</td>
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* 7 of the points are used for the cyclic boundary condition
b. Coarse Resolution

Before examining how well the nonlocal diffusion at coarse resolutions can represent the convective boundary layer of 10 January 1998, a simulation was conducted at coarse resolution with just standard K-theory diffusion (using equations 2.2 and 2.3). This simulation was identical to the CR-simulation of Tripoli, with the exception of the degradation of the horizontal resolution from 400m to 2000m. This simulation was run to give a basic understanding of the abilities of standard K-theory diffusion at the coarse resolution. As expected, the coarse resolution simulation with K-theory diffusion (here after referred to as CK-simulation), was a deterioration compared to the CR-simulation. (Table 3 contains abbreviations used to refer to simulations.)

<table>
<thead>
<tr>
<th>Abbreviation</th>
<th>Simulation Description</th>
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<tr>
<td>CR-simulation</td>
<td>Cloud resolving simulation using fine resolution and K-theory diffusion (Tripoli, in prep.)</td>
</tr>
<tr>
<td>CK-simulation</td>
<td>Coarse resolution simulation using K-theory diffusion</td>
</tr>
<tr>
<td>CNL-simulation</td>
<td>Coarse resolution simulation using nonlocal term in diffusion</td>
</tr>
<tr>
<td>FNL-simulation</td>
<td>Fine resolution simulation with nonlocal term in diffusion</td>
</tr>
<tr>
<td>CNLL-simulation</td>
<td>Coarse resolution simulation with increased land between the western absorbing layer and the western lakeshore</td>
</tr>
<tr>
<td>CNLRV-simulation</td>
<td>Coarse resolution simulation with nonlocal transport applied to only theta, and not moisture</td>
</tr>
</tbody>
</table>
A cloud first appeared around 2760 seconds (Figure 9). The initial cloud formed over the land and then grew into the flow towards the windward shore. The fetch of Lake Michigan (~90km) is large enough to expect the cloud to form before reaching the leeward shore. The CK-simulation cloud grew to a depth of over 1600m (not shown). Radiosondes launched on 10 January 1998, on the leeward side of Lake Michigan, showed a boundary layer depth of about 1.3km. The CK-simulation resulted in a boundary layer cloud deeper than the observations for the boundary layer on that day. The western extent of the cloud was less than that of the CR-simulation (Figure 10). Analogous with the cloud structure, the greatest
amount of snowfall in the CK-simulation was over the land east of Lake Michigan (Figure 11).

The thermal and moisture structure in the convective boundary layer were examined for the coarse resolution simulation in order to gain a more complete understanding of the boundary layer produced by K-theory at coarse resolutions. Several unrealistic features in the convective boundary layer demonstrated that the cloud location and growth was not the only problem. An east-west cross section of the Richardson Number (averaged in north-south direction) showed a boundary

Figure 10: Cloud structure at 6 hours for CK-simulation (A) and CR-simulation (B). The CR-simulation produced a cloud coverage with greater western extension.
layer characterized by superadiabatic lapse rates (Figure 12). While the superadiabatic lapse rates were consistent with how local K-theory diffusion is able to transport entropy ($\theta$), it was inconsistent with a realistic atmosphere. In the atmosphere, the instability associated with an absolutely unstable layer would result in spontaneous overturning. This instant mixing prevents absolutely unstable layers from existing, with the exception of close to the surface in the presence of large heat fluxes. The CR-simulation developed a superadiabatic layer as well, but the superadiabatic layer in the CR-simulation disappeared when explicit convection occurred (Figure 13). The CK-simulation lacked the explicit convection needed to

Figure 11: Total precipitation (only solid precipitation) at final time (6 hours) for CK-simulation (A), CR-simulation (B) and CNL-simulation (C). Contours have increments of 0.2 cm.
remove the superadiabatic lapse rates. There was a drop in the vertical extent of the negative Richardson Numbers in the eastern portion of the domain for the CK-simulation. The drop in depth of the negative Ri coincides with the western extent of the cloud deck.

![Richardson Number Dry at 3 - 10](image)

**Figure 12:** East-west cross section of meridionally averaged Richardson number, for CK-simulation. Dashed lines represent negative Richardson numbers, indicative of superadiabatic lapse rates. Valid at 6 hours.

Not only the thermal structure, but also the moisture structure in the convective boundary layer of the CK-simulation showed unrealistic characteristics. Figure 14 is a sounding taken from a location in the middle of Lake Michigan, three hours into the CK-simulation. The sounding shows an atmosphere that was nearly saturated throughout the entire depth of the boundary layer. A boundary layer that
is both saturated and superadiabatic is extremely unrealistic. Convectively driven saturated atmospheric layers are characterized by temperature profiles that are

Richardson Number Dry at J= 27
6HR Forecast Valid 13 UTC 10 JAN 98 (Model Time 21600.0 s) Grid 1

![Richardson Number Dry at J= 27](image)

Figure 13: Meridionally averaged Richardson number for the CR-simulation, at final time (6hours). Dashed lines are negative values.

moist adiabatic (Adams, in press).

The one physical process that the coarse resolution with K-theory simulation somewhat captured was the gravity waves observed in the satellite image of 10 January 1998. The gravity waves are observable in the east-west cross section of potential temperature (Figure 15) near the top of the boundary layer. The potential temperature gradient in the cross section showed consistency with the Richardson
number cross section, and the sounding, once again illustrating the unrealistic instability above the lake.

Figure 14: Sounding taken over lake 3 hours into CK-simulation, notice the deep absolutely unstable temperature profile (notated with bracket).

The CK-simulation, as expected, was a poor representation of the convective boundary layer. Superadiabatic lapse rates dominated the entire boundary layer and with the inability of explicit convection to occur, the superadiabatic lapse rates were not removed. Compared to the CR-simulation the CK-simulation 1. had a cloud deck with a poor western extension, 2. the superadiabatic lapse rates continued further to the east across the lake, and 3. had a snowfall maximum over the land rather than the lake.
c. Nonlocal Transport at Coarse Resolution

With the failures of K-theory at coarse resolutions known due to the CK-simulation, a coarse resolution simulation employing the new nonlocal diffusion scheme was conducted (referred to as the CNL-simulation). The CNL-simulation showed some vast improvements when compared to the CK-simulation. There were improvements to the location and growth of the cloud deck, the temperature profile, the moisture profile, and the stability within the convective boundary layer.
The nonlocal terms' ability to transport moisture and temperature quickly away from the surface resulted in a more realistic boundary layer structure.

Figure 16: Appearance of first cloud for CNL-simulation (500 seconds). Notice that the cloud forms over the water.

The cloud deck first began to form around 500 seconds into the simulation. This was over half an hour earlier than the first cloud in the CK-simulation. The cloud that formed did so over nearly the entire width of Lake Michigan (Figure 16). While the location of the cloud was much better than that of the initial simulation, there were a few problems with this cloud. The cloud first appeared less than 600 seconds into the simulation. With a wind of 5ms\(^{-1}\) near the surface, it is unrealistic to believe that the air parcels coming off of the shore had had time to be sufficiently
modified by the lake and transported to the top of the boundary layer by convection in less than 10 minutes time. However, if the nonlocal transport is thought of as representing the mean effect of having large convective eddies, rather than representing the path of individual parcels within the eddy, the almost spontaneous formation of the cloud is acceptable.

The ability of the nonlocal diffusion to transport entropy in the absence of a superadiabatic lapse rate, allowed the Richardson Numbers in the convective boundary layer to remain more realistic (Figure 17). Superadiabatic lapse rates were
contained to the first few hundred meters of the convective boundary layer. While the depth of the superadiabatic lapse rates was still somewhat deeper than would exist in the real atmosphere, this was a vast improvement over the local diffusion scheme. On the eastern side of Lake Michigan, negative Richardson numbers reached only 300m above the surface. Considering the vertical resolution of 100m, this means only the first three grid points contain a superadiabatic lapse rate. The nonlocal transport by large eddies is responsible for transporting only a portion of

![Figure 18: Sounding over Lake Michigan, 3 hours into the CNL-simulation. Notice that the boundary layer does not exhibit a deep unstable layer, unlike the CK-simulation (Figure 13).](image)

the heat and moisture fluxes coming off of the warm lake. Smaller eddies still exist, and their down gradient transport represented within K-theory, allowed a
superadiabatic layer in the first few grid points above the surface. It is possible that a higher vertical resolution near the surface would confine the superadiabatic layer to an even shallower depth, but that was not tested in this work. The lapse rates of the CNL-simulation are an improvement over both the CK-simulation and the CR-simulation. Recall, the CR-simulation removed the superadiabatic lapse rates only in locations where the explicit convection occurred. The nonlocal transport was able to remove the superadiabatic lapse rates over most of the lake.

A sounding taken from over the lake about 3 hours into the CNL-simulation (Figure 18) showed a different boundary layer than the one produced by the CK-simulation (Figure 14). The boundary layer in this sounding does not show the absolute instability of the CK-simulation. The temperature profile in Figure 18 is almost moist adiabatic. This is interesting in that moist adiabatic temperature profiles, that aren’t saturated, often form in vicinities of convection due to the subsiding motion between the clouds. The convective cloud rolls are not resolved at this resolution (Figure 19a), and the subsiding motion was not being explicitly represented. The moisture profile also differed from the CK-simulation sounding. The CK-simulation resulted in a nearly saturated boundary layer (Figure 14), while the CNL-simulation had saturation only at the level where the boundary layer cloud existed (Figure 18). This moisture profile is much more realistic. As mentioned in Chapter 4, the moisture transport by large eddies is not counter-gradient, or even
nonlocal. However, the simulation that employed the "nonlocal" transport, in order to parameterize the large eddies, did produce a more realistic moisture profile.

The ability of the nonlocal diffusion to quickly transport heat and moisture upward allowed the cloud deck to form much closer to the windward shore (Figure 19a). The cloud deck should be closer to the shoreline in order to completely agree with satellite observations (Figure 5), nonetheless this is a noteworthy improvement over the CK-simulation (Figure 10a) and the CR-simulation of Tripoli (Figure 10b). The distance of the cloud from the shore will be discussed more in a following section. Corresponding to the location of the cloud, the location of the snowfall
improved for the CNL-simulation compared to the CK-simulation. Recall that the CK-simulation kept the maximum in snowfall over the land in Michigan. The CNL-simulation produced a snowfall maximum over the eastern half of the lake (Figure 12c). The western extent of the snowfall for the CNL-simulation was similar to that of the CR-simulation (Figure 12b). The CR-simulation had a banded snowfall pattern, which matched its banded cloud structure. The snowfall produced by the CNL-simulation had virtually no north-south variation in amount, but considering the lack of banded cloud structure this is reasonable. The important characteristic of the snowfall area produced by the CNL-simulation is that it is virtually the same area as the CR-simulation. Both of the coarse resolution simulations (CK-simulation and CNL-simulation) produced less total precipitation than the CR-simulation.

The potential temperature cross section taken west to east across the middle of the domain at the final time (Figure 20) was once again different from the CK-simulation. There was a lack of strong gravity waves evident at the top of the boundary layer. The top of the boundary layer was characterized by a much smoother growth, and did not undulate. Compared with potential temperature cross sections of past lake effect simulations (Hoggatt 1996) the depth of the superadiabatic layer was significantly less. The slight lowering of the boundary layer height on the windward shore, due to the change in surface friction, was captured by the nonlocal diffusion. The boundary layer growth across the width of the lake was consistent with observations taken on 10 January 1998.
Figure 20: Potential temperature cross section of CNL-simulation at final time (6h). The boundary layer top exhibits a smooth growth, and the superadiabatic lapse rates remain close to the surface.

The CNL-simulation showed improvements to the CK-simulation in terms of cloud location and growth, moisture profile, temperature profile, and stability with the convective boundary layer. Of even more interest, are the improvements of the CNL-simulation over the CR-simulation. At the coarse resolution the nonlocal term was better able to eliminate unrealistic superadiabatic layers than the CR-simulation of Tripoli. The CNL-simulation, while unable to resolve the roll cloud structure, was able to create a more realistic cloud cover than the CR-simulation. The clouds in the CNL-simulation are considerably closer to the western shore of Lake Michigan compared to the clouds in the CR-simulation.
d. High Resolution with Nonlocal Diffusion

The success of the nonlocal diffusion at coarse resolution (CNL-simulation) prompted the testing of nonlocal diffusion at fine resolution. A simulation at exactly the same resolution as the CR-simulation, including the nonlocal transport, was conducted (hereafter referred to as FNL-simulation). The FNL-simulation showed one of the downfalls of the nonlocal parameterization in that while the cloud deck did develop much closer to the windward shore than the CR-simulations (Figure 21), the cloud deck showed none of the roll convection structure.
resolved in previous simulations at the same resolution (CR-simulation). Both cloud structures in Figure 21 capture the gravity waves parallel to the shore. In agreement with the CNL-simulation, the FNL-simulation through the nonlocal transport term was able to develop clouds much earlier. The lack of a roll structure to the clouds questions the resolution for which the nonlocal transport parameterization is applicable.

![Figure 22: Dashed line represents the Dry Rayleigh Number computed across the lake (90km) for the a FNL-simulation (A) and the CR-simulation (B) at the final time (6hr). Solid lines represent different wavelengths of overturning. The scale on the left is from $10^2$ to $10^6$. Note that for the FNL-simulation the Rayleigh number remains stable.]

At 400m resolution the 6km spacing of observed clouds (Figure 5), is within the resolvable range, but the FNL-simulation failed to resolve the rolls. The reason that the rolls remain unresolved is that the initiation of rolls depends on the
Rayleigh-Bernard instability that the nonlocal diffusion greatly reduced. The explicit convection seemed to be initiated by the growth of the internal boundary layer (IBL) to the top of the PBL where the cold air was tapped. The only way for the IBL to grow to the PBL with K-theory is through superadiabatic lapse rates. The explicit convection in the CR-simulation tapped into the instability that local diffusion allowed to build. The CR-simulation eliminated the deep superadiabatic layer only when explicit overturning occurred due to convection. The importance of instability in forming these rolls is illustrated by a plot of the Rayleigh number across the width of the lake (Figure 22).

Figure 22 is based on the definition of a Rayleigh number by Houze(1993), and uses the modeled eddy viscosities for heat and momentum in place of molecular viscosity. The Dry Rayleigh Number is kept smaller than the various wavelengths plotted in the case of the nonlocal diffusion. This means that the boundary layer was stable relative to dry roll convection of 4km, 6km, and 8km wavelengths, and thus roll convection never developed. This is not the case for the local diffusion simulation (CR-simulation). In the case of local diffusion (CR-simulation), the UW-NMS needed superadiabatic lapse rates to initiate the roll convection. The FNL-simulation was unstable for moist convection (Figure 23), however the moist convection needs the dry convection to initiate it, and dry convective rolls could not develop.
Cross sections of meridionally averaged Richardson Numbers (Figures 13 & 24) showed that the nonlocal diffusion, even at fine horizontal resolution, kept the superadiabatic layer confined to the first few vertical grid points, while the local diffusion scheme allowed the instability to grow to the depth of the boundary layer. In the CR-simulation that produced the roll convection, the negative Richardson Numbers dropped to a lower vertical extent only when the explicit convection developed. The roll convection, as mentioned earlier, seemed to be the mechanism for stabilizing the upper part of the convective boundary layer. The nonlocal
diffusion kept the mixing done by explicit convection from being necessary to stabilize the boundary layer in the FNL-simulation. At cloud resolving scale, both types of diffusion had benefits. The nonlocal diffusion kept lapse rates more realistic, while having only the local diffusion allowed explicit convection to organize the clouds in the correct roll structure.

The application of the nonlocal term at high resolutions is questionable when the purpose of the term is considered. The term is intended to represent the effect of large eddies. When these large eddies can be at least partially explicitly resolved, the nonlocal term is trying to parameterize the very thing that is trying to be
explicitly resolved. This creates a theoretical "gray" area, for there is a double accounting of those large eddies when they are both parameterized and partially resolved.

e. Addition of Land Upstream

Previous cloud resolving simulations of 10 January 1998 (Tripoli, in prep) have shown that the clouds could get closer to the western shore by moving the absorbing layer further upstream. While the simulations with nonlocal diffusion (CNL-simulation and FNL-simulation) moved the clouds closer to the western shore than simulations by Tripoli, the clouds were still further offshore than satellite observations showed. In an attempt to reduce the distance between the western shore and the cloud deck the absorbing layer was moved upstream (Figure 8b), and an additional 40 points were added to the domain (here after referred to as CNLL-simulation). The grid spacing remained the same as the previously discussed coarse resolution simulations ($\Delta x = \Delta y = 2000m$).

The number of grid points that act as a horizontal absorbing layer remained the same; they were just farther upwind. By increasing the amount of land upstream, and moving the absorbing layer further upstream, turbulence over the land was allowed to grow. In this simulation, the flow approaching the shore was not completely laminar. The amount of turbulent kinetic energy (TKE) over the land upstream of the lake increased with the absorbing layer further to the west. The
CNLL-simulation, at 3600 seconds, had a maximum value for TKE of approximately 3.6 m²s⁻², compared to a maximum value of 1.6 m²s⁻² at the same time in the CNL-simulation. The increase in the TKE of the flow approaching the lake resulted in higher eddy diffusivity coefficients over the western edge of the lake. This increase in TKE can be attributed to the destabilizing effects of differential cold air advection upstream brought on by the slowing of the flow. When the absorbing layer is used, the differential advection destabilization tendency is opposed and the TKE finds a lower equilibrium. The maximum eddy diffusivity \((K_m)\), at 3600 seconds, for the CNL-simulation and the CNLL-simulation were 35.59 m²s⁻¹ and 41.82 m²s⁻¹ respectively. The transport done by both the local gradient \((\partial C/\partial z)\) and the nonlocal term \((-\nu\nabla^2C)\) are multiplied by the eddy diffusivity (equation 2.12). The larger eddy diffusivities obtained by moving the absorbing layer further to the west allowed for larger transports of entropy and moisture.

The increase in vertical transport of moisture and entropy near the western shore of Lake Michigan resulted in a cloud deck that formed closer to the western shore. (Figure 19b). The distance offshore at which the clouds formed was in good agreement with satellite observations (Figure 5). The absorbing layer was initially applied to keep simulations consistent with the work of Tripoli (in preparation). The results obtained by moving the absorbing layer further upstream imply that flow coming offshore is not laminar, and the turbulent structure of that flow plays a role in forming the convective boundary layer over the water. It is possible that the
higher TKE and eddy diffusivity values are simply the result of model numerics, and not a physical representation of the flow off of the land. The result does support the conclusions of Tripoli (in prep.) that the flow off of the land should be studied more.

The increase in land upstream not only improved the distance from shoreline to cloud, but also the stability within the convective boundary layer. The superadiabatic layer still existed at depths up to 300m (Figure 25), but the values of the Richardson number were less negative. This implies that the transport being done by the K-theory, was less than the previous case. While a 300m superadiabatic
layer remains somewhat unrealistic, it is important to note that the superadiabatic layer is barely superadiabatic.

f. Nonlocal Transport of Moisture

As mentioned in Chapter 4, and earlier in Chapter 5, the term "nonlocal" is not really applicable to moisture. In the case of moisture, in the convective boundary layer, the nonlocal term will often be transporting moisture in the same direction as the local gradient. This brings to light the question of whether the nonlocal transport term should be applied to moisture, since there is not a disagreement with the local gradient transport. A simulation similar to the CNL-simulation was run, except that moisture was transported only through K-theory diffusion (referred to as CNLRV-simulation).

In the CNLRV-simulation entropy could be transported away from the surface quickly due to the nonlocal term, but the moisture was forced to slowly diffuse its way up in the boundary layer one grid box at a time. This resulted in the lower part of the boundary layer becoming unrealistically moist. The first signs of a cloud appeared over the lake at 2,400 seconds, but the cloud was seemingly touching the surface (Figure 26).

While the transport of moisture by large eddies in the atmosphere is not counter-gradient, or nonlocal, it is still occurring. The nonlocal transport term is intended to represent the mean effect of large eddies. Those large eddies are able to
Figure 26: When the nonlocal transport term is applied for heat alone, and not moisture (CNLRV-simulation), the result is fog at the surface. Figure valid at 2400 seconds.

transfer moisture through the depth of the boundary layer faster than the small eddies that K-theory represents. Quickly transporting the heat away from the surface dropped the saturation vapor pressure, and without also transporting away the high surface moisture, the air near the surface became saturated. In effect it created a fictitious source of moist entropy. This simulation shows that while the moisture is not being transported against the local gradient, the "nonlocal" transport term must be applied to remain consistent with the mean transport of large eddies in the convective boundary layer.
However, when thinking about the nonlocal transport term for moisture physically there are some problems with how the nonlocal term works for transporting moisture. As with other passive scalars, the transport of moisture is dependent on only the thermal structure of the boundary layer, and the surface flux of moisture. This means that moisture will be transported upward as long as the surface flux of moisture is positive, and the thermal structure is conducive to convective overturning. While the moisture gradient in the planetary boundary layer is almost always directed upward, the nonlocal formulation has no bounds on the moisture term to keep the air aloft from getting more moist than the surface.

Figure 27: The dashed arrows represent how deep the given quantity (heat or moisture) should be mixed given the structure of that quantity. Notice that for a boundary layer characterized by moisture aloft, the nonlocal term could possibly make the air aloft more moist than the air above the surface.
(Figure 27). This problem has a self-corrector when dealing with entropy ($\theta$). If the air aloft starts to get warmer than the surface, then the boundary layer depth will be diagnosed as lower, and the nonlocal term will cease to mix entropy up that high. But with moisture, it is irrelevant how moist it is aloft, and nonlocal transport could actually make the air aloft much more moist than the air right above the surface, disobeying the second law of thermodynamics. While the CNLRV-simulation showed that a parameterization for the transfer of moisture by large eddies is necessary, the parameterization needs to be reformulated so that it must obey the second law of thermodynamics.
6. Conclusions and Future Work

a. Conclusions

A term representing the nonlocal transport of moisture ($q_v$) and entropy ($\theta$), by large eddies in the convective boundary layer was implemented in the UW-NMS diffusion scheme. The simulations conducted showed that adding a nonlocal term to standard K-theory diffusion allows a coarse resolution simulation to produce comparable results to a cloud resolving simulation with only K-theory diffusion. Nonlocal transport eliminated many of the problems that simple K-theory produced at coarse resolutions. Simulations employing the new diffusion scheme showed improvements in the boundary layer profiles of moisture and potential temperature. The new diffusion scheme eliminated the unrealistic superadiabatic lapse rates that coarse resolution simulations employing K-theory allowed to build through the depth of the boundary layer. K-theory diffusion at coarse resolutions could not produce the cloud coverage that the cloud resolving simulation could.

Coarse resolution simulations with the new diffusion scheme produced cloud coverage not only comparable to the cloud resolving simulation, but in some respects better than the cloud resolving simulation. The coarse resolution simulation with nonlocal transport (CNL-simulation), while unable to resolve the roll structure of the clouds, produced a cloud deck closer to the western shore than the CR-simulation. The areal coverage of the CNL-simulation was more in agreement with satellite observations than the CR-simulation.
The results of the new diffusion scheme at coarse resolutions prompted the testing of the scheme at cloud resolving resolution (FNL-simulation). The FNL-simulation identified a "double accounting" problem at cloud resolving resolutions. Analogous to cumulus parameterization, there seems to be a "gray" area for the parameterization of large convective eddies. At resolutions where the large convective eddies can be partially resolved, both the nonlocal term and explicit mixing are trying to represent the same eddies. At cloud resolving resolutions the addition of the nonlocal term prevents the formation of roll convection, while standard K-theory is able to initiate the roll convection through the build up of unrealistic deep superadiabatic layers. Whether the diffusion scheme with nonlocal transport or the standard K-theory diffusion should be employed at small horizontal scales depends on whether the goal of the simulation is to produce a realistic thermal structure in the boundary layer or correct cloud structure.

The success of the nonlocal transport in the TKE based diffusion scheme of the UW-NMS disagrees with the work of Belair et al. (1999) who found little improvement with the addition of a nonlocal transport term to a TKE based boundary layer model. Most implementations of a nonlocal transport term have been done in models with similarity theory based boundary layers, and thus Belair et al.'s implementation of a nonlocal term was the most similar to that of the UW-NMS. Implementation of the nonlocal correction term on moisture transport in the
UW-NMS showed results similar to implementation in other models (Holtslag and Boville 1993).

b. Future Work

The addition of a nonlocal diffusion scheme to the UW-NMS has applications beyond the convective boundary layer observed during lake effect snow events. It is intended to implement the nonlocal diffusion into the operational version of the UW-NMS. Unrealistically deep superadiabatic layers near the surface have been a source of concern in the UW-NMS. The use of the nonlocal diffusion scheme presented in this thesis may alleviate some of this concern. While there still remains a question as to which resolutions the nonlocal diffusion is applicable, the resolution of the operational UW-NMS is coarse enough to ignore this issue. The nonlocal diffusion may prove quite beneficial in accurately depicting the convective boundary layer that develops in the summer in areas of localized heating. Extensive localized surface heating creates large surface heat fluxes similar to the heat fluxes over the Great Lakes during an arctic air outbreak.

The lack of a limit on the moisture transport by the nonlocal term must be further evaluated. Future work will attempt to find a way to limit the moisture transport so that the second law of thermodynamics is always obeyed. Large convective eddies not only transport moisture up from the surface but also can mix down the drier air in a capping inversion. The nonlocal term is currently formulated to represent the mean properties of large eddy transport, but is based on the thermal
structure and thus does not capture the downward mixing of dry air. Large eddies that transport moisture and entropy may also transport momentum. It is planned to implement a term to represent the nonlocal momentum transfer. This term is more complicated than that for entropy and moisture, in that the tilt of the convective eddies, relative to the ambient shear, plays a role in the momentum transport.

What remains the most daunting task to be accomplished, is determining a way to transition between scales where the large eddies can not be resolved to those where they can be partially resolved to those where they are completely resolved. The "double accounting" of the large eddies needs to be eliminated. The UW-NMS is intended as a completely scalable weather prediction model and thus the diffusion scheme should ultimately be completely scalable as well.
7. References


Grell, Georg, Jimy Dudhia, and David Stauffer, A Description of the Fifth-Generation Penn State/NCAR Mesoscale Model (MM5), NCAR/TN-398+STR


Appendix I.

A variable step topography, unique to the UW-NMS, is used. In this method, the lowest grid box has a variable depth that exactly matches surface elevation. The surface boxes are finite differenced implicitly to insure stability. The step boundary conditions conserve vorticity and momentum. This allows the UW-NMS to handle even the most subtle topography while having no slope restrictions to severe topography. The UW-NMS can represent elevation changes as small as 1m.