ARPA ORDER NO.: 189-1
5P10 Information Processing Techniques Office

R-1654-ARPA
March 1975

Parameterization of the Planetary Boundary Layer in Atmospheric General Circulation Models—A Review

Chandrakant M. Bhumralkar

A Report prepared for

DEFENSE ADVANCED RESEARCH PROJECTS AGENCY

Rand
SANTA MONICA, CA. 90406

The research described in this Report was sponsored by the Defense Advanced Research Projects Agency under contract No. DAHC15-73-C-0181. Reports of The Rand Corporation do not necessarily reflect the opinions or policies of the sponsors of Rand research.

PREFACE

The latest versions of the atmospheric general circulation models (GCMs) incorporate many physical processes that govern the larger scales of atmospheric circulation, although there is still need to improve their fidelity. This need has been considered to be important with respect to the parameterization--or physical representation--of sub-grid scale turbulent processes, and specifically the parameterization of the fluxes within the surface boundary layer.

This report reviews the parameterization techniques currently being used in the GCMs and suggests a series of experiments to test various parameterization schemes in the Rand two-level GCM. It is hoped that the findings from these tests, when completed, will help us to devise a scheme that represents the boundary layer processes more realistically.

This work has been done as part of the Rand Climate Program, under support from the Defense Advanced Research Projects Agency. Related Rand publications on the dynamics of climate are R-877-ARPA, *A Documentation of the Mintz-Arakawa Two-level Atmospheric General Circulation Model* (December 1971); R-1511-ARPA, *Numerical Experiments on the Computation of Ground Surface Temperature in an Atmospheric Circulation Model* (May 1974); and R-1653-ARPA, *A Survey of the Research on the Treatment of the Planetary Boundary Layer in Atmospheric Circulation Models* (in preparation).

Published by The Rand Corporation

SUMMARY

The thermal and dynamic interaction between the atmosphere and underlying surface occurs through the boundary layer. While these interactions have generally been ignored for short-term prediction of large-scale atmospheric circulations, they are quite important for long-range forecasting and in studies related to general circulation and climate dynamics. In this case a consideration must be given not only to the supply of energy but also to the dissipation of kinetic energy, as well as to the vertical transport of heat and moisture within the boundary layer. Since the intensity of small-scale processes is affected by large-scale processes, incorporation of the boundary-layer dynamics constitutes an essential part in studying the physical principles of long-range forecasting of large-scale processes and climatic changes.

Numerical general circulation models have been adopted as important tools for understanding the physical basis of climatic changes. These models vary greatly in many respects, especially with respect to the degree of detail in their treatment of those physical processes that cannot be resolved by the grid spacing of the model viz the boundary-layer processes.

The boundary-layer parameterization in a general circulation model is usually dependent on the vertical resolution of the model and is related to the determination of four factors in terms of the variables predicted by the model:

(a) surface fluxes of momentum, heat, and moisture,
(b) vertical profiles of the turbulent fluxes within the boundary layer,
(c) height of the boundary layer,
(d) vertical velocity at the top of the boundary layer.

There are several approaches to the boundary-layer parameterization aimed at incorporating the boundary-layer processes in general

circulation models. Some of these techniques are based on so-called K-theory while others are based on the similarity theory. Considering the varying degrees of vertical resolution of different general circulation models, the determination of the surface fluxes is perhaps the most important aspect of the boundary-layer parameterization in a general circulation model.

At present there is not sufficient evidence to determine which particular boundary-layer parameterization scheme is most satisfactory. Hence it is suggested that a systematic sensitivity test of various schemes be carried out to determine the best approach.

ACKNOWLEDGMENTS

The author is thankful to W. L. Gates and W. S. King for reviewing the manuscript.

CONTENTS

LIST OF SYMBOLS

C_D	drag coefficient
C_E	moisture transfer coefficient
C_H	heat transfer coefficient
E	moisture flux
E_N	entrainment rate
$\vec{F}$	net frictional force
GW	measure of ground wetness
h	height of boundary layer
$K_M/K_H/K_E$	eddy diffusion coefficients for momentum, heat, and moisture
ℓ	mixing length
p	pressure
q	mixing ratio
Q	heat flux
R	gas constant
Ri	Richardson number
Ri_B	bulk Richardson number
S	stability parameter
S_θ	net diabatic heating
S_q	net moisture addition rate
T	temperature
U_*	friction velocity
$\vec{V}$	vector horizontal wind
W	vertical velocity
Z	vertical coordinate
Z_o	roughness parameter
Z_s	height of surface layer (anemometer level)
α_C	empirical constant

α angle between the surface stress and the free atmospheric wind

γ_{CG} counter-gradient heat flux

λ longitude

ϕ latitude

ρ density

θ potential temperature

$\vec{\tau}$ stress

I. INTRODUCTION

PARAMETERIZATION AND NEED FOR IT

In setting up numerical models of atmospheric processes, the scales of primary interest are included within the available resolutions of grid size. For example, for large-scale motions the domain of interest is the entire globe, and the large-scale processes themselves can be resolved by prescribing a horizontal grid width of a few hundred km and a vertical grid mesh size of 50 m in the lower atmosphere and several kilometers in the free atmosphere. However, resolutions of such magnitude cannot account for significant interactions with the scales that have been truncated. Whereas such scales have been neglected for short periods ($\sim$ a day or so), for longer periods they are considered to be quite important, both as a source as well as a sink of energy for the explicitly resolved scales. And since these processes (called sub-grid or sub-resolution processes) cannot be explicitly handled in the relevant numerical model, their effects are usually incorporated by expressing their statistical effect on the large- (explicitly resolved) scale processes in terms of large-scale parameters. This technique of relating interactions between resolvable and unresolvable processes is called *parameterization*, and it involves empirically related parameters which are determined either on the basis of observations or from theoretical considerations.

PROBLEMS IN PARAMETERIZATION

Perhaps the most important problem in large-scale modeling is to first identify which sub-grid scale processes are of importance. This job is complicated because in the atmosphere various processes (on different scales) tend to occur simultaneously, making it difficult to isolate a particular scale process and determine its role meaningfully, especially with respect to its interaction with the resolved scale processes. In view of the complexities and interdependence of atmospheric processes on different scales and also the lack of knowledge about their dynamics, parameterization of some sub-grid scale processes is generally intuitive until the scheme is actually included and tested in a large-scale numerical model. Since the parameterization techniques are not unique and can be approached in many different ways, it is possible to get involved in a lot of details; this should be avoided unless, of course, the fundamental physics is involved.

The first and most critical step in developing a parameterization technique is to be able to define the problem precisely. It is important to determine which of the several different smaller scale processes (present simultaneously) have important interactions with the large-scale process. This can be done by a combination of steps:

- We can use all the available observations of the relevant sub-grid scale process to obtain a realistic description of the phenomena and thus determine universal values for empirical constants.
- We can define parameterization hypotheses on the basis of observed physical features. These hypotheses can be tested by incorporating the parameterization in an actual general circulation model (GCM) and calibrating it until a realistic simulation of the "modeled" process is attained. For this purpose, all existing *explicit* models of the sub-grid scale processes can be used to get a better insight into the physics and dynamics of the process. This is especially true in the parameterization of the atmospheric boundary layer in a GCM.

We have seen that the atmosphere, which physically is a continuum, is treated in numerical models as a fluid with finite numbers of degrees of freedom. The number of degrees of freedom is determined in the horizontal by a distance d between discrete grid points and in the vertical by a number of levels n. In most large-scale models, $d \geq 300$ km and $n \leq 20$. As a consequence, the theoretical model equations cannot describe sub-grid scale boundary-layer turbulent processes, and it becomes necessary to develop certain relations for stresses and turbulent heat and moisture fluxes in terms of external (to the boundary layer) variables resolved and computed by the model.

The interaction of the atmosphere with the underlying surface takes place via the atmospheric boundary layer within which the model equations should incorporate the predominantly vertical turbulent fluxes of momentum, heat, and moisture. These fluxes are usually large in the vicinity of the underlying surface and tend to decrease with height. Considering the vertical resolution of current global circulation models, it appears desirable to neglect the detailed vertical structure of the boundary layer and restrict parameterization of boundary-layer processes to the incorporation of near-surface values for horizontal components of momentum, heat, and moisture fluxes in the model equations. Thus the parameterization methods should be able to calculate these fluxes in terms of the large-scale output of the model as well as the parameters of the underlying surface. It may be noted that most of the parameterization techniques used in GCMs are based on the *local* considerations of the boundary layer, whereas the GCMs themselves deal with data averaged for horizontal distances of the order of several hundreds of km. While the effects of such averaging is not likely to be very large over oceans (Fleagle et al., 1967), they can be very significant over land.

Thus the purposes of this report may be stated as:

1. To discuss the interactions between the atmospheric boundary layer and the large-scale atmospheric circulation.

2. To discuss the parameterization techniques currently being used in GCMs.

3. To discuss the testing of various parameterization schemes in a specific GCM in order to estimate the advantages and deficiencies of various approaches.

II. ATMOSPHERIC BOUNDARY LAYER AND LARGE-SCALE ATMOSPHERIC CIRCULATIONS

It is generally acknowledged that the development of research on atmospheric boundary-layer (b.l.) dynamics has a direct bearing on many problems relating to micrometeorological processes within the b.l. For example, the turbulent state of the boundary layer plays a decisive role in the phenomenon of the diffusion of atmospheric pollutants, and also the dissipation of the kinetic energy of the atmosphere which occurs to a considerable extent in the boundary layer. It is through the boundary layer that there is thermal and dynamic interaction between the atmosphere and the underlying surface. Thus a knowledge of various features of the b.l. is essential for numerical weather forecasting and also for simulating the dynamics of climate. Whereas the dissipative role of boundary-layer turbulence is generally ignored for the *short-range* forecasting of *large-scale* atmospheric circulations, this cannot be done in the case of long-range forecasting, general circulation studies, and studies of climate. This is because in the latter cases (unlike short-range forecasting) consideration must be given not only to the supply of energy, but also to dissipation of kinetic energy within the b.l. and to the vertical transport of heat and moisture through the boundary layer. Since the intensity of micrometeorological processes is, in turn, affected by the large- (macro-) scale processes, the study of b.l. dynamics must constitute an essential part in studying the physical principles of long-range forecasting of large-scale phenomena and the simulation of climate.

We attempt to delineate the *interaction* between the b.l. and the free large-scale atmosphere through answers to some specific questions.

Q. *What is the basic information that relates the large-scale processes to the boundary-layer processes?*

The interaction between the atmosphere and the underlying surface occurs through the b.l., and the vertical transport of momentum, heat, and moisture appears to be the basic mechanism of exchanges between the atmosphere and the earth. Whereas the small-scale turbulent fluxes of the quantities are continuous and reach their maximum within the b.l., they are intermittent in the free atmosphere, beyond the b.l. The exchange between the b.l. and the free atmosphere takes place mainly through regular vertical transports caused by evolution of meteorological fields on synoptic scales and frictional convergence in the boundary layer. Also, the exchange caused by meso-scale cumulus convection, although relatively random, occurs through the boundary layer and is essential for larger scale atmospheric dynamics. Figure 1 shows the temperature spectrum in the surface boundary layer. It can be seen that regions of the spectrum corresponding to the small-scale turbulence and relatively large period variation of temperature are divided by a gap of periods of the order of 10 minutes (Koleshnikova and Monin, 1965). In other words, the interaction between the atmosphere and the underlying surface occurs in two stages. First there are fluxes through small-scale turbulence from the earth to the boundary layer, and at the next stage, these are transformed from the boundary layer to the free atmosphere by comparatively regular vertical motions.

Thus the basic information required for describing the large-scale processes (in terms of the boundary-layer processes) relates to the

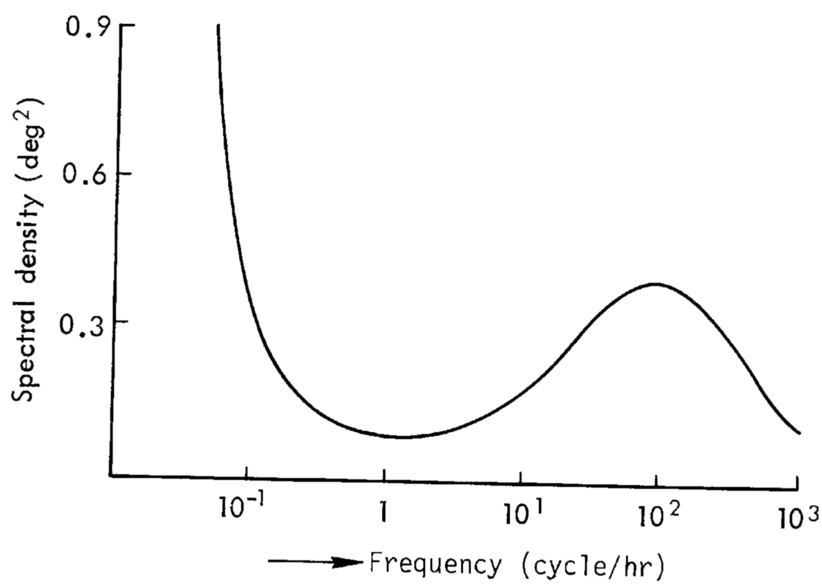

Fig. 1--The temperature spectrum in the surface boundary layer (Koleshnikova and Monin, 1965)

turbulent fluxes of momentum, heat, and moisture near the underlying surface.

Q. *What is the important link between the b.l. and the free atmosphere?*

One important link between the boundary layer and the free atmosphere is the *vertical velocity* at the top of the boundary layer. Since boundary layers are of limited vertical extent, it is vertical velocity that transports momentum, heat, and moisture *out* of the boundary layer *into* the free atmosphere. This velocity, typically of the order of mm/sec, is largely produced by frictionally modified flow near the surface and has a significant effect on the production of cloud and the general development of large-scale circulations. However, the mechanism by which the boundary layer and the free atmosphere interact is not understood adequately. It is necessary to study *temporal variations* of the boundary layer to understand the mechanism by which a previously unstable boundary layer becomes stable or vice versa. This could occur by advection and/or by entrainment. The latter term is more applicable near the top of the boundary layer where the intensity of turbulence and/or stability of the air above the boundary layer determines the extent of incorporation of the free atmosphere air with that from the boundary layer (or vice versa).

The pressure that determines the gradient wind, and the precipitation which falls through the atmosphere to the surface, are the other links between the free atmosphere and the b.l.

Q. *What is the effect of underlying surface roughness characteristics on the large-scale atmospheric circulations?*

The planetary boundary layer is defined as the layer of air in which surface frictional effects are significant. Studies of the b.l. in micrometeorology have examined the variations in momentum flux due to spatial variations in surface roughness characteristics in great detail. Elliot (1958), who studied the effect of abrupt changes in surface roughness on momentum flux, suggested that an internal boundary layer separates the flow that is in equilibrium with the upwind surface from the flow that is approaching equilibrium with the downwind surface. Panofsky and Townsend (1964) generalized Elliot's model and also came to the conclusion that there is a zone of transition (in the vertical) that separates flow above that has not "felt" the new surface, from flow below that is nearly in equilibrium with the new surface roughness. In addition, various other studies have clearly demonstrated the importance of incorporating variations in surface roughness in local micro-meteorological studies.

There is considerable interest in the mechanisms of energy transfer in the atmosphere due to increased interest in long-range weather forecasting. Lettau (1959), by estimating the mean reservoir of atmospheric mechanical energy and the *global* mean of energy dissipation, has shown that there is a renewal period for kinetic energy of about three days. This would suggest that any long-term prediction scheme for a period greater than one or two days must include a proper specification of mechanisms for frictional dissipation. As a consequence it is necessary to obtain a fairly precise knowledge of both spatial and temporal variations of dissipation effects on (large) scales. Lettau has attempted, both empirically (1959) and theoretically (1962), to solve the general problem of including surface fluxes in the large-scale systems by relating them to the external parameters controlling the flow. He introduced the concepts of geostrophic drag coefficient and surface Rossby number, both of which depend on, among other factors, the surface roughness length. Kung (1963) subsequently used Lettau's model to estimate climatological patterns of energy dissipation over the entire Northern Hemisphere. His calculations were based on his assessments of the surface roughness parameter, which in turn were based on *land use* statistics at various (360) locations. Holopainen (1963) has also estimated frictional dissipation over the British Isles for a 3-month period.

On a global scale the surface characteristics can vary over a wide range, as over mountains of varying sizes and shapes, forests, etc. As indicated by Sawyer (1959), large-scale circulations of the atmosphere are affected by the characteristics of the underlying surface through different mechanisms which depend on the physical extent of surface characteristics. For example, large-scale motion can be affected by:

- o Surface roughness over land features whose dimensions may range from a few cm to a few hundred meters, as well as those ranging from a few km to several hundreds of km.
- o Differences between the roughness characteristics of land and ocean.

In each of these cases it may be necessary to consider the effect of local surface characteristics. For example, in a global circulation study, it may be desirable to incorporate the effect of a mountain *as well as* the local "roughness" characteristics of the mountain itself.

Q. *How does the surface drag within the boundary layer modify the large-scale atmospheric circulations?*

Charney and Eliassen (1949), who were the first to pose this question, have described the principal dynamical mechanism involved. It is argued that motion with the friction (surface) layer occurs with approximate balance among the horizontal pressure gradient, Coriolis acceleration, and the eddy stress. This results in the Ekman drift across the isobars toward low pressure, which causes horizontal convergence of air within the friction layer into regions of cyclonic vorticity and divergence from regions of anticyclonic vorticity. Consequently, a compensating net divergence must occur in the free atmosphere. There is a positive correlation between divergence and vorticity in the free atmosphere which leads to deceleration of the winds. This theory, of course, would fail in case of deep convection where the effective eddy stress is not confined to a relatively shallow layer, and also over the oceans where the eddy stress in the free atmosphere may be comparable with the surface stress (Sheppard et al., 1952). Over *land* the theory is quite useful since eddy stress in the surface layer is much greater than that in free atmosphere. This theory has been used to determine the most important link between the b.l. and free atmosphere, namely the vertical velocity (W_h) at the top of the b.l. in terms of free atmospheric variables, such as geostrophic vorticity at the top of the b.l. Mahrt (1974) has extended Charney and Eliassen's theory to depict W_h as a function of geostrophic relative vorticity, drag coefficient, and boundary-layer depth.

Q. *To what height does the b.l. extend into the atmosphere and what are the factors which determine this height?*

The b.l. thickness varies not only in time but also in space. The study of boundary-layer thickness and its variation from hour to hour, from day to day, and from season to season has become the subject of extensive research. The boundary-layer thickness depends on many factors, including the following:

- stability of the surface layer which in turn depends upon the time of the day,
- entrainment of free atmosphere air into the boundary layer near its top,
- large-scale vertical motions at the top of the boundary layer,
- horizontal advective changes near the boundary-layer top (which control development of boundary layer over sea),
- radiative heat fluxes and latent heat effects which are significant when there are clouds present within the boundary layer,
- the lapse rate of potential temperature in the air above the boundary layer and the intensity of the inversion in case of mixed layers.

Figure 2 shows the diurnal variation of the *mean* boundary-layer thickness under strongly convective daytime conditions. It can be seen that it varies from a few tens of meters at night to nearly 2000 m during the afternoon. By late afternoon there is no further thickening of the boundary layer, and around sunset the boundary layer becomes quite shallow and grows slowly through the night to complete the diurnal cycle. Thus there are significant discontinuities in the b.l. depth near sunrise and sunset. A detailed quantitative treatment of determining the height of the b.l. has been discussed by Bhumralkar (1974) in a separate survey report.

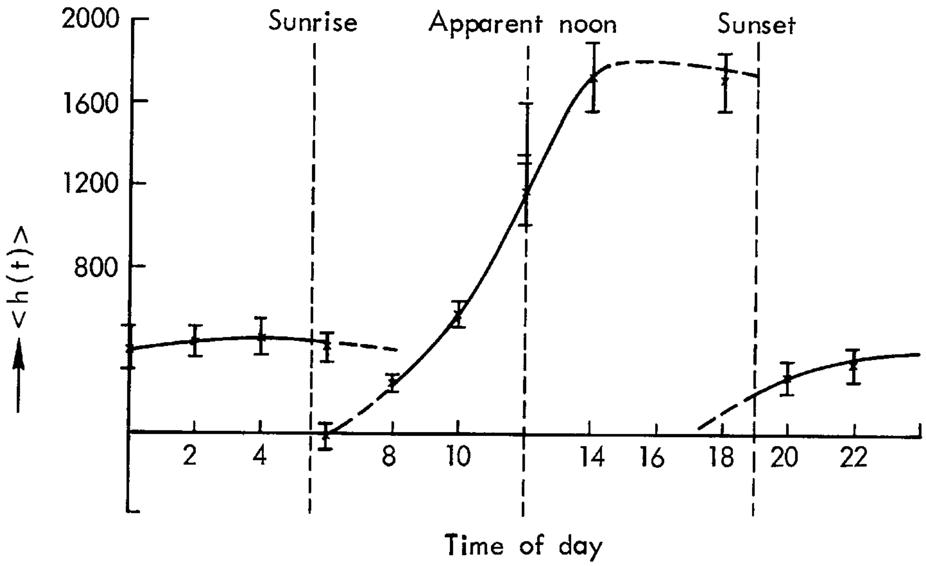

Fig. 2--The mean boundary-layer thickness, <h(t)>, deduced for the O'Neill data and plotted with standard errors as functions of time of day, t, in Mean Solar Time (Carson, 1973)

From the above it is clear that treatment of the boundary layer for studies of the general circulation of the atmosphere presents much more formidable problems than those encountered in smaller (micro/meso/synoptic) scale studies. The most important problems pertain to the wide range of variations in the b.l. characteristics over the entire globe and the lack of a "universal" theory to study these.

III. GOVERNING EQUATIONS OF THE GENERAL CIRCULATION MODEL RELEVANT TO THE BOUNDARY-LAYER PROCESSES

INTRODUCTION

The dynamical (numerical) general circulation models (GCMs) have been developed with the immediate objective of understanding the general circulation of the atmosphere. They have also been adopted as important tools for the longer range objective of understanding the physical basis of climatic changes.

During the last decade or so there has been significant progress in the physical and theoretical basis for the development of numerical modeling of the general circulation due to our increased understanding of physical processes in the atmosphere. However, most of these processes have been incorporated in GCMs through parameterization techniques only. The type and degree of parameterization, of course, depends on the horizontal and vertical resolution of the particular model. For example, on one hand the most simple models are the vertically and zonally averaged models (Sellers, 1973; Budyko, 1969) which parameterize the transport of heat in terms of mean zonal variables; on the other hand, there are models of very high resolution which are capable of resolving the details of cyclone-scale motions. However, even the models of high resolution are required to parameterize certain physical processes through empirical or statistical representation of nonresolved (sub-grid scale) processes in terms of the resolved parameters.

All GCMs are based on the fundamental hydrodynamical equations that govern the large-scale behavior of the atmosphere. The governing equations of a GCM consist of the momentum, hydrostatic, thermodynamic energy, continuity of mass, and water vapor continuity equations. They are:

$$\frac{\partial \vec{V}}{\partial t} + \vec{V}\cdot\nabla\vec{V} + w\,\frac{\partial \vec{V}}{\partial z} + 2\Omega \times \vec{V} + \frac{1}{\rho}\,\Delta p = \vec{F}\ , \tag{1}$$

$$\frac{\partial p}{\partial z} + \rho g = 0\ , \tag{2}$$

$$\frac{\partial \theta}{\partial t} + \vec{V} \cdot \nabla \theta + w \frac{\partial \theta}{\partial z} = S_\theta \quad , \tag{3}$$

$$\frac{\partial \rho}{\partial t} + \nabla \cdot \rho \vec{V} + \frac{\partial}{\partial z} (w\rho) = 0 \quad , \tag{4}$$

$$\frac{\partial q}{\partial t} + \vec{V} \cdot \nabla q + w \frac{\partial q}{\partial z} = S_q \quad , \tag{5}$$

$$p = \rho RT \ . \tag{6}$$

Here $\vec{F}$, S_θ, and S_q represent, respectively, the sources and sinks of momentum, heat, and moisture due to various physical processes in the atmosphere. The system of governing equations ($1 \rightarrow 6$) is closed by parameterizing $\vec{F}$, S_θ, and S_q in terms of the large-scale (dependent) variables $\vec{V}$, θ, q, and ρ. GARP Publications Series No. 14 (June 1974) has compiled the current status of global models and has given details of the GCMs that are in use all over the world. It is evident that the models vary considerably in many respects, particularly with regard to the degree of detail in their treatment of the sub-grid scale physical processes. At present there is neither sufficient experimental nor theoretical evidence to determine which particular treatment (parameterization) of the unresolved physical processes included in $\vec{F}$, S_θ, and S_q is the most satisfactory.

GOVERNING EQUATIONS RELEVANT TO THE BOUNDARY LAYER

As stated earlier (Sec. I) the most important problem in large-scale modeling is to first identify which sub-grid scale processes are of importance. Here we attempt to identify the contribution of the b.l. processes to the large-scale general circulation and write the governing equations of the GCM as if other physical processes were nonexistent.

Physically, the term $\vec{F}$ (Eq. 1) represents the *net* frictional force consisting of the frictional drag at the earth's surface ($\vec{F}_S$) as well as the internal friction in the free atmosphere ($\vec{F}_I$). The former is the part related to the boundary layer. The flowchart in

Fig. 3 shows schematically the various components of $\vec{F}_s$ (Box numbers 3 to 5) which are to be parameterized in a GCM for incorporating the physical effects of the boundary-layer friction.

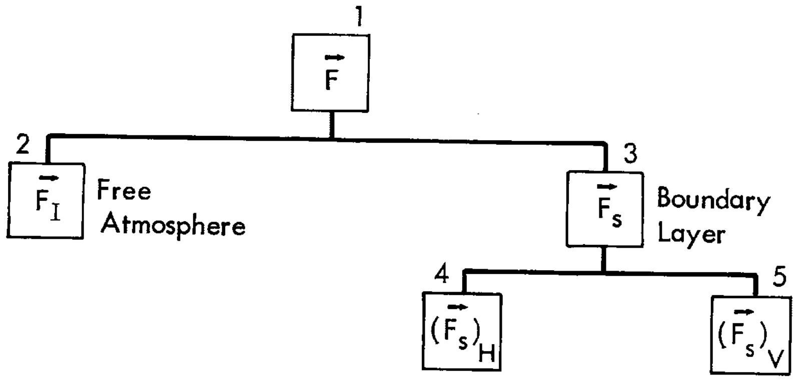

Fig. 3--Flow diagram showing the components of $\vec{F}_s$, where $(\vec{F}_s)_H$ and $(\vec{F}_s)_V$ represent, respectively, the horizontal and vertical components of the frictional drag at the earth's surface.

The term S_θ (Eq. 3) represents the net diabatic heating rate and consists of three parts:

- S_A: heating/cooling due to radiation,
- S_C: heating due to release of latent heat during condensation,
- S_D: heating/cooling due to turbulent heat transports.

It is evident that the last part represents boundary-layer effects. The flow diagram (Fig. 4) shows the components of S_θ (box numbers 5 to 9) which are relevant to the boundary-layer processes and should be parameterized in a GCM.

The terms S_{DV}, S_{DL}, and S_{DR} are evaluated only at levels within the b.l., and the term S_{DH} is evaluated at all levels of the GCM.

Finally, the term S_q (Eq. 5), which represents the net moisture addition rate, consists of the difference between the evaporation rate and the condensation rate. The evaporation can take place both from the underlying surface as well as clouds and precipitation in the free atmosphere. For the boundary-layer processes, we will be mainly

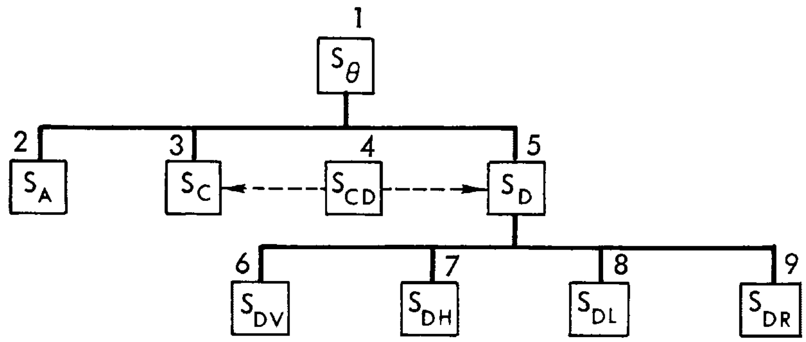

Fig. 4--Flow diagram showing components of S_θ, where S_A= radiation effects (short wave + long wave); S_C= consensation in free atmosphere (cumulus scale + large scale); S_{CD}= interaction between the b.l. and cumulus scale processes; S_D= turbulent transport of heat (boundary layer effects); S_{DV}= convergence of vertical flux of sensible heat; S_{DH}= horizontal diffusion of sensible heat; S_{DL}= condensation effect when fog/stratus clouds are present within the boundary layer; S_{DR}= radiation cooling within the boundary layer when it contains fog/stratus clouds.

concerned with the evaporation from the underlying surface as well as from fog/stratus clouds which may be present within the boundary layer. This can be brought about by the turbulent flux of moisture, both in the vertical and horizontal directions. The flow diagram (Fig. 5) illustrates the parts of S_q *which are related to the boundary layer*. These are indicated by box numbers 7 to 10.

In view of the above considerations, we can rewrite the governing equations for momentum, thermodynamic energy, and water vapor *for the part related to the boundary layer* as follows:

$$\frac{d\vec{V}}{dt} = \ldots\ldots + (\vec{F}_s)_V + (\vec{F}_s)_H \quad , \tag{7}$$

$$\frac{d\theta}{dt} = \ldots\ldots - S_{DV} + S_{DH} + S_{DL} - S_{DR} \quad , \tag{8}$$

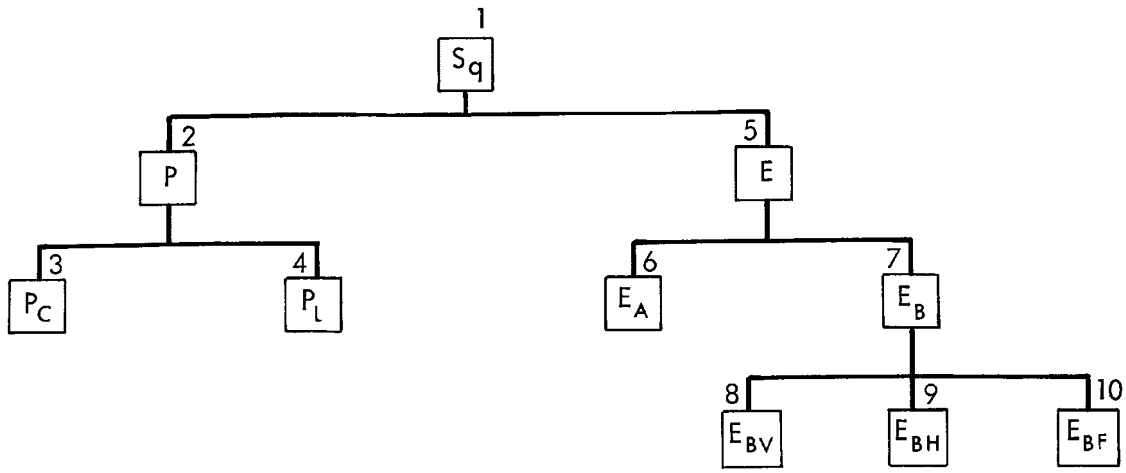

Fig. 5--Flow diagram showing the components of S_q, where P= precipitation; P_C= cumulus scale precipitation; P_L= large-scale precipitation; E= evaporation; E_B= boundary layer effects; E_A= evaporation from clouds and precipitation in free atmosphere; E_{BV}= vertical water vapor flux through the b.l. (evaluated in the b.l. only); E_{BH}= horizontal diffusion of water vapor within the b.l. (evaluated at all levels in the GCM); E_{BF}= evaporation from fog/stratus clouds within the boundary layer. (Note: Interaction between the b.l. and cumulus scale processes and the effect of precipitation falling through the b.l. should also be considered.)

$$\frac{dq}{dt} = \ldots\ldots - E_{BV} + E_{BH} - E_{BF} \quad , \tag{9}$$

where the terms on the right-hand sides of Eqs. (7)-(9) have been identified above in the flow diagrams for $\vec{F}$, S_θ, and S_q. It may be mentioned here that in these equations we have not included contributions due to interactions between the b.l. and the cumulus scale processes, and consequently the terms on the right-hand side are strictly for the part related to the b.l. processes only.

The problem of parameterization of the b.l. processes in a GCM thus reduces to the determination of the terms $\vec{F}_V$, $\vec{F}_H$, S_{DV}, S_{DH}, S_{DL}, S_{DR}, E_{BH}, E_{BV}, and E_{BF} in terms of the large-scale variables provided by the GCM.

IV. PARAMETERIZATION OF THE BOUNDARY-LAYER EFFECTS IN CURRENT ATMOSPHERIC GENERAL CIRCULATION MODELS

INTRODUCTION

There are various atmospheric models with primitive equations that are either hemispheric or global and with varying degrees of vertical resolution. The ways of treating the various physical processes differ considerably depending on the purpose for which the models are designed. Most present models are applied to weather forecasting for 1 to 5 days, while some are being developed as tools to study atmospheric changes beyond a few weeks or longer, i.e., to study climatic variations. In this report we consider this latter type of GCMs only. These are the GCMs developed by

- Geophysical Fluid Dynamics Laboratory (GFDL), NOAA, USA,
- National Center for Atmospheric Research (NCAR), USA,
- The Rand Corporation, USA,
- Goddard Institute for Space Studies (GISS), NASA, USA,
- University of California at Los Angeles (UCLA), USA,
- Meteorological Office (UK-I) and Universities Modeling Group (UK-II), UK,
- Hydrometeorological Research Center, USSR.

We will now consider specific formulations which have been used in various GCMs to determine the terms on the right-hand side of Eqs. (7)-(9). Whereas all the GCMs considered in this report evaluate the vertical eddy fluxes of momentum, heat, and moisture (represented by $\vec{F}_V$, S_{DV}, and E_{BV}, respectively), only some incorporate horizontal sub-grid scale mixing ($\vec{F}_H$, S_{DH}, and E_{BH}) through use of the nonlinear eddy diffusion coefficient proposed by Smagorinsky (1963). The importance of *interaction* between the b.l. and cumulus scale processes has been emphasized both observationally (Gray, 1972) and theoretically (Arakawa and Schubert, 1974). However, this interaction (term S_{CD}, Fig. 4) is a complicated problem requiring much further research, and we will not consider it in our report. The other physical effects of b.l. which have not been generally included in GCMs pertain to condensational and radiative effects (terms S_{DL} and S_{DR}, Eq. 8) within the b.l. Again these have been included only in the UCLA GCM, which considers boundary layers with stratus clouds as well as fog.

From the above discussion it appears that the terms that have been parameterized to incorporate b.l. effects in GCMs are those which represent the vertical turbulent fluxes of momentum ($\vec{F}_V$), heat (S_{DV}), and moisture (E_{BV}), which of course include surface fluxes. Thus each of the terms $\vec{F}_V$, S_{DV}, and E_{BV} consists of two parts: (1) the flux ($\vec{F}_o$) within the surface layer (usually considered to be a constant flux layer) extending from the underlying surface to a few tens of meters above, and (2) the flux ($\vec{F}_u$) within the so-called Ekman layer that lies from the top of the surface layer to the top of the b.l. Thus we have

$$\begin{aligned} \vec{F}_V &= \vec{F}_o + \vec{F}_u \ , \\ S_{DV} &= S_o + S_u \ , \\ E_{BV} &= E_o + E_u \ . \end{aligned} \tag{10}$$

Any b.l. parameterization in a GCM must be able to treat the b.l. processes in a physically realistic manner and relate the calculated variables (within the GCM) to the following four factors:

(1) the surface values of the turbulent fluxes of momentum, heat, and moisture;
(2) vertical profiles of the turbulent fluxes (as well as of the meteorological parameters) within the b.l.;
(3) the height of the b.l.;
(4) the vertical velocity at the top of the b.l.

We now consider the determination of the above factors by the current GCMs. Considering the varying degrees of vertical resolution of different GCMs (which may or may not enable determination of actual profiles of fluxes), determination of the surface fluxes is perhaps the most important aspect of b.l. parameterization in GCMs.

DETERMINATION OF SURFACE FLUXES

The actual method of parameterization mainly depends on whether or not the b.l. is included explicitly in the numerical model. It also depends on the closeness of the first internal GCM level, at which the complete set of model-governing equations is solved, to the underlying surface. For example, if the first GCM level is such that the whole b.l. is placed below it, then *only* the surface/near-surface values of the turbulent fluxes of momentum, heat, and moisture are relevant for the b.l. parameterization, and these are determined diagnostically. This can be achieved by the use of either some resistance, heat transfer, and moisture transfer laws or by some bulk transfer formulas. These techniques are also applicable when the lowest GCM level is placed within the surface layer (or constant stress layer) at a height of about ten or a few tens of meters.

Case I. Bulk Type Formulas

The surface fluxes are calculated by the bulk formulas:

$$\vec{F}_o = \vec{\tau}_o = \rho_s C_D V_s |\vec{V}_s| \quad ,$$

$$S_o = Q_o = - \rho_s C_p C_H |\vec{V}_s| (T_b - T_s) \quad , \tag{11}$$

$$E_o = E_o = - \rho_s (GW) C_E |\vec{V}_s| (q_b - q_s) \quad .$$

(The above equations are solved together with the heat budget equation.) Here $\vec{\tau}_o$, Q_o, and E_o are the surface stress, heat flux, and moisture flux, respectively. $\vec{V}_s$, T_s, q_s, and ρ_s are wind vector, temperature, moisture, and density at the level s. T_b, q_b are temperature and moisture at the underlying surface, and GW is the ground wetness ($0 \rightarrow 1$).

Case II. Monin-Obukhov Similarity Theory Formulas

According to this theory the surface values of the turbulent fluxes are given by α, which is the angle between the surface stress ($\vec{\tau}_o$) and $\vec{V}_s$, and by the values of U_*, Q_o, and E_o. The formulations are

$$\vec{\tau}_o = \rho_s U_*^2 \quad ,$$

$$Q_o = - \rho_s C_p C_H |\vec{V}_s| \Delta\theta \quad , \tag{12}$$

$$E_o = - \rho_s (GW) C_E |\vec{V}_s| \Delta q \quad .$$

Here $\Delta\theta$, and Δq are the differences between potential temperature and the mixing ratio, respectively, at the *boundaries* of the surface layer.

In order to evaluate $\vec{\tau}_o$, Q_o, and E_o in GCMs (from Eqs. (11) and (12)) the knowledge of the following parameters is essential:

- o height of the top of surface layer (for Eq. (11)) and height of the b.l. (for Eq. (12));
- o $\vec{V}_s$, T_s, and q_s, or the wind, temperature, and moisture at level s;
- o T_b, q_b, the temperature and moisture at the underlying surface;
- o GW, the wetness measure of the ground ($0 \rightarrow 1$);
- o values of the coefficients C_D, C_H, and C_E;
- o Z_o, the roughness parameter (Eq. (12) only);
- o α (Eq. (12) only), the angle between the surface stress and $\vec{V}_s$;
- o U_* (Eq. (12) only), friction velocity.

Table 1 gives a description of the values of the parameters involved in formulations of both Case I and Case II and used in various GCMs. It can be seen that out of eight GCMs considered, four use bulk aerodynamic formulations (Eq. (11)) to evaluate surface fluxes whereas three others use Eq. (12). The technique used by the UCLA GCM is

Table 1

COMPUTATION OF SURFACE FLUXES IN VARIOUS GCMS

Model	Fluxes Calculated from Eq. (11) or Eq. (12)	No. of Levels Coordinate System	Location of Level s Lowest GCM Level/Layer	Values for $C_D/C_H/C_E$	Determination of $\vec{V}_s$	Determination of T_s q_s	Determination of T_b	Determination of q_b	Remarks
1. GFDL (1971) Holloway & Manabe (1971)	Case I Eq. (11)	9/σ	75 m 75 m	$C_D = C_H = C_E$ (i) $C_D = 2 \times 10^{-3}$ (both land and sea) (ii) (C_D)ocean $= 1.1 \times 10^{-3}$ (C_D) land $= 4.3 \times 10^{-3}$	All three calculated from prognostic equations at the level s		Sea: specified Land: heat budget equation at the surface	Saturated q at T_b	GFDL (1974): 18 levels in the vertical; level s at 36 m
2. NCAR (1971) Kasahara & Washington (1971)	Case I Eq. (11)	6/Z	No explicit level s in the GCM Lowest GCM level = 3 km	$C_D = C_H = 3 \times 10^{-3}$ $C_E = 0.7\ C_D$ to reduce evaporation	All three calculated diagnostically by equating surface fluxes to the fluxes in layer from s to the lowest GCM level (at 3 km) (see text)		Same way as GFDL	Same as GFDL	
3. RAND (1971) Gates et al. (1971)	Case I Eq. (11)	2/σ	No explicit level s in the GCM Lowest GCM level = 800 mb = σ = .75 layer = 600 mb = σ = .50	For neutral surface layer (C_D)ocean $= \min[(1 + .07\lvert\vec{V}_s\rvert)\ 10^{-3}, .0025]$ (C_D)land $= .002 + .006 \left(\frac{Z_s}{5000\ m}\right)$ also $C_D = C_H = C_E$	$\vec{V}_s$ obtained by linear extrapolation of predicted velocities at levels 1 & 3 to the lower boundary	T_s, q_s are obtained by equating surface fluxes of sensible heat and moisture to the same fluxes from s to the lowest GCM level (at σ = .75	Prognostic equation	Saturated q at T_b	
4. GISS (1974) Somerville et al. (1974)	Case I Eq. (11)	9/σ	No explicit level s in the GCM Lowest GCM level ~ 1 km layer ~ σ = .9	For neutral surface layer same as in RAND GCM For non-neutral conditions (for all surfaces) $C_D = (C_D)\text{neutral}\ \frac{1}{1 - 7\ \Delta T/\lvert\vec{V}_s\rvert^2}$ for $\Delta T < 0$ (stable) $C_D = (C_D)\text{neutral}\ (1 + \sqrt{\Delta T/\lvert\vec{V}_s\rvert^2}$ for $\Delta T > 0$ (unstable) $\Delta T = T_b - T_s$ also $C_D = C_H = C_E$	$\vec{V}_s$ obtained by extrapolating velocities in layers 8 and 9 to the lower boundary	T_s, q_s same as in RAND except the surface fluxes are equated to fluxes from 5 to σ = .9, the lowest GCM level	Prognostic equation	Saturated q at T_b	
5. UCLA (1974) Arakawa and Mintz (1974)		3/σ	No explicit level s in the GCM Lowest GCM level σ = 8/9 Layer σ = 7/9	C_D and C_H have a functional dependence on bulk Richardson number and (h/Z_o) h is the depth of b.l. Z_o is roughness: $\begin{cases} .45\ \text{m(land} \\ 2.5 \times 10^{-4}\ \text{m(oceans)} \end{cases}$	See text				

Table 1 (continued)
COMPUTATION OF SURFACE FLUXES IN VARIOUS GCMS

Model	Fluxes Calculated from Eq. (11) or Eq. (12)	No. of Levels Coordinate System	Location of Level s Lowest GCM Level/Layer	Values for $C_D/C_H/C_E$	Determination of $\vec{V}_s$	Determination of T_s q_s	Determination of T_b	Determination of q_b	Remarks
6. UK I (1972) (Similarity theory) Corby et al. (1972)	Case II Eq. (12)	5/σ	No explicit level s in the GCM Lowest GCM level at 1 km (also the top of the b.l.)	$C_D = C_H = C_E$ α Land: $\begin{bmatrix} 0.3 \times 10^{-3} \text{ stable} & \pi/6 \\ 4.0 \times 10^{-3} \text{ unstable} & \pi/18 \end{bmatrix}$ Ocean: $\begin{bmatrix} 0.2 \times 10^{-3} \text{ stable} & \pi/9 \\ 2.0 \times 10^{-3} \text{ unstable} & 0 \end{bmatrix}$ α is the angle between the surface stress and the wind at level s. Z_o: roughness parameter Z_o = 10 cm for land; 1 cm for sea	Computed prognostically at 1 km. $\Delta\theta_e = (\theta_e)_s - (\theta_e)_b$ $\Delta q = (q)_s - q_b$	Uses equivalent potential temperature (see text) $Q_o = \rho C_p C_D \left[\lvert\vec{V}_s\rvert + A\left(\frac{\Delta\theta_e}{\theta_e}\right)^{1/2} \right] \cdot \Delta\theta_e$ where θ_e = equivalent potential temperature. $E_o = \rho C_D L \left[\lvert\vec{V}_s\rvert + A\left(\frac{\Delta\theta_e}{\theta_e}\right)^{1/2} \right] \cdot \Delta q$	Prognostic equation	RH = $\begin{pmatrix} 75\% \text{ over ocean} \\ 40\% \text{ over land} \end{pmatrix}$	Formulation for Q_o is modified in order to establish a near moist adiabatic lapse in lowest km over typical oceans (as is observed) and at the same time allow for a steeper lapse rate to exist over treated land areas (see text).
7. UK II (1974) (Similarity theory) (Only ocean-covered planet considered) Pearce (1974)	Case II Eq. (12)	5/σ	No explicit level s in the GCM Lowest level at 850 meters	$C_D = \left(\frac{U_*^2}{\lvert\vec{V}_s\rvert^2}\right)$, C_H, C_E, α are obtained from nomograms (Clark, 1970) which are plots of U_*, angle α, C_H, C_E as functions S and R_o.	Given by prognostic equation at 850 m. (This is required to obtain Q_o and E_o). (i) Surface stress given by $\tau_o = \rho U_*^2$ (ii) τ_o, Q_o and E_o are zero when $\lvert\vec{V}_s\rvert < 5 \frac{cm}{sec}$	T_s, q_s are referred to the lowest GCM level, i.e., at 850 m to obtain $\begin{cases} \Delta T = T_s - T_b \\ \Delta q = (q)_s - q_b \end{cases}$	Prescribed as function of latitude	Saturated q at T_b	S = stability parameter $= (g \frac{\Delta T}{T} + \varepsilon g \Delta q)/\lvert f\rvert \lvert\vec{V}_s\rvert$ $\Delta T = T_o - T_h$ R_o = surface Rossby No. = $\frac{\lvert\vec{V}_s\rvert}{\lvert f\rvert Z_o}$ θ = potential temperature is used instead of Temperature f corresponds to 21° latitude between 21° and equator
8. USSR (1974) Fux-Rabinovich (1974)	Case II Eq. (12)	5/ξ	No explicit level s in the GCM Upper limit of the b.l. = 1.5 km fixed	C_D, C_H, C_E and α determined from nomograms (same as in UK II above (7)).	Details not available				ξ, coordinate in vertical: takes into account the global orography $S = \frac{g}{T} \frac{(T_o - T_h - h\gamma_h}{f\lvert\vec{V}_s\rvert}$ h: b.l. top (first GCM level) γ = lapse rate at h = 6°C/km

discussed below. While most GCMs use Eq. (12) to determine the surface fluxes, UK-I (Corby et al., 1972) has modified it as follows:

1. Since UK-I GCM does not predict surface wind, the surface stress is prescribed to act in a different direction from that of the wind at the lowest GCM level (i.e., lowest free atmosphere wind). As a consequence, if α is the angle by which surface wind is backed from lowest free wind, then the λ and ϕ components of surface stress are given by (instead of $\vec{\tau}_o$ of Eq. (12))

$$\begin{aligned} \tau_{\lambda o} &= \rho_s C_D |\vec{V}_s| (U \cos\alpha - V \sin\alpha) \quad , \\ \tau_{\phi o} &= \rho_s C_D |\vec{V}_s| (V \cos\alpha + U \sin\alpha) \quad . \end{aligned} \tag{13}$$

Also, C_D and α are functions of stability and surface roughness with the values given in Table 1.

2. Q_o (of Eq. (12)) is replaced by

$$Q_o = -\rho C_p C_H \left[|\vec{V}_s| + A\left(\frac{\Delta\theta_e + \varepsilon_1}{\bar{\theta}_e}\right)^{\frac{1}{2}} \right] (\Delta\theta + \varepsilon_2) \quad , \tag{14}$$

where $A = 0$ for stable conditions. This result is based on discussion given by Sutton (1953) and others and is called a "pragmatic" approach. Here θ_e is the equivalent potential temperature given by

$$\theta_e = \left(T + \frac{L_q}{C_p}\right)\sigma^{-k} \quad ,$$

and $\Delta\theta_e$ is the difference between the values of θ_e at the surface and the lowest GCM level; $\bar{\theta}_e$ is the average of the θ_e at these two levels.

The main purpose of using Eq. (14) is to be able to establish (through Q_o) a near moist adiabatic lapse rate in the lowest km as is usually observed over tropical oceans, and at the same time to allow a steeper lapse to exist over land areas. The UK-I GCM uses

$$A = 10^3 \text{ m/s} \quad \text{for unstable case, i.e., if } (\Delta\theta_e + \varepsilon_1) > 0,$$

$$= 0 \quad \text{for other cases,}$$

$$\left.\begin{aligned} \varepsilon_1 &= 2°K \\ \varepsilon_2 &= 2°K \end{aligned}\right\} \quad \text{based on experience.}$$

A limit is imposed on the rate of heating under unstable conditions such as when extremely cold air flows over warm sea such that

$$A \left[\frac{\Delta\theta_e + \varepsilon_1}{\overline{\theta}_e}\right]^{\frac{1}{2}} \leq 200 \text{ m/sec} \quad .$$

The surface humidity mixing ratio is calculated on the assumption that relative humidity at the ocean surface is 75 percent and that at the land surface is 50 percent. This enables Eq. (14) to provide a realistic discrimination between the lapse rates over a land and sea.

Analogous to Eq. (14), the latent heat flux at the surface is given by

$$E_o = -\rho C_E \left[|\vec{V}_s| + A \left\{ \frac{\Delta\theta_e + \varepsilon_1}{\overline{\theta}_e} \right\}^{\frac{1}{2}} \right] \Delta q \quad , \tag{15}$$

when C_E, A, and ε_1 have the same values as those in Eq. (14).

This model gives unrealistic results since strong lapse rates may be established spuriously without instability. As a consequence, one may have dry air at 1 km in equilibrium with the underlying surface, which is clearly not possible over desert regions. This may be obviated by distinguishing between different types of land surfaces.

Determination of Surface Fluxes in UCLA-GCM (3-Level Version)

The parameterization of surface eddy fluxes in the UCLA GCM is based on the method suggested by Deardorff (1972). This method determines surface fluxes of momentum, heat, and moisture as functions of stability (bulk Richardson number), surface roughness (Z_o), and the *total b.l. depth* (h) obtained through a rate equation. Deardorff derived his formulations by combining the explicit formulations for the surface layer and the b.l. deficit formulations for the layer above the surface layer. This technique enabled him to eliminate the use of the values of the parameters at the so-called anemometer level s, and express the formulations in terms of the bulk properties of the entire b.l.

In actual practice, the UCLA GCM computes the surface fluxes in the same way as those GCMs which use Eq. (12) except for the following differences:

o The surface transfer coefficients C_D/C_H are obtained from Fig. 6 which shows the functional dependence of these coefficients on the bulk Richardson number (Ri_B), and the ratio $(h - Z_s)/Z_o$ of the effective b.l. depth to the roughness parameter. C_E is assumed to be equal to C_H. The figures are obtained from the following formulas for C_D/C_H:

$$C_D\left(Ri_B, \frac{h - Z_s}{Z_o}\right) = \frac{U_*}{U_M} ,$$

$$C_H\left(Ri_B, \frac{h - Z_s}{Z_o}\right) = \frac{Q_o}{U_* \Delta\theta} , \qquad (16)$$

$$C_E = C_H .$$

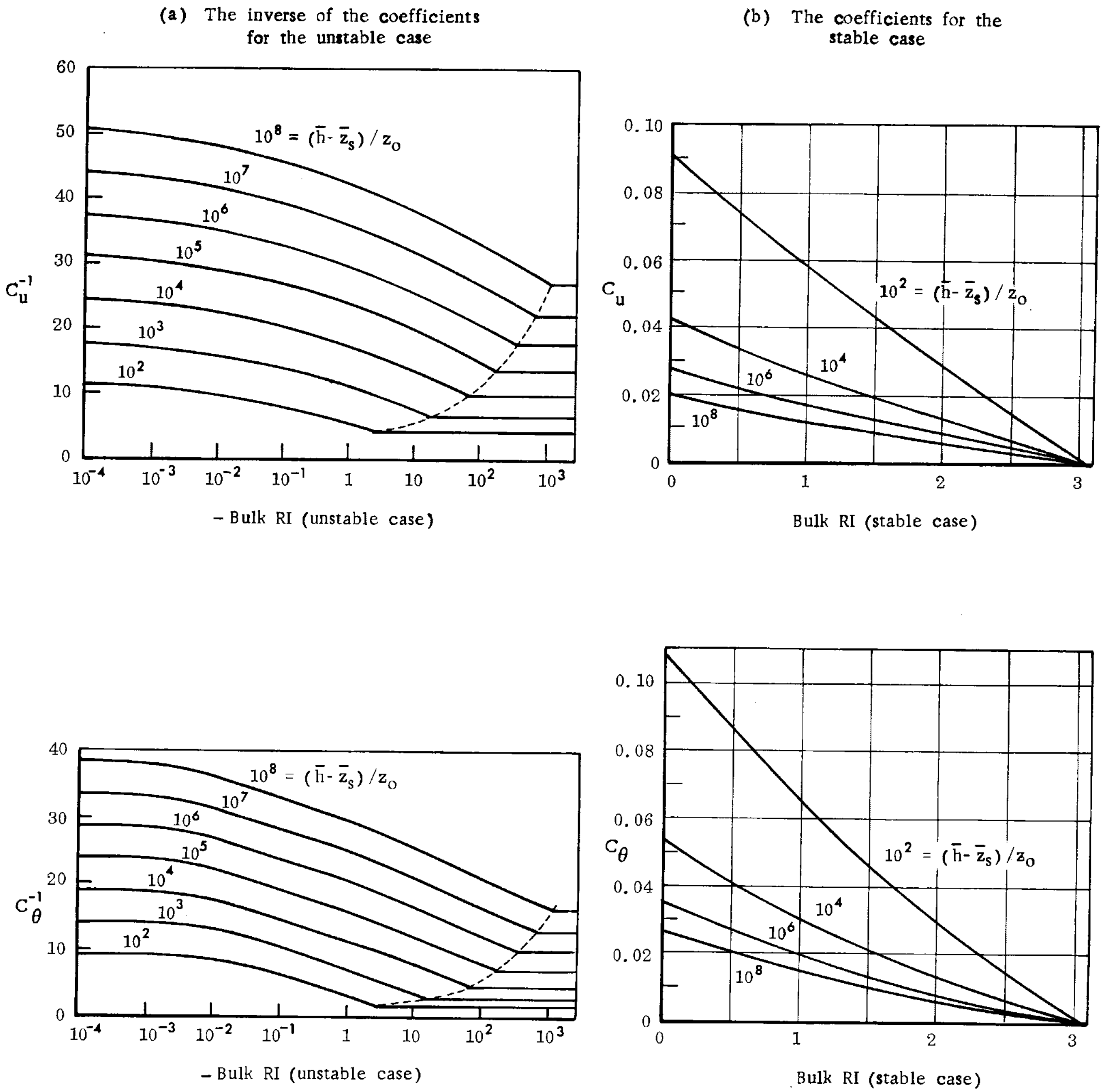

Fig. 6--The surface transfer coefficients (C_D and C_H) as function of the bulk Richardson number (after Deardorff, 1972).

Here

$$Ri_B = -\frac{g}{C_p \theta_M} \frac{(h - Z_s)(\theta_M - \theta_s)}{U_M^2} \quad .$$

o The model uses static energy instead of temperature (θ).

o The difference $\Delta\theta$ (Eq. (12)) represents in this case the effective difference between the vertically averaged static energy of the b.l. and its value at the ground. Similarly, Δq is the difference between the value of q at the ground and the mean value within the b.l.

o This method also considers the occurrence of saturated b.l. through its entire depth as well as the presence of fog.

o h, the top of the b.l., is not fixed but instead is predicted by a prognostic equation.

o Z_o = 0.45 m at all land points, and
= 2.5×10^{-4} m at all ocean points.

VERTICAL PROFILES OF THE TURBULENT FLUXES

As indicated earlier, the actual method of parameterizing b.l. in GCM depends on the number of levels (in the vertical) incorporated within the b.l. In global models which cannot resolve the b.l. explicitly and in which the lowest GCM level is well *above* the b.l. top, only the surface fluxes are of any significance. In some global models which have several levels within the b.l., the vertical structure of the b.l. in terms of the vertical profiles of turbulent fluxes can be determined. The lowest GCM level is usually placed at the top of the surface layer (presumed to be a constant flux layer) and the other levels are placed in the so-called Ekman layer above. We have described above the techniques used by various GCMs to evaluate surface fluxes. In this section we concentrate on the determination of turbulent fluxes within the Ekman layer.

The formulations generally employed to compute the turbulent fluxes in the layer above the surface layer are

$$\text{stress} = \vec{\tau} = \rho K_M \frac{\partial \vec{V}}{\partial Z} \quad ,$$

$$\text{heat flux} = Q = -\rho C_p K_H \left(\frac{\partial \theta}{\partial Z} - \gamma_{CG} \right) \quad , \tag{17}$$

$$\text{moisture} = E = -\rho K_E \frac{\partial q}{\partial Z} \quad ,$$

where γ_{CG} is a counter-gradient flux (Deardorff, 1966) and K_M, K_H, and K_E are the eddy diffusion coefficients for momentum, heat, and moisture, respectively.

The actual use of Eq. (17) in GCMs depends on the representation of the b.l. in the numerical model, and these can be considered for the following three cases:

Case I. Implicit b.l. (NCAR, GISS, RAND GCMs)

NCAR, GISS, and RAND GCMs, which do not incorporate the b.l. explicitly, use Eq. (17) to compute the values of $\vec{V}_s$, T_s, and q_s, which are required to evaluate surface fluxes (Eq. 11). The method of computing $\vec{V}_s$, T_s, and q_s is based on the assumption that there is no flux divergence in the b.l. and thus the surface fluxes can be equated to the fluxes in the layer from level s to the first internal GCM level in the model. For example, equating surface stress to the stress in the Ekman layer we get

$$\rho_s C_D |\vec{V}_s| V_s = \rho_s K_M \frac{\vec{V}_1 - \vec{V}_s}{Z_1 - Z_s} \quad . \tag{18}$$

Here $\vec{V}_1$, the large-scale wind velocity at the first GCM level, is known from the GCM, and if C_D and K_M are specified, Eq. (18) can be solved for the unknown $\vec{V}_s$. Similar sets of expressions are used for obtaining T_s and q_s, which are then used to determine surface fluxes for heat and moisture from Eq. (11).

Specification of $K_M/K_H/K_E$

It is assumed that K_M, K_H, and K_E are equal. The formulation for K_H is given by

$$K_H = \begin{cases} d\left[A_1 + A_2\left\{1 - \exp\left[A_3\left(\frac{\partial\theta}{\partial Z} - \gamma_{CG}\right)\right]\right\}\right] & \text{for neutral/unstable case, i.e., } \frac{\partial\theta}{\partial Z} \leq \gamma_{CG} \ , \\ d\left[\frac{A_1}{1 + A_4 Ri} + A_5\right] & \text{for stable case, i.e., } \frac{\partial\theta}{\partial Z} > \gamma_{CG} \ , \end{cases} \tag{19}$$

where

$$Ri = \text{Richardson number} = \frac{\frac{g}{T}\left(\frac{\partial\theta}{\partial Z} - \gamma_{CG}\right)}{\left[\left(\frac{\partial U}{\partial Z}\right)^2 + \left(\frac{\partial V}{\partial Z}\right)^2 + A_6\right]} \quad ;$$

A_1, A_2, A_3, A_4, A_5, and A_6 are empirical constants; and γ_{CG} is the counter-gradient heat flux. d is a factor introduced to reduce the value of K_M over the mountains where the thickness of the lowest GCM layer is reduced considerably.

The values of the above empirical parameters used in the NCAR GCM are given by

$$A_1 = 10^5 \text{ cm}^2/\text{sec},$$

$$A_2 = 10^6 \text{ cm}^2/\text{sec},$$

$$A_3 = 1.2 \times 10^5 \text{ cm}/°\text{C},$$

$$A_4 = 40,$$

$$A_5 = 2 \times 10^4 \text{ cm}^2/\text{sec},$$

$$A_6 = 10^{-12},$$

$$\gamma_{CG} = \begin{cases} 5 \times 10^{-5}\ {}^\circ\mathrm{C/cm} & \text{at} \quad Z = s \;, \\ 1 \times 10^{-5}\ {}^\circ\mathrm{C/cm} & \text{otherwise} \;, \end{cases}$$

$$d = \begin{cases} 1 - \dfrac{H}{\Delta Z} & \text{for} \quad H < \Delta Z \;, \\ 2 - \dfrac{H}{\Delta Z} & \text{for} \quad \Delta Z \leq H \leq 2\Delta Z \;, \end{cases}$$

where H is the height of orography. The GISS GCM uses the same values for A_E, A_1, A_2, A_3, A_4, A_5, and A_6 as in the NCAR GCM, but it uses values of 0 and 1 for γ_{CG} and d respectively. The RAND GCM uses $K_M = K_E = K_H = 1|\vec{V}_s|$.

Case II. Explicit b.l. (GFDL/UK-II) GCM

Both these GCMs place the lowest GCM level at the top of the surface layer. GFDL has two more GCM levels up to the top of the b.l., and the turbulent fluxes in the layer above the surface layer are computed from Eq. (17) through finite difference formulations. The eddy diffusion coefficients are assumed to be the same for momentum, heat, and moisture fluxes and are given by

$$K_M = \begin{cases} \ell^2 \left|\dfrac{\partial \vec{V}}{\partial Z}\right| (1 - \alpha_c S) & \text{for unstable case} \quad \left(\dfrac{\partial \theta}{\partial Z} < 0\right) \;, \\ \ell^2 \left|\dfrac{\partial \vec{V}}{\partial Z}\right| (1 + \alpha_c S)^{-1} & \text{for neutral/stable case} \quad \left(\dfrac{\partial \theta}{\partial Z} \geq 0\right) \;. \end{cases} \tag{20}$$

Here ℓ is the mixing length, S is a stability parameter, and α_c is an empirical constant ($\alpha_c = 18$). The b.l. is explicitly included in the model to the extent that a mixing length given by ℓ applies. Table 2 shows the values of ℓ and S used by the GFDL and the UK-II GCMs; the formulations have also been used by Estoque & Bhumralkar (1969).

Case III. Explicit b.l. Treatment: UCLA GCM (3-Level)

The parameterization technique used in the UCLA GCM is essentially based on the one proposed by Deardorff (1972) and is designed for GCMs

Table 2

VALUES FOR ℓ AND S IN GCMS

Model	Total Levels in Vertical	Levels in b.l.	ℓ	S	Remarks
UK-II	5	1 (850 m)	30 meters	0(neutral)	$\gamma_{CG} = 0$ (Eq. 17)
GFDL	(i) 9(1969)	3	(i) 30 meters	0(neutral)	
	(ii)18(1974)	4	(ii) $= 0.4Z$ for $Z \le s$; $= 0.4 \dfrac{s(B - Z)}{B - Z_s}$ for $s < Z < B$; $= 0$ for $B < Z$ ($Z_s = 75$ m, $B = 2.5$ km)	0(neutral)	Does not include vertical heat flux Q. Also $\gamma_{CG} = 0$ in all cases. B is the level at which $\ell = 0$
	(iii) 9(1971)[a]	3	(iii) $\dfrac{0.4\ Z}{1 + \dfrac{1480\ fZ}{\lvert\vec{V}\rvert}}$	$S = \dfrac{\sqrt{g\ell}\ \dfrac{\partial\theta}{\partial Z}}{\theta \left\lvert \dfrac{\partial \vec{V}}{\partial Z} \right\rvert}$	

[a]Not used in long-term integrations, but used by Delsol et al. (1971) to test b.l. parameterizations.

that do not resolve the b.l. However, the method involves solving *prognostic* equations for

- the height of the b.l. (h), and
- the "strength" of the discontinuity in momentum, heat, and moisture at the b.l. top (ΔU, ΔV, $\Delta\theta$, Δq) to maintain consistency between the b.l. parameterization and the GCM.

The first item has been discussed in considerable detail by Bhumralkar (1974) and incorporates the *entrainment* hypothesis which in turn depends, among other things, on the flux of the quantity at the b.l. top as well as its discontinuity there. Thus, neglecting radiative effects,

$$(E_N) = -\frac{Q_h}{\Delta\theta} \quad , \tag{21}$$

where Q_h is the heat flux at the top of the b.l., and $\Delta\theta$ is the inversion "strength" at h.

The prognostic equations for computing the "strength" of the discontinuity at the top of the b.l. require evaluation of the eddy flux divergence within the b.l. (in addition to advection and interaction with cumulus scale processes). Thus

$$\frac{\partial}{\partial t}(\Delta\theta) \approx -\left(\frac{\partial Q}{\partial Z}\right)_h \quad . \tag{22}$$

It can be seen from Eqs. (21) and (22) that it is necessary to know the vertical distribution of eddy fluxes in the b.l. This is also necessary for the following two reasons:

(a) Since the b.l. depth is variable (and determined prognostically), the b.l. is permitted to completely fill one or more GCM layers and thus the eddy fluxes *between* such layers must be determined.

(b) The parameterization method considers the formation of new boundary layers; for this it is required to determine the rate of conversion of the buoyant potential energy into the turbulent kinetic energy ($B = g/C_p \int_s^h (Q/T)dZ$), which in turn depends on the knowledge of vertical distribution of the flux of heat (say) within the b.l.

The model estimates the flux of heat within the b.l. from the equation

$$Q = Q_h + [Q_s - Q_h]\left[\frac{Z - h}{h - s}\right] + K\, E_N\, \mathrm{Sin}\left[\pi\left(\frac{Z - h}{h - s}\right)\right] \frac{\partial\theta}{\partial Z} \quad , \qquad (23)$$

where K is a constant, and E_N is the rate of entrainment given by Eq. (21), which equals 0 at the surface s and the b.l. top h. Equation (23) can be used to compute eddy flux convergence terms at h, which are then used to determine the "strengths/discontinuities" at the b.l. top (reference Eq. 22). The last term in Eq. (23)--the mixing term--is added to obviate the inconsistencies which arise when a simple linear profile of h is considered.

An equation similar to (23) can be written for moisture flux within the b.l. However, at this time (August/September 1974), research is continuing and no conclusive discussion can be given.*

HORIZONTAL DIFFUSION TERMS IN GCMs

There is, in general, some disagreement regarding the necessity to include explicit horizontal eddy diffusion terms in GCMs. Until recently there was a general denial of its physical significance; many felt that a parameterized horizontal diffusion was purely mathematical gimmickry to suppress computational instability (Smagorinsky, 1970). However, apart from the development of differencing methods which apparently obviate the need for these terms, the experience of the GFDL GCM with nonlinear viscosity designed to prevent the -5/3 power law

*For details see Randall and Arakawa (1974).

near the limit of resolution has been quite satisfactory. Whereas the GFDL/NCAR/UK-I GCMs use explicit non-linear lateral diffusion terms (Smagorinsky, 1963) in their governing equations, the RAND, GISS, and UCLA GCMs do not. However, it may be mentioned that a horizontal averaging scheme has been introduced in the RAND GCM (and also the UCLA GCM) in the higher latitude regions. The UK-II GCM uses a constant diffusion coefficient to evaluate lateral diffusion.

V. EXPERIMENTS TO TEST DIFFERENT BOUNDARY-LAYER PARAMETERIZATIONS IN AN ATMOSPHERIC GENERAL CIRCULATION MODEL

NEED FOR EXPERIMENTS

As discussed in the preceding section, several approaches to the b.l. parameterization are aimed at incorporating the b.l. processes in GCMs. However, in the absence of any systematic sensitivity tests it is difficult to determine the specific advantages or disadvantages of each of the approaches. There has been no estimate of errors of parameterization in relation to the basic assumptions of the theory on which a technique is based. The different GCMs, which have parameterized the b.l. processes either explicitly or implicitly, do not specifically discuss the impact of the particular parameterization technique on the global results. Thus, there has been no attempt to test the degree of acceptability of the simplifications/assumptions in evaluation of turbulent fluxes of momentum, heat, and moisture in GCMs. In fact, there have been no studies even to determine whether the incorporation of b.l. physics in a GCM is justified or not. As far as is known to the author at this time only Delsol et al. (1971) have studied (in a numerical GCM) the effects of parameterization of various processes related to the b.l. physics. They evaluated the comparative effects by repeating the integrations and by gradually increasing the complexity of the model in terms of the use of

- different roughness over land and sea,
- Monin-Obukhov treatment of the surface layer turbulence,
- stability dependent eddy diffusion coefficient in Ekman layer,
- diurnal variation of insolation, and
- conduction of heat into the soil.

Their results, based on 14 days of real-time prediction, showed that the sophistication of the b.l. physics *does not* produce a particularly large effect on large-scale prediction *until about 7 days*. They speculate (without evidence) that it may become large after 10 days.

Their results have prompted Pasquill (1972) to suggest that (if these are true) it may be *unnecessary* to incorporate much of the detail of the b.l. effects in GCMs.

In the author's opinion, this conclusion is not quite justified because the experiments of Delsol et al. (1971) are *not conclusive* due to various factors. For example:

(1) They assume a mixing length of 30 m, and yet their first GCM level is at 75 m, the next at 640 m, and so on. This does not enable the model to resolve the vertical eddy size in both the constant flux layer and the Ekman layer.

(2) The formulations used in the Ekman layer are not quite valid in low-latitude/equatorial regions.

(3) No comparison with observed data was attempted.

In view of the above, and because of the increasing importance of the use of GCMs for climatic studies, it is evident that a comparison of various b.l. parameterization schemes is desirable in order to select the "best" approach. This view has also been expressed at various international meetings held to discuss the problem of b.l. parameterization (GARP Publications Series, No. 8, October 1972, and Report No. 5, July 1974, of the GARP Programme on Numerical Experimentation).

PARAMETERS INVOLVED IN COMPARISON TESTS

Before any experiments are performed to compare different b.l. parameterization schemes in a GCM, it is necessary to specify parameters or groups of parameters in relation to which the comparisons should be performed. It appears that it may be useful to integrate a GCM by including different b.l. parameterization schemes (one at a time) and then compare these specified parameters as functions of the parameterization schemes. For a consistent and meaningful comparison it is necessary that for each integration the general prognostic scheme and initial conditions should remain fixed. The comparison may be performed, over the globe, with respect to the following parameters under all stability conditions:

(1) Values of the turbulent surface fluxes of momentum, heat, and moisture.

(2) Vertical velocities at the top of the b.l.

(3) Height of the b.l.

(4) Global heat flux distribution in the vertical.

(5) Vertical profiles of the turbulent fluxes as well as of the meteorological variables within the b.l.

(6) Variability of surface parameters (roughness, soil moisture content, etc.) at subgrid scale. This involves use of commonly used aerodynamic roughness length vs the effective roughness (Fiedler and Panofsky, 1972).

It is envisaged that in order to determine the applicability of a b.l. parameterization scheme to a GCM, the results pertaining to the above quantities will be compared between the different models themselves, as well as with the observational data for some situations for which there are sufficient data available.

Another important factor which could influence the determination of the quality of a parameterization scheme (in a GCM) is the real time *duration of integration* of the GCM. Zilitinkevich (1972) has suggested that it may be reasonable to compare results of different integrations for 3 to 4 days. However, as stated earlier, the experience of Delsol et al. (1971) indicates that even 7 days of integration may not be sufficient to bring out significant differences. Thus it appears that the integrations may have to be performed for at least a two-week period or even longer. It may, however, be noted that due to lack of experience, either the specification of integration period or the list of "comparable" parameters (above) is not final and these may have to be modified.

PROPOSED EXPERIMENTS

We propose to perform a series of experiments to compare some b.l. parameterization schemes on the basis of discussions given in the preceding two subsections. We expect to use an existing version of the Rand 2-level GCM as a tool to perform these experiments. It is evident that, due to the coarse vertical resolution of the Rand GCM, we will not be able to determine the vertical structure of the b.l. in respect to the profiles of the fluxes as well as of the meteorological variables. Consequently, the emphasis will have to be on the determination and comparison of surface fluxes with observations.

The proposed experiments may involve comparison of parameterization schemes which are based on similarity theory as well as K-theory concepts. For a given configuration of the GCM prognostic scheme and initial conditions, we propose to integrate the model by incorporating the following succession of parameterization schemes:

(1) K-theory scheme

The formulations similar to the one used by the NCAR/GISS GCM may be used. This may be performed for different values of the eddy coefficient K and of the drag coefficient.

(2) Similarity theory scheme

(a) We can use the formulations in their original form which consider the scale height of b.l. to be given by U_*/f. The disadvantage of using f in low latitude/equatorial regions can be obviated by specifying a critical latitude ϕ_c and using f for this across the equatorial region between the critical latitudes on either side of the equator.

(b) Instead of U_*/f, the height, h, of the b.l. can be computed by a prognostic equation. The latter involves use of entrainment hypotheses and can be patterned after Deardorff (1972).

(c) Instead of U_*/f and a prognostic equation for h which may give rise to singularities under certain conditions, we can also test the use of "interpolation" prognostic equations, such as that proposed by Deardorff (1974), which has been designed to overcome the singularities.

(d) The above three procedures involve a given set of so-called "universal" functions. It may be found necessary to experiment with other sets of values for these functions.

(e) The effect of the use of the roughness parameter (z_o) can also be investigated. In this work it may be desirable to test the formulations suggested by Zilitinkevich (1969) and Clark (1970). The concept of effective roughness (Fiedler and Panofsky, 1972) can also be investigated.

(3) Variable boundary-layer thickness

In addition to the above we can perform an experiment using the parameterization scheme developed by UCLA for their 3-level GCM. In the Rand 2-level GCM, however, we can treat only that case which considers an entirely sub-grid scale boundary layer. This is because in this model it may be more realistic to assume that the b.l. depth is always confined to the lower layer (surface to 600 mb) and is thus not resolved by the GCM.

The results and conclusions of such comparisons, together with specific details in respect of the parameterization schemes used, will form the subject matter of a subsequent report.

VI. CONCLUDING REMARKS

It is recognized that we can hope to make a quantitative assessment of the general circulation and climatic changes only through the use of reasonably calibrated numerical models. With this end in view, there have been concerted efforts to improve the fidelity of GCMs by improving (among other physical processes) the parameterization of boundary-layer (b.l.) processes, by using schemes that are not unique and can be approached in a variety of ways.

At present there are several approaches to the b.l. parameterization in GCMs, and it is fair to expect that the number of these formulations will increase in time. However, so far there have been no studies aimed at the comparison of the various parameterization schemes in GCMs in order to select the best approach and to estimate errors of the parameterizations. It is evident that the limitations associated with the theories used in parameterizations will have a controlling influence on the results. For example, in case of similarity theory, it is not clear to what extent the assumptions of this theory as regards horizontal homogeneity of the underlying surface (in the steady state) must be satisfied to permit utilization of the conclusions of the theory. Also, the heterogeneity of the underlying surface is currently taken into account by using the roughness parameter (Z_0) as a function of the character of the underlying surface. It is yet to be determined, however, whether this is the only possible kind of heterogeneity on a sub-grid scale. The use of a rate equation to obtain the height of the b.l. has also to be investigated before the method is universally applied in practice.

What we have suggested in this review is a systematic approach to determine the most suitable b.l. parameterization for the Rand GCM.

Through comparison of the simulation responses of the GCM, the experiments can be designed to determine the relative role and the sensitivity of the various formulations.

REFERENCES

Arakawa, A., and Y. Mintz, 1974: *The UCLA Atmospheric General Circulation Model*, Dept. of Meteor., Univ. of Calif., Los Angeles, Calif.

Arakawa, A., and W. H. Schubert, 1974: Interaction of a Cumulus Cloud Ensemble with the Large-Scale Environment, Part I, *J. Atmos. Sci.*, 31, 674-701.

Bhumralkar, C. M., 1974: *A Survey of the Research on the Treatment of the Planetary Boundary Layer in Atmospheric Circulation Models*, The Rand Corporation, R-1653-ARPA (in preparation).

Budyko, M. I., 1969: The Effects of Solar Radiation Variations on the Climate of the Earth, *Tellus*, 21, 611-619.

Carson, D. J., 1973: The Development of a Dry Inversion-Capped, Convectively Unstable Boundary Layer, *Quart. J. Roy. Meteor. Soc.*, 99, 450-467.

Charney, J. G., and A. Eliassen, 1949: A Numerical Method for Predicting Perturbations of the Middle-Latitude Westerlies, *Tellus*, 1, 38-54.

Clark, R. H., 1970: Recommended Methods for the Treatment of the Boundary Layer in Numerical Models, *Aust. Meteor. Mag.*, 18, 51-73.

Corby, G. A., A. Gilchrist, and R. L. Newson, 1972: A General Circulation Model of the Atmosphere Suitable for Long Period Integrations, *Quart. J. Roy. Meteor. Soc.*, 98, 809-833.

Deardorff, J. W., 1966: The Counter-Gradient Heat Flux in the Lower Atmosphere and in the Laboratory, *J. Atmos. Sci.*, 23, 503-506.

Deardorff, J. W., 1972: Parameterization of the Planetary Boundary Layer for Use in General Circulation Models, *Mon. Wea. Rev.*, 100, 93-106.

Deardorff, J. W., 1974: Three-Dimensional Numerical Study of the Height and Mean Structure of a Heated Planetary Boundary Layer, *Boundary-Layer Meteor.*, 7, 81-106.

Delsol, F., K. Miyakoda, and R. H. Clark, 1971: Parameterized Processes in the Surface Boundary Layer of an Atmospheric Circulation Model, *Quart. J. Roy. Meteor. Soc.*, 97, 181-208.

Elliot, W. P., 1958: The Growth of the Atmospheric Internal Boundary Layer, *Trans. of Am. Geophy. Union*, 39 (6), 1048-1054.

Estoque, M. A., and C. M. Bhumralkar, 1969: The Flow Over a Localized Heat Source, *Mon. Wea. Rev.*, 97, 850-859.

Fiedler, F., and H. A. Panofsky, 1972: The Geostrophic Drag Coefficient and the Effective Roughness Length, *Quart. J. Roy. Meteor. Soc.*, 98, 213-220.

Fleagle, R. G., F. I. Bagley, and Y. Hsueh, 1967: Calculations of turbulent fluxes by integral methods, *J. Atmos. Sci.*, 24, 356-373.

Fux-Rabinovich, M. S., 1974: Global Five-Level General Circulation Model, Hydrometeorological Research Center, Moscow, USSR, Modeling for the First GARP Global Experiment, *GARP Publications Series No. 14*, WMO, Geneva, Switzerland, 101-112.

Gates, W. L., E. S. Batten, A. B. Kahle, and A. B. Nelson, 1971: *A Documentation of the Mintz-Arakawa Two-Level Atmospheric General Circulation Model*, The Rand Corporation, R-877-ARPA, 408 pp.

Gray, W. M., 1972: A *Diagnostic Study of the Planetary Boundary Layer Over the Oceans*, Dept. of Atmos. Sci., Paper No. 179, Colorado State Univ., Fort Collins, Colorado, 95 pp.

Holloway, J. L., Jr., and S. Manabe, 1971: Simulation of Climate by a Global General Circulation Model. I. Hydrologic Cycle and Heat Balance, *Mon. Wea. Rev.*, 99, 335-370.

Holopainen, E. O., 1963: On the Dissipation of Kinetic Energy in the Atmosphere, *Tellus*, 15, 26-32.

Kasahara, A., and W. M. Washington, 1971: General Circulation Experiments with a Six-Layer NCAR Model, Including Orography, Cloudiness, and Surface Temperature Calculations, *J. Atmos. Sci.*, 28, 657-701.

Koleshnikova, V. N., and A. S. Monin, 1965: On the Spectra of Meteorological Fields Oscillations, *Izv. Atmos. Ocean. Phys.*, 7, 653-669, (English Ed.).

Kung, E. C., 1963: Climatology of Aerodynamic Roughness Parameter and Energy Dissipation in the Planetary Boundary Layer Over the Northern Hemisphere, *Studies of the Effects of Variations in the Boundary Conditions on the Atmospheric Boundary Layer*, Annual Report, Dept. of Meteor., Univ. of Wisconsin, USAERDA, Ft. Huachuca, Ariz., 37-96.

Lettau, H., 1959: Wind Profile, Surface Stress and Geostrophic Drag Coefficients in the Atmospheric Surface Layer, *Advances in Geophysics*, 6, Academic Press, N.Y., 241-256.

Lettau, H., 1962: Theoretical Wind Spirals in the Boundary Layer of a Barotropic Atmosphere, *Beitr. Phys. d. Atmos.*, 35, 195-212.

Mahrt, L. J., 1974: Time-Dependent, Integrated Planetary Boundary Layer Flow, *J. Atmos. Sci.*, 31, 457-464.

Panofsky, H. A., and A. A. Townsend, 1964: Change in Terrain Roughness of the Wind Profile, *Quart. J. Roy. Meteor. Soc.*, 90, 147-155.

Pasquill, F., 1972: Some Aspects of Boundary Layer Description, *Quart. J. Roy. Meteor. Soc.*, 98, 469-494.

Pearce, R. P., 1974: Five-Level General Circulation Model, Universities Modeling Group, U.K., Modeling for the First GARP Global Experiment, *GARP Publications Series No. 14*, WMO, Geneva, Switzerland, 127-140.

Randall, D. A., and A. Arakawa, 1974: *A Parameterization of the Planetary Boundary Layer for Numerical Models of the Atmosphere*, Notes distributed at the Workshop of the UCLA Atmospheric General Circulation Model, March 25-April 4, 1974, Dept. of Meteor., Univ. of Calif., Los Angeles, Calif.

Sawyer, J. S., 1959: The Introduction of the Effects of Topography into Methods of Numerical Forecasting, *Quart. J. Roy. Meteor. Soc.*, 85, 31-43.

Sellers, W. D., 1973: A New Global Climatic Model, *J. Atmos. Sci.*, 12, 241-254.

Sheppard, P. A., H. Charnock, and J. R. D. Francis, 1952: Observations of Westerlies Over the Sea, *Quart. J. Roy. Meteor. Soc.*, 78, 563-582.

Smagorinsky, J., 1963: General Circulation Experiments with the Primitive Equations, I. The Basic Experiment, *Mon. Wea. Rev.*, 91, 99-164.

Smagorinsky, J., 1970: Numerical Simulation of the Global Atmosphere, in *The Global Circulation of the Atmosphere*, G. A. Cosby, ed., Royal Meteor. Soc., London, 24-41.

Somerville, R. C. J., P. H. Stone, M. Halem, J. E. Hansen, J. S. Hogan, L. M. Druyan, G. Russell, A. A. Lacis, W. J. Quirk, and J. Tenenbaum, 1974: The GISS Model of the Global Atmosphere, *J. Atmos. Sci.*, 31, 84-117.

Sutton, O. G., 1953: *Micrometeorology*, McGraw Hill, New York, 333 pp.

Zilitinkevich, S. S., 1969: On the Computation of the Basic Parameters of the Interaction Between the Atmosphere and the Ocean, *Tellus*, 21, 17-24.

Zilitinkevich, S. S., 1972: Turbulent Boundary Layer: Parameterization of Sub-Grid Scale Processes, *GARP Publication Series No. 8*, WMO, Geneva, Switzerland, 69-80.