COMMENTS ON THE SENSITIVITY OF NUMERICAL SIMULATIONS TO DIFFERENT PARAMETERIZATIONS OF THE BOUNDARY-LAYER PROPERTIES AND PROCESSES.

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1. INTRODUCTION

The atmospheric boundary-layer is one of the major energy sources and momentum sinks in the atmosphere and plays a vital rôle in the exchange of heat, moisture and momentum between the earth's surface and the 'free' atmosphere. Until fairly recently designers and users of global atmospheric circulation models have in general not been concerned with the details of boundary-layer structure and processes in their own right but for their influence on atmospheric structure and circulation on a larger, perhaps global, scale. This attitude is changing however as global models continue to be developed for climate and forecasting studies and more direct interest is now being shown in the characteristics of the simulated near-surface atmosphere.

There are several aspects to the general problem of parameterizing boundary-layer processes which have to be considered by the modeller who must contend with the atmospheric boundary-layer as a 'lower boundary condition'. These are identified by the following somewhat loosely posed questions:

(1) What are the main features of the atmospheric boundary-layer that need to be modelled? (IDENTIFICATION PROBLEM)

(2) What boundary-layer theories and ideas are available and suitable for the specified modelling objectives? (SELECTION PROBLEM)

(3) What practical difficulties are met when constructing and using a parameterization based on a particular theory? (APPLICATION PROBLEM)

(4) What impact does the boundary-layer parameterization have on the model simulations and how sensitive are these simulations to changes in the formulation? (ASSESSMENT PROBLEM)

Although none of these questions can be answered satisfactorily in isolation from the others, nevertheless the emphasis in this paper centres on the fourth and least well documented aspect, that of identifying and assessing the response of the simulated large-scale flow to different specifications of atmospheric boundary-layer and surface properties and processes.

The assessment of the impact of a new physical parameterization within the framework of a general circulation model is acknowledged to be a very difficult problem in general. To judge the significance of a response to a subtle change in a parameterization would, in general, require a carefully conducted and often unrewarding series of lengthy and costly integrations. Even if a response is judged to be statistically significant it may not be possible to judge if it is a change for the better. It is therefore understandable that there is a dearth of commendable literature on this topic. The most comprehensive series of tests to examine the response of the large-scale flow to the parameterization of the boundary-layer processes within the framework of a global model appears to have been carried out by Miyakoda and his colleagues at GFDL (Princeton). In order of increasing complexity of the studies I recommend the papers of Miyakoda (1975), Delsol et al (1971) and Miyakoda and Sirutis (1976).

It is not the intention in this extract to review the subject or to present in a systematic manner results from a comprehensive and rigourously conducted series of sensitivity experiments. The results presented at the Workshop attempted to convey some personal reflections on a variety of experiments made with the Meteorological Office 11-layer general circulation model. The sensitivity of simulations to changes in the surface roughness length over the sea is the specific example chosen to be discussed here in order to illustrate some of the more general aspects of the ASSESSMENT PROBLEM.

2. THE METEOROLOGICAL OFFICE 11-LAYER ATMOSPHERIC GENERAL CIRCULATION MODEL

The version of the 11-layer model used in the experiments discussed below follows to a large extent that described by Saker (1975) and has the following general properties.

Type Finite difference. Global.

Horizontal Grid

Irregular and non-staggered.

Quasi uniform, except close to the poles.

Grid lines are 2^o latitude apart (N=45 i.e. 45 rows between pole and equator). Grid length approx. 220 km.

180 points at equator (1^oN and S) reducing to minimum of 16 near poles.

10,352 grid points on globe.

Vertical Resolution

11 fixed layers of variable thickness.

Vertical coordinate, $\sigma = p/p_s$ where p is atmospheric pressure and p_s is its value at the surface.

Numerical Scheme

Spatial Second-order, energy-conserving, finite difference scheme.

Temporal Explicit time integration, with $7\frac{1}{2}$ minute time-step (except close to the poles).

The scheme is centred for advection and forward for eddy viscosity and physical parameterizations.

A weak time-filter is used at each time-step to prevent time-splitting.

Physical Schemes

One of the prime research uses of the 11-layer model is to test different methods of parameterizing the main sub grid-scale physical processes. Radiative processes and related cloud, convection, the turbulent exchange of momentum, sensible heat and water vapour at the surface and through the boundary-layer, and precipitation are all represented in the standard version of the model. Because of its particular importance in the discussion that follows, the standard boundary-layer structure used in the 11-layer model is described briefly in the next section. Other specific aspects of the physical parameterizations are mentioned only where it is appropriate to do so.

Surface Properties

In the standard version of the model, soil moisture and snow depth are allowed to vary throughout the integrations as determined by budget equations for water, snow and heat. Land- and ice-surface temperatures are also normally allowed to vary according to a surface energy balance equation. Sea-surface temperatures are fixed throughout the integrations at their initial values. For a full discussion of the parameterization of the land-surface processes see, for example, the review by Carson (1982).

3. THE STANDARD BOUNDARY-LAYER STRUCTURE IN THE 11-LAYER MODEL

Figure 1 is a schematic representation of the boundary-layer structure in the 11-layer model. The model's fixed layers of different thickness were chosen to provide the highest resolution close to the surface. The surface and Ekman-layer turbulent fluxes of momentum, heat and moisture are parameterized such that they can be determined from the primary model variables, namely, the zonal and meridional wind components (u and v, respectively), the temperature (T) and specific humidity (q) which are 'carried' at the full levels at each grid-point (i.e. without staggering), and from values of surface properties such as surface roughness, temperature and wetness.

The general philosophy of the approach is that of Method I of Clarke (1970). Let F_χ represent the mean vertical turbulent flux of some conservative quantity whose mean value is χ, i.e. $F_\chi = \overline{w'\chi'}$ in conventional turbulence notation is the eddy covariance of χ and the vertical velocity component w.

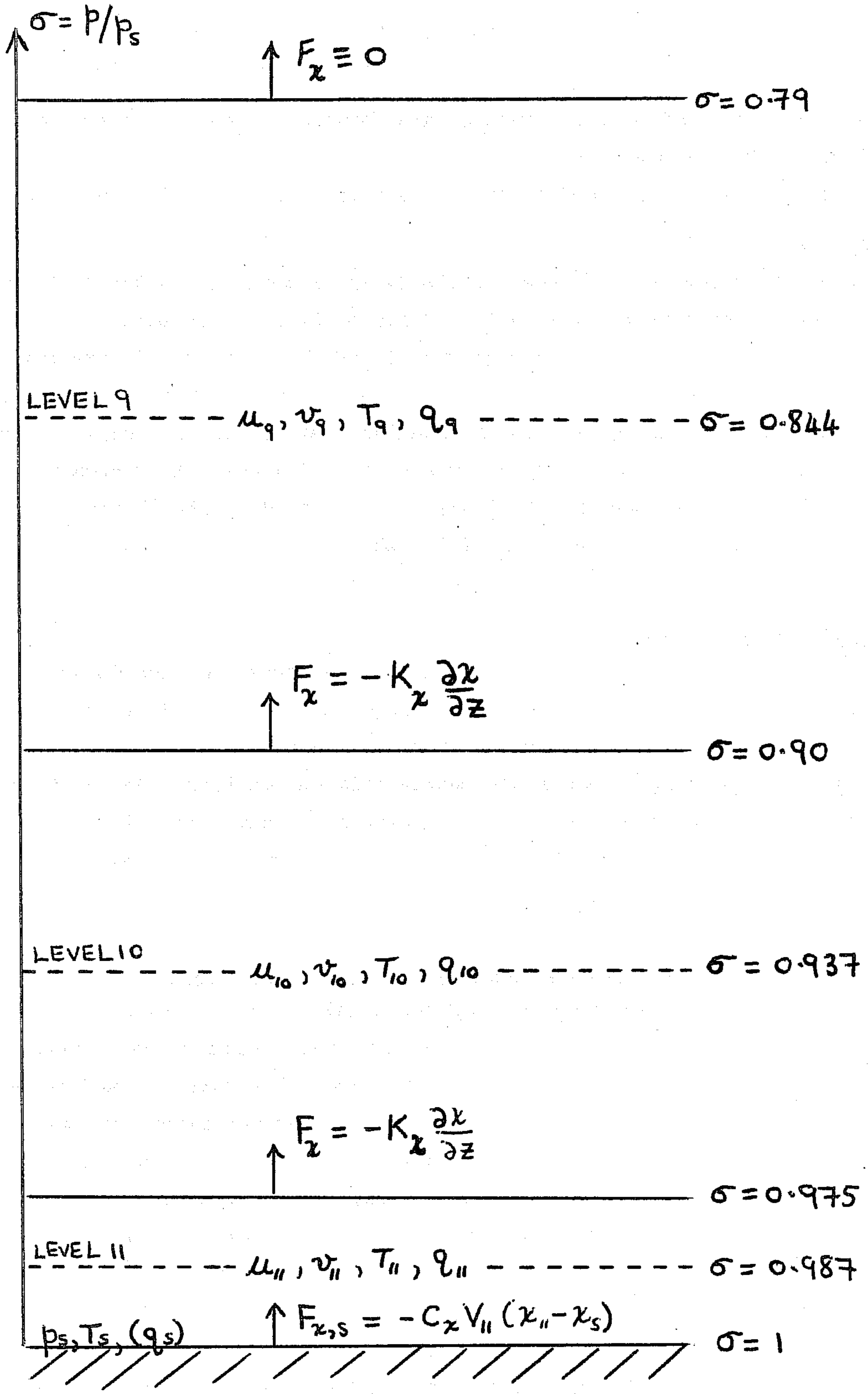


Figure 1. Schematic representation of the standard boundary-layer structure in the Meteorological Office 11-layer model. See text for details.

3.1 Surface turbulent fluxes

It is standard practice at the surface ($\sigma = 1$) to express F_χ by

$$F_{\chi,s} = -C_\chi V_{11} (\chi_{11} - \chi_s) \qquad (1)$$

where V_{11} is the mean wind speed at level 11 ($\sigma = 0.987$), χ_s is the surface value of χ and C_χ is the so-called bulk transfer coefficient. Theoretical and field observational studies indicate that, in general, C_χ is different for each χ and is a complicated function of height, atmospheric stability, surface roughness and other physical and physiological characteristics of the surface vegetation. The surface layer bulk transfer coefficients for momentum (C_D), heat (C_H) and moisture (C_E) transfer used in the 11-layer model are illustrated in Figure 2 and have been derived from Monin-Obukhov similarity theory. They are expressed as functions of the surface roughness length, z_0 (assumed to be 0.1 m over land and 10^{-4} m over sea), and a suitably defined stability parameter (the bulk Richardson number, Ri_B, for the layer approximately 100 m thick, from the surface to level 11). For a full description of the general method and the assumptions implicit in it see, for example, Carson and Richards (1978). For further particular details for the 11-layer model see Carson (1982).

3.2 Ekman-layer turbulent fluxes

The boundary-layer is assumed to extend throughout the bottom three layers of the 11-layer model to $\sigma = 0.79$ where all the vertical turbulent fluxes are assumed to vanish. At the remaining two model layer boundaries ($\sigma = 0.975$ and 0.90, respectively) within the atmospheric boundary-layer the turbulent fluxes are parameterized using a flux-gradient approach. Thus, in the theoretical framework of a continuous height coordinate, z,

$$F_\chi = -K_\chi \frac{\partial \chi}{\partial z} \qquad (2)$$

where K_χ, the eddy diffusivity for the quantity χ, can be expressed as

$$K_\chi = \ell_m \ell_\chi \left| \frac{\partial \underset{\sim}{V}}{\partial z} \right| = \Lambda_\chi^2 \left| \frac{\partial \underset{\sim}{V}}{\partial z} \right| \qquad (3)$$

where $\underset{\sim}{V}$ is the horizontal wind velocity, ℓ_χ is a mixing-length appropriate for the vertical transfer of χ and ℓ_m is the corresponding mixing-length for momentum transfer, thereby defining the mixing-length $\Lambda_\chi = \surd(\ell_m \ell_\chi)$.

In reality the specification of appropriate functions for K_χ (or Λ_χ) is very difficult. However, in practice in the original version of the 11-layer model the mixing-lengths were assumed to vary very simply with height and stability as detailed in Method I of Clarke (1970). Where mixing-lengths are used for the

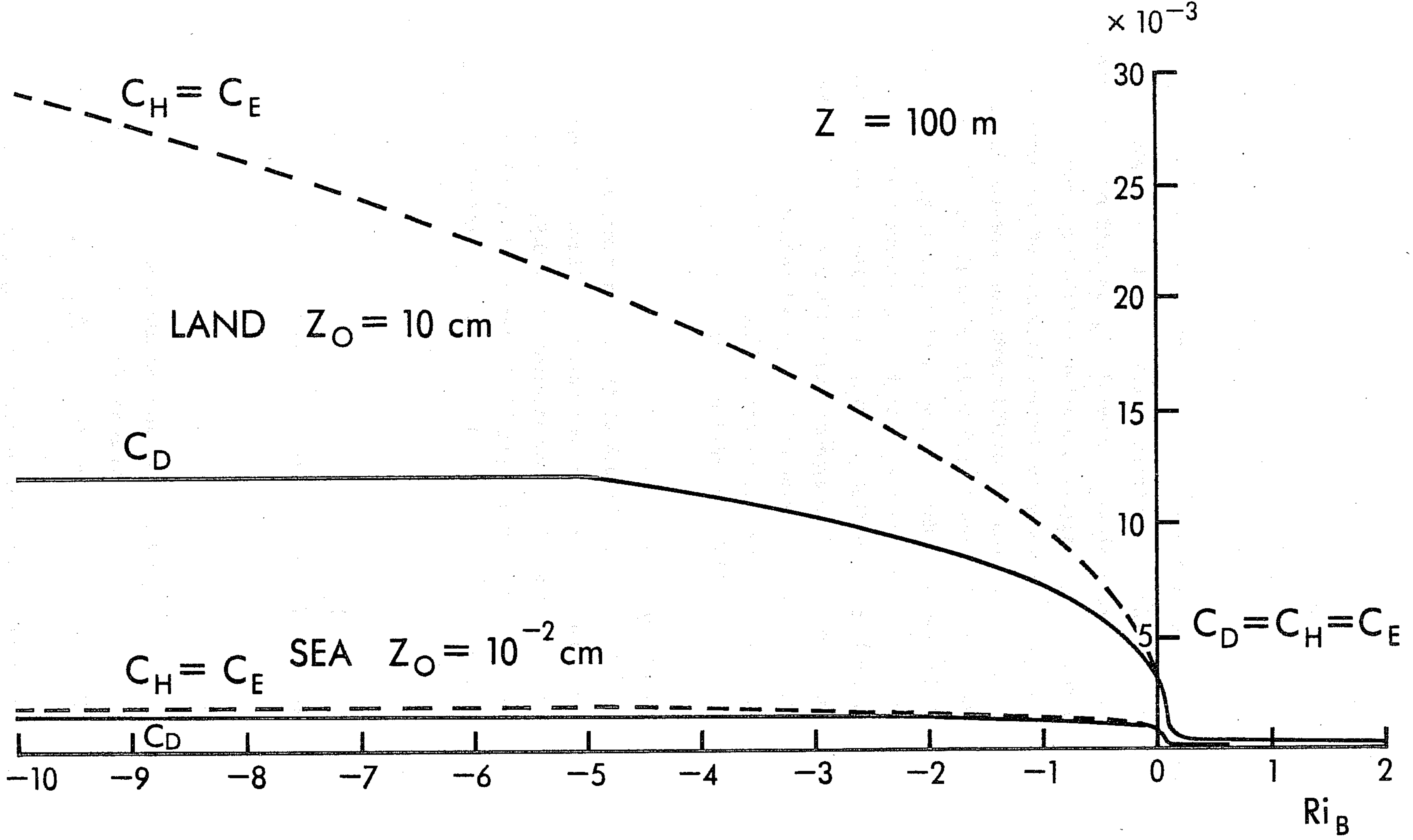


Figure **2** Surface layer bulk transfer coefficients derived from Monin-Obukhov similarity theory and used in the MO 11-level model.

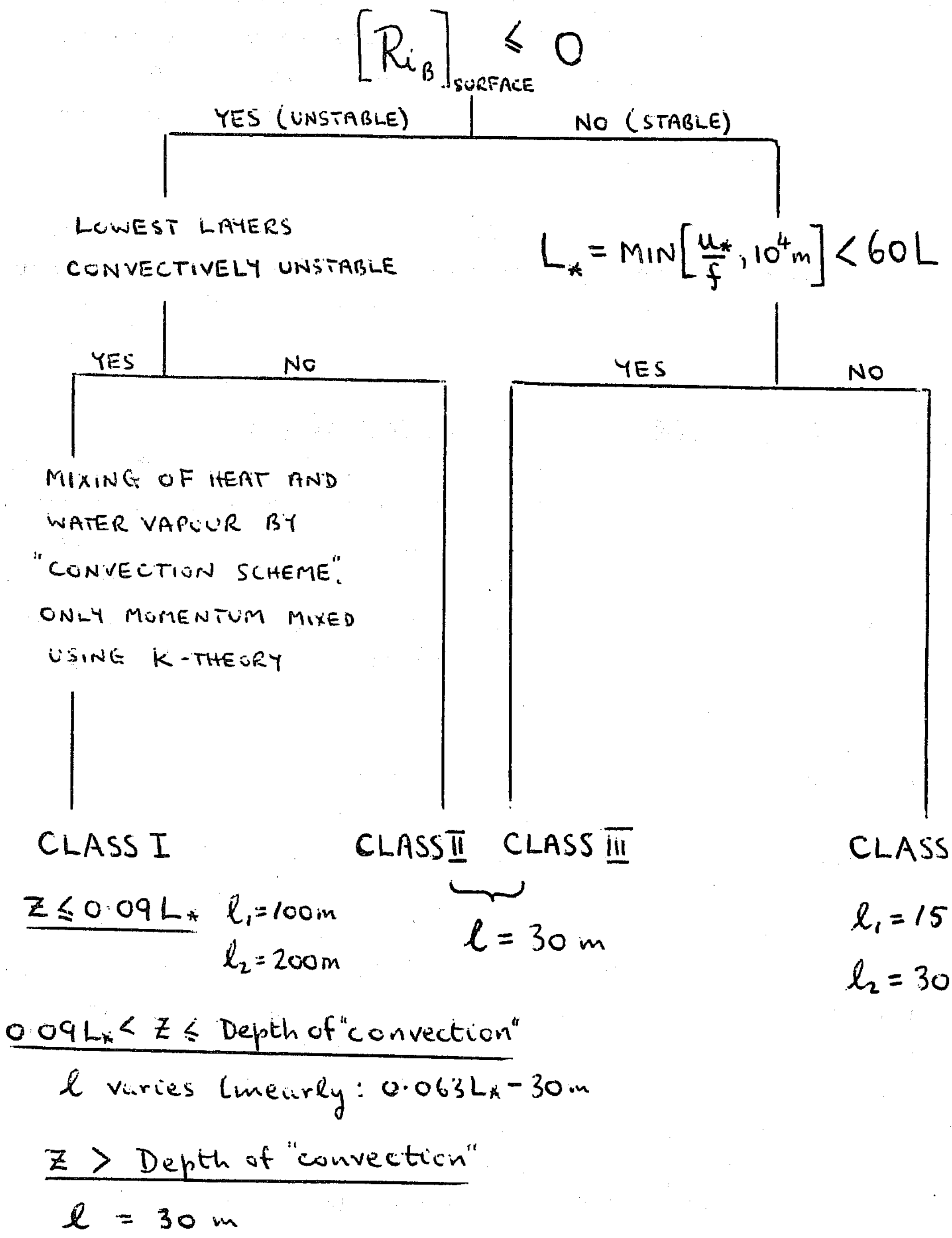


Figure 3. Simplified, schematic description of Method I of Clarke (1970) for estimating the Ekman-layer turbulent fluxes, as employed in the Meteorological Office 11-layer model. L is the surface Monin-Obukhov length, u_* is the surface friction velocity and f is the Coriolis parameter. The other symbols are defined in the text.

turbulent transfer of heat and water vapour no distinction is made between them and that used for momentum. In suitably unstable conditions the vertical mixing of heat and water vapour is catered for by whichever convection scheme is employed in the model. Figure 3 gives a simplified and schematic description of Clarke's proposals as used in the 11-layer model. It is assumed that $\sigma = 0.975$ corresponds to $z_1 \simeq 200$ m and $\sigma = 0.90$ to $z_2 \simeq 830$ m with corresponding mixing-lengths, ℓ_1 and ℓ_2.

A comprehensive study of the eddy diffusivity approach to the parameterization of the turbulent transfer processes in the boundary-layer has been made by Richards (1980) who developed and tested more detailed functional forms for the mixing-lengths for use in the 11-layer model.

4. SURFACE TURBULENT FLUXES OBTAINED WITH THE STANDARD SCHEME

One of the main aims (debatably the main aim) of a 'boundary-layer scheme' is to provide realistic estimates of the surface fluxes of heat, moisture and momentum. Various diagnostics of the surface fluxes, ranging from time-series of globally-averaged values to detailed time-step by time-step representations of the diurnal cycle at selected grid-points have been derived from integrations of the 11-layer model with the standard boundary-layer scheme described above. A few examples only are provided in Figures 4-7.

Figure 4 gives the zonal averages of sensible heat, evaporation and the corresponding precipitation meaned over the last thirty days of two 50-day global integrations, one a January simulation, the other for July. It is not easy to obtain reliable data, particularly on a global basis, for verifying the surface fluxes. The 'observed' values shown here were taken from the climate data catalogues of Schutz and Gates (1971, 1972a, b). Perhaps all that can be claimed is that the major latitudinal and seasonal differences evident in the 'observations' are captured reasonably well by the model. However, the model's precipitation rates are generally too low compared with Möller's (1951) estimates as given in Schutz and Gates (1972a, b). The need to exercise extreme caution when interpreting such data is emphasised for example by the discrepancy between modelled and 'observed' values of the sensible heat fluxes north of 50°N in January. The climatological, zonally-averaged estimates of Schutz and Gates at these latitudes are based on very few values mainly over the sea where the fluxes are in general of larger magnitude and often of different sign to those over the neighbouring land and sea-ice.

Global maps of these time-meaned surface fluxes (not shown here) further indicate the seasonal differences and also show the expected general contrasts between values over land and sea. At least in a qualitative sense then the main geographical and seasonal variations appear to be simulated with some degree of success in the model. Such variations are illustrated for the magnitude of the surface momentum flux, $\underset{\sim}{\tau}$, in Figures 5 and 6.

Figure 5 shows the distribution of $|\underset{\sim}{\tau}|$ over the Northern Hemisphere at Day 6 Hour 12 and Day 7 Hour 00 of a hemispheric integration run from analyses for 00Z, 4 January 1974. Figure 6 shows the corresponding fields from an integration run from data for 00Z, 1 July 1975. The contour interval is 0.1 Nm^{-2}. Note first of all the major seasonal contrasts in the general character of the values over land and sea. Note also the evidence of large diurnal variation over land in both seasons with a marked decrease in the momentum transfer at the surface at night when the surface layer is stable. Obtaining realistic estimates of the turbulent fluxes in stable conditions over areas represented by the horizontal grid-boxes or elements of a typical general circulation model remains a problem for the numerical modeller. The influence in the models of stable surface conditions, particularly over land, is probably greater than originally appreciated. Modelling evidence suggests that the simplistic view that the surface fluxes in stable conditions are generally so small relative to those in neutral and unstable conditions that it may often be sufficient to set them all to zero (perhaps beyond some specified degree of stability), may not be a satisfactory approximation. See Carson and Richards (1978) for further discussion.

Figure 7 shows time series of the magnitude of the surface momentum flux, τ , the surface sensible heat flux, H, and the surface latent heat flux, $Q = L_eE$, where E is the surface evaporation rate and L_e is the latent heat of evaporation. The instantaneous fluxes are averaged over all land points and all sea points, respectively, in the Northern Hemisphere and the values at 6-hourly intervals are plotted for the first seven days of the hemispheric integrations run from 00Z, 4 January 1974 and 00Z, 1 July 1975, respectively. Note again the contrasts between land and sea, day and night, and summer and winter. Note also the significant temporal trends over the first four days or so of the integrations particularly in the fluxes over the sea. These reflect, in particular, a steady readjustment and increase of the surface winds over the sea during this period.

More detailed analyses have been made (but are not shown here) of the individual fluxes as functions of grid-point characteristics (e.g. latitude and longitude; land or sea or ice; orographic height; snow-free or snow-covered; wet or dry), season of the year, time of day, and the prevailing synoptic weather type. For other examples see, for example, Delsol et al (1971).

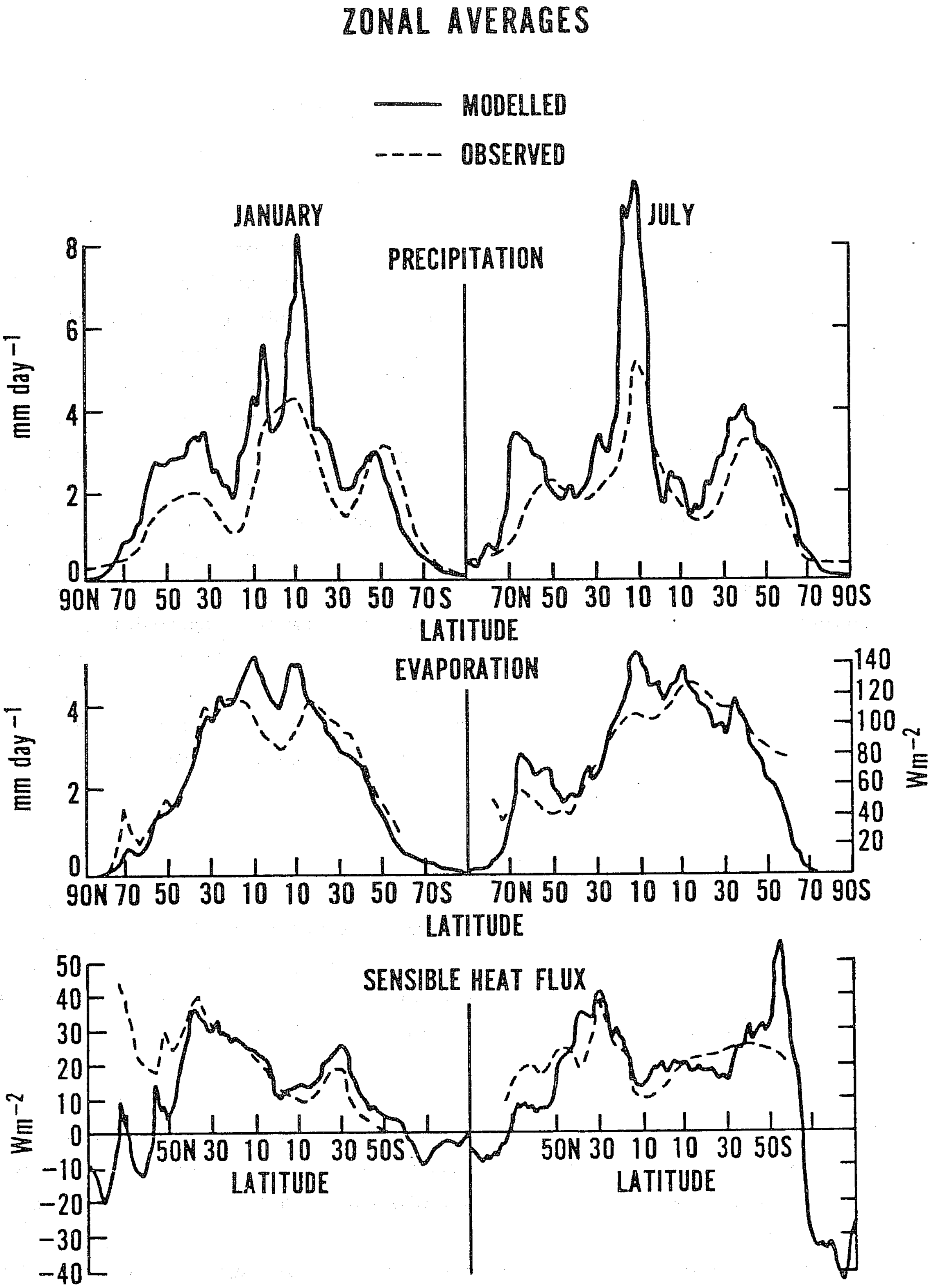


Figure 4. Zonal averages of sensible heat flux, evaporation and precipitation meaned over Days 21–50 of one January and one July integration, respectively, made with the Meteorological Office 11-layer general circulation model. The 'observed' data are from Schutz and Gates (1971, 1972a, b).

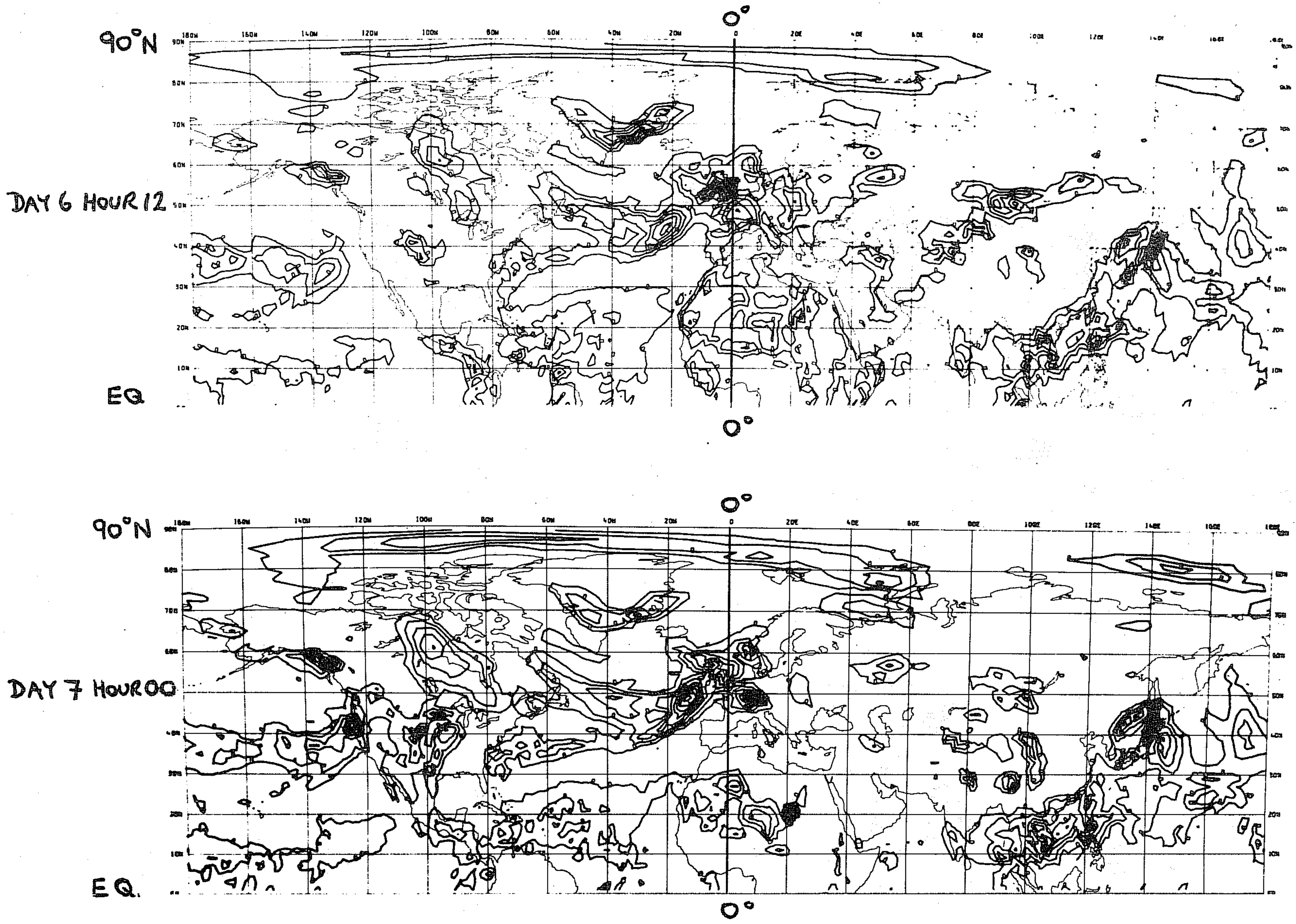


Figure 5. Maps of the magnitude of the surface shearing stress (momentum flux), $|\underline{\tau}|$, at Day 6 Hour 12 and Day 7 H00 of a hemispheric integration run from data for 4 January 1974. The contour interval is 0.1 Nm^{-2}.

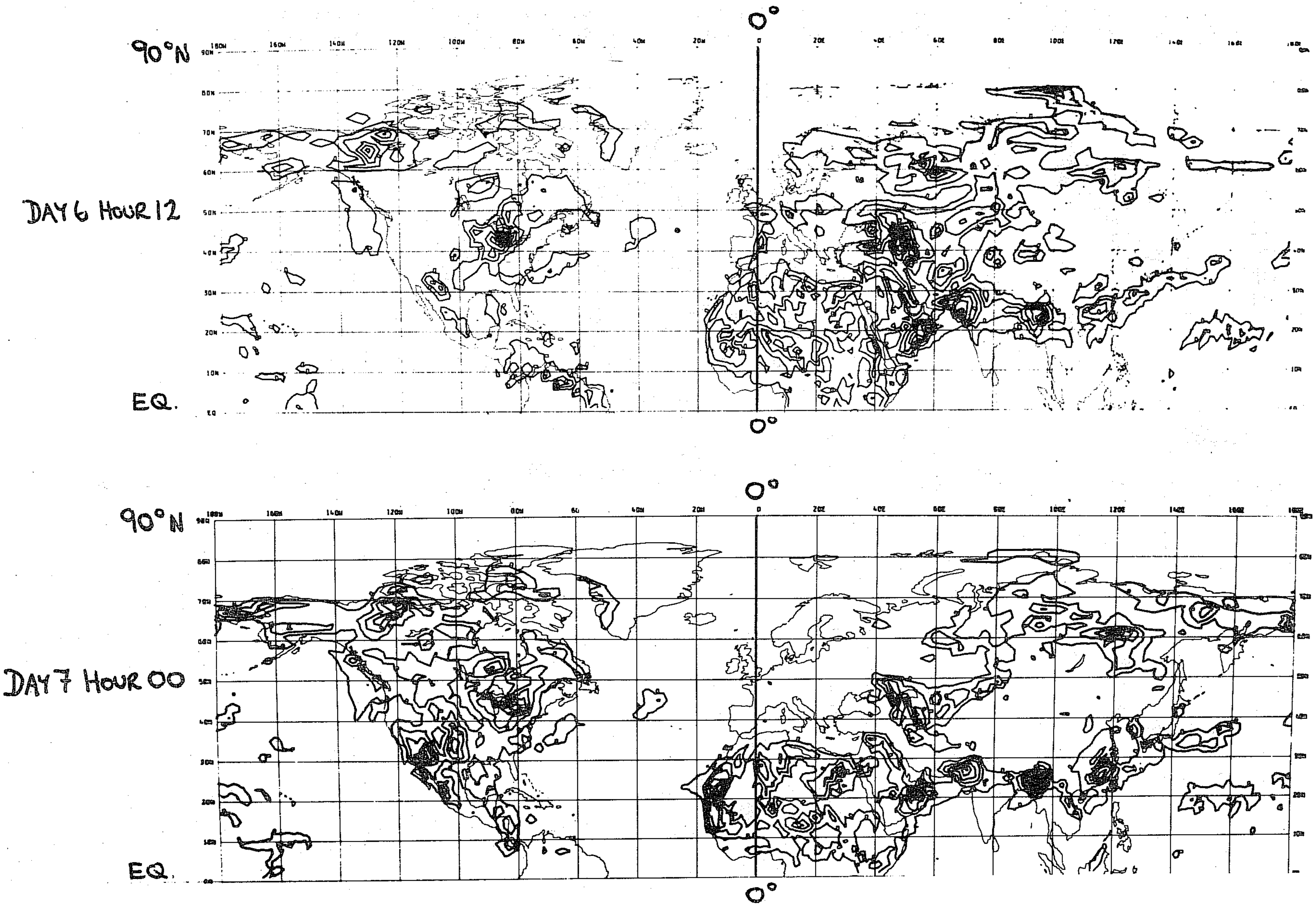


Figure 6. $|\underset{\sim}{\tau}|$ as for Figure 5 but from an integration run from data for 1 July 1975.

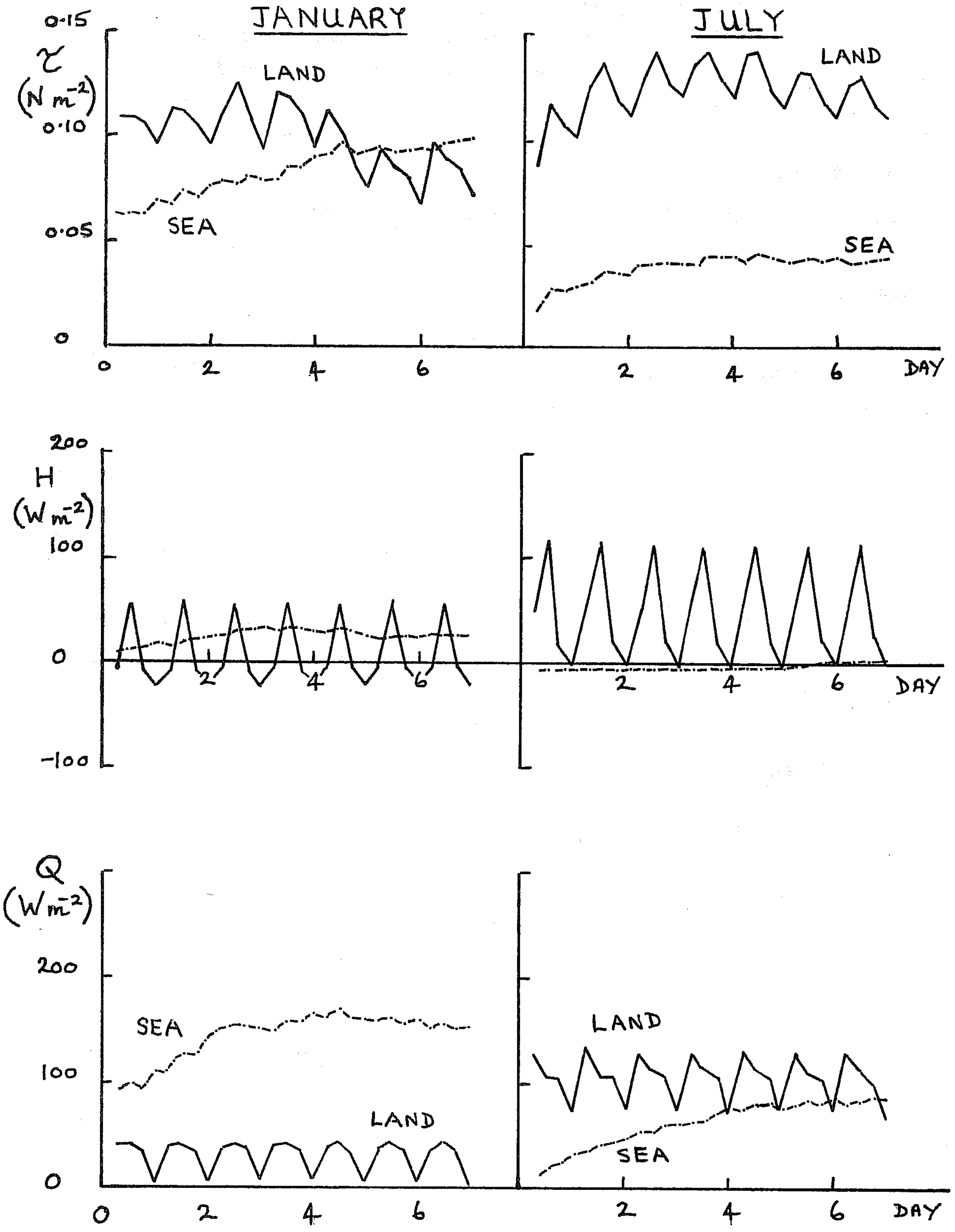


Figure 7. Time series of surface fluxes averaged over all land and sea points, respectively, in the Northern Hemisphere, from a January and a July integration.

5. RESPONSE OF THE 11-LAYER MODEL TO DIFFERENT SPECIFICATIONS OF THE SEA-SURFACE ROUGHNESS LENGTH

5.1 Three different specifications of z_o (sea)

The most substantial effects on 14-day, January, hemispheric forecasts noted by Delsol et al. (1971) were caused by varying the surface drag coefficient between land and sea. Several forecasts have been made with the 11-layer model to assess its sensitivity to three different specifications of the surface roughness length, z_o, over the sea.

Method 1 In the standard version of the model z_o (sea) is fixed at 10^{-4}m and z_o (land) at 0.1 m.

Method 2 z_o (sea) = z_o (land) = 0.1 m was imposed.

Method 3 In this formulation z_o (land) was fixed, as always, at 0.1 m but z_o (sea) was allowed to vary essentially with the surface wind speed according to the so-called Charnock (1955) formula,

$$z_o \text{ (sea)} = \frac{M u_*^2}{g} = \frac{M}{g} \cdot C_D V^2 \tag{4}$$

where u_* is the surface friction velocity, $g = 9.81$ ms^{-2} is the acceleration due to gravity, C_D is the drag coefficient appropriate to the height, z, at which the wind speed V is measured and M is a 'constant' estimated to be about 0.012 by Charnock but later by others to be in the higher range 0.035–0.05. In strictly neutral conditions

$$C_D = \frac{k^2}{(\ln z/z_o)^2} \tag{5}$$

where $k = 0.4$ is von Kármán's constant, and so

$$V^2 = \frac{g}{Mk^2} \cdot z_o \, (\ln z/z_o)^2 \tag{6}$$

$$= \frac{g z}{M C_D} \exp(-k \, C_D^{-\frac{1}{2}}) \tag{7}$$

Therefore, for given z, z_o and C_D can be expressed as functions of $V(z)$ only. For $z = 100$ m and $z_o = 10^{-4}$m, equation (5) gives

$$C_D \equiv C_{D,o} = 0.84 \times 10^{-3} \tag{8}$$

and further with $M = 0.035$ equation (6) gives $V(100) = 5.8$ ms^{-1}.

From the above equations and the constraint that we wish z_o (sea) $\geqslant 10^{-4}$m the following empirical relationship between C_D, $C_{D,o}$ and $V(100)$ was found to apply very well in strictly neutral conditions.

$$C_D = \begin{cases} C_{D,o} & V < 5.8 \text{ ms}^{-1} \\ C_{D,o}(0.74 + 0.046\,V) & 5.8 \leqslant V \leqslant 16.8 \text{ ms}^{-1} \\ C_{D,o}(0.94 + 0.034\,V) & V > 16.8 \text{ ms}^{-1} \end{cases} \qquad (9)$$

Although (9) is strictly only applicable in neutral conditions it was adopted for ease of doing some sensitivity experiments to modify any surface layer bulk transfer coefficient, $C_{\chi,o}$, say, over the sea according to the value of the level 11 wind speed. $C_{\chi,o}$ is the value obtained from the standard Monin-Obukhov similarity approach with z_o (sea) = 10^{-4}m as depicted in Figure 2.

5.2 Results from general forecasting experiments

The three methods described in 5.1 were used to make 7-day forecasts with a hemispheric version of the 11-layer model from initial data for both winter and summer conditions. Differences between the integrations were more noticeable for the winter cases particularly over the sea. Presumably the model is more sensitive to changes in z_o (sea) in winter because of the higher surface winds and generally much larger fluxes in January than in July. The forecast error grows rapidly with all the methods and it is not long before it is difficult to judge which simulation is best. Assessment at about Day 4 is often sensible because it gives a long enough period for the different methods to take effect but is not so far into the integrations that the model's natural tendency for solutions to diverge becomes a major problem. However it is often very difficult to judge which method, if any, gives the best or most realistic simulation without a careful study of the full sequences of synoptic evolutions. Verification of forecasting skill is usually not a very useful measure in sensitivity studies of this sort. From the earliest stages there is usually a clear distinction, particularly over the sea, between methods so crude and diverse as Method 1 (z_o (sea) = 10^{-4}m) and Method 2 (z_o (sea) = 0.1 m). It is usually much more difficult to discriminate between Method 1 and the variable sea-surface roughness case (Method 3).

Figure 8 shows the three forecasts and the corresponding analysis of mean sea-level pressure (PMSL) for Day 7 Hour 00 of integrations run from 00Z, 4 January 1974. The models used differ only in their formulation for z_o (sea). The 'convection scheme' used in all these experiments has been described by Saker (1975) and is an adaptation of the partial-mixing scheme described by Corby et al (1972). The 'radiation scheme' is the simple, climatological scheme which was described by Richards (1977). At Day 7 there is little point in assessing the integrations as deterministic forecasts. There are clearly differences between the results from Methods 1 and 2. However, even at this stage, the differences between Methods 1 and 3 are relatively small. Figures 9 and 10 show the instantaneous differences in PMSL and 500 mb heights at Day 7 between pairs of experiments.

The impact of the different specifications of z_0 (sea) on the synoptic character of the simulations is illustrated in a simple way in Figures 11 and 12 which show the evolution of the mean errors in the forecast PMSL and 500 mb heights averaged over the area north of 30°N. Also displayed are the corresponding mean errors for a forecast made with the boundary-layer processes removed from the model and for a persistence forecast. Note the consistent separation of the graphs from the zero-roughness (no boundary-layer) to the extreme-roughness (z_0 = 0.1 m everywhere) case. Recall that the minimum roughness in the variable-roughness case is z_0 (sea) = 10^{-4}m. Differences between Method 1, Method 2 and the 'no boundary-layer' case are apparent from the early stages of the integrations and are maintained throughout. The results with variable z_0 (sea) (Method 3) show signs of a steady but gradual deviation from those of Method 1. The difficulty of assessing the influence of a change in a particular parameterization by simply objectively verifying a forecast is clearly illustrated in Figure 12 where 'climate drift' of the model is seen in all four model forecasts. The evolutions of the standard deviations of the forecast errors in PMSL and 500 mb heights, averaged over the same domain, are much less distinct and are not shown here.

In summary, integrations of this type show in general that:

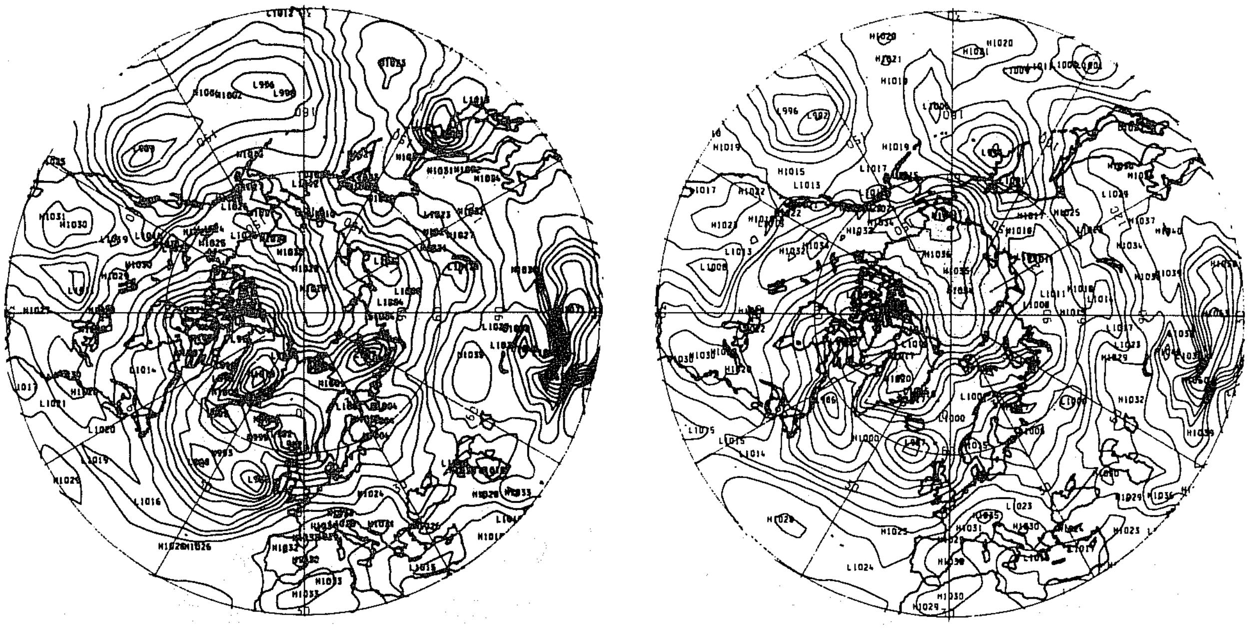
(1) The forecast errors beyond short-range are large for all three methods. Differences between the forecasts in the range of deterministic predictability are generally small compared to the forecast errors.

(2) There are clear differences in character between forecasts using Method 1 and those using Method 2, although it is often not possible to discriminate between them in terms of which gives the better forecast. The synoptic fields are at least sensitive to the differences between Methods 1 and 2.

(3) Differences between Methods 1 and 3 are much smaller even out to Day 7 of a forecast. It is difficult to assess from a very small number of general integrations what the significance of these small differences might be. It is conceiveable that the differences between Metods 1 and 3 become more important locally and under certain conditions, for example, when wind speeds are very high and developments over the sea are particularly intense.

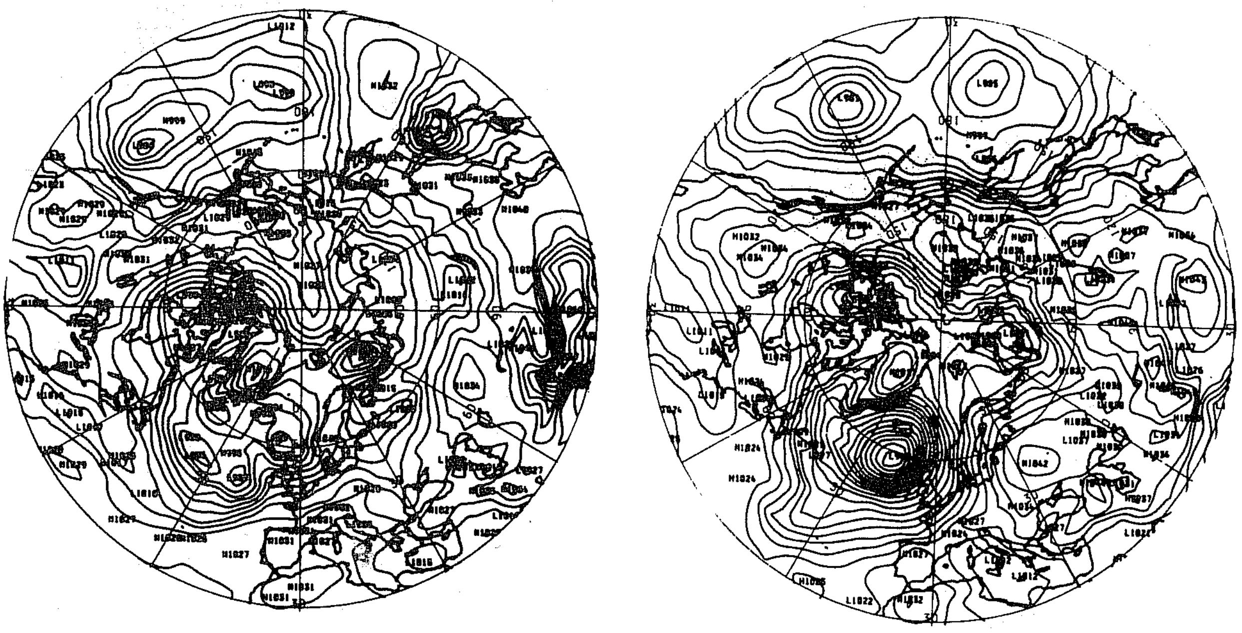
5.3 Sensitivity studies related to the North Sea Storm Surge of 11-12 January 1978

In response to questions related to the Meteorological Office's commitment to the



Method 1 (z_o (sea) = 10^{-4}m)

Method 2 (z_o (sea) = 0.1 m)



Method 3 (z_o (sea) variable)

Analysis for 00Z, 11 January 1974

Figure 8. Three 7-day forecasts and the corresponding analysis of the mean sea-level pressure for 00Z, 11 January 1974. The isobaric interval is 4 mb.

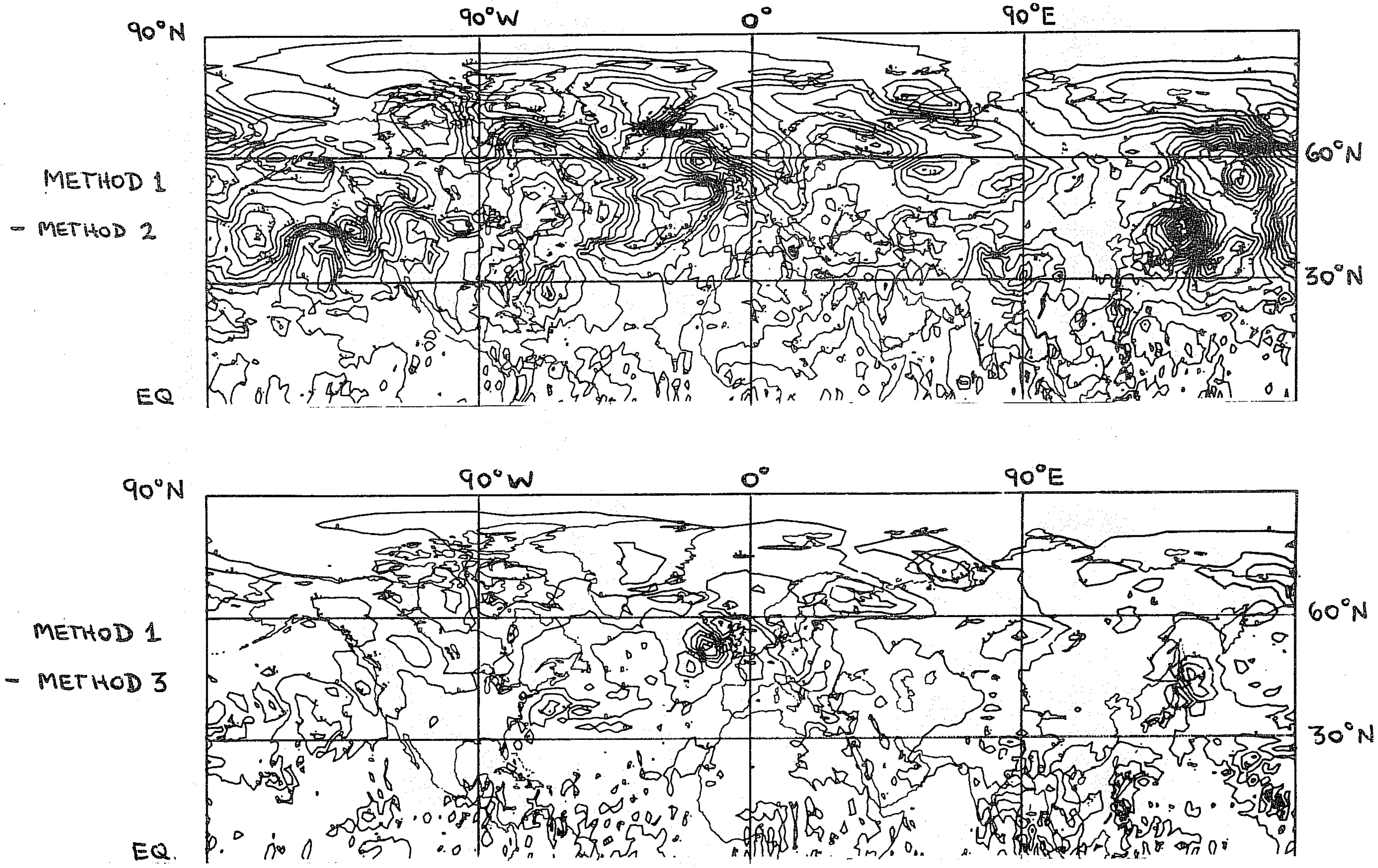


Figure 9. Differences in mean sea-level pressure for Methods 1 and 2 (upper) and Methods 1 and 3 (lower) at Day 7 Hour 00 of integrations from 4 January 1974. The isobaric interval is 3 mb.

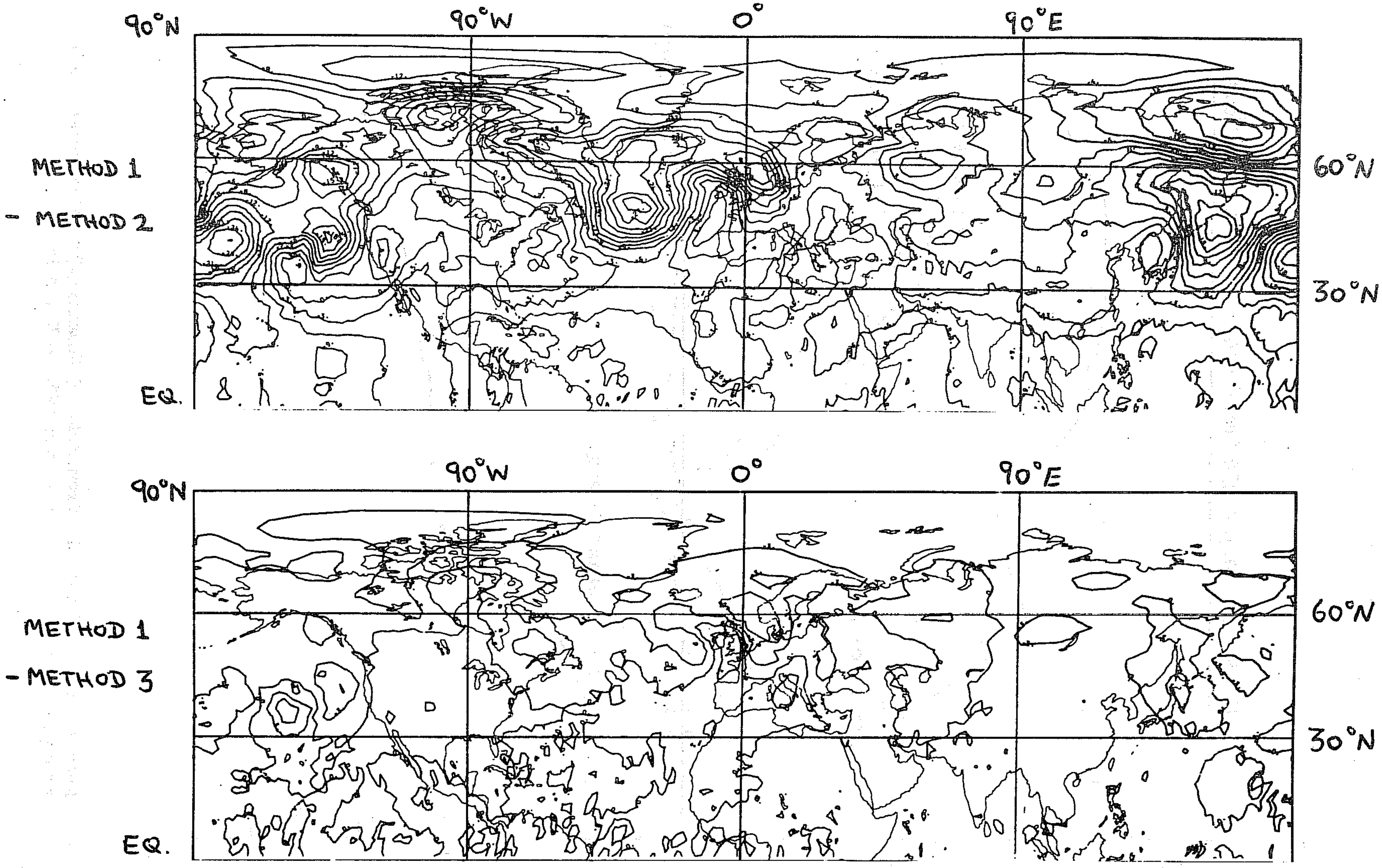


Figure 10. As for Figure 9 but for the 500-mb heights. Contour spacing is 3 dam.

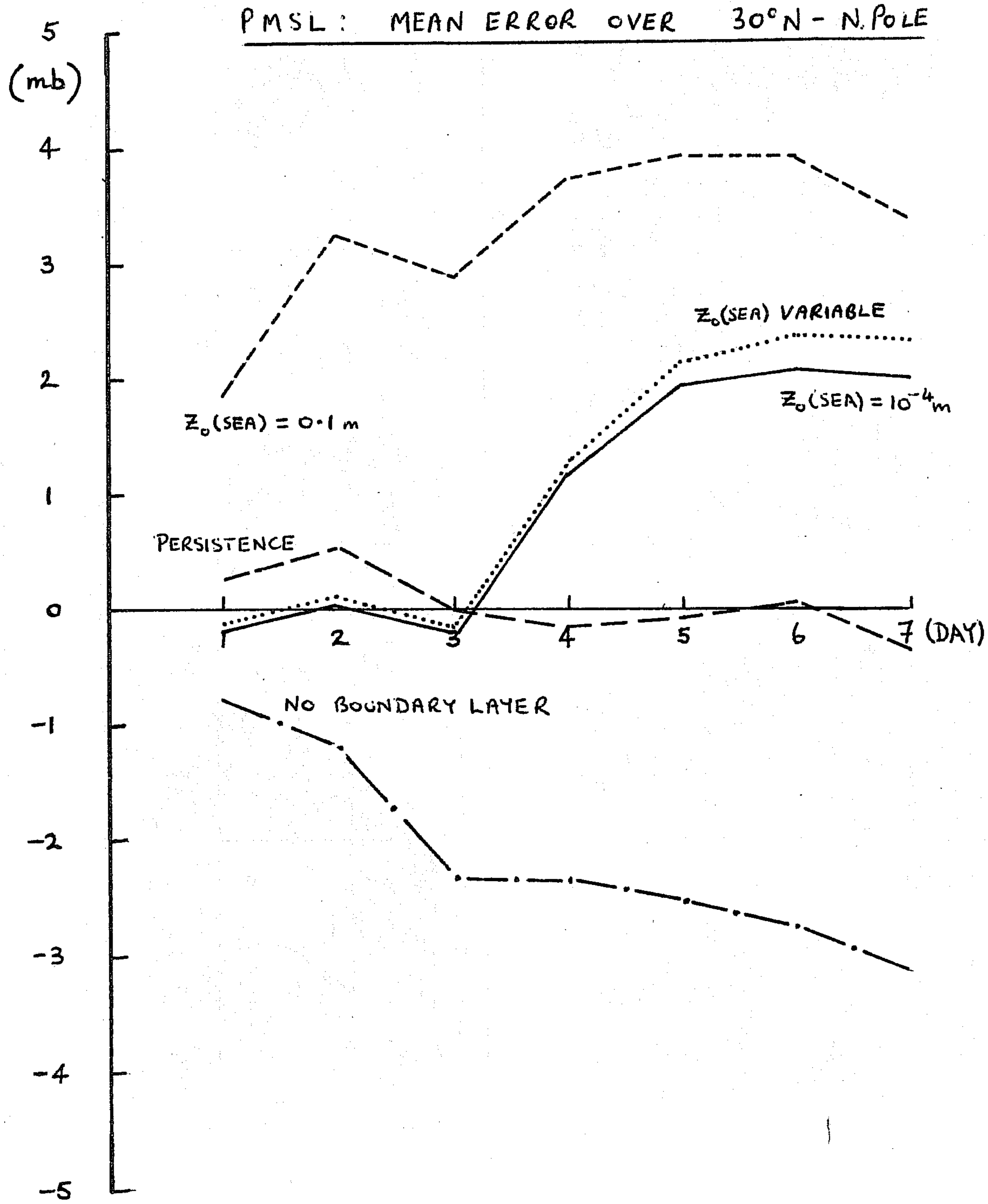


Figure 11. Mean errors in forecasts from 00Z, 4 January 1974 of mean sea-level pressure averaged over the region 30°N - North Pole.

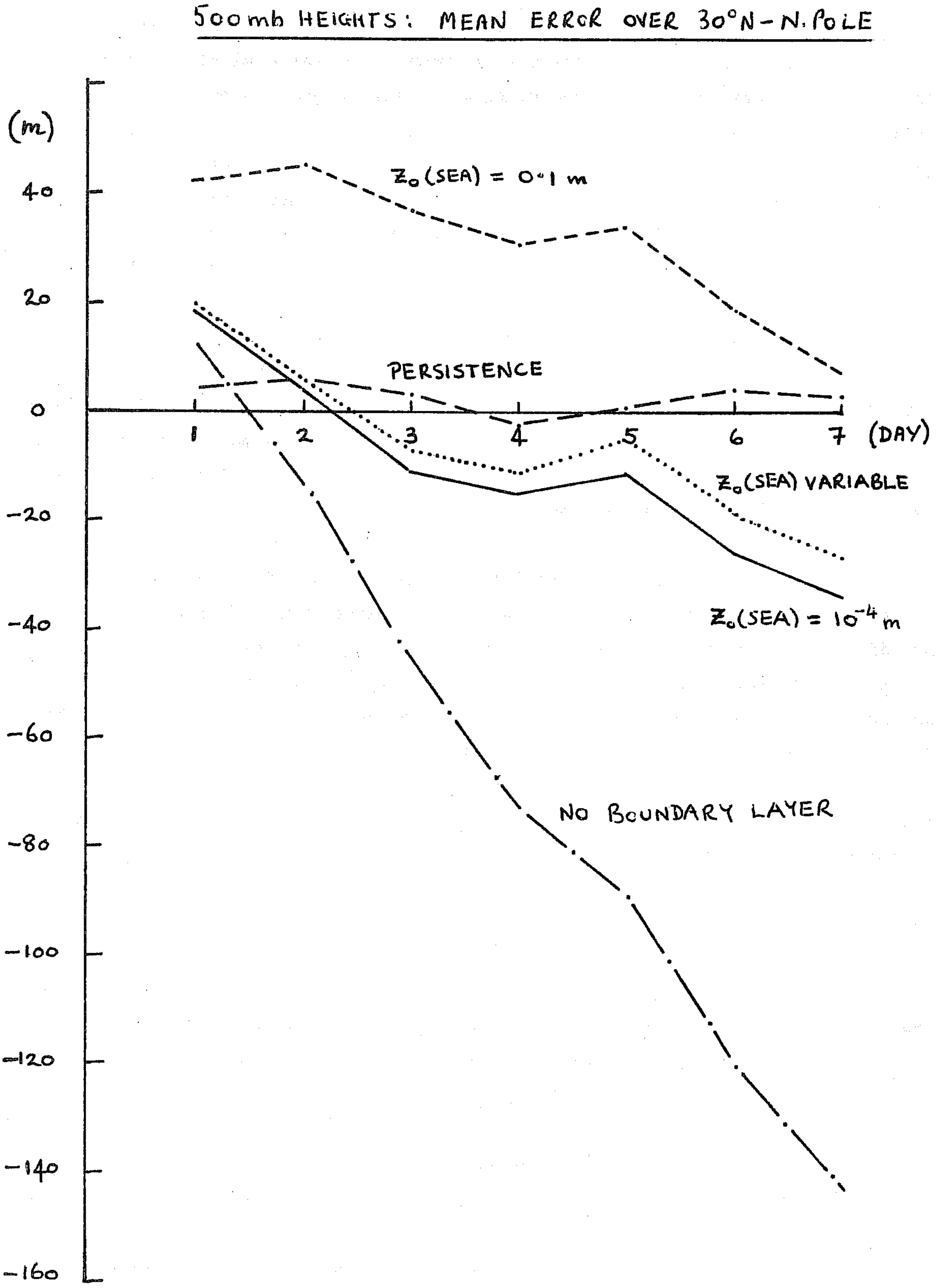


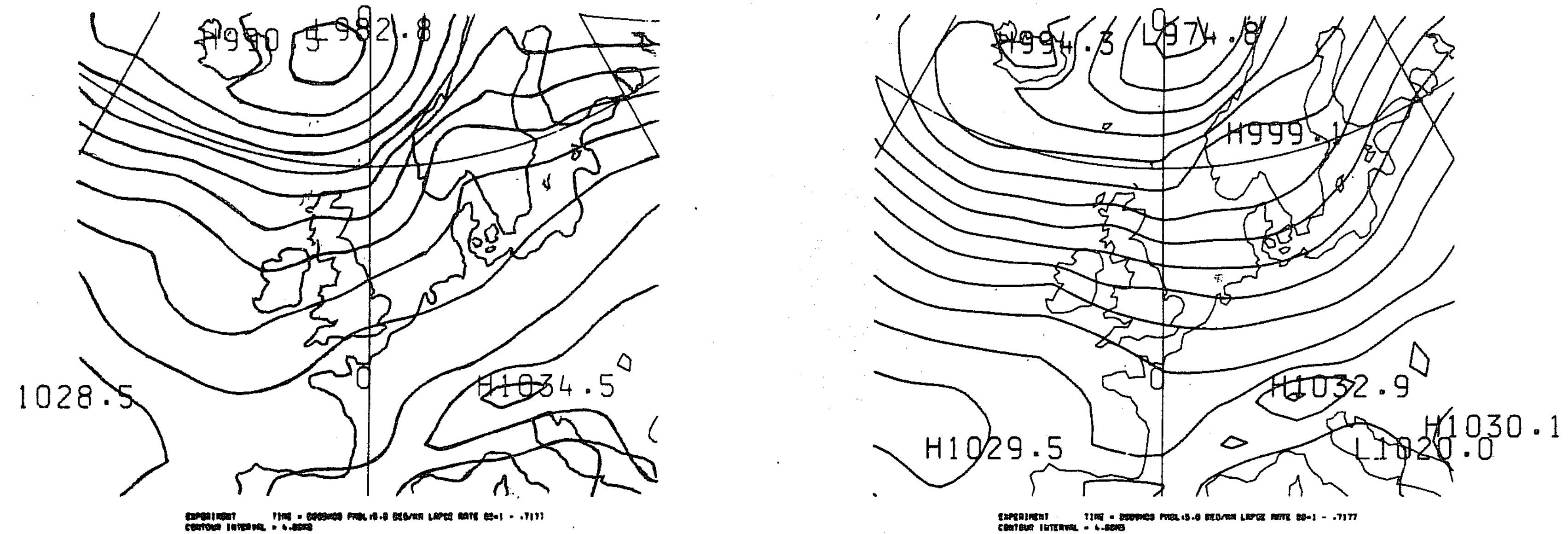
Figure 12. Mean errors as for Figure 11 but for 500 mb heights.

operational prediction of storm surges, a series of nine hemispheric forecasts ranging from three to five days was made with the 11-layer, research model for the period of a significant storm-surge event in the North Sea on 11-12 January 1978. (For a discussion of recent developments in operational storm-surge forecasting using numerical marine and atmospheric models and, in particular, for a fuller description of the storm-surge of January 1978 see, for example, Flather (1979) and other references cited there.) Integrations to 14 January were made with Methods 1-3 for specifying z_0 (sea) using initial conditions from 00Z on each of 9, 10 and 11 January. The series affords a comparison of the impact on the forecasts of changing z_0 (sea) with that of changing the initial conditions.

Figure 13 shows the sequence of PMSL analyses at 24 hr intervals from 00Z 9 January - 00Z 12 January, 1978. On 9 January there is a broad south-westerly airstream over the U.K. with an embedded trough of low pressure over Ireland and Scotland. Over the next 24 hr the deep low east of Iceland moves steadily north-east leaving a trough of low pressure extending to the south of Iceland. Pressure remains high over south-eastern Europe. Troughs of low pressure move east across U.K. leaving it under the influence of a fresh to strong westerly airstream. By 00Z, 11 January a low has moved steadily eastwards from S.E. Iceland and has filled. During the next 24 hr a further low develops over U.K. and moves steadily south east into the southern North Sea bringing a very cold northerly airstream into most districts. By 00Z, 12 January the low has turned north east and is centred over Southern Denmark. Pressure rose quickly to the west of the depression and the strong pressure gradient gave north to north-easterly gales across all parts of the U.K. and over the western part of the North Sea from midday on 11 January to mid-morning on 12 January. The resulting storm surge coincided with a spring tide to produce sea levels on the north-east coast of England in excess of those experienced during the major surges of 31 January 1953 and 3 January 1976 (Flather, 1979).

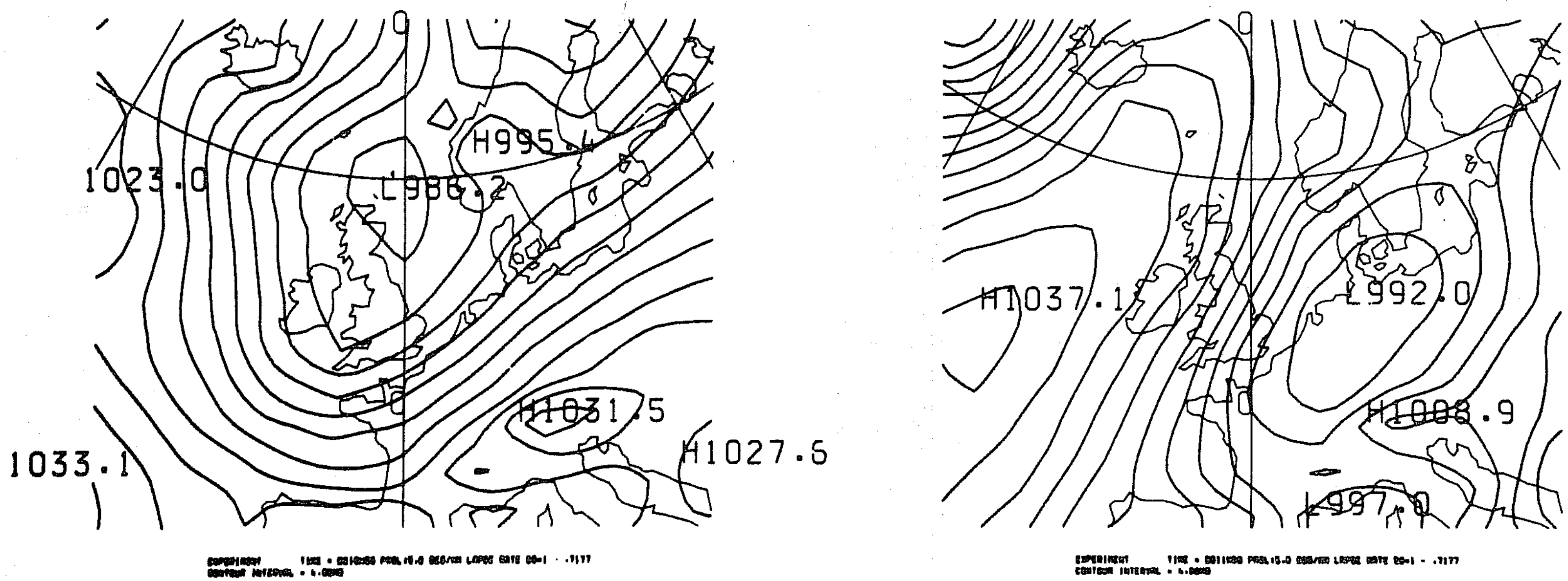
Figure 14 shows the three 72-hour forecasts of PMSL from 00Z, 9 January 1978 and the verifying analysis for 00Z, 12 January 1978. It is quite evident that Method 1 produces large forecast errors but nevertheless there is skill in the sense that the nature of many of the observed changes has been forecast (recall from Figure 13 the initial conditions for 00Z, 9 January). Method 2 has produced just as large forecast errors and a simulation more similar in general character to that from Method 1 than to the analysed PMSL. However differences are already apparent. Indeed the details of the evolution of the forecast tend to favour Method 1. Gradients, for example, tend to maintain a more realistic strength with Method 1. The forecast with Method 3 (variable z_0 (sea)) is very similar to that from Method 1.

Figure 13. Sequence of analyses of mean sea-level pressure for 00Z, 9–12 January, 1978. Isobaric spacing is 4 mb.



00Z, 9 January 1978

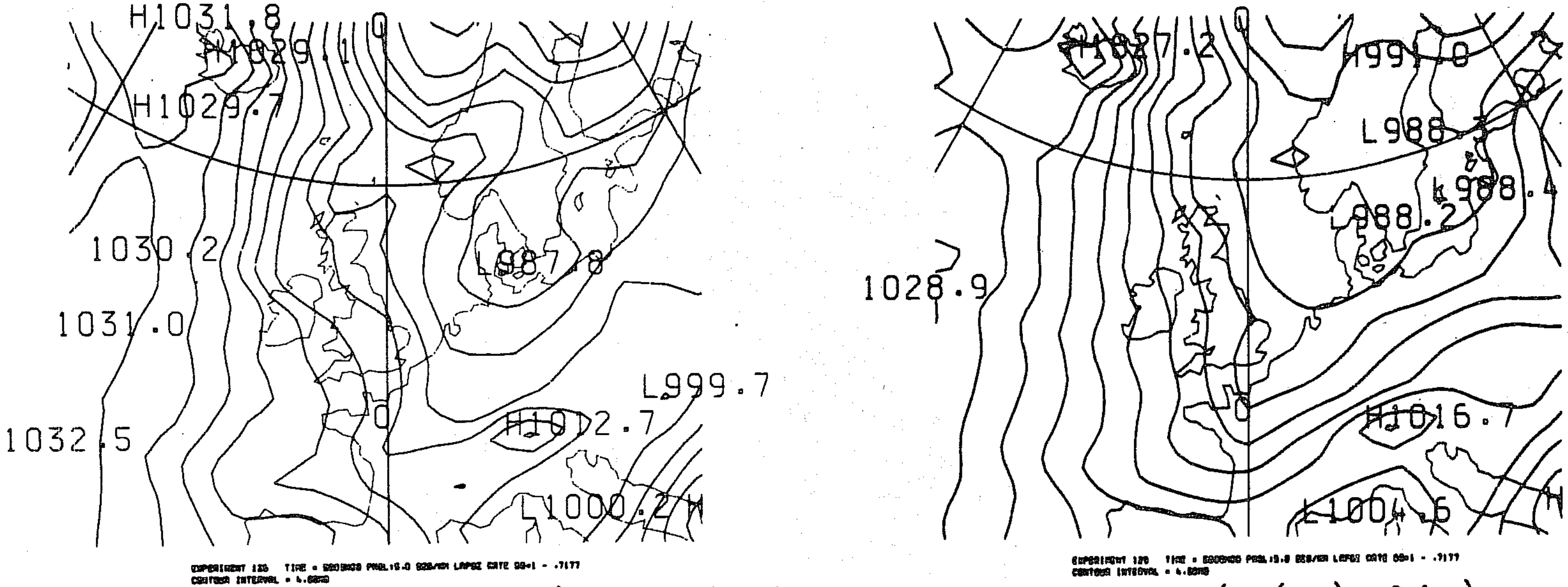
00Z, 10 January 1978



00Z, 11 January 1978

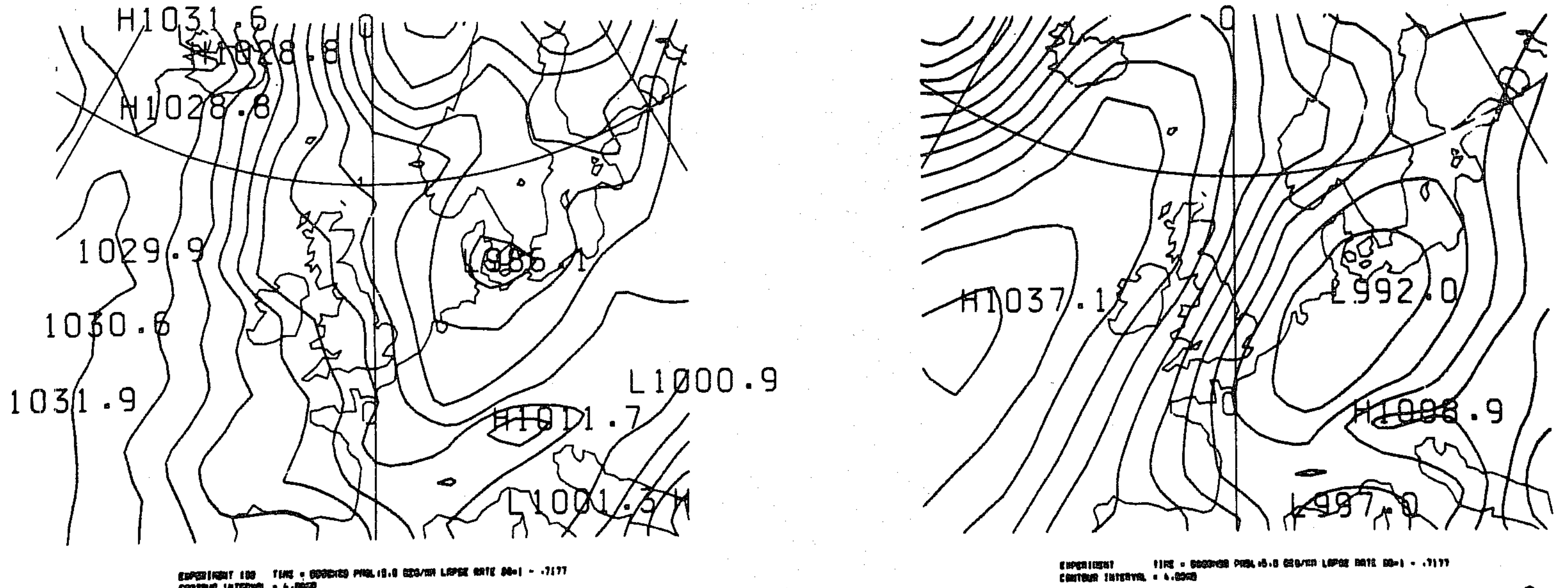
00Z, 12 January 1978

Figure 14. Three 72-hour forecasts of mean sea-level pressure and the corresponding analysis for
12 January 1978. Isobaric spacing is 4 mb.



Method 1 $(z_o\ (\text{sea}) = 10^{-4}\text{m})$

Method 2 $(z_o\ (\text{sea}) = 0.1\ \text{m})$



Method 3 $(z_o\ (\text{sea})$ variable)

Analysis for 00Z, 12 January 1978

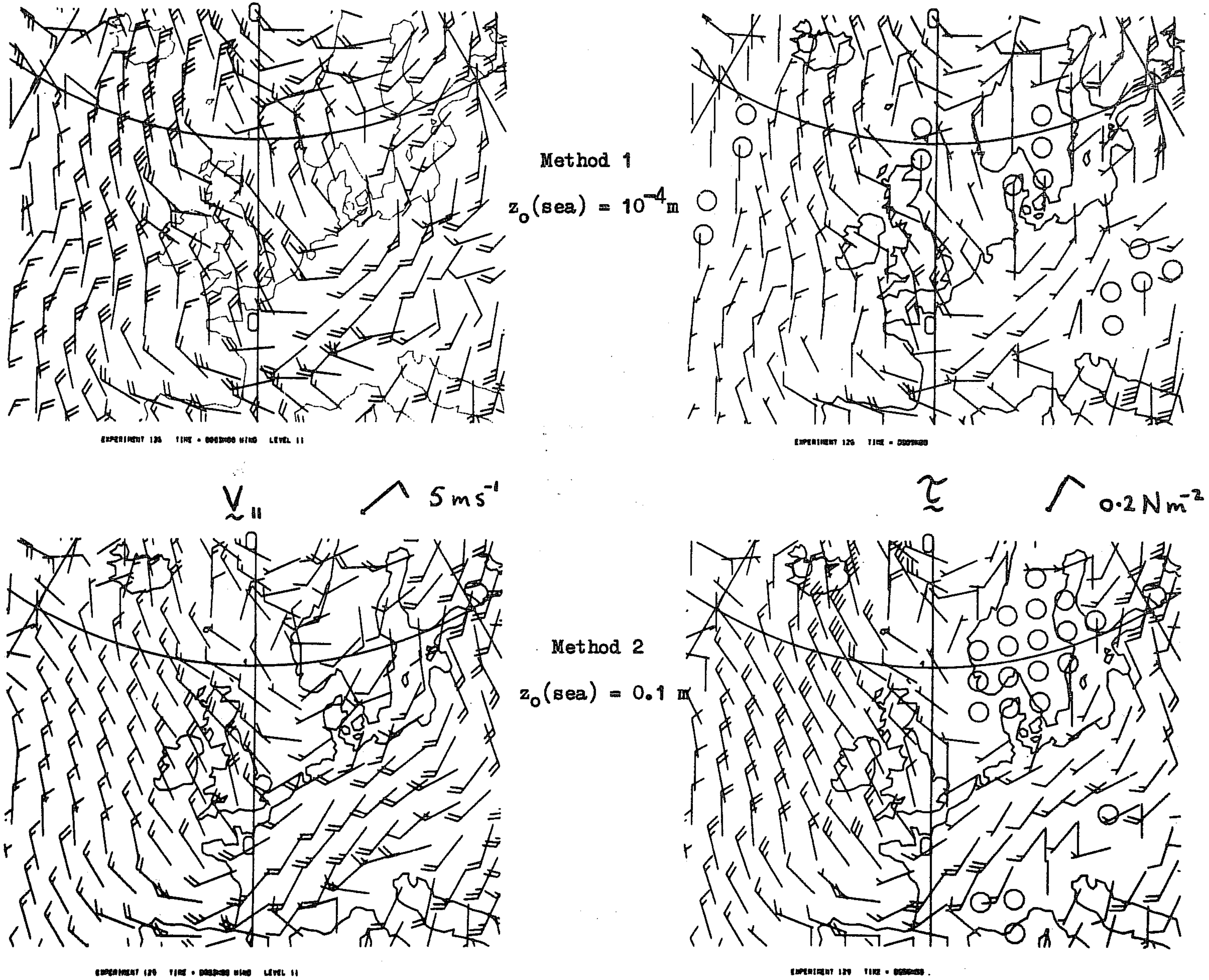


Figure 15. 72-hour forecast low-level winds, V_{11}, (left) and surface momentum fluxes, $\underset{\sim}{\tau}$, (right) for 00Z, 12 January 1978.

Although certain synoptic developments are clearly sensitive, even at such short range, to the three formulations used, such differences are generally small compared to those obtained by moving the time of initialisation of the forecast nearer to that of the events of interest and also compared to the general level of forecast error over the time-scales considered here. The winds used to initialise each of these integrations were derived from a simple linear balance equation. It can be seen from Figure 18 that the impact on the quality of the 24-hour forecast for 00Z, 12 January 1978 when Method 1 was rerun with initial winds derived from a non-linear balance equation far outweighed the effects at that stage of changing z_0 (sea). It should be noted however that, although perhaps not of significance for the synoptic-scale atmospheric developments at these time-scales, nevertheless local differences in low-level winds and surface stresses may be very important if such data are being extracted from an atmospheric forecasting model to initialise or provide boundary conditions for a dynamical marine model. Such differences could be important for example in forecasting the actual elevations of the storm surges.

6. SOME GENERAL AND PARTICULAR INFERENCES FROM THE SENSITIVITY EXPERIMENTS

The following general and specific inferences are tentatively drawn from sensitivity experiments such as those described above to study the response of the Meteorological Office 11-layer model to different specifications of the sea-surface roughness length:

(1) The boundary-layer characteristics and processes which need to be modelled and the degree of accuracy required depend, among other things, on what the model is being used to predict and, for example, the temporal range of the integrations. See, for example, the discussion in Smith and Carson (1977).

(2) Theoretical understanding and observational evidence for the specification of z_0 (sea) are still inadequate, especially for the high-wind regime. In view of the large uncertainty in reducing the complex behaviour of z_0 for a sea-surface to a simple formula, it is thought to be unwarranted at this stage to use, for example, the Charnock (1955) formula and for many modelling purposes it should suffice to use simply $z_0 \simeq 10^{-4}$m everywhere over the sea. Note that this assumption may not be adequate for example for storm-surge or hurricane modelling. Also, changes not obviously important for short-range forecasting may become important in extended-range forecasting, general circulation and climate studies. For the time-scales and the problems being tackled with climate models much more sophisticated couplings between atmosphere and surface may be required, in particular with regard to air-sea interactions. For further discussion and references see, for example, Delsol et al (1971), Augstein (1976) and Smith and Carson (1977).

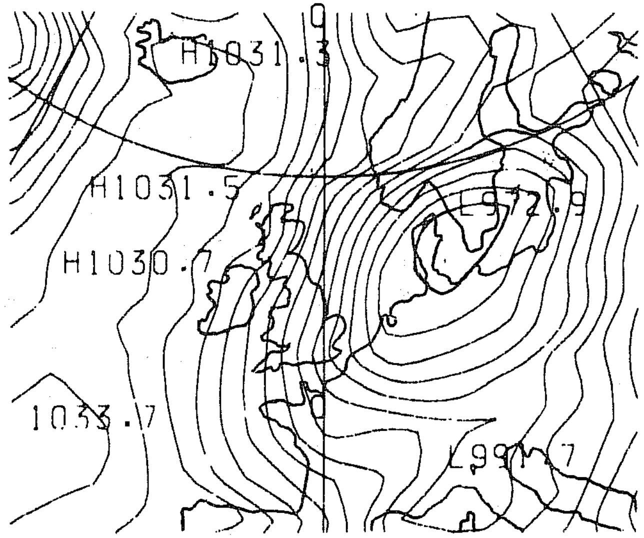
(3) Model sensitivity to different physical parameterizations needs to be determined more systematically and thoroughly than is often the case. It is often

Other variables have also responded to the different formulations although not to the extent of altering significantly the quality of the PMSL forecast at this stage. Figure 15 shows the level 11 (i.e. 100 m) horizontal wind field and the surface shearing stress (momentum flux) field from Day 3 Hour 00 of the forecasts from 00Z 9 January 1978 with Methods 1 and 2. Although the winds are clearly stronger in general over the sea with Method 1, nevertheless the difference in the drag coefficients associated with the different roughnesses (compare, for example, the neutral values for land and sea in Figure 2) more than compensates for this and Method 2 produces the generally larger sea-surface stresses. The wind and stress fields from Methods 1 and 3 are very similar and are not shown here.

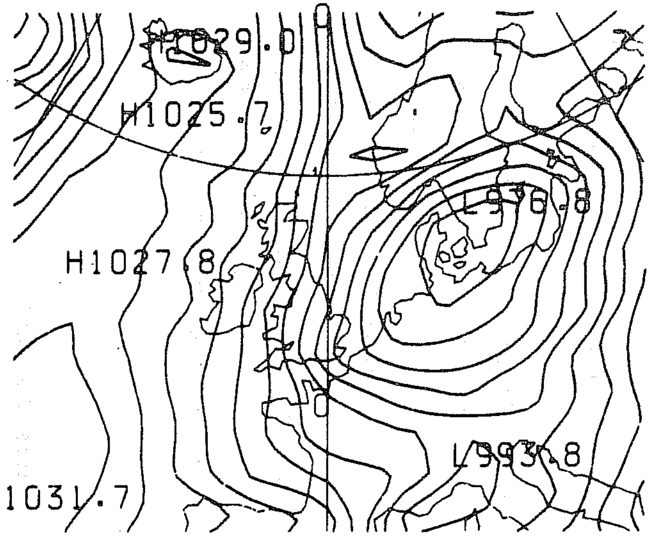
Let us now turn to the 24-hour forecasts for 00Z, 12 January run from analyses for 00Z 11 January, 1978. Figure 16 shows the three forecasts of PMSL and the analysed field for verification. Note that in each case, as expected, the character of the 24-hour forecast for 00Z, 12 January is much better than that of the corresponding 72-hour forecast. Even after only 24 hours there are differences between the forecasts with Methods 1 and 2. There are however still substantial forecast errors which make it difficult to choose between them as forecasts. Although I believe that Method 1 gives the better representation of the nature of the development a simple objective measure such as a root-mean-square error might well favour Method 2. Even at this stage then it could be important from the local forecasting point of view to distinguish between sea and land roughnesses. In contrast, there are at this stage no differences in the PMSL fields produced by Methods 1 and 3 which could be considered significant.

The low-level winds and the surface shearing stresses also reflect the changes in the quality of the forecasts with all three methods (see Figures 17a, b). Again even after only 24 hours obvious differences exist. The northerly gales are better represented by Method 1 than by Method 2. The winds produced by Method 3 are very similar to those of Method 1 but are slightly weaker where the winds are stongest, for example in the North Sea. The surface stresses are also generally much stronger than achieved in the corresponding 72-hour forecasts with again the biggest values over the sea for the case with z_0 (sea) = 0.1 m (Method 2). The surface shearing stresses in the North Sea are generally stronger with the variable roughness (Method 3) than with fixed z_0 (sea) = 10^{-4}m (Method 1), again in spite of the noticeably lighter winds in the former case. This again reflects that allowing the drag coefficient to increase with increasing wind speed according to equation (9) more than compensates for the resulting general decrease in wind speed.

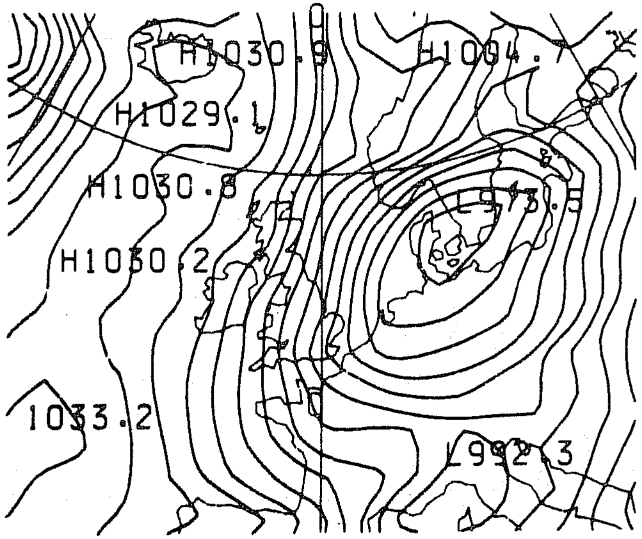
Figure 16. Three 24-hour forecasts of mean sea-level pressure and the corresponding analysis for 00Z, 12 January 1978. Isobaric spacing is 4mb.



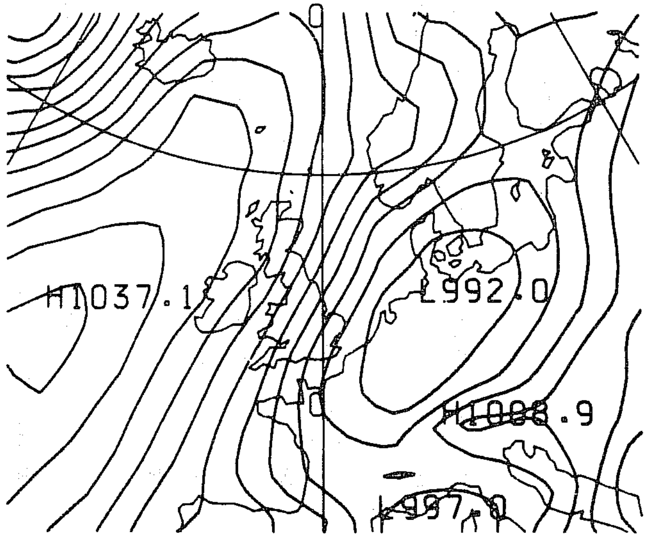
Method 1 (z_o (sea) = 10^{-4}m)



Method 2 (z_o (sea) = 0.1 m)



Method 3 (z_o (sea) variable)



Analysis for 00Z, 12 January 1978

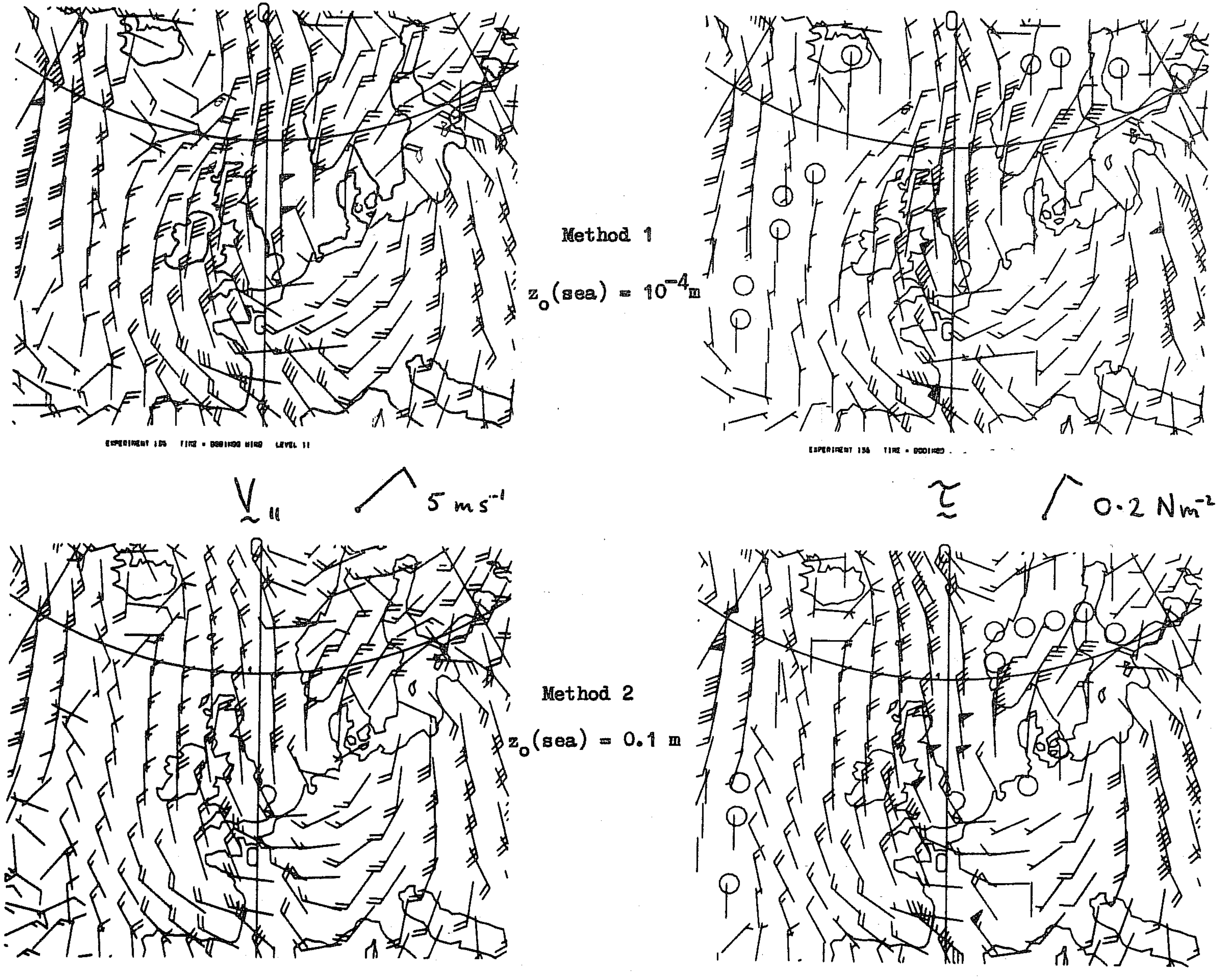


Figure 17a. 24-hour forecast low-level winds, V_{11}, (left) and surface momentum fluxes, $\underset{\sim}{\tau}$, (right) for 00Z, 12 January 1978.

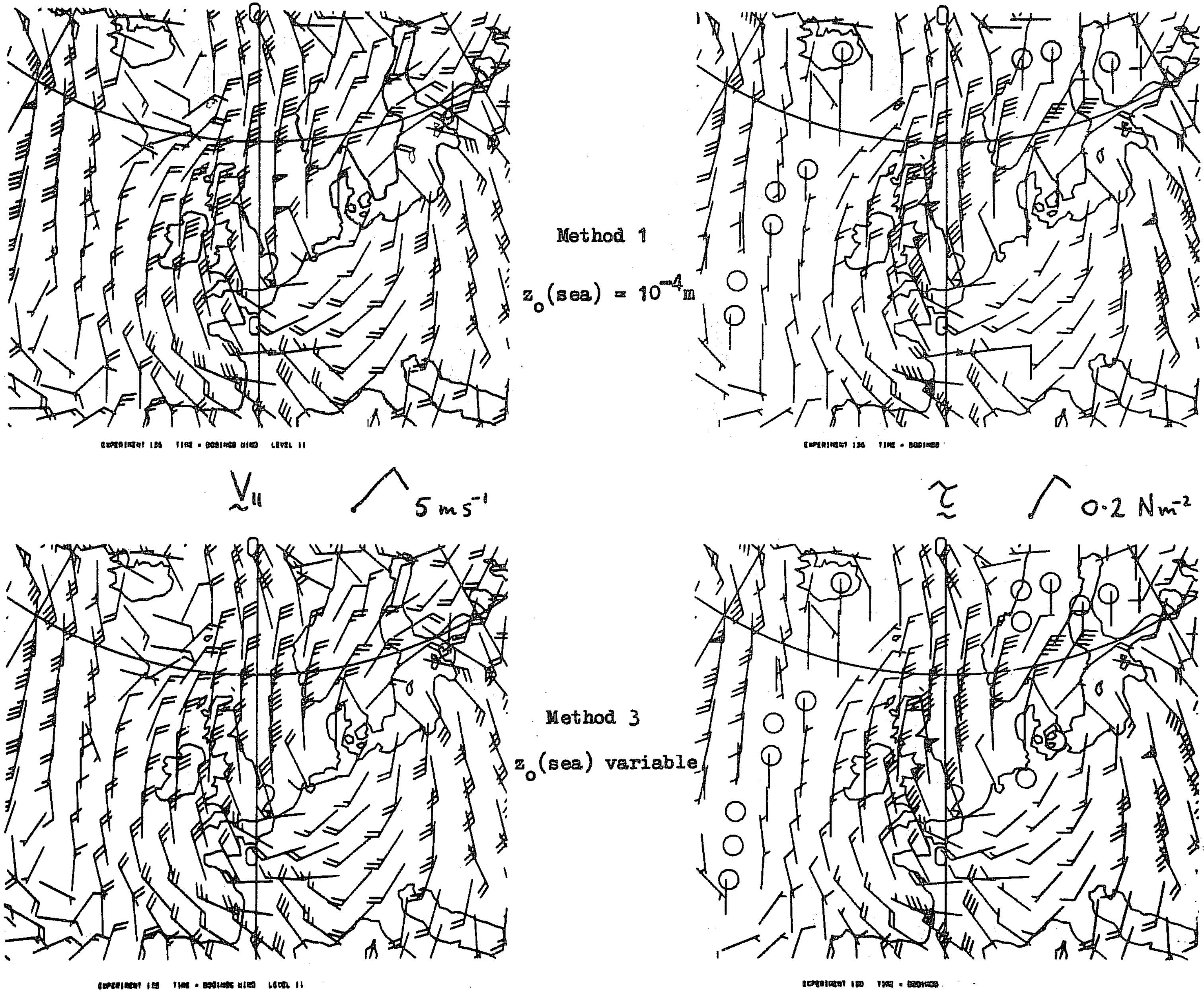
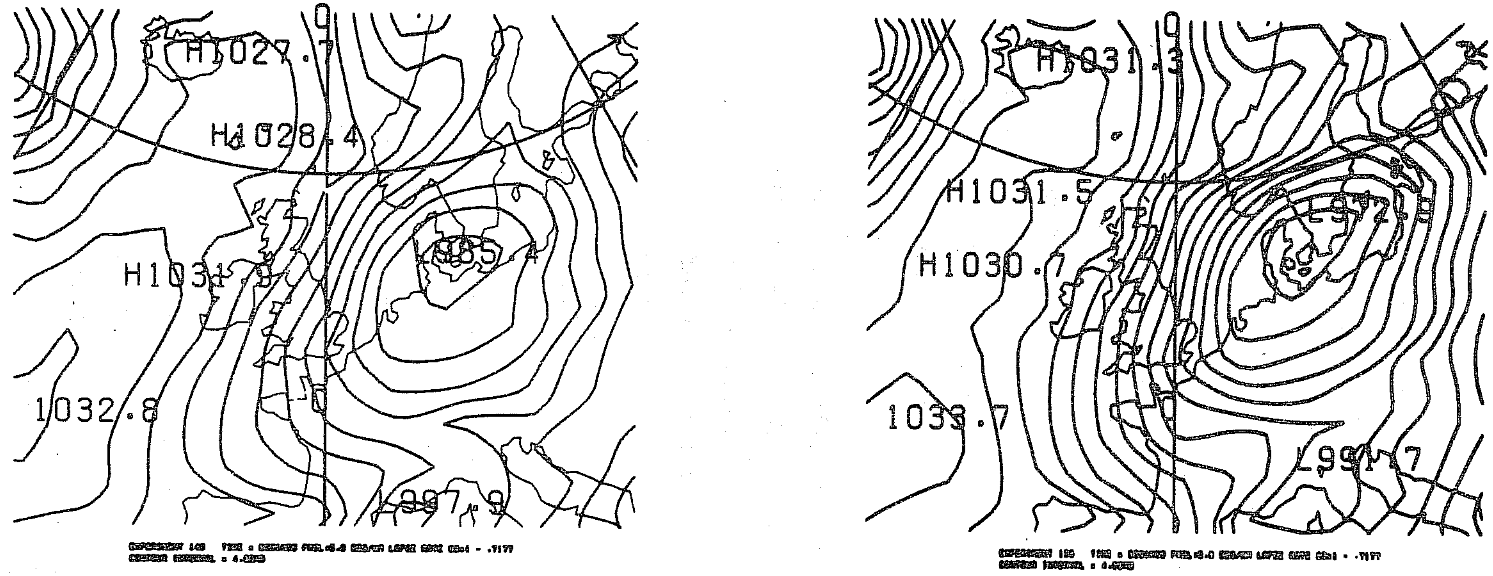
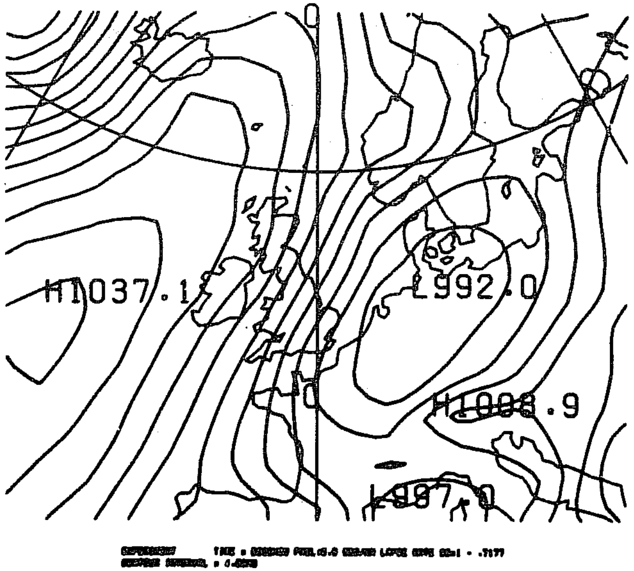


Figure 17b. As for Figure 17a.



Initial winds from non-linear balance equation

Initial winds from linear balance equation



Analysis for 00Z, 12 January 1978

Figure 18. Two 24-hour forecasts of mean sea-level pressure using Method 1 (z_o (sea) = 10^{-4}m) and the corresponding analysis for 00Z, 12 January 1978. Isobaric spacing is 4 mb. The upper left forecast was initialised with winds from a non-linear balance equation. The upper right forecast used initial winds from a linear balance equation.

shown that models are sensitive to certain changes but are they sensitive to the right degree and in the right sense? For economic and other practical reasons it is not always feasible to run and analyse a sufficient number and variety of cases to determine the statistical significance of a response to a particular model change. The models themselves are often too simple to provide the evidence to justify subtle changes. We can certainly discriminate between z_0 (sea) = 10^{-4}m and 0.1 m but it is much more difficult to judge the need or worth of allowing z_0 (sea) to vary with wind speed.

(4) It may not only be very difficult but also quite inappropriate at times to try to determine an optimum physical parameterization on the basis of synoptic forecasts. Forecast errors, as we have seen, can often be very large compared to the differences obtained with different parameterizations of a particular physical process or property. It should be stressed also that the magnitude and character of a response to a parameterization change often depend on the quality of the control state. If the control model already contains the essence of a good parameterization of boundary-layer structure and physics then the 'signal' from a change to a more sophisticated scheme may be weak or absent. If the control state is poor then a strong response to the change might be diagnosed. One should always be wary of using an apparent improvement in a forecast as a reliable guide to a better parameterization unless this is accompanied by an understanding of the physical causes responsible.

(5) Coupled ocean-atmosphere models are still in the early stages of development and remain one of the most outstanding requirements for climate modelling. The magnitude and sign of a modelled climatic change may depend critically on our ability to represent very precisely certain highly interacting, non-linear, physical processes. Air-sea interaction processes may have to be parameterized much more precisely than at present. Undoubtedly there are certain features of ocean modelling which are more sensitive to the surface processes than are the meteorological features normally studied with the atmospheric models. For example, storm-surge predictions by dynamical methods appear at present to be very sensitive to the specification of z_0 (sea) and imply very strongly a relation between z_0 (sea) and the surface wind speed. Simple parameterizations of the surface stress may not be adequate in short-range forecasting models which are 'linked' to marine models for predicting storm surges.

7. CONCLUDING REMARKS

A few particular experiments with the Meteorological Office 11-layer model to assess its sensitivity to changes in the sea-surface roughness have been selected here to illustrate also some of the more general aspects of the ASSESSMENT PROBLEM as it was defined in the INTRODUCTION. Many other such sensitivity studies relating to the parameterization of the boundary-layer properties and processes have been, and are still being, conducted with the 11-layer model. Some of these were described briefly at the Workshop.

The most recent studies revolve around three boundary-layer schemes now available for the 11-layer model. One is the standard 'Clarke-type K-scheme' as described above in Section 3. Of the other two schemes, both of which were developed and partially tested by Richards (1980), one is a more advanced K-approach. The third method follows essentially the philosophy of the bulk or slab method discussed, for example, by Randall elsewhere in these Proceedings. In particular it represents our first attempt (at least in the 11-layer model) to incorporate a prognostic equation for the depth of the boundary-layer.

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