INFLUENCES OF LAKE ONTARIO INTERFACE TRANSPORT PROCESSES ON ATMOSPHERIC CONVECTION

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SUMMARY

The Great Lakes exert important mesoscale influences on the climate of the surrounding land areas. These influences originate in the transport of momentum, mass, and heat across the air-lake interface. It is proposed that Colorado State University and the National Center for Atmospheric Research jointly study the interface transport processes of Lake Ontario and utilize calculated fluxes in three numerical models for the purpose of investigating atmospheric influences of the Lake. A two-dimensional mesoscale convection model, similar to the Lavoie mesoscale model, will be developed and used as the primary tool in studies of Lake climatic influences, focusing particularly on lake-effect storms. The domain of the model will be a rectangular area covering the entire Lake Ontario drainage basin. Providing suitable and sufficient field data are gathered during the IFYGL Field Year, two three-dimensional models will also be applied in studies of Lake climatic influences. One, a deep convection model, will be used to investigate regions of intense convection for both lake-effect storms and summer thunderstorms. The other, a lake breeze model, will be used to study interactions among the phenomena of lake breezes, urban heat island effects, and the climatic influences of the drainage basin topography. Interface transport processes will be investigated using micro-meteorological lake tower data. The data will be used to evaluate bulk aerodynamic formulas, which in turn will be applied to data from the IFYGL lake buoy-tower network for the purpose of determining input fluxes at the lower boundaries of the three respective model domains. Land climatological station data will be used with similar formulas to obtain corresponding input fluxes over land areas. Quantities at the other boundaries of the respective domains, as well as initial conditions, will be determined from available meteorological data.

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INTRODUCTION

One of the primary objectives of the International Field Year for the Great Lakes (IFYGL) is to investigate a number of problems associated with the meteorology and physical limnology of Lake Ontario. Colorado State University, Fort Collins, Colorado and National Center for Atmospheric Research (NCAR), Boulder, Colorado jointly propose to investigate transport processes at the air-water interface of Lake Ontario, and, with the aid of numerical atmospheric convection models, determine how these processes affect the meteorological system over the Lake and its drainage basin. The proposed study falls within the primary objectives of the IFYGL and into two of its four general programs, Lake Meteorology and Energy Balance.

The purpose of this report is to discuss the approach to the proposed study that a preliminary investigation has determined to be most feasible and potentially productive. In the preliminary investigation, methods of obtaining momentum, mass, and heat fluxes at an air-water interface and experimental comparisons with the results of such methods were appraised. In addition, techniques for modelling mesoscale convection systems were evaluated as to their suitability for the Lake Ontario region.

The report is organized to first present background on mesoscale influences of the Great Lakes and justification of a convection model analysis approach utilizing a modified version of the Lavoie mesoscale model. Following this, three numerical convection models that will be applied to the atmosphere above the Lake Ontario region are described and interface transport processes and the relevant bulk aerodynamic theories used to calculate fluxes are discussed. Finally, the proposed plan of study is presented.

MESOSCALE INFLUENCES OF THE GREAT LAKES

General Lake Influences

The Great Lakes modify the climate of the land area which surrounds them. One of the main reasons for this modification is the thermal influence of the lakes. Kopec [1965, 1967] found that Lake Superior is the most influential because of its great surface area and depth, while Lake Erie is the least effective because it is relatively shallow and usually freezes over in the winter. Other thermal influences of the lakes given by Kopec are:

1. The temperature variations along the lake shores are less than the temperature variations over land away from the lakes;
2. The thermal influence is greatest in the winter when the lakes are warmer than the land;
3. The influence is much less in the summer when the lakes are cooler than the land;
4. There is relatively little thermal influence during the seasons of the equinox;
5. April is the month of thermal transition when the land rapidly heats up and the lakes begin to assume their summer role as cooling agents;
6. By October the lakes are warmer than the land;
7. Areal patterns of thermal influence are basically symmetrical with the lake shapes but skewed to the East; the intensity and extent of these influences are related primarily to the surface area and depth, the existing contrast in temperature between the surface water and the surrounding land mass, and the prevailing wind patterns.

Because of these thermal conditions, the lakes modify precipitation over the water and adjacent land areas, Eichenlaub [1970]. In the fall and winter the lakes are heat sources. Hence, there is a transfer of moisture and sensible heat from the lakes into overpassing cold air masses. This produces instabilities which greatly increase cloudiness and precipitation. In the spring and summer the lakes are heat sinks and the development of convective cloudiness and precipitation is restricted over the water surfaces. Therefore, the net effect of the lakes on precipitation depends on the balance between summer curtailment and winter augmentation.

A heat source such as the Great Lakes in the winter time produces a horizontal convergence in the atmosphere and this results in a cyclonic circulation around the source, Petterssen and Calabrese [1959]. Hence, the lakes produce a local system of cyclonic vorticity which alters the pressure configuration and the flow field. For example, during the storm of February 8-14, 1958, the air warmed by the Great Lakes appeared to account for a central pressure deficiency of about 6 mb.

Probably, the greatest climatic effect of the Great Lakes is the amount of snow which falls in their drainage area. There are four causes of snow storms in the Great Lakes area, Muller [1966]. They are the warm front storms which are the most wide spread, smaller storms produced by passing cold fronts, orographic storms, and last, but certainly not least, the post cold front lake squalls or snowbursts. Often times, these snowbursts are very intense storms which are due to the energy and moisture exchanges between polar continental air masses and the relatively warm waters of the Great Lakes. They are most pronounced during late autumn and early winter when the temperature differences between warm lake water and cold polar continental air are greatest. Wiggin [1950] found that the snowbursts, lake squalls, or lake-effect storms are characterized by or require the following:

1. Long fetches for the cold air blowing over the lake water;
2. A flow pattern in the atmosphere which possesses cyclonic curvature and a shear sufficient to produce the longitudinal convective cells, or snow bands;
3. A strong flow of cold polar air with a lee shore lapse rate which is adiabatic or greater to over 5000 feet;
4. Lakes which are relatively free of ice;
5. Bands of snowfall which are 2 to 80 km wide and 40 to 160 km long.

Lake Ontario Influences

The influence of Lake Ontario on the eastern area of the lake and its drainage area was discussed in detail by Falconer, Lansing and Sykes [1964]. They found that in the spring the water warms very slowly; as a result, the islands, promontories and lake shores warm up slowly. Hence, the shore vegetation is two to three weeks behind in leaf production as compared with trees 15 miles inland. As an example of temperature difference, for June 1964, the mean monthly Lake Ontario water temperature was 62°F, while the mean temperature at the Syracuse Airport was 78°F. Also, during this season of the year, the air above the water is in thermal equilibrium with the water if the winds are calm to light. For example, for the spring of 1964, measurements gave a shoreside temperature of 75°F, a lake water temperature of 47°F, and an air temperature over the lake of 49°F.

During the summer, the lake continues to warm up, but not until August is the lake as warm as the surrounding air. During July, the average surface temperature over northeastern Lake Ontario is 65°F to 68°F. In fact, the mean maximum July temperature at Galloo Island (10 miles offshore) is 15°F less than the mean maximum at Syracuse Airport. Hence, during the summer the lake is relatively cool with respect to the surrounding air mass. This produces stable air in the lower levels and there is almost a complete lack of convective clouds over the lake in this season. On many summer days there may be towering cumulus clouds over the eastern shore of Lake Ontario while at Galloo Island the skies are absolutely free of any clouds. Another phenomenon which may occur during this period is a thunderstorm moving from the northwest to the southeast out of Canada and into the Lake Ontario region. The cool lake will cause this storm to dissipate over the water unless there exists a strong cold front. The lack of air mass thunderstorm activity produces a deficiency of water over the lake. For example, for the period June 1 to August 31, 1964, Galloo Island received 5.87" of precipitation, while Boonville (55 miles southeast of Galloo Island) received 12.31".

As fall approaches, there are outbreaks of cold polar air masses from Canada which flow over the warmer water. The warm water produces lapse rates in the air at lower levels which are dry adiabatic or steeper. Hence, the lower levels are unstable and the air begins to rise. Eventually, stratus, strato-cumulus, and heavy cumulus clouds are formed over the water. These drift to the lee shores and precipitation falls over the lake and/or the rising land masses on the lee shores.

As fall advances into winter, colder and colder continental air masses advance over the lake while the surface water cools. Since the lake is still much warmer than the air, stratus and strato-cumulus clouds are formed which produce snow flurries and often heavy snow squalls. The maximum effect of these storms is felt from 15 to 25 miles inland from the eastern end of Lake Ontario. The distance these storms travel inland is a direct function of the wind speed. In the late winter ice often covers much of Lake Erie and small parts of eastern Lake Ontario. Since the temperature contrast between air and water is not so great during this time, the lake-effect storms tend to taper off and be less frequent. However, under the proper conditions, these storms may occur even in late March.

From this seasonal outline of the meteorological influences of Lake Ontario, we see three major weather phenomena. These are the absence of thunderstorm activity over the lake during the summer, summer lake breezes due to the cool lake and the warm land masses, and the fall and winter lake-effect storm or snowburst. Of these, the lake-effect storm is the most dramatic.

Lake-Effect Storm Influences

As a consequence of the fact that Lake Ontario has a west-east alignment which is, in general, parallel to the prevailing air movement under flow of Arctic air, the snowbelt region is east of the lake in New York State, Muller [1966]. This region extends from Syracuse northward to Watertown, and eastward to the Adirondack Mountains. In this snowbelt there is more snowfall than any other place in the United States, east of the Rocky Mountains. The seasonal snowfall in this area (the Tug Hill Plateau) is greater than 160" and the maximum recorded mean seasonal snowfall of 209" (for 1951-1960) is for the Boonville station, whose elevation is 1575 feet above mean sea level.

Lake effect storms have a bandlike structure and thus the area affected in the snowbelt is dependent upon the direction of the prevailing wind, Lansing [1951] and Johnson and Mook [1953]. If the winds are from the southwest, the Adams-Watertown area gets snow; from the west, the Pulaski-Sandy Creek area gets snow; and from the northwest, the Fulton-Syracuse area is affected. During the storm of January 28-30, 1953, the winds changed direction during the storm such that all three areas received intense snowfalls.

From the storm analyses of Peace and Sykes [1966] and Sykes [1966], several other characteristics of lake-effect storms can be derived. Some of the characteristics of these storms over the Lake Ontario region are as follows:

1. The storms may be rather local in character or they may be nonlocal, covering many thousands of square miles over the eastern shores of Lake Ontario;
2. The primary control of the snow bands seems to lie with the conditions aloft;
3. The lake-effect storms tend to form parallel to Lake Ontario because of the need for the longer trajectories over water;
4. These storms produce heavy snowfall rates, but they are shallow storms (usually less than 10,000 feet) and the moisture is concentrated into narrow convergence zones;
5. The lake adds heat and moisture to the cold air as it flows over the water;
6. Changes in wind can occur over very small distances and the wind can shift as much as 40° or more in a period of five minutes;
7. A reliable large scale indicator of lake-effect storms is a surface pressure trough over the Great Lakes caused by the heating of cold air flowing over the warm lake water;
8. A given lake-effect storm may have several distinct phases due to changing synoptic conditions.

Hence, any mesoscale model which attempts to simulate the above phenomena must be capable of following changes in wind velocity and direction over relatively short times and distances, account for surface effects, account for conditions aloft, possess a precipitation mechanism, and allow the boundary conditions in the problem to be time-varying to account for changing synoptic conditions.

Meteorological Questions Concerning the Great Lakes

Nearly 35 million people live within the drainage basin of the Great Lakes. About half of these depend directly upon the Great Lakes for their water supplies. Since fresh water is so important to human life, everything possible must be done to improve the quality and quantity of this important resource. The Great Lakes are also used by man in other ways, such as, for transportation, fishing, "natural waste treatment plants," heat sinks for power plants and other industrial operations, and recreation. The previous section elaborated on the Great Lakes as weather and climate modifiers. Because of the enormity of the lakes and because their waters and the air above are so strongly interrelated, many questions concerning man's uses of the lakes can be answered through detailed meso-scale numerical models of the airflow over the lakes and their drainage areas. In the following list are questions which can be either partially or wholly answered through the use of such models.

1. How much moisture falls directly on the lake surface and where does it come from? If we knew the quantity and quality of the precipitation falling over the lakes, then we could assess man's contribution to the lake pollution problem from other sources, such as, pollution from boats and streams, Bruce, Lane and Weiler [1968].
2. How can we obtain better wind information so that we can make more reliable wave, set-up, and seiche forecasts? Knowledge of water levels are very important for navigation, water supply and power development, Bruce, Lane and Weiler [1968].
3. How can we improve on evaporation estimates over the lake surface? The knowledge of evaporation is very important because of the great amount of water lost at given periods by this mechanism. For example, the highest evaporation loss over Lake Ontario is during October and the rate is about 25 percent of the flow rate of the Niagara River, Richards and Rogers [1964] and MacDowall [1970].
4. How can we better define the climatic effects of the lakes? Knowledge of climatic effects is an important aid in the planning and development of shoreline areas, MacDowall [1970].
5. How can local weather forecasts, especially the forecast of lake-effect storms, be improved?
6. There will be many conventional and nuclear power plants on the shores of the Great Lakes in the next few decades, Frye [1970]. These plants will use lake water for cooling purposes and will return this cooling water to the lakes 20°C warmer than when it

was taken from the lake. Because there are thermal bars and coastal jets in the lakes during some seasons of the year, this warm cooling water may be concentrated in the nearshore areas. Hence, several questions come to mind as one considers the above phenomena, Bruce, Lane and Weiler [1968], Stewart [1969] and Charlier [1970].

(a) Will ice development be retarded?

(b) If so, will lake-effect storms continue later into the winter?

(c) As the warmer nearshore water interacts with the overlying air, will local evaporation rates increase enough to reduce the abnormally high water temperature?

(d) At what stage would increased evaporation measurably affect water levels in the Great Lakes?

(e) What local climate changes might be brought about through the modification of the nearshore sensible and radiant heat exchanges?

7. What are the effects of the urban heat islands in the Lake Ontario basin on the mesoscale circulation, Munn, Hirt and Findlay [1969] and Tag [1969]?

8. What are the effects and influences of lake breezes? The lake breeze influences the local climatology of regions bounding the Great Lakes. It affects cloud formations, mass transfer rates, and water circulations within the lakes, and is an important consideration in air pollution over large population centers, Moroz [1967].

NUMERICAL MODELS

A major effort of the present study will be the construction of a mesoscale numerical convection model. The model will incorporate the essential features of the Lavoie mesoscale model and will be capable of simulating the important characteristics of lake-effect storms summarized in a preceding section. It will be used to investigate the meteorological causes and effects of lake-effect storms. In addition, providing there is sufficient data on deep convective activity over the Lake Ontario drainage area during the data year, a three-dimensional, deep convection model will be applied to the data and attempts will be made to predict and analyze the convective activity. Finally, a three-dimensional, lake breeze model will be constructed to study the combined effects of lake breezes and urban heat islands.

The Lavoie Mesoscale Model

Lavoie [1968] developed a three-layer, mesoscale model to aid in the analysis of lake-effect storms over the eastern part of Lake Erie and its drainage basin. His is a dry model and the surface influences which are included are friction, heating and topography. The three layers in the model are a surface layer (I), a homogeneous layer (II), and a deep upper layer (III). Layer I is characterized by a superadiabatic lapse rate in which there is an upward transfer of heat and a downward transfer of momentum. Lavoie assumes that this lowest layer is 50 m thick. In Layer II, the lapse rate is dry adiabatic since the layer is assumed to be kept well-mixed by strong winds and surface heating. The upper limit of this layer is determined by soundings which show the existence of zero or first order discontinuities in temperature. Finally, Layer III is a deep stratum possessing a constant, stable lapse rate.

This three-layer thermal structure has been observed experimentally and applied analytically by various investigators. George [1940] used this thermal structure in a qualitative discussion of the mechanisms responsible for lake-effect storms. Burke [1945] made use of the structure in his explanation of the transformation of polar continental air to polar maritime air by a warm ocean. Asai [1965], also, considered this structure in his study of cold air outbreaks over the Sea of Japan. In Roll [1965], daytime oceanic trade winds are described as having a well-mixed layer below 2 km where the potential temperature and wind fields are homogeneous in the vertical and the layer is capped by an inversion. Hence, when these trade winds blow over a heated island, the three-layer thermal structure occurs. Lavoie [1968] and others have found that there exists a well-mixed atmospheric layer capped by an inversion during lake-effect storms over the Great Lakes. Therefore, the three-layer model is also generated when cold continental air masses undergo surface heating due to their passage over the warm lake water.

In the Lavoie mesoscale model the pressure field is determined by the density distribution alone without reference to the motion fields; hence, the pressure field is always in hydrostatic balance, Ogura [1963]. Estoque [1962a] justified the hydrostatic approximation in a sea breeze model through an order of magnitude argument for the vertical component of the momentum equation. Hovermale [1965] justified the hydrostatic balance for gravity waves induced by corrugated terrain if the square of the horizontal wavelength is much greater than the square of the vertical wavelength. Hovermale's analysis is applicable to lake-effect storms and his criterion is satisfied because the characteristic wavelength in the horizontal is tens of kilometers while the characteristic wavelength in the vertical is five or less kilometers.

The initial set of equations for the dry Lavoie model is made up of the horizontal equations of motion for the wind field (u, v), the hydrostatic pressure equation, the thermodynamic energy equation for the potential temperature θ, the equation of state for dry air, and the conservation of mass equation. These equations are only applied in Layer II. Since Layer II is homogeneous in the vertical, the above equations are integrated vertically from the top of the surface friction layer, Layer I, to the top of the well-mixed layer, Layer II. The altitude of the top of Layer II is denoted by h. Hence, the final equations are a set which govern the behavior of the mixed layer as a whole; the dependent variables are the wind field (u, v), potential temperature θ, and the altitude h. The average vertical velocity in Layer II can be obtained from the continuity equation. These variables are functions of the horizontal coordinates (x, y) and time t. For a given set of boundary and initial conditions, this system is very similar to the shallow water equations of oceanography.

The averaged momentum equations in Layer II contain the following terms: local time derivatives, advection terms, coriolis terms, horizontal pressure gradients and surface drag terms. The service drag term represents the momentum transfer from Layer II into Layer I due to surface friction. Since this stress in the bottom layer is parameterized by the bulk aerodynamic method, the surface drag term contains a momentum drag coefficient, C_f, Sutton [1953] and Priestley [1959]. The horizontal pressure gradient terms account for the interaction between Layers II and III. If Layer II is perturbed, adiabatic density changes occur in Layer III which, in turn, induce pressure changes in Layer II.

The average energy equation contains a term which accounts for the transfer of heat from Layer I into Layer II. Again, the bulk aerodynamic method is used; hence, this transfer term contains a heat "drag coefficient," C_H, see Roll [1965].

Lavoie [1968] used a forward-upstream differencing technique to numerically integrate the above system. That is, all advection terms are upstream differenced while all other spatial derivatives are centrally differenced; for local time changes, forward extrapolation is used. Lavoie claims that this technique is physically acceptable because the gradients are always evaluated in the fluid region that is being advected over the grid points during the forecast time interval. Also, this technique damps out the shortest waves (wavelengths less than or equal to four times the mesh size) most effectively and may then control nonlinear instability, Phillips [1959].

Lavoie [1968] used this dry model to predict and analyze lake-effect storms over Lake Erie. The technique of applying this model is to specify time-invariant boundary conditions and simple initial conditions. Then the equations are stepped along in time until a steady state solution is obtained. This steady state solution is the desired result. Hence, the model, when used in this manner, is not capable of predicting unsteady storm conditions due to changing synoptic conditions on the boundaries. This model also cannot describe convective motions driven by buoyancy forces.

Spelman [1969] used the Lavoie dry model to represent the flow of oceanic trade winds over a tropical island. The model island was a segment of a paraboloid of revolution whose vertex, or crest, extended to one-half of the height of the trade wind inversion. The circular cross-section at the water level had a 26 km radius. Since the island's surface temperature was 5°C warmer than the sea, the island influenced the airflow through surface heating, surface roughness and topography. However, Spelman found that the dominant atmospheric responses were generated by the topography and the heating, not the roughness; in fact, most of his solutions exhibited a "hydraulic jump" in the inversion surface at the lee of the island.

Spelman used both forward-upstream differencing and the Lax-Wendroff scheme to numerically solve for the steady state values of the horizontal wind field, potential temperature and the depth of the mixed layer. He used a 3-km mesh size over the island and increased the mesh exponentially near the boundaries of his domain of computation. The dry model of Lavoie gave acceptable results for this application, although it was not too accurate in the neighborhood of the lee jump.

Huang [1969] applied the Lavoie dry model in a study of airflow over the Black Hills. The objective of this study was to attempt to give a quantitative explanation of the deformation of the inversion surface which, in turn, is probably related to the generation of clouds over the Black Hills area. As in the above studies, the effects of surface heating, surface friction and topography were considered.

Weickmann, Jiusto, McVehil, Pilie and Warburton [1970] conducted a modification experiment on lake-effect storms over the Buffalo area. They hypothesized that, if the number of snow crystals in a storm could be increased by seeding, then the fall velocities of the new crystals would be smaller than the unseeded crystals. Hence, the airflow around and through the storm would redistribute the snowfall without affecting the total precipitation. Thus, ideally, Buffalo would receive less intense snowfalls and the downwind areas would receive more. The airflow pattern over the eastern end of Lake Erie and over Buffalo was computed through the use of the Lavoie dry model. This computed airflow was acceptable for computing the trajectories of the snow crystals.

Dirks, Marwitz and Veal [1970] studied the possibility of snow pack augmentation from orographic clouds (cap clouds) over Elk Mountain in Wyoming. In order to understand how to more effectively seed these clouds, the airflow pattern over Elk Mountain and through the cap clouds was required. Since the thermal structure of the Lavoie dry model was satisfied in 50 percent of the cases studied, it was used to give this flow field. The authors neglected surface heating and included a release-of-latent-heat term in the thermodynamic energy equation. For this application, the results were reasonable.

The work of Huang [1969] was extended by Chang [1970] and Chang and Orville [1970] to include the processes of evaporation and condensation for the model of airflow over the Black Hills. The thermal structure of the dry model was retained but the introduction of moisture effects produced many changes in the dry model equations. In the superadiabatic layer, Layer I, heat and water vapor are now transported upward and momentum is transported downward. This transfer of water vapor appears as a bulk aerodynamic term, containing a "drag coefficient," C_E, for water vapor, in a new equation describing the conservation of total water. The pressure gradient term in the motion equations is modified to account for the introduction of water vapor due to evaporation and advection. The temperature equation is modified by the latent heat flux due to condensation of water vapor, and it includes a source term which accounts for condensation or evaporation due to perturbations in the inversion surface. Clouds which form in this model have bases determined from the equation giving the change of dew point with height, have tops which terminate at the inversion surface, and have pseudoadiabatic liquid water contents within the clouds. However, no precipitation mechanism is specified.

Several sets of computations were run. They included dry cases, vapor effect cases where no phase changes occurred but only a change in air density due to the presence of water vapor, and moist cases with no precipitation. There was not sufficient field data to check the computed results but the model was reasonably successful and the work will continue.

Lavoie, Cotton and Hovermale [1970] also considered a moist model. This model was used to study the lake-effect storms over the eastern part of Lake Erie. The major moisture effects incorporated into the dry model were evaporation from the lake, areal mean convective precipitation and latent heat release. In order to account for these effects, the following considerations were made: in the conservation-of-specific-humidity equation, there is a term which accounts for the moisture flux through the inversion surface due to convective mixing; the rate of production of hydrometeors in cumulus clouds is given by an empirical formula which depends on the vertical wind speed at 500 m and the depth of the clouds; and the latent heat of condensation or fusion is accounted for in the energy equation. Also, the continuity of mass equation was modified to account for atmospheric compressibility effects. The computed results for certain lake-effect storms agree fairly well with field data. It appears that the Lavoie, et al., moist model is of great value in the analysis of lake-effect storms; with some modifications, it should prove to be a good forecasting tool.

Proposed Mesoscale Model

The model proposed for development in the present study will basically be that of Lavoie, et al. [1970] and Chang [1970], with some modifications that will allow the model to be a better forecasting tool. It is expected that the model will be applicable over Lake Ontario and its drainage basin for the analysis of lake storms from late September to mid-April. Also, the model may be applicable during the summer months for mesoscale lake breeze studies. In regions of intense convection, both for lake-effect storms and summer thunderstorms, it is proposed that a three-dimensional, deep convection model be used to give better vertical resolution. The regions of intense convection may be indicated by the mesoscale model. For better vertical resolution in the lake breeze problem, it is proposed that a three-dimensional, lake breeze model be used. In the remainder of this section, the mesoscale model will be discussed.

The horizontal area which will be covered by the model will be rectangular in shape and will cover Lake Ontario and its drainage area. The sides of this rectangle will lie approximately along the latitude lines of 41°50' and 45°25' and the longitude lines of 74°30' and 80°10'. Hence, the dimensions of the rectangle are about 460 km in the east-west direction and 400 km in the north-south direction. Besides the Lake Ontario drainage basin, this region includes part of the St. Lawrence River and its drainage basin, and the eastern ends of Lake Erie and Georgian Bay. Major cities included in the region are Toronto, Hamilton, Niagara Falls, Buffalo, Rochester, Syracuse and Erie.

The model computation grid network covering the region will be made-up of squares 2.5 kilometers on a side. This fine spatial resolution will place several grid squares over the larger cities and will permit the study of heat island effects. Computational time steps will be about one minute, giving the capability to study small, temporal features of meteorological phenomena. It is anticipated that, with the CDC 7600 computer available at NCAR, continuous computational studies can be made of weather phenomena that span three or four days of time.

Boundary conditions for the five unknowns (wind components, temperature, height of the inversion surface and humidity) are required along the six sides of the model domain and initital conditions are necessary at every internal point. At the lower boundary of the domain, input to the model will consist of momentum, mass (humidity) and heat fluxes, calculated from the buoy-tower and climatological station network data using bulk aerodynamic formulas. A description of interface transport processes and the methods to be used in obtaining the flux input is given in the next major section of the report. Boundary conditions at the remaining five sides of the domain and initial conditions will be specified on the basis of meteorological data made available by other IFYGL participants.

During the "tooling-up" phase of the mesoscale model, several investigations and modifications must be performed. Most of the previous investigations used the forward-upstream differencing technique to solve the system of equations in the dry and moist models. It is well known that this technique introduces a great deal of dissipation which is due to the differencing and not to the physics of the problem. There have been many extensive studies concerning the difference techniques for solving the primitive equations for barotropic fluids and the shallow water equations, Gary [1964], Richtmyer and Morton [1967], Gramneltvedt [1969] and Gustafsson [1970]. From these papers and from numerical experiments with the Lavoie, et al., moist model, the best overall difference scheme for the Lake Ontario study will be determined.

All of the previous applications of the Lavoie models considered time-invariant boundary conditions along with simplified initial conditions. Then, through the use of a marching process, the steady state solution was calculated. Hence, every former example was treated as a steady-state phenomenon. However, in the previous discussion of lake-effect storms, it is seen that the snow bands may shift over relatively short periods of time and the character of the storms is strongly dependent upon synoptic conditions which occur on the boundaries of the domain of computation. Hence, it is desirable that the vertical boundary conditions vary with time in order that they be consistent with external synoptic changes for the period being analyzed. Also, it is desirable that the surface conditions and the "drag" coefficients vary with time due to changing sea-state conditions and cloud cover. In order to account for each phenomena as thermal pollution, heat islands, changes in roughness, etc., all surface conditions and "drag" coefficients should be treated as functions of the horizontal coordinates, x and y. The models of Taylor [1969] should be helpful in formulating some of these spatial relationships.

There are two ways to account for the time-varying boundary conditions in this model. One, if there is sufficient meteorological data collected over the lake and the drainage basin, the data can be initialized so that it is consistent with the governing equations, Nitta and Hovermale [1969]. Then, for these consistent initial and boundary conditions, a solution for the time evolution of the mesoscale storm conditions can be obtained. Ideally, each new time step will produce real weather effects. However, this technique may prove to be unsuitable because of inadequate field data over the lake and basin. Another possibility is to start with time-invariant boundary conditions and simple initial conditions and then obtain the steady state solution. This steady state solution is now assumed to be the new initial conditions, and the new boundary conditions are initially equal to the original time-invariant boundary conditions; however, as time progresses, the new boundary conditions slowly change with time due to changing synoptic conditions. Therefore, the time dependent solution of this new system ideally represents true weather changes over the lake and basin.

In conjunction with this last technique of handling time-varying boundary conditions, there is a need to investigate the possibility of obtaining steady-state solutions by a method that is more efficient than a marching process. This matter will also be studied.

Through the use of a simplified domain of computation, the Lavoie terms for the pressure influence of Layer III on Layer II, for the moisture flux through the inversion surface, for the net condensation rate, and for the rate of production of hydrometeors will be investigated. If improvements upon this work can be made, new formulations will be used. If not, Lavoie's expressions will be used. Also, an investigation will be made of the effect of Chang's [1970] modification of the pressure terms due to the presence of water vapor in Layer II.

Three-Dimensional Deep Convection Model

At NCAR, a three-dimensional, deep convection model (Ogura and Phillips [1962]) which will eventually evolve into a comprehensive hail-producing thunderstorm is in the process of being formulated. The "final" model will, indeed, be very complex due to the dynamics of turbulent convection, the interacting microphysical processes, the interaction of microphysics and dynamics, and the interaction between the storm itself and larger scale phenomena. For the data year of the IFYGL program, it is expected that a tested and working model of three-dimensional, deep, turbulent, moist convection will be available. The moist part of the convection will be made up of the variables of water vapor, a distribution for cloud droplets, and a distribution for raindrops. Also, a precipitation mechanism will be included; however, detailed microphysics will probably not be incorporated in the model by 1972.

During lake-effect storms there are, occasionally, examples of intense convective activity. Since the Lavoie model cannot follow buoyant convection, although it should be capable of indicating where this convection occurs, the deep convection model will be applied in these regions to give better vertical resolution of the activity. It is also proposed that the deep convection model be applied during the summer periods if sufficient field data on thunderstorm activity is obtained during the IFYGL data year.

To date, there have been only a few three-dimensional, turbulent numerical calculations reported in the literature. Probably the most important of these calculations are those of Deardorff [1969a,b,c; 1970a,b,c,d]. In Deardorff [1969a, 1970a], the three-dimensional, primitive equations of motion have been integrated numerically in time for the case of turbulent, plane Poiseuille flow at very large Reynolds numbers. The grid used in the model was 24 x 20 x 14, or 6720 uniform grid intervals, with subgrid scale effects simulated with eddy coefficients proportional to the local velocity deformation, Deardorff [1970b]. Although the grid size is rather crude in this model, the agreement of calculated statistics against measured results ranges from good to marginal. From this work of Deardorff, it is concluded that the numerical approach to the problem of turbulence at large Reynolds numbers is profitable and the accuracy will increase with an increase in numerical resolution.

In Deardorff [1969b,c], the three-dimensional, nonlinear equations of motion were integrated numerically in time for a region of xyz-space of volume 3h x h x h, where h is a height slightly above the level where the wind first attains the geostrophic flow direction. The case treated in [1969b] consisted of a horizontal lower boundary, a neutrally stable atmosphere, horizontal homogeneity of all dependent mean variables except the mean pressure, and a statistically steady-state flow field. Deardorff investigated the relative directions of stress, wind shear and wind velocity, differences in Ekman wind spirals for the neutral case and a stable case, profiles of dimensionless turbulence statistics, effect of allowing the mean density to be either constant or to decrease with height, effect of the wind direction upon the turbulence intensities, and the characteristic structure of the eddies in the planetary boundary layer. The numerical model used by Deardorff is limited with respect to questions concerning eddies smaller than about two grid intervals in length or eddies longer than the region modelled. However, the eddies much longer than the model length appear in the zero wave-number mode. Also, the energy and momentum contained by the small-scale eddies are roughly accounted for by the small-scale eddy-coefficient assumption, while the small fraction which should be contained in the eddies somewhat larger than the model length is assumed to be carried by the calculated eddies. Both of these limitations can be gradually relaxed as computer speeds and storage are improved.

In Deardorff [1970c,d], certain preliminary results from the numerical integrations of an unstable planetary boundary layer are reported. This unstable boundary layer is devoid of clouds, mountains, or mean horizontal density gradients. The Boussinesq approximation is used and potential temperature is now included as a dependent variable. Other improvements over the previous models are the addition of a vertically distributed heat sink, a doubling of the number of grid points utilized (total being 16,000), the use of an exact solution for the pressure fluctuations rather than over-relaxation, and the use of a larger proportionality constant in the formulation of the nonlinear, subgrid scale, eddy coefficient of momentum.

The only three-dimensional, time dependent, moist, deep convection calculations of which the authors are aware are those of Shafrir, Kaniel and Shkoller [1970]. However, in this work constant eddy coefficients were used and the calculations were not of the type considered by Deardorff. That is, the only place that turbulence was considered was in the constant eddy coefficient terms; the turbulent structure of the flow was not, otherwise, considered.

It seems clear that the fluctuating quantities in a moist, turbulent, convection flow field are very important in the consideration of the interaction between microphysics and dynamics in large scale convective storms. Hence, the three-dimensional, deep convection model which will be constructed at NCAR will incorporate the "numerical" turbulent dynamics developed by Deardorff. The first step in the construction of this model, and the only part which will be ready for the IFYGL study, will be to expand the computation domain of Deardorff to thunderstorm size and to include the above-mentioned moisture variables.

Three-Dimensional Lake Breeze Model

A three-dimensional, mesoscale, lake breeze model will be developed at NCAR during the IFYGL study year. This model will be useful in studying the interactions between lake breezes, heat islands, and the topography of the lake coastline. Since the model will give more vertical resolution than the Lavoie model, it can be used as a "back-up" model in storm situations in which the Lavoie model is not applicable. If the quality and quantity of data collected during the IFYGL data year is sufficient to warrant the effort, some of the lake-breeze - urban-heat-island interactions will be analyzed. In addition, some of the thermal pollution effects on the mesoscale airflow will be analyzed in greater detail than is possible by the Lavoie model.

Some of the best numerical work in the sea-breeze area has been done by Estoque [1961, 1962a]. Magata [1965] extended the model of Estoque to include the effects of the release of latent heat by condensation and the effects of evaporation, and considered a condition of heat balance at the earth's surface by taking into account insolation, nocturnal radiation, conduction and eddy transfer of heat. McPherson [1970] extended the work of Estoque to include the three-dimensional effects of sea breezes.

McPherson's model considers a 50 m surface layer in which there are large vertical gradients of wind and temperature, a transition layer of 3.85 km in height in which the numerical calculations are carried out, and a horizontal area of 56 km by 276 km. The prognostic equations governing the behavior of the transition layer are the horizontal component equations of motion and the thermodynamic equation for the potential temperature. The values of the pressure and vertical velocity are computed from the hydrostatic equation and continuity equation, respectively. The exchange coefficients for the eddy transport processes in the transition zone are assumed to decrease exponentially with height. Other formulations of the exchange coefficients are discussed in a recent paper by O'Brien [1970]. Also, McPherson discussed in great detail the boundary and initial conditions which he used in his analysis of the effects of Galveston Bay on the sea breezes blowing in from the Gulf of Mexico.

The NCAR model will be a combination of the McPherson model and the Magata model. That is, a three-dimensional Estoque sea-breeze model with moisture effects included will be constructed. Another possible improvement to this model will be to add the effects of interactions between the atmospheric and soil boundary layers. This last type of calculation has been carried out by Estoque [1962b], Tag [1969] and Sasamori [1970] in models which do not contain the nonlinear advection terms. Hence, the objective of this proposed lake-breeze model will be to develop a three-dimensional, shallow, mesoscale model which can be used for lake-breeze studies, air and thermal pollution studies, and urban-heat-island studies, and can indicate the effects of certain air-water and air-soil interactions.

INTERFACE TRANSPORT PROCESSES

Effects on weather of a lake and its surrounding drainage basin are determined largely by the interface processes whereby momentum, mass, and heat are transported between the water or land and the atmosphere. It is therefore crucial to the success of the proposed mesoscale model and the two three-dimensional models that the transport rates, or fluxes, of these quantities be accurately evaluated for input to the models. As mentioned previously, these flux inputs must be calculated from synoptic data using bulk aerodynamic formulas, which past investigations have shown to be only approximate and to apply only under certain conditions. An important phase of the proposed study will be the analysis and evaluation of the relevant formulas based on micro-meteorological data taken over Lake Ontario and its drainage basin. Relatively accurate methods for determining fluxes from such experimental data will be discussed in the next section. Following that will be a section on the methods of calculating fluxes from synoptic data using the bulk aerodynamic formulas, and finally, a section on the technique to be used in assigning fluxes at model domain gridpoints based on the fluxes calculated at the data network sites.

Experimental Methods for Determining Fluxes

The *direct* or *cross-correlation method*, because it is relatively free of assumptions, is generally regarded as the most accurate means for obtaining the momentum, mass, and sensible heat fluxes, Hasse [1970]. The method involves obtaining coincidental time series records of horizontal velocity u, vertical velocity v, specific humidity q, and temperature T at some point in the boundary layer. Means are removed from the records leaving only the time fluctuating quantities u', v', q', and T'. The time series of u', q', and T' are each convolved with the time series of v' to obtain the time-averaged products $\overline{u'v'}$, $\overline{q'v'}$, and $\overline{T'v'}$. Flux values are then calculated directly from the expressions

$$\tau = -\rho\,\overline{u'v'} \qquad \text{Momentum flux}$$

$$E = \rho\,\overline{q'v'} \qquad \text{Mass flux}$$

$$H = c_p\rho\,\overline{T'v'} \qquad \text{Heat flux}$$

where ρ is the mean density of air at the point of observation and c_p is the specific heat of air at constant pressure. In order to obtain accurate results by this method, high response instruments are required so that the data satisfactorily represent actual fluctuating quantities over those portions of the respective power spectra where energy is significant.

The *profile method* involves measuring vertical distributions of mean horizontal velocity $\overline{u}$, specific humidity $\overline{q}$, and potential temperature $\overline{\theta}$. At a given elevation the vertical gradients of these quantities are obtained and the fluxes calculated from the relationships

$$\tau = \rho K_M \frac{\partial \overline{u}}{\partial z}$$

$$E = -\rho K_E \frac{\partial \overline{q}}{\partial z}$$

$$H = -c_p \rho K_H \frac{\partial \overline{\theta}}{\partial z}$$

where the respective coefficients K_M, K_E, and K_H are eddy viscosity, eddy diffusivity, and eddy conductivity. To a first approximation, the eddy coefficients are simply proportional to the distance from the surface and to the shear velocity $u_* = (\tau/\rho)^{\frac{1}{2}}$. However, for non-neutral boundary layer stratification, the coefficients also depend upon a stability parameter, Monin and Obukhov [1954]. This method for obtaining the fluxes is quite sensitive to environmental conditions and requires careful analysis of the data in order to account for these effects.

The *spectrum method* for determining momentum flux was recently investigated in considerable detail by Clancy [1970] and found to be significantly more accurate than the profile method. The method utilizes only the high wave number range of the downwind, horizontal velocity component power spectrum. In addition, the results of the method appear to be relatively insensitive to air-sea temperature differences. Crucial to the application of the method are the assumptions that an inertial turbulence subrange exists and flow conditions are statistically stationary and isotropic. Clancy's conclusions as to the accuracy of the method stand even where these assumptions were not fully realized in his experiments.

Bulk Aerodynamic Methods for Determining Fluxes

Synoptic meteorological data give information on the differences between properties of the air flow at one level above the water surface and properties at the surface itself. Such information can be used in semi-empirical bulk aerodynamic formulas to yield values of flux:

$$\tau = \rho C_f(a)\bar{u}^2(a) \tag{1}$$

$$E = \rho C_E(a)\bar{u}(a)[q(w) - q(a)] \tag{2}$$

$$H = \rho c_p C_H(a)\bar{u}(a)[\theta(w) - \theta(a)] \tag{3}$$

where $x(a)$ indicates the value of a parameter at elevation a, $x(w)$ is the value at the water surface, and the respective coefficients are C_f, surface friction or drag, C_E, Stanton number for water vapor transport, and C_H, Stanton number for heat transport. The coefficients are functions of stability in the boundary layer. Previous observations have shown that the variation of the fluxes over the first few decameters of the atmospheric boundary layer is at most a few percent. Consequently, theoretical treatment in this region is justified in assuming that the fluxes are constant with elevation.

Momentum flux, τ, can be expressed in the form of a shear velocity u_*, in which case the drag coefficient C_f is then given by the expression

$$C_f(a) = [u_*/\bar{u}(a)]^2 . \tag{4}$$

Application of the Monin and Obukhov similarity theory to boundary layers over the sea surface gives a velocity profile of the form

$$\bar{u}(a) = \frac{u_*}{\kappa}\left[\ell n\, \frac{z}{z_o} - \Psi(\frac{z}{L})\right] \tag{5}$$

where κ is the von Karman constant of 0.4, z_o is a roughness parameter, L is a stability length, and Ψ is a stability function dependent on the ratio z/L. Observations indicate that this equation is a good description of actual profiles for conditions ranging from stable to unstable. In terms of the Monin and Obukhov velocity profile, the drag coefficient is given by

$$C_f(a) = \frac{\kappa^2}{\left[\ell n\, \frac{a}{z_o} - \Psi(\frac{a}{L})\right]^2} . \tag{6}$$

Cardone [1969] obtained an empirical expression for z_o that assumes a dependence only on u_* and spans the range of aerodynamically smooth to rough flow:

$$z_o = C_1/u_* + C_2 u_*^2 + C_3 \qquad (7)$$

where the constants are dimensional. Based on the experimental results of Charnock [1955] and Kitaigorodskii and Volkov [1965], and consistent with the work of several other investigators, Cardone found the constants to be $C_1 = 0.684$, $C_2 = 4.28 \times 10^{-5}$, and $C_3 = -4.43 \times 10^{-2}$. In addition, he calculated the relationship between $C_f(a)$ and $\overline{u}(a)$ for a value of 10 meters and for neutrally stable conditions, i.e. $\Psi(a/L)$ equal to zero. Similar calculations can in general be done for other values of a and non-neutral stability conditions.

Many other investigations of the drag coefficient have been made, Smith [1970], Miyake, et al. [1970], Adelfang [1969], Hasse [1970], Weiler and Burling [1967], and Zubkovski and Kravchenko [1967]. Most of the results are not inconsistent with drag coefficient values calculated from the Monin and Obukhov velocity profile using the Cardone formula for z_o

Mass or *water vapor flux,* E , has been analyzed by several theoretical approaches. Rigorous analysis is not possible at present owing to the lack of complete and detailed knowledge of the transfer characteristics in the layer of air immediately above the air-water interface. The respective theories are based on differing assumptions as to the nature of these characteristics. The following discussion treats only those theories that deal with aerodynamically rough flow, the most usual condition occurring over natural bodies of water.

Sverdrup [1937] hypothesized that, even though the flow over a sea interface is aerodynamically rough, there exists a viscous sublayer next to the air-water interface within which transport of water vapor takes place by molecular diffusion. Above this diffusion layer, the transport of water vapor takes place by turbulent diffusion, and the eddy diffusivity, K_E , is a linear function of the height above the sea and depends on the roughness of the interface:

$$K_E = \kappa(z + z_o)u_* \quad . \qquad (8)$$

With these assumptions, he obtained the following expression for evaporation:

$$E = \rho u_* [q(w) - q(a)]/(\frac{1}{\kappa}\, \ell n \frac{a + z_o}{\delta_\ell + z_o} + \frac{\delta_\ell u_*}{D}) \tag{9}$$

where δ_ℓ is the thickness of the diffusion layer and D is the molecular diffusivity of water vapor in air. Based on a few field measurements by Montgomery, Sverdrup deduced the thickness of the diffusion layer in terms of u_* and ν, the kinematic viscosity of air, to be $\delta_\ell = 27.5\nu/u_*$. Comparing equations (2) and (9), the Stanton number for mass transport is

$$C_E(a) = \sqrt{C_f(a)}/(\frac{1}{\kappa}\, \ell n \frac{a + z_o}{\delta_\ell + z_o} + Sc\delta_\ell^*) \tag{10}$$

where $Sc = \nu/D$ is the Schmidt number and $\delta_\ell^* = \delta_\ell u_*/\nu$. This theory was tested, using $\delta_\ell^* = 30$, in a study of water vapor transport over Lake Hefner, Marciano and Harbeck [1954], and good agreement with observation was found. However, it was reported in a study of Lake Mead by Harbeck, et al. [1958] that the theory seemed to under-predict evaporation. Recently, Mangarella (1970] has found that a value of $\delta_\ell^* = 15$ gives a good fit to his laboratory experimental data.

Sverdrup [1946] also proposed a simpler theoretical model based on the assumptions that the turbulent boundary layer extends completely to the sea surface, and that in the boundary layer the eddy diffusivity of water vapor equals the eddy viscosity, which increases linearly with distance from the sea surface. With these assumptions, the vertical flux of water vapor through the air-sea interface is

$$E = \rho\kappa u_* [q(w) = q(a)]/\ell n \frac{a + z_o}{z_o} \quad . \tag{11}$$

Comparison of equations (2) and (11) yields

$$C_E(a) = C_f(a) \quad . \tag{12}$$

This is, indeed, the result of the Reynolds analogy. Evidence of the correctness of this theory was found in the fact that observations made on board the Indianapolis by Lt. F. L. Black, U. S. N., in 1937 were in good agreement with the theoretical results. However, comparison of the theory with observations from the study at Lake Hefner by Marciano and

and Harbeck [1954] indicates that the theory overpredicts the experimental data. The same conclusion has also been drawn by Mangarella [1970] based on his measurements in the laboratory. On the other hand, the theory under-predicts the results obtained at Lake Mead, Harbeck, et al. [1958].

Sheppard [1958] proposed a simple model which assumes there is no distinct layer of exclusively molecular transfer; molecular and turbulent exchanges are, rather, supposed to exist simultaneously. He also assumed that $K_E = \kappa u_* z$ by neglecting z_o on the argument that it is negligible compared to values of z normally used in evaluating K_E. Thus, he deduced an expression for evaporation which is confined to the fully turbulent layer:

$$E = \rho \kappa u_* [q(w) - q(a)] / \ell n \frac{\kappa u_* a}{D} \quad . \tag{13}$$

From equations (2) and (13),

$$C_E(a) = \sqrt{C_f(a)} / \frac{1}{\kappa} \ell n \frac{\kappa u_* a}{D} \quad . \tag{14}$$

Roll [1965] compared this theory with observations by Takahashi [1958] and indicates that Takahashi's evaporation measurements appear to be represented by Sheppard's formula with appreciable reliability. Recently, laboratory experimental results obtained by Lai [1969] have also been found to be in good agreement with Sheppard's theory. However, Mangarella [1970] points out that Sheppard's theory tends to over-predict his experimental data.

Kitaigorodskii and Volkov [1965] proposed a two-layer model similar to Sverdrup's but differing significantly in the diffusion layer hypothesis. They assumed that above the diffusion layer the turbulent and molecular Schmidt numbers are equal. Furthermore, in the fully turbulent region, molecular viscosity is considered unimportant for water vapor transport and eddy viscosity is taken to be $K_M = \kappa u_* z$. With the above assumptions, the evaporation can be expressed as

$$E = \rho \kappa u_* [q(w) - q(a)] / Sc (\ell n \frac{a}{\delta_\ell} + \kappa \delta_\ell^*) \tag{15}$$

and

$$C_E(a) = \sqrt{C_f(a)} \, / Sc (\frac{1}{\kappa} \ell n \frac{a}{\delta_\ell} + \delta_\ell^*) \tag{16}$$

To determine δ_ℓ , Kitaigorodskii and Volkov point out that the mobility of the roughness and the transfer of wind energy to the troubled wave surface must be considered in the transfer mechanism. Therefore, they incorporated the wave field properties in their diffusion layer hypothesis and arrived at an expression for δ_ℓ :

$$\frac{c\delta_\ell}{\nu} = \frac{1}{16P}\frac{\rho_w}{\rho}\left(\frac{gh}{u_*^2}\right)^2\left(\frac{gx}{u_*^2}\right)^{-1} \tag{17}$$

where c is the phase velocity of waves, ρ_w is the density of water, g is the gravitational acceleration, h is the wave height, x is the fetch distance, and P is a constant defined by $u(z) = P\,\frac{c^2 z}{\nu}$. They found that $P = 2.5 \times 10^{-4}$, using the data on the wave parameters and x given by Takahashi [1958]. Mangarella [1970] found that a value of $P = 2.3 \times 10^{-3}$ gives a best fit to his laboratory observations.

Heat flux , H , can be directly related to the water vapor flux. Kays [1966] has shown that the governing mass and energy conservation equations and their boundary conditions are identical for a number of common applications. Under the assumptions of molecular Lewis number and turbulent Lewis number being unity, the heat and mass transfer Stanton numbers are equal. These assumptions are quite valid for a broad range of field conditions. The molecular Lewis number, $Le = Pr/Sc$, is 1.16 for an air-water vapor mixture with low water vapor concentrations. Paulson [1967] has shown experimentally that the turbulent Prandtl and Schmidt numbers each appear to be close to unity; therefore, the turbulent Lewis number, $Le_t = Pr_t/Sc_t$, must also be close to unity. That is, the heat and mass transfer are governed by essentially similar processes and $C_H = C_E$. This result has been experimentally confirmed by Mangarella [1970]. It is possible, therefore, to predict the heat flux on the basis of the computed water vapor flux. It should be noted here that the above-mentioned theories for mass and heat transfer are confined to nearly neutral stratified conditions. For non-neutral conditions, stability effects must be taken into account in order to obtain reliable results.

Transport phenomena over bodies of water play a key role in large scale meteorological processes, yet the mechanisms governing heat and mass transfer through a wavy air-water interface over which turbulent wind is blowing have not been well established. Various formulas for estimating the vertical fluxes across an air-water interface from meteorological data have been offered by many authors and each has its individual weaknesses. The available empirical evidence is too scanty and inconsistent to decide what assumptions may be acceptable. The formulas mentioned above will be evaluated in the present study on the basis of micro-meteorological data. The formula or modified formula giving best results will then be applied to calculate mass and heat flux from synoptic data.

Technique for Assigning Fluxes to Model Gridpoints

Once fluxes are obtained for a given time at the many, irregularly distributed lake and land data stations, values must be assigned to the much larger number of model gridpoints. This will be accomplished by the successive-approximation technique (SAT) developed by the Joint Numerical Weather Prediction Unit (JNWP) of the U. S. Weather Bureau, Gilchrist and Cressman [1954] and Cressman [1959]. It is an interative procedure for using observations to correct successive guess fields so that the representation of observed conditions is increasingly improved. The initial-guess field may be obtained by the nearest-observation-to-a-gridpoint technique in which the value assigned to a gridpoint is the value observed at the nearest station. For each reporting station, an interpolated value of the initial-guess field is computed by fitting a curvilinear surface to the four surrounding gridpoints. Based on the difference between the observation itself and the interpolated value at the observation point, and on the distance-weighting function which is dependent on the influence radius and distance between the observation and a gridpoint, corrections to the guess field are computed for all gridpoints within a radius of influence of the observation. Multiple corrections result at those gridpoints where circles of influence overlap and must be combined to form a single overall correction. To obtain the first approximation to the analysis field, these gridpoint corrections are applied to the corresponding initial-guess values and, if desired, the resulting first-pass analysis field may be smoothed to remove small-scale perturbations. The smoothing operator improves the quality of the analysis between observations but at the expense of a close fit to the observations.

Subsequent approximations may be obtained by repeating the correction procedure and using the previous approximation as the new guess field. However, with each iteration, the value of the influence radius should be reduced appropriately. The reduction of the influence radius permits successive introduction of smaller-scale features.

Although SAT in its present form produces an entirely satisfactory analysis, it has one drawback that might become critical in an operational system. The problem is that when large radii are used to provide corrections in sparse-data regions, the number of gridpoints to be corrected becomes quite large, with a corresponding increase in machine time required for computations. A more efficient analysis method, Harris, Thomasell and Welsh [1966], would be one in which the curvilinear interpolation is performed deliberately and in a straightforward manner, resulting in the optimum analysis for a given set of data and a saving in computation time. Curvilinear interpolation may be accomplished very easily by requiring that gridpoint values in no-data regions be solutions of Laplace's equation. The solution in this case is determined entirely by the boundary values and is obtained through relaxation procedures. Observations translated in some appropriate manner to gridpoints are used as internal boundary points. A variation on this method requires solutions of Poisson's equation at gridpoints and uses a forcing function computed from an initial guess or a previous approximation to the desired analysis. A smoothing operator is required to smooth the analysis near isolated observations and to permit the computation of a more representative forcing function. As with SAT, multiple passes through the data are probably needed to produce the best analysis.

PLAN OF STUDY

The general scope of the proposed plan of study has been indicated in the foregoing sections. Here, the various segments of the plan will be drawn together and the comprehensive scheme of investigation explained.

Development of Numerical Models

A *mesoscale numerical convection model* will be constructed and used as the primary tool in a study of atmospheric influences of Lake Ontario. The model will be patterned after the Lavoie mesoscale model and will incorporate certain improvements that will permit more accurate simulation of lake-effect storms. The model should be applicable to the analysis of lake storms from late September to mid-April and may be suitable for mesoscale lake breeze studies during the summer months.

The mesoscale model domain will be a rectangular region covering the entire Lake Ontario drainage basin. Spatial grid squares will be 2500 meters on a side and calculation time steps will be about one minute. In the vertical direction, the model domain will be the well-mixed layer extending from 50 to 3000 meters above the basin. The model will assume that atmospheric conditions within this layer can be represented by parameters at a single level. Thus, the model will be two-dimensional in the sense that spatial resolution of the variables is possible only in the two components of horizontal direction. Free atmospheric boundary conditions and initial conditions for model calculations will be obtained from Field Year data made available by other IFYGL participants. Conditions at the lower boundary of the model domain will be calculated from synoptic data gathered during the Field Year by the buoy-tower and climitological station network planned for the Lake and its drainage basin.

During 1971, several tests of the mesoscale model will be made using a model domain that is substantially smaller than the Lake Ontario drainage basin. Determinations will be made of the best differencing scheme for the solution of the equations and of the best method for treating time-dependent boundary conditions. Evaluations will be made of the pressure gradient terms, the precipitation mechanism, and the moisture flux terms of the Lavoie and Chang models. Finally, a study will be made of model sensitivity to heat island effects, thermal pollution, and changes in surface roughness, temperature, and humidity.

In 1972, a comprehensive geometrical model for the Lake Ontario drainage basin will be constructed, incorporating the optimum numerical techniques determined in the 1971 tests. During the latter part of 1972 and early in 1973, sufficient field data should be available to run the model for several real storm cases. The results of these cases should be ready for publication during the last part of 1973.

A *three-dimensional deep convection model* incorporating turbulence and moisture effects is being developed at NCAR as part of a general effort to simulate hail-producing thunderstorms. A working model should be available by late in the IFYGL data year. The model will be used to study regions of intense convection, indicated by the mesoscale model, for both lake-effect storms and summer thunderstorms. Application of the model will be fruitful only if sufficient data on deep convective activity over the Lake Ontario area is made available.

A *three-dimensional lake breeze model* particularly suitable for mesoscale applications to lake breeze studies will be developed at NCAR by late 1972. The model will provide more vertical resolution than the mesoscale convection model and hence can be used in storm situations in which the latter model is not applicable. In addition, the Lake breeze model will be useful in studying interaction among the phenomena of lake breezes, heat island effects of cities, and the climatic influences of the drainage basin topography. This model, too, will only be applied if sufficient data is made available to warrant its use.

Evaluation and Application of Flux Formulas

The accuracy of bulk aerodynamic formulas for calculating fluxes at the air-water interface of Lake Ontario will be evaluated using comprehensive micro-meteorological data taken over the Lake. All fluxes will be calculated by the cross-correlation and profile methods. Momentum flux may also be determined from the spectrum method. Comparison among these experimental results for various conditions of stratification in the atmospheric boundary layer and wave activity on the Lake will permit assessment of the accuracy and consistency of the methods. Synoptic-type information will be extracted from the same data and used to calculate fluxes using the bulk aerodynamic formulas: the momentum flux method of Cardone [1969]; the mass flux methods of Sverdrup [1937] and [1946], Sheppard [1958], and Kitaigorodskii and Volkov [1965]; and the heat flux methods paralleling those of mass flux, i.e. for the mass and heat transport Stanton numbers equal to each other. The validity of this last condition will also be checked. Flux results from the cross-correlation, profile, and spectrum methods will be compared with results from the bulk aerodynamic methods and an evaluation of the latter made for the entire measured range of Lake and atmospheric conditions. The evaluation will result in a method or methods deemed most appropriate for calculating lake surface fluxes from buoy and tower synoptic data for use in the numerical models.

There are similar bulk aerodynamic formulas available for calculating fluxes over land, based on two-level synoptic data. Such methods will be examined using preliminary information from land climatological stations scattered over the Lake Ontario drainage basin. Should the data necessary for such calculations be lacking, a more approximate parameterization approach will be substituted.

Once fluxes are obtained for a given time at the many, irregularly distributed lake and land data stations, values must be assigned to the much larger number of model grid points. This will be accomplished by a successive-approximation technique (SAT), an iterative procedure for using observations to correct successive guess fields so that the representation of observed conditions is increasingly improved.

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