SIMULATION OF CLOUDINESS WITH THE U.K. METEOROLOGICAL OFFICE 11-LAYER AGCM

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Summary: Results are presented from a two-year integration of the Meteorological Office 11-Layer Atmospheric General Circulation Model (AGCM) using a simple cloud-prediction scheme. The ability of such a scheme to represent the global distribution of cloudiness is assessed. Errors in the low cloud field over the sub-tropical oceans are of particular concern as the model is being coupled to an Ocean GCM for long-range climate sensitivity experiments. The reasons for these errors and proposed changes to the model to remove them are discussed.

1. INTRODUCTION

The Meteorological Office 11-Layer AGCM is being used to continue various climate and extended range prediction studies, notably of the impact of sea surface temperature anomalies (Palmer and Mansfield 1984) and of increasing carbon dioxide concentrations (Mitchell 1983,1984). An important element of current research is the development of a reliable cloud-prediction scheme which is based on sound physical principles and is as free as possible from artificial constraints tuned to the present climate. A particular requirement is that the distribution of cloud over the oceans should be realistic, as the model is also being coupled to dynamical ocean and thermodynamic sea-ice models for climate simulations over several decades. In such a coupled model, errors in the cloud cover over the oceans can lead to errors in the sea surface temperature and sea-ice distributions which in turn can degrade the atmospheric circulation.

Some results from an integration with model-predicted cloudiness are presented here and assessed. In some respects the results are encouraging, but there are systematic errors in the low-cloud distribution over the sub-tropical oceans. It is argued that these are due to shortcomings in the representation of the interaction between the boundary layer and the free atmosphere, rather than in the cloud scheme itself. Relatively minor modifications to the model to improve the realism of this interaction are discussed here. In the accompanying paper (Smith 1985), more substantial changes are proposed as part of an attempt to make the prediction of cloud and of its effects on atmospheric thermodynamics more closely integrated with the rest of the model.

2. THE 11-LAYER AGCM

The version of the model used here is that described by Slingo and Wilderspin (1984a), which differs from that used by Rowntree (1984) as follows. The horizontal grid is 2.5 degrees latitude by 3.75 degrees longitude (i.e. 72 x 96 points). The radiation scheme is that described by J.M. Walker (1977), as modified by Slingo and Wilderspin (1984a,b) to remove systematic errors in the longwave fluxes and incorporate revised cloud radiative properties and surface shortwave albedos. The diurnal cycle of radiative heating is included by calling the radiation scheme at each grid-point every 3 hours of model time.

The cloud-prediction scheme is derived from the work of J.M. Slingo (1980). In the present version, convective cloud is still obtained from the saturated mass flux calculated in the convection scheme. The dependence of the low cloud amount on the lapse rate has been removed, however, so that layer cloud amounts (in three categories: high, medium and low) are now determined solely by relative humidity. This relative humidity is an adjusted value which corresponds to that of the environment between any convective clouds at that level. The layer cloud amounts are given by a quadratic dependence on relative humidity, rising from zero at the threshold relative humidity of 85% to full cover at saturation. Some tests of the scheme have been carried out in short forecast integrations as part of the work initiated by WGNE, which indicated that the scheme can reproduce the cloud cover associated with major synoptic features such as mid-latitude depressions and tropical convective systems.

3. RESULTS

3.1 Global means

The results shown here are from a recent integration of the model for over two simulated years, using FGGE data for 25 July 1979 as the initial conditions. Some globally-averaged cloud and radiation diagnostics from each year of the run are summarised in Table I. There is very little inter-annual variability in these data, primarily because the model is constrained by the imposed sea surface temperatures and sea-ice extents, which repeat exactly each year. The total cloud cover averaged over the year is in satisfactory agreement with climatological data, although since the latter estimates vary between 48% (London 1957) and 60% (Berlyand et al, 1980) this is hardly a severe test. The seasonal variations in the

TABLE 1 Globally-averaged cloud and radiation diagnostics from an integration of the 11-Layer AGCM

Diagnostic	YEAR 1			YEAR 2		
	JAN.	JULY	YEAR	JAN.	JULY	YEAR
Shortwave Albedo (%)	34.0	31.0	32.8	34.2	30.5	32.8
Outgoing Longwave Radiation (Wm^{-2})	232.4	239.2	234.7	232.0	240.2	235.0
Net Radiation (Wm^{-2})	1.5	-9.9	-4.2	1.2	-9.1	-4.4
Cloud Cover (%)	57.4	55.2	56.8	58.7	54.1	56.9

components of the radiation budget at the top of the atmosphere (the shortwave albedo and outgoing longwave radiation) show good correspondence with satellite data (Stephens et al 1981, Ohring and Gruber 1983, Jacobowitz et al. 1979, 1984). The annual-mean outgoing longwave radiation is within the range of the satellite measurements, but the mean albedo is higher than the observed value of about 30%. Care must be taken in such comparisons, however, as from physical considerations the annual-mean net radiation should be close to zero, whereas in the satellite datasets the components have not been adjusted to ensure that this is the case. The small negative value from the model would be removed by a reduction in the shortwave albedo of about 1%.

3.2 Precipitation

Maps of data from the first January of the integration are shown on the next few pages to illustrate some of the successes and failures in representing clouds and associated quantities. It is appropriate first to examine the precipitation distribution, as not only is this an important variable but it also shows some errors which are present even when prescribed cloudiness is used, which indicates that some of the deficiences in the cloud distributions shown later are due to shortcomings in other parts of the model rather than in the cloud-prediction scheme itself. The total precipitation (rainfall plus snowfall) is compared with Jaeger's (1983) climatology in Figure 1. The modelled distribution is generally good in equatorial regions, with the maxima and minima well positioned and comparable in magnitude to the observed. Note, however, that the minima over the eastern sides of the sub-tropical oceans do not extend far enough to the west. For example, the 1mm/day contour in Jaeger's data extends from central America almost to Hawaii, whereas in the model the precipitation rate is several times the climatological value by the time this longitude is reached. This is symptomatic of an important model error which will be discussed later. In middle and high latitudes there are again many satisfactory features, such as the minima over the continental areas and maxima over the north Atlantic and Pacific storm tracks. However, these are more intense than in the climatology and the Atlantic maximum extends too far across Europe and Asia. This is indicative of probably the most serious error in this version of the model when run at this horizontal resolution; excessively strong westerly winds in northern mid-latitudes in winter, associated with low surface pressure over the north pole and insufficient blocking over the northern continents (S.D.C.B. 1984, Slingo and Wilson 1984).

3.3 Cloud Cover

The amounts of high, medium and low layer cloud and the convective cloud diagnosed from the convection scheme are shown in Figures 2 and 3.

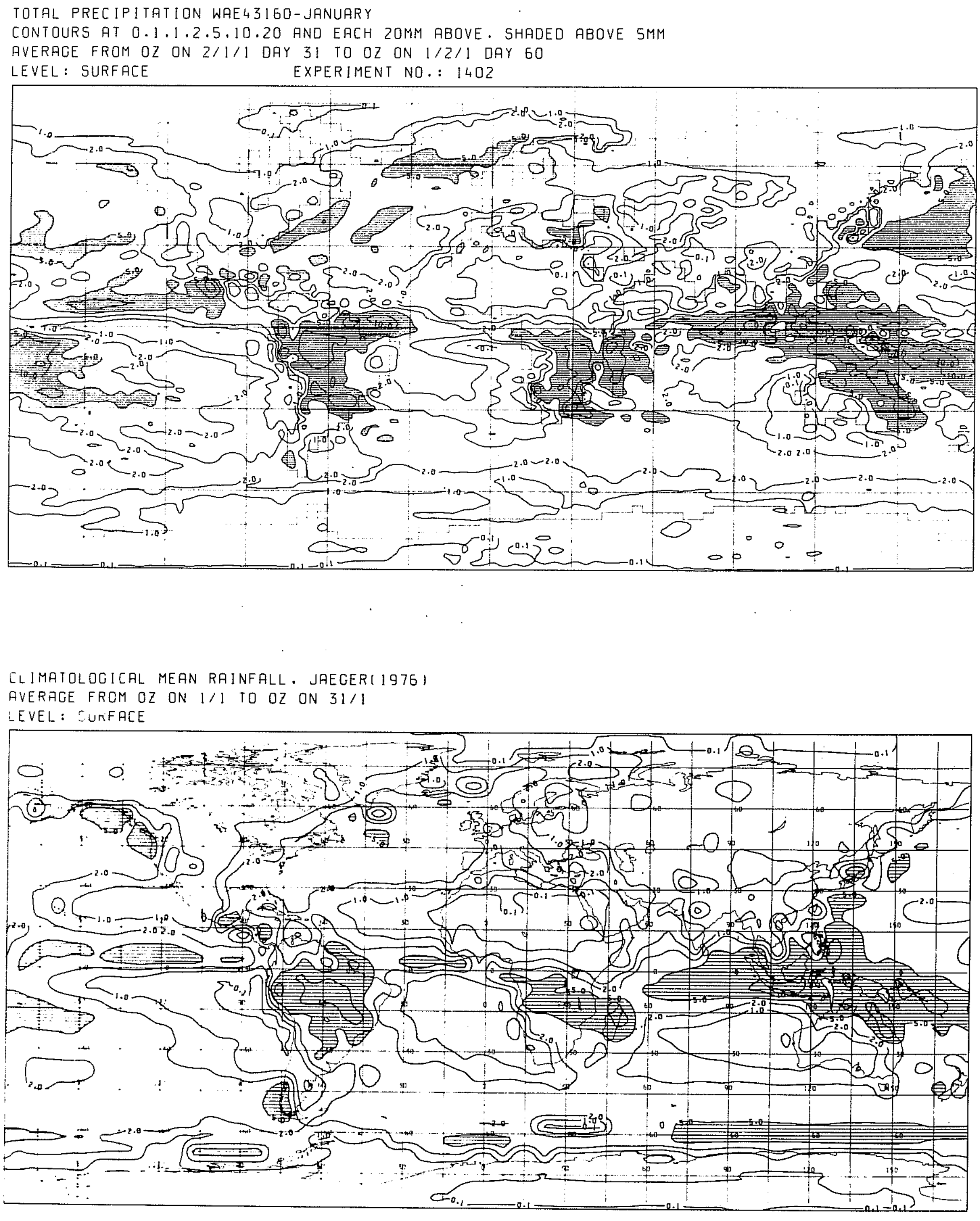


Fig.1. Distribution of Total Precipitation (Rain plus Snow) from the first January of a recent annual-cycle integration of the 11-Layer AGCM with interactive cloud and the climatological distribution derived by Jaeger (1983).

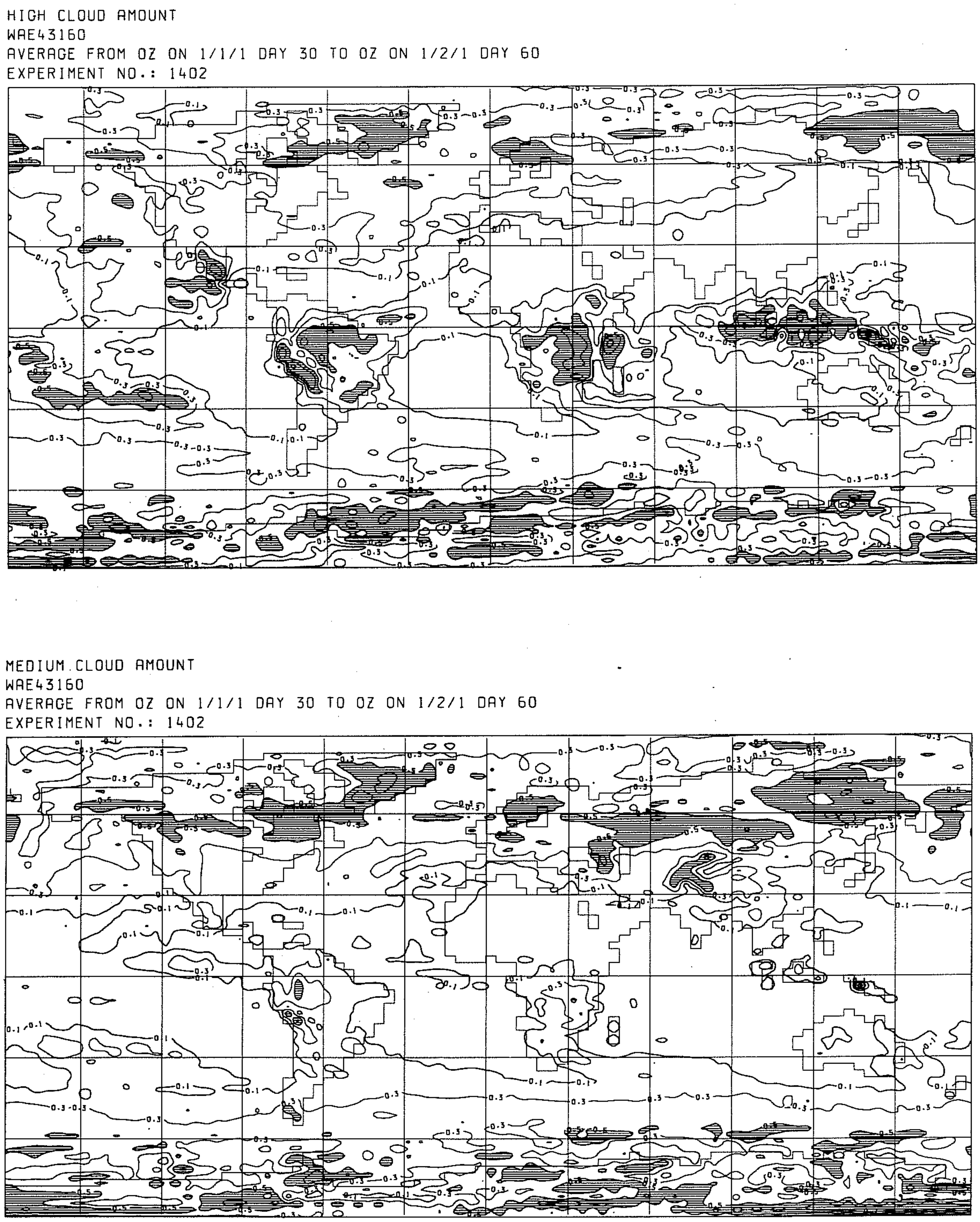
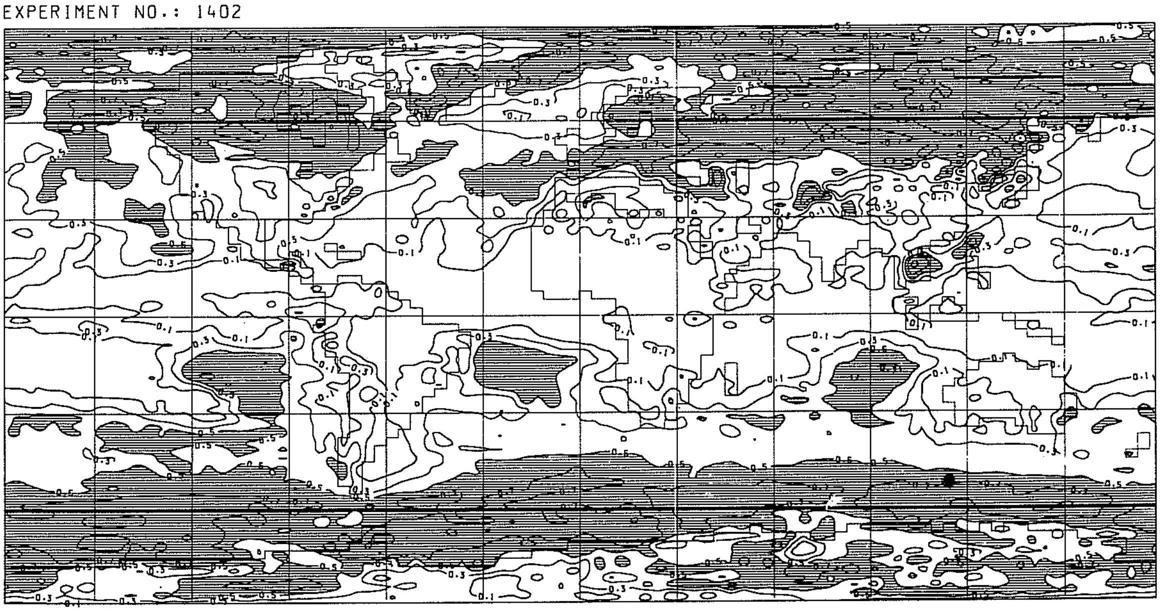


Fig.2. High and Medium cloud amounts for the same period as Fig.1. Contours begin at 10% with a contour interval of 20% and shading above 50%.

LOW CLOUD AMOUNT
WAE43160
AVERAGE FROM 0Z ON 1/1/1 DAY 30 TO 0Z ON 1/2/1 DAY 60
EXPERIMENT NO.: 1402



CONVECTION CLOUD AMOUNT
WAE43160
AVERAGE FROM 0Z ON 1/1/1 DAY 30 TO 0Z ON 1/2/1 DAY 60
EXPERIMENT NO.: 1402

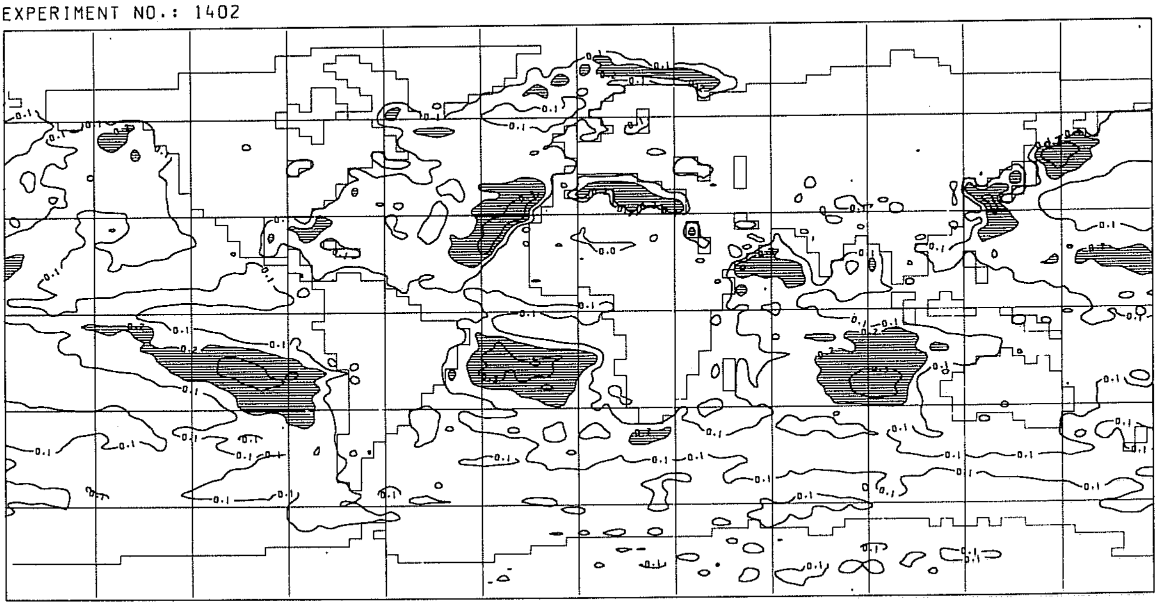


Fig.3. Low and Convective cloud amounts for the same period as Fig.1. Contours begin at 10% with for Low cloud a contour interval of 20% and shading above 50% and for Convective cloud a contour interval of 10% and shading above 20%.

Consider first the cloud distribution in the tropics. There is close correspondence between the high cloud and precipitation distributions from the model, as is observed. Detrainment from the convection scheme at upper tropospheric levels in areas of deep convection leads to high relative humidities and hence the prediction of layer cloud, representing the extensive cirrus seen on satellite images. Detrainment at middle levels from less vigorous convection leads to the prediction of medium cloud, which may co-exist with high cloud (as for example over southern Africa) or may be the dominant cloud type (e.g. over the eastern equatorial Pacific). In some areas, however, there are substantial precipitation totals with only small amounts of cloud predicted, such as over the Pacific Ocean at about 10°N 150°W and 10°S 165°E. It is also clear from Figure 3 that the convective cloud field completely misses the areas of deep convection and instead only picks out the shallow convective cloud over the oceans. Data for the point at 11.2°S 166.9°E illustrate both these shortcomings. The tephigram (Fig.4) shows instability throughout the troposphere and an average rainfall rate of over 26 mm/day. The upper troposphere is extremely moist, which is a characteristic feature of the convection scheme, leading to the prediction of about 40% cover of high cloud. However, the amounts of medium, low and convective cloud are each less than 10 per cent, which seem much too low. For such regions it is clearly physically reasonable to use the information from the convection scheme to calculate the cloud cover, but the method used in this integration predicts only small amounts of deep cloud. This is probably because, for various programming reasons, the convective cloud cover was diagnosed from a separate call to a version of the convection scheme from within the radiation scheme. Since this precedes the point at which the surface fluxes are applied to the lowest model layer, the magnitude of the diagnosed convective activity was in most areas much less than that actually occurring in the model. This also accounts for the lack of convective cloud over the land in Fig.3. The scheme is being modified to obtain the convective cloud amount directly from the call to the convection scheme used by the model.

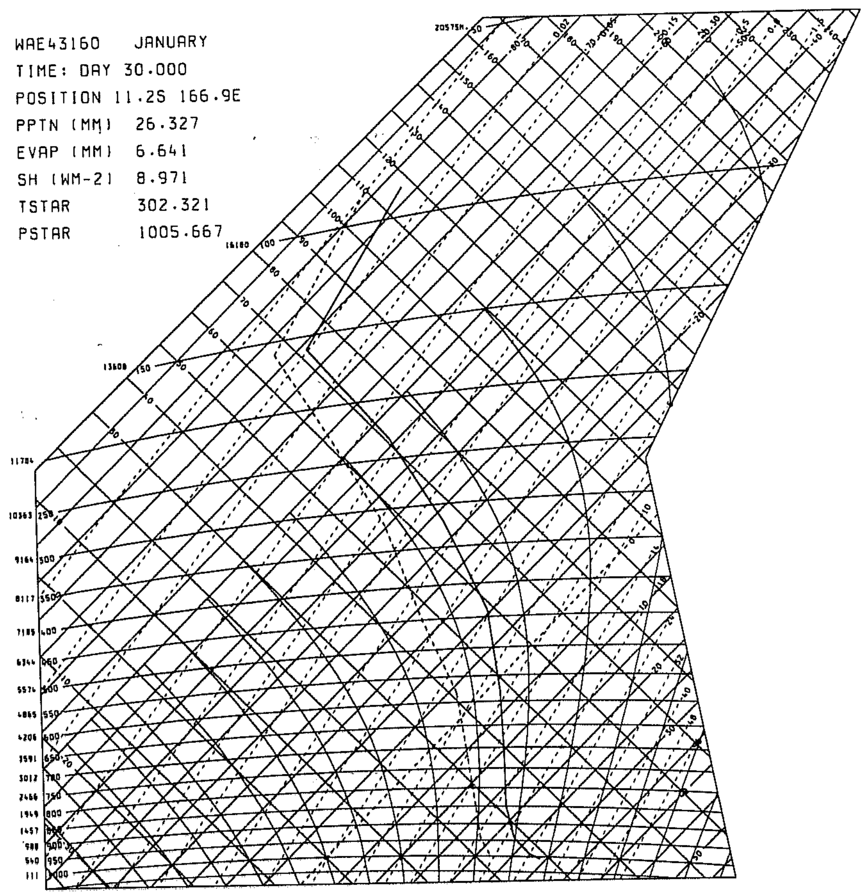


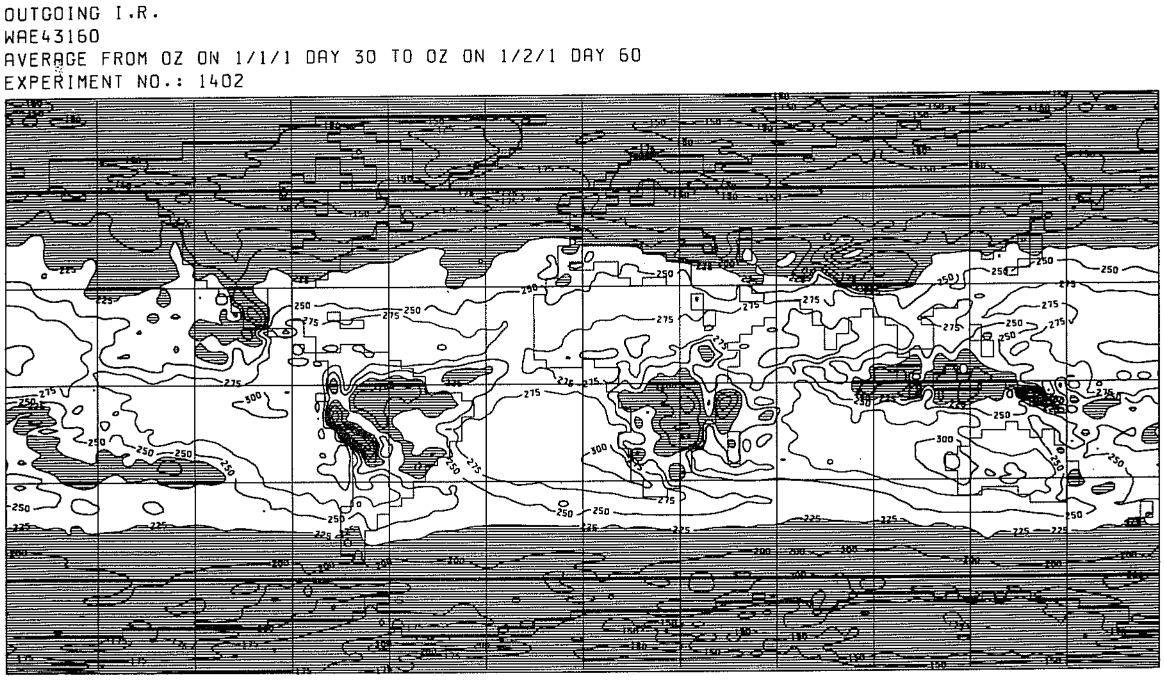
Fig.4. Tephigram from the model for the same period as Fig.1 for a point in the tropical east Pacific.

The distribution of low cloud over the sub-tropical oceans shows that the model is capable of maintaining extensive areas of boundary layer cloud, but these are positioned too far to the west compared with the observed locations over the cold ocean currents on the eastern sides of the Pacific and Atlantic oceans. This error is extremely important as will be discussed later. It will also be seen that in January there is too much low cloud over the northern continental areas, whereas in reality at this time of year there is less cloud there than over the oceans (Gordon et al. 1984, Figure 4a).

3.4 Radiation budget

Maps of the components of the radiation budget at the top of the atmosphere are shown in Fig. 5. At low latitudes, the patterns are strongly influenced by the cloud distribution. The outgoing longwave radiation compares well with the observed (e.g. Ohring and Gruber 1983), because in most areas of the tropics the modelled high cloud follows the precipitation distribution, although the regions noted earlier where there is too little high cloud are clearly visible. Over cloud-free areas of the tropical oceans the values exceed 300 Wm^{-2}, which is higher than the observed. This is probably due to the modelled humidities being too low, as comparisons with other longwave schemes for standard profiles show no such bias (Slingo and Wilderspin 1984b).

In contrast, the modelled albedo distribution at low latitudes is poor, with strong peaks over the oceanic low cloud maxima and minima where these peaks should be found. There is also very little evidence of maxima over the areas of tropical convection, despite the strong features in the outgoing longwave map caused by the high cloud. This is a result of the use of prescribed cloud radiative properties, the values being chosen to represent the "mean" properties for each cloud type. For high cloud the longwave emissivity is taken to be 0.75 and the shortwave albedo is 0.2. These values are consistent with the data reviewed by Stephens and Webster (1981), although it is known that the properties are highly variable, particularly for cirrus (eg Platt et al 1984). Observations during GATE showed that deep convection in the ITCZ is concentrated in cloud clusters, associated with which are extensive middle-to-upper level cloud shields



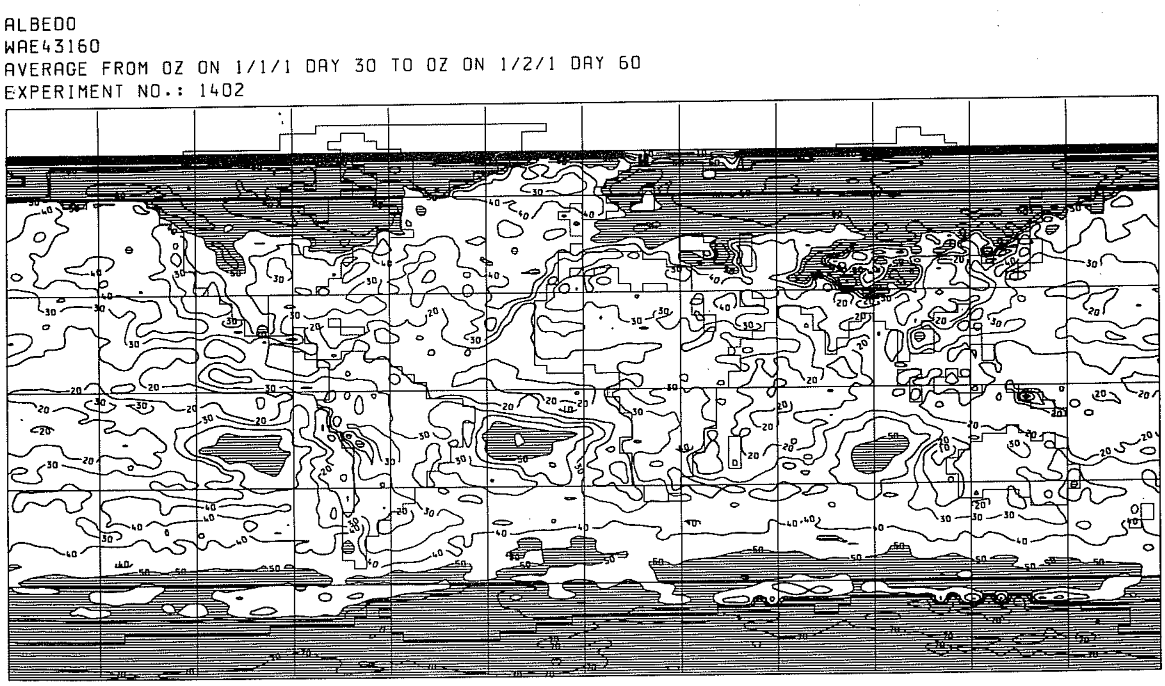


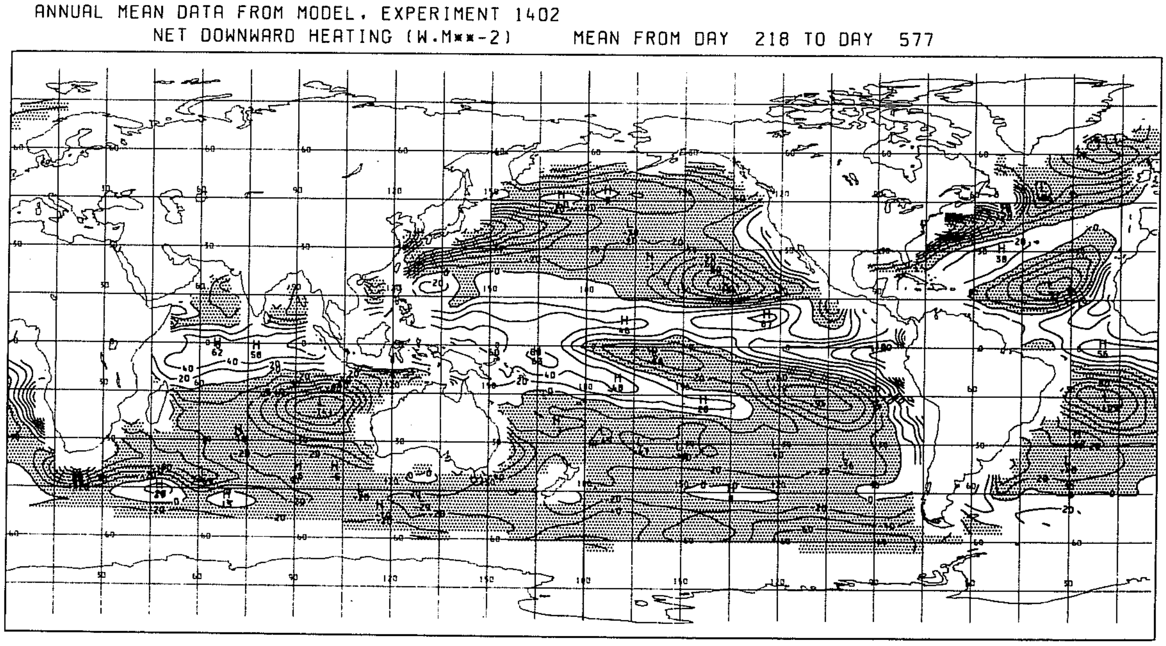
Fig.5. Outgoing Longwave Radiation (Contour interval 25 Wm^{-2} with shading below 225 Wm^{-2}) and Shortwave Albedo (Contour interval 10 per cent with shading above 50 per cent) at the top of the atmosphere for the same period as Fig.1.

("anvils") which may be several km deep (Houze and Betts 1981). It is of course these highly-reflecting anvils which appear on satellite visible images rather than the convective elements themselves. In order to represent these features, the model albedo for high cloud in the ITCZ should be much higher than 0.2. However, high values would not be appropriate for cirrus in the extra-tropics, where liquid (or ice) water contents are lower. This suggests that a re-formulation of the radiative properties in terms of the liquid (or ice) water content of cloud is required. The water content could be parametrized from the model variables or obtained from the approach described by Smith (1985). The latter is an attractive possibility as it represents a way in which the radiative, thermodynamic and turbulent properties of clouds could be represented in a consistent way throughout the model.

At high latitudes, the distribution of outgoing longwave radiation is in reasonable agreement with satellite data. The modelled values are up to about 25 Wm^{-2} lower than the NOAA values, but the latter are known to be too high by up to this amount at these latitudes (Ohring and Gruber 1983) and have been reduced recently following corrections to the analysis method (Janowiak et al. 1985). The albedo distribution is also generally satisfactory, although values are too low over the Atlantic Ocean north of about 50° latitude, due in part to unrealistically small amounts of low cloud (see Fig. 3).

4. LOW CLOUD OVER THE SUB-TROPICAL OCEANS

The importance of a realistic distribution of low cloud over the sub-tropical oceans from a model which is to be coupled to an ocean GCM is shown very clearly by Fig.6. This compares the estimates by Esbensen and Kushnir (1981) of the annual-mean net heating of the ocean surface with the data from one year of the annual-cycle integration. The observed estimates demonstrate that in the annual-mean the tropical oceans receive an excess of energy, which is transported by the oceanic general circulation to the high latitudes, where it is released to the atmosphere. The largest negative areas occur over the western boundary currents in northern mid-latitudes, but there are also minima due to the reflection of shortwave radiation to space by semi-permanent cloud features such as the sub-tropical stratocumulus in the eastern north Pacific and south-east



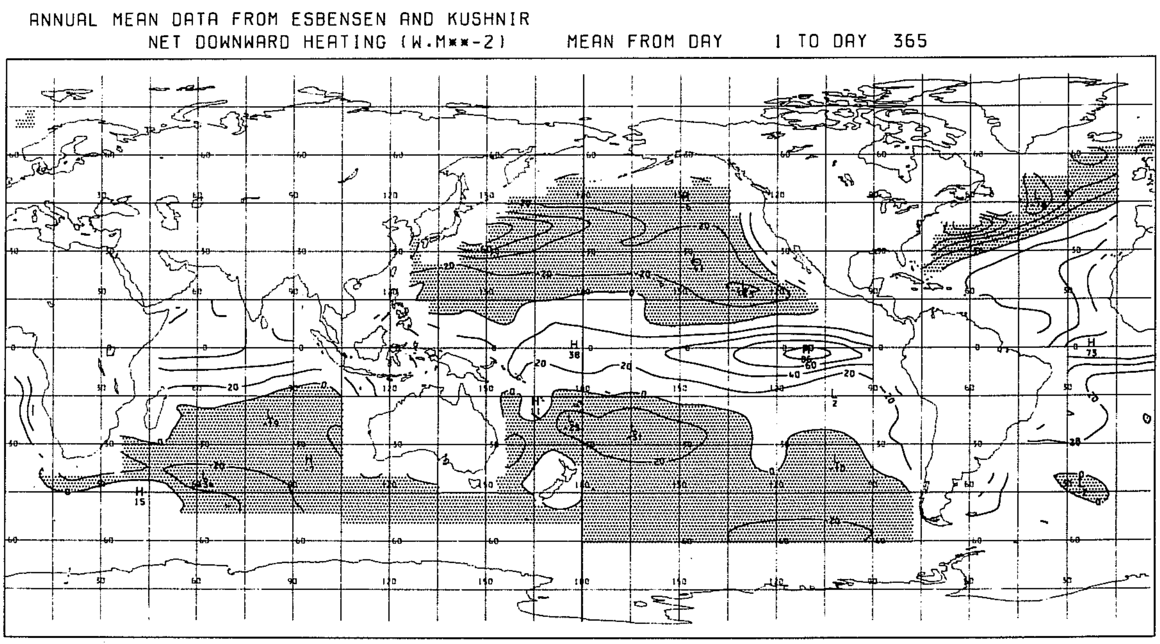
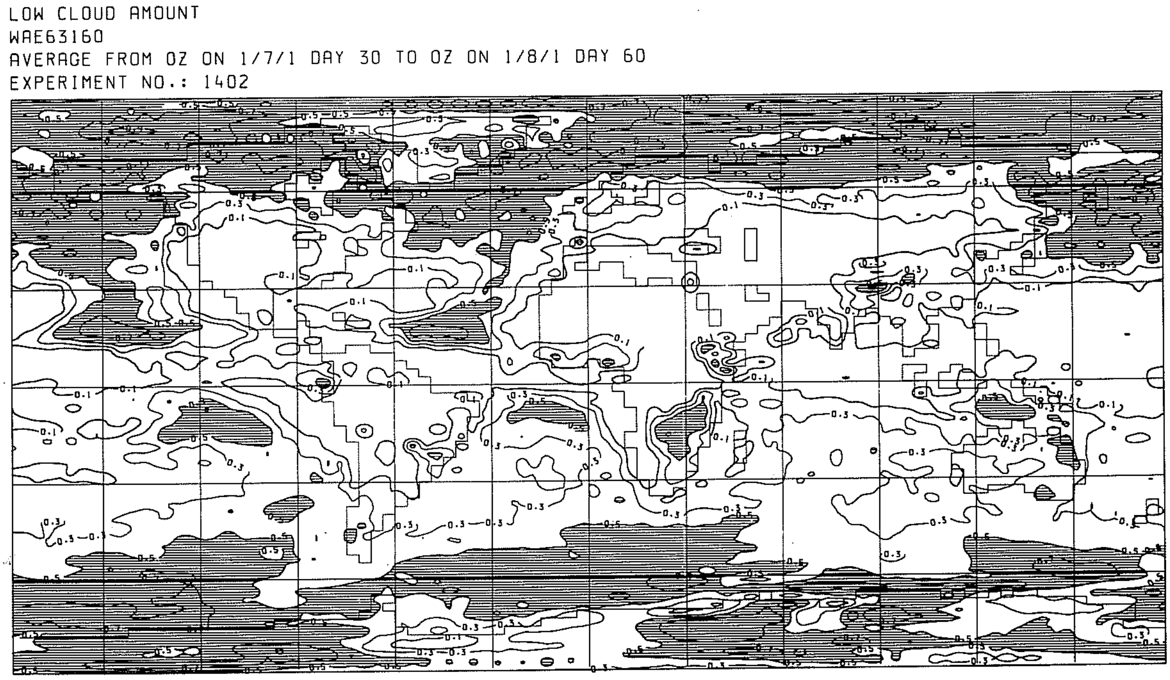


Fig.6. Annual-mean net surface heating (Shortwave heating minus Longwave cooling minus sensible and latent heat fluxes) from one year of model data compared with the estimates by Esbensen and Kushnir (1981). Contour interval is 20 Wm^{-2} with negative areas (i.e. net surface cooling) shaded.

Atlantic and the cloud in the South Pacific Convergence Zone. In the model, the distribution is seriously distorted by errors in the low cloud field. In the southern hemisphere the effects of the maxima in the low and convective cloud fields (Fig.3) are clearly visible. In the northern hemisphere the maxima in the low cloud amount in July (upper map on Fig.7) also have counterparts in errors in the net surface heating. The pattern of surface heating is altered so radically that if this version of the model were to be coupled to an ocean GCM one would expect large errors in the sea surface temperature field, with consequent errors in both the oceanic and atmospheric general circulations. Efforts to improve the cloud distribution thus have a high priority in current research and some work in progress will now be discussed.

The boundary layer scheme used in this integration treats neither the exchange of mass between the boundary layer and the free atmosphere by turbulent entrainment nor the variations in boundary layer depth caused by this and other processes (Carson 1982), whereas both are known to be important. This has been replaced by a more general scheme (Richards 1980), also used in the operational numerical weather prediction model (Dickinson and Temperton 1984), in which the number of model layers within the boundary layer is allowed to vary. Turbulent entrainment is not included in this scheme, but analysis of the model data indicates how important this might be in influencing the distribution of boundary layer cloud. The lower map on Figure 7 shows the difference between the equivalent potential temperature (θ_e) for the model layer just below the top of the boundary layer and that for the layer just above. Values lower than -2K are shaded, the significance being that this is (roughly) the criterion for Cloud Top Entrainment Instability (Deardorff 1980, Randall 1980). This occurs when a parcel of inversion air above a cloud layer is sufficiently dry that when it is entrained into the cloud the resulting evaporation of cloud water into the parcel cools it so much that it becomes negatively buoyant with respect to the surrounding air, thus promoting further entrainment which can dry out the boundary layer and dissipate the cloud. Comparison of the two maps on Figure 7, which are for the same period, shows that the areas of excessive low cloudiness are highly unstable to this mechanism, so that if it were included one might expect



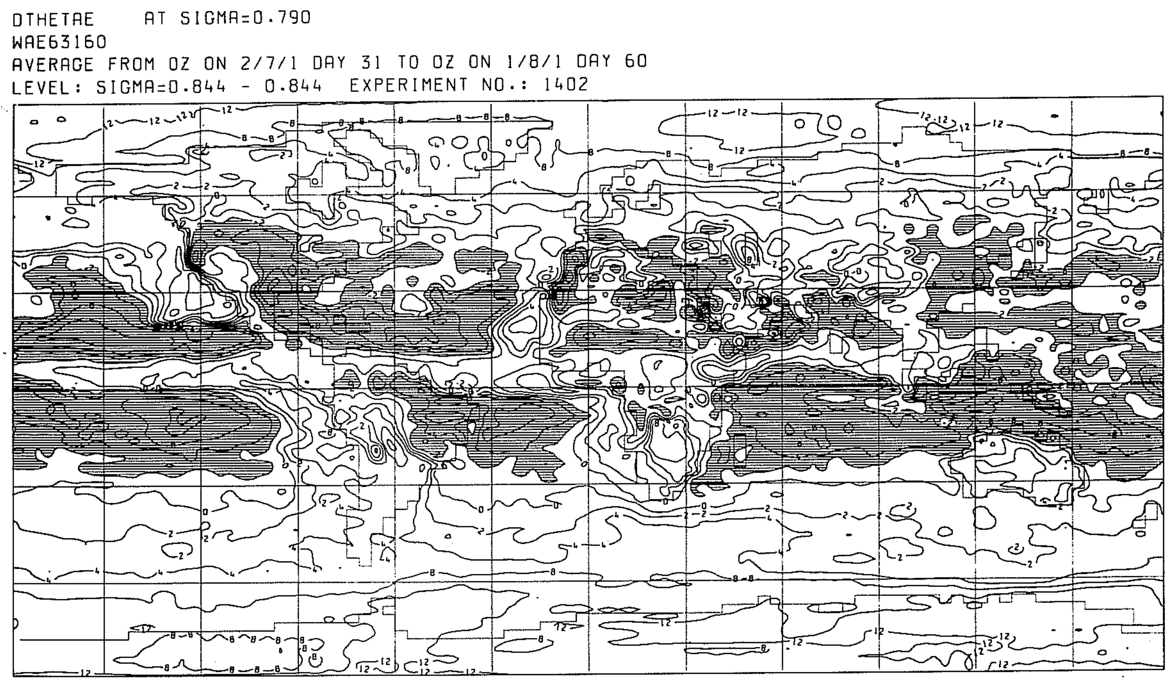


Fig.7. Low cloud amount (Contouring as in Fig.3) from the first complete July of the model integration and the difference in equivalent potential temperature between the model layer just below the top of the boundary layer and that just above (Contour interval 2K with values lower than -2K shaded).

the model to produce less cloud in these regions. The boundary layer in these areas is certainly too moist with precipitation amounts in excess of 2 mm/day (Fig.1), due not only to large-scale saturation at the top of the boundary layer but also to shallow, non-penetrative convection (hence the correspondence between the low and convective cloud fields as shown by Fig.3). It is noteworthy that in the UCLA model this mechanism is an important factor in controlling the distribution of boundary layer cloud (Randall et al. 1984).

The recognition of the importance of cloudy processes in turbulent transport in the boundary layer has led to a project in which the model is being re-formulated in terms of variables which are conserved during phase changes (Smith 1985). One advantage of this approach, as shown by Smith, is that entrainment instability is implicitly accounted for without the need for further parametrization.

Given that entrainment instability may be important in limiting the extent of stratiform clouds, a simple formulation is being tested both in the full model and a single column version. However, it would obviously be wrong to use a layer cloud instability mechanism to remove the spurious maxima in the low cloud field, as in reality at these locations the cloudiness is not stratiform but trade-wind cumulus. Various observational papers have demonstrated the importance of cumulus clouds in determining the structure of the trade-wind boundary layer (e.g. Augstein 1979). These transport boundary layer air into the free atmosphere and vice versa and thus prevent large-scale saturation from occurring. Tiedtke (1984) has shown that including a simple parametrization of such "shallow convection" in the ECMWF model leads to substantial improvements not only in the local thermodynamic structure but also in the extra-tropical flow. One might have expected that the convection scheme used in the 11-Layer Model would also treat this convective interchange, given the significant amounts of shallow convective cloud shown on Fig.3. However, a feature of this scheme (Rowntree 1984) is that when convective parcels encounter a stable layer, such as the Trade-Wind inversion, which they cannot penetrate to rise to the next model level, they detrain completely (with slight excess buoyancy) at the lower level. Shallow convection within the boundary layer thus leads to no convective interchange with the free atmosphere if such an inversion is present, so the boundary layer remains too moist. A minor modification to the scheme to model such interchange by allowing a fraction of the parcel to reach the next level is being tested. Preliminary results are encouraging and suggest that with this modification the scheme may be able to simulate the effects of Trade-Wind cumulus on the boundary layer. With a drier boundary layer the amounts of stratiform cloud diagnosed in these regions should thus be reduced substantially.

5. CONCLUDING REMARKS

In the previous section virtually no reference is made to the cloud-prediction scheme. Instead, shortcomings in other physical parametrizations are identified as the reasons for errors in the low cloud fields and solutions sought. Our attitude has been that only when no such errors can be found and no significant differences between the modelled and observed atmospheric structures identified is it legitimate to tune the cloud prediction equations so that the model reproduces the observed cloud fields.

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