CLIMATOLOGY OF THE ATMOSPHERIC CIRCULATION SIMULATED BY THE NCAR COMMUNITY CLIMATE MODEL

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1. INTRODUCTION

The NCAR Community Climate Model is derived from the spectral general circulation model developed by William Bourke and collaborators. The model is described in Bourke et al. (1977) and also in McAvaney et al. (1978), in which its original climate simulation properties were discussed. Change have been made in the radiation and cloud parameterisations and in the topography in the model. Some coding changes have also been made to increase the computational speed on the CRAY-1 computer. The changes in physical processes are discussed in Pitcher et al. (1982) and in Ramanathan et al. (1982).

Using fixed boundary conditions (sea surface temperatures, albedo, soil moisture, etc.) two 1200-day simulations have been run, one each for perpetual January and perpetual July conditions. Most of the material I will discuss today is an average of 120 days from the first 200 days of these runs. More details may be found in Pitcher et al. (1982).

Table 1 is a list of some of the model's characteristics. Fig. 1 shows the topography used in the model. This topography was chosen after a modest amount of experimentation. Details of the derivation of topography can be found in Pitcher et al. (1982).

Table 1

Summary of Model Characteristics

- SPECTRAL DYNAMICS
  - M=15 TRUNCATION (48x40 GRID)
  - SEMI-IMPLICIT TIME INTEGRATION SCHEME
  - VECTORIZED FFT
- 9 VERTICAL LEVELS, σ COORDINATES
- REALISTIC LAND-SEA-ICE DISTRIBUTION
- REALISTIC TOPOGRAPHY (SMOOTHED)
- NCAR RADIATION PACKAGE
  - SOLAR AND IR RADIATIVE TRANSFER
  - PREDICTED CLOUDINESS
  - SURFACE ALBEDO VARIES GEOGRAPHICALLY
- CONVECTIVE ADJUSTMENT AND CONDENSATION
- BULK PARAMETERIZATION OF SURFACE STRESS, SENSIBLE AND LATENT HEAT
- HORIZONTAL AND VERTICAL DIFFUSION
- SURFACE TEMPERATURE OVER LAND FROM ENERGY BALANCE; FIXED OVER OCEAN
- NO SURFACE HYDROLOGY

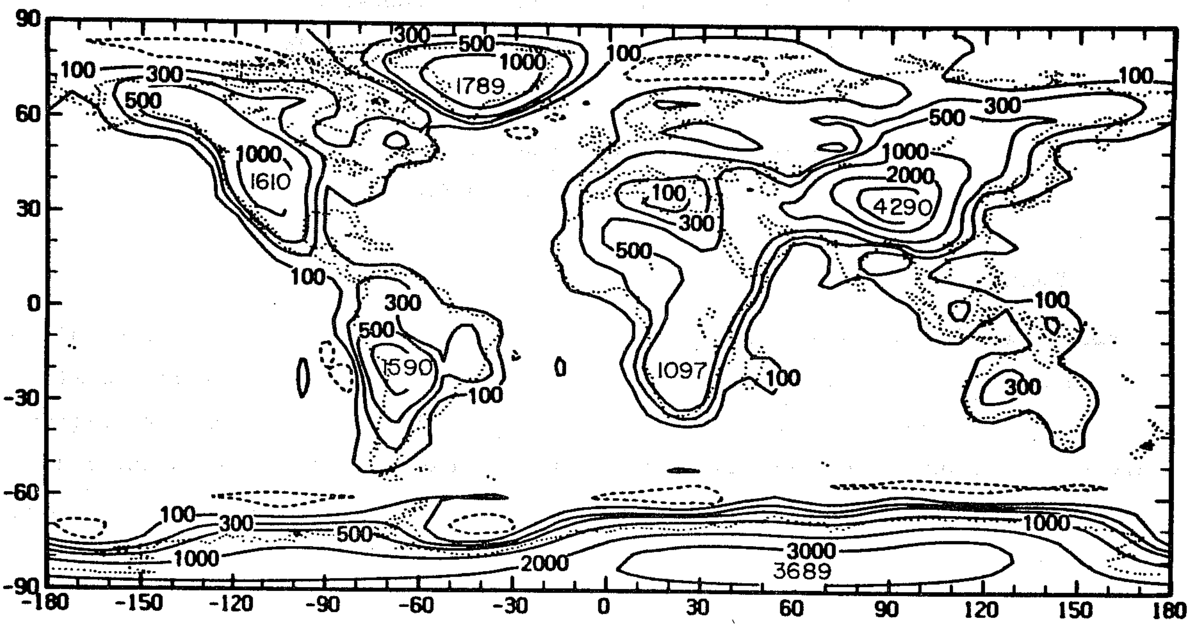

Fig. 1. Topographic heights (m) used in the model simulations.

2. SIMULATION OF AVERAGE FIELDS

In Fig. 2 are shown the January and July simulations of the zonal average of the zonal wind, plus corresponding observations from Newell et al. (1972). The mid-latitude tropospheric jets generated by the model are well positioned in both latitude and height. The winter hemisphere jets are of the same magnitudes as observations, while the summer hemisphere jets are too weak by about 5 m s^{-1}. A notable improvement over previously reported simulations is the polar night stratospheric jet. This jet is distinctly separated from the tropospheric jet, and near 30 km, the jet maximum in January is about 30 m s^{-1}, compared to the observed value of about 35 m s^{-1}. Typical previous results, such as in McAvaney et al. (1982) give a value of about 70 m s^{-1} in this region. The reasons for this improvement are discussed in detail in Ramanathan et al. (1982). There are deficiencies in the simulations also, most notably the excessive speeds of the tropical easterlies by about 5 m s^{-1} near the surface and of about 10-15 m s^{-1} in the stratosphere.

In Fig. 3 we show the zonal average of the temperature for January and July and observations from Newell et al. (1972). The January polar stratosphere temperatures are perhaps 5K warmer than observed, in contrast to typical previous results where simulated temperatures are 20-25K too cold. This is, of course, connected to the improvement in the zonal wind simulation from the thermal wind equation. The temperatures generated by the model, however, are 3-7K too cold near the tropical tropopause and in the troposphere.

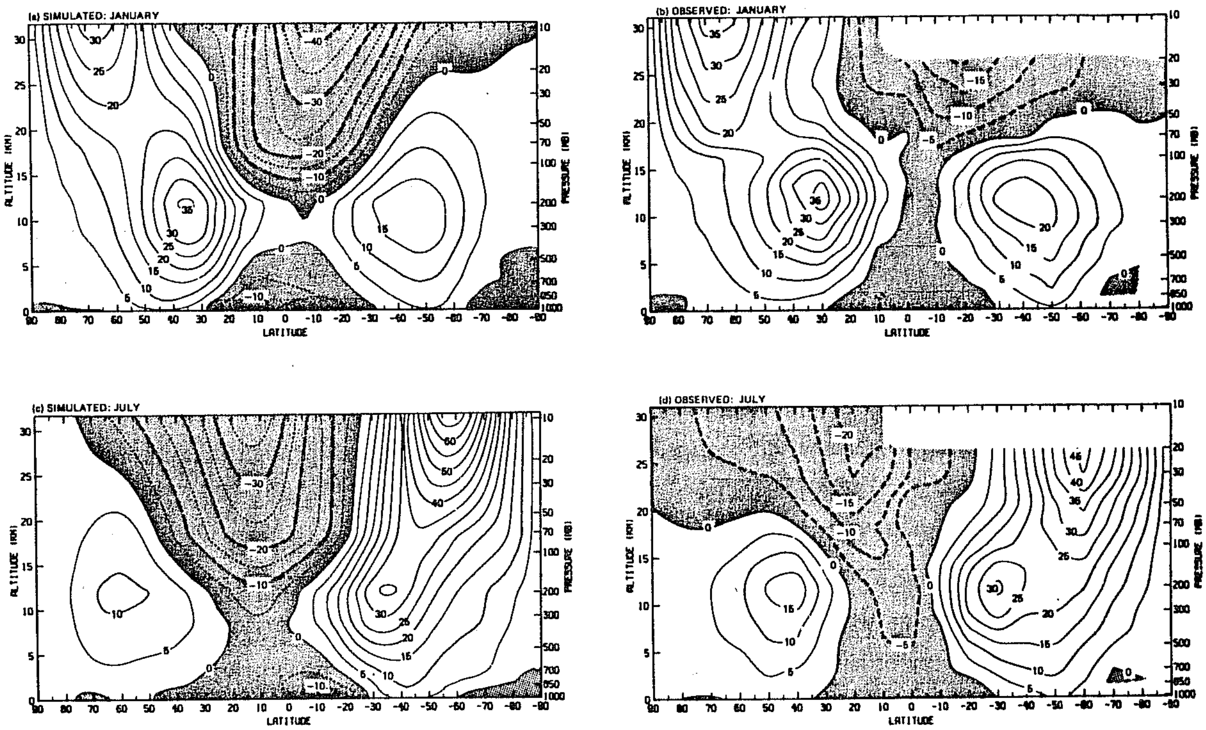

Fig. 2. Latitude-height section of the zonal average of the zonal wind (m s^{-1}) for January and July.

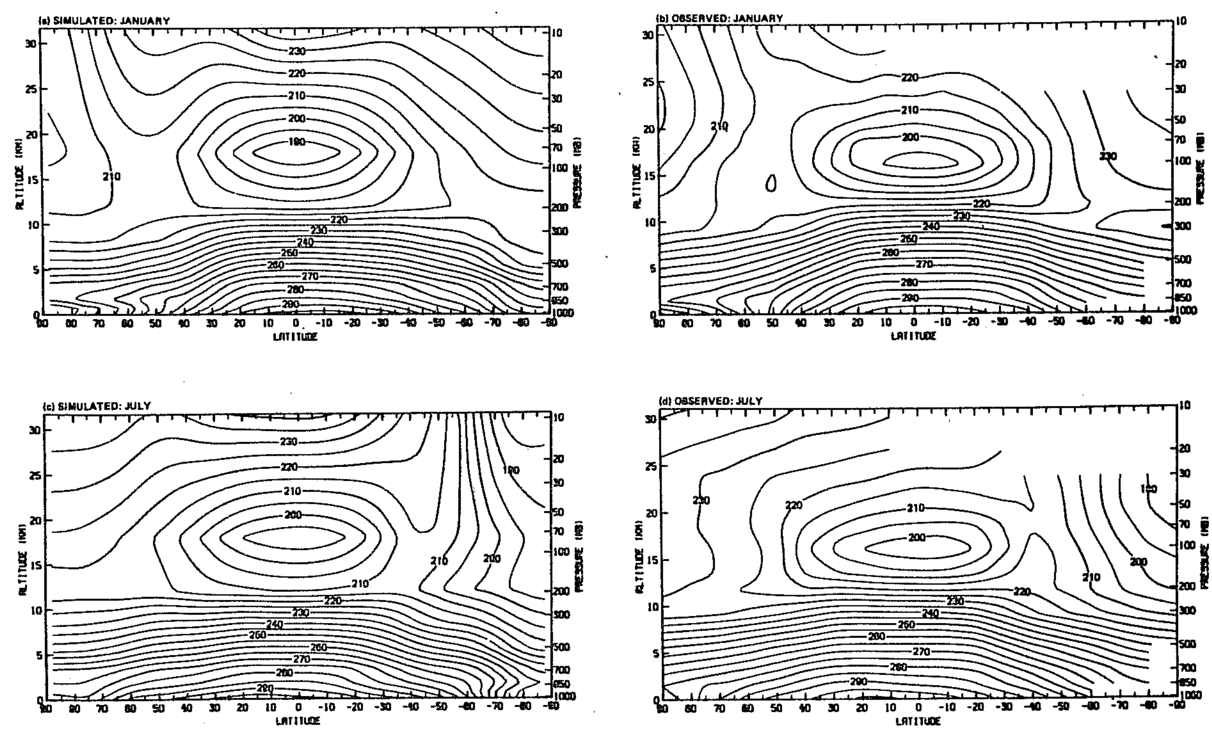

Fig. 3. Latitude-height section of the zonally averaged temperatures (K) for January and July.

Fig. 4 compares the simulated sea level pressure fields with observations from Shutz and Gates (1971, 1972) for January and July. The January simulation is successful in capturing most of the features appearing in the climatological chart. The strength and position of the Icelandic low compare favorably with the observations, with its simulated eastwards extension somewhat south of the observed trough axis. The Aleutian low is well simulated, appearing near the international dateline as a single centre of low pressure, while the observations suggest an elongated minimum in the pressure field centred about this mean position. The Siberian high present in the model is well positioned in longitude and about 7° south of the observed centre. It is too intense by about 5 mb. The westwards extension of the high-pressure ridge into the Middle East and North Africa is not in agreement with the observations. (A similar feature may be seen in McAvaney et al., 1978 and Manabe and Hahn, 1981.) The model does capture the broad equatorial belt of low pressure, and positions the subtropical high-pressure systems to the east of continents in the Southern Hemisphere. The climatological low-pressure troughs over South America and Africa are evident in the simulation but to a much lesser extent. We believe that this last deficiency is a direct result of the prescription of surface evaporation and shall return to this in the next section. The model properly locates the circumpolar Antarctic trough, but underestimates its intensity by some 10-15 mb.

In the July simulation the centres of the North Atlantic and Pacific high-pressure systems are properly located west of the principal land masses, but the centres are 5-10° too far north and 5 mb too high. The model generates an Arctic zone of high pressure which is not observed. The Indian monsoon is well developed in the model, perhaps even more vigorous than the observations would indicate, while the centre of the low is displaced several degrees to the east of its climatological position. Over South America and Africa are found regions of excessive high pressure in the simulation. These features appear to be counterparts to the continental subtropical high pressures occurring in the Northern Hemisphere January simulation, and their existence remains unexplained. Central pressures associated with the Antarctic circumpolar trough are closer to that observed in the July simulation, yet the centres are about 7° too far north.

In Fig. 5 is shown the simulated and observed surface air temperatures for January and July. Since the ocean surface temperatures are fixed, we expect to find close agreement between observations and simulations over the oceans. However, the temperature of land surface is calculated and here we find surface air temperatures to be systematically too cold. This is true even in the high Arctic in the January simulation where the temperature is 5-10K too low. Other studies (McAvaney et al., 1978 and Manabe and Hahn, 1981) using similar models show temperatures in this region to be too warm by a comparable amount.

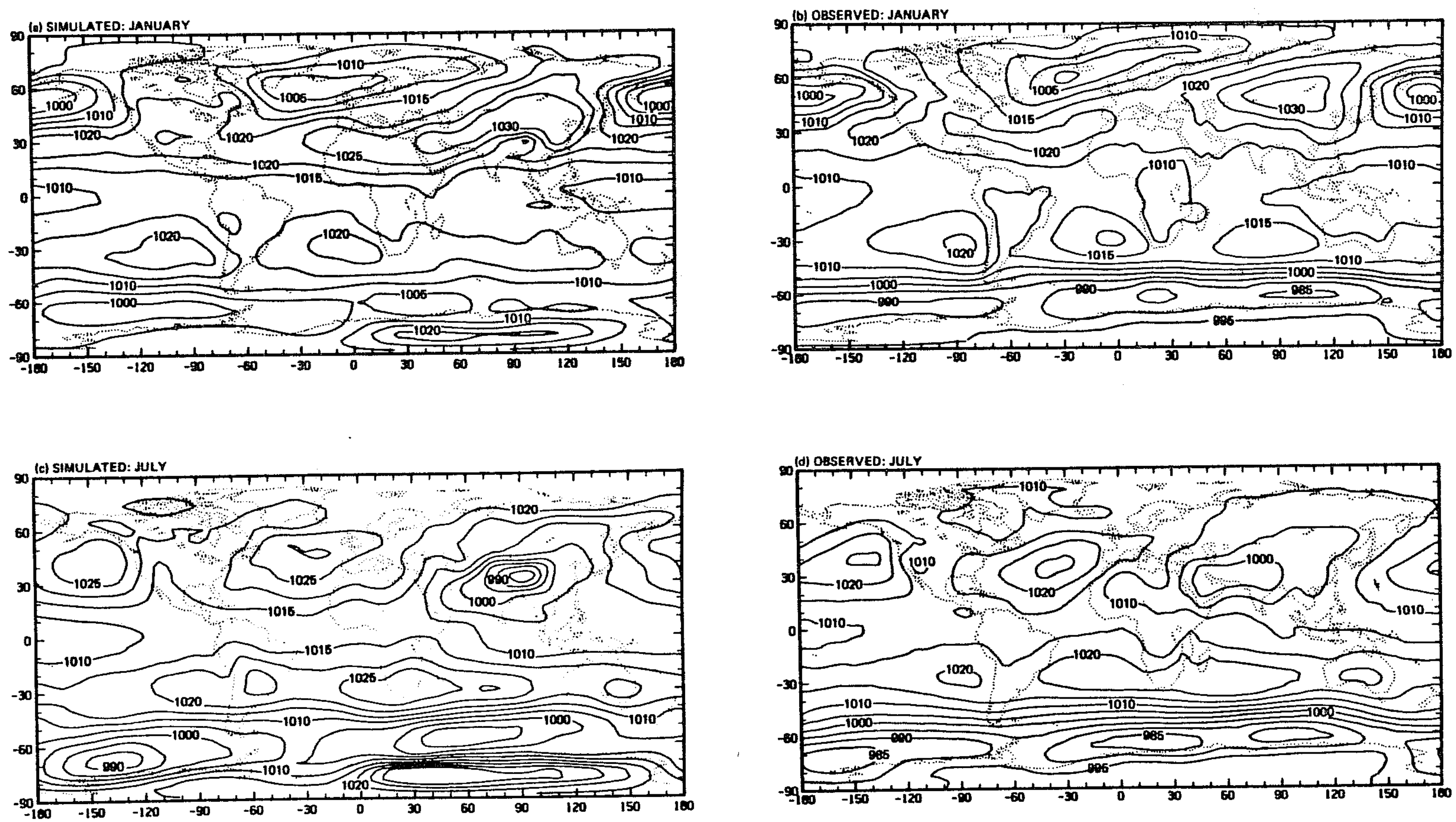

Fig. 4. Comparison of simulation and observed sea level pressure (mb) for January and July.

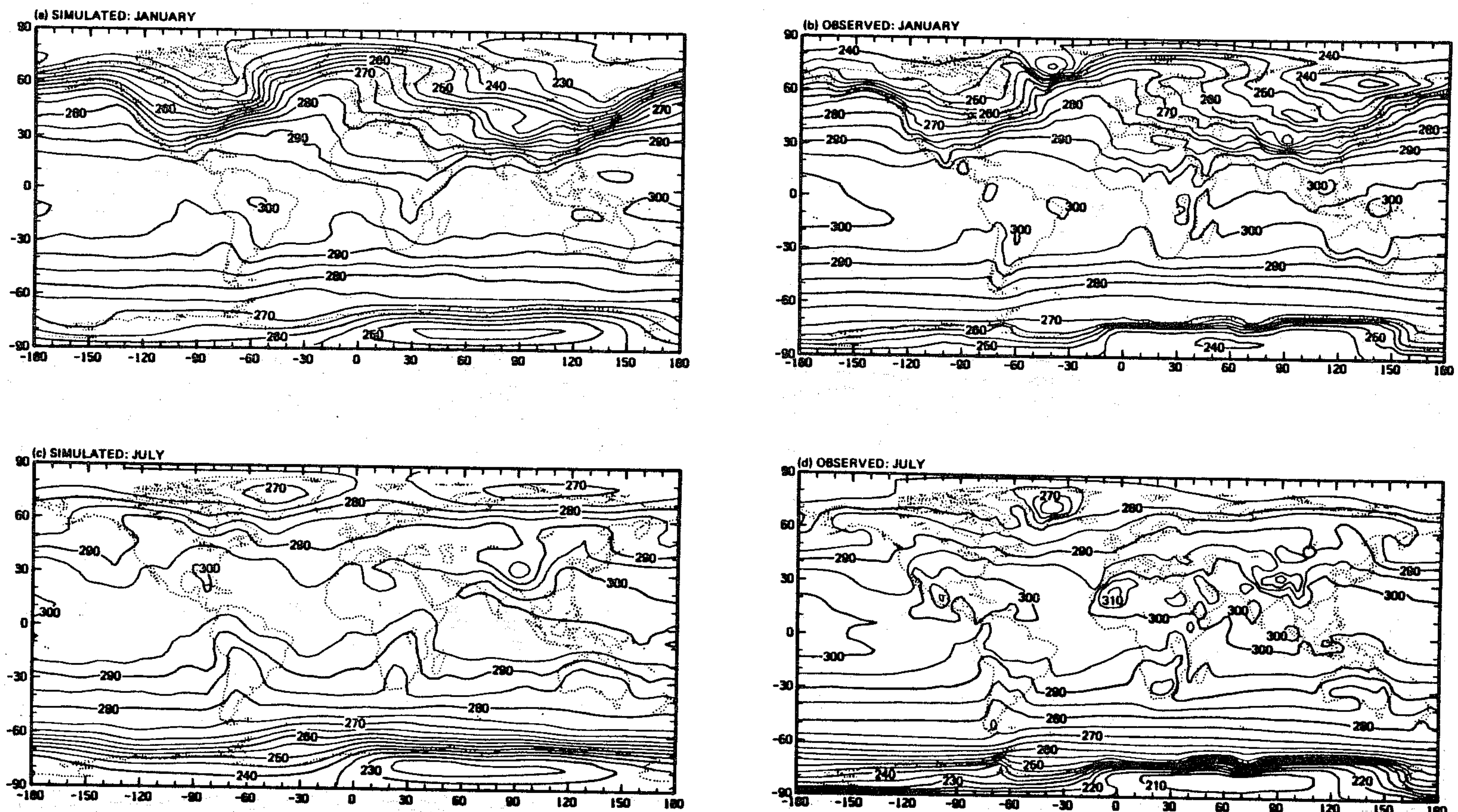

Fig. 5. Comparison of simulated and observed surface air temperatures for January and July.

One of the model characteristics which produces the biases in the surface temperature calculation is the use of a constant "wetness factor." Evaporation for the land surface is taken to be 0.25 times the value that would have resulted had the underlying boundary been ocean instead. This leads to evaporation which is too large over arid regions and too small over tropical rain forests. A simple experiment was performed which set the "wetness factor" to zero in the latitude belt 11-39° in both hemispheres. This produced increases in the surface air temperatures over land surfaces in the affected regions of the summer hemisphere of 7-10K. This suggests that the simulations of surface temperatures could be improved by making the "wetness factor" a more realistically geographically varying field. Such "tuning" has not been attempted.

In Fig. 6 we show the mean January 500 mb geopotential heights for the Northern Hemisphere for the model and observations (Lau et al., 1981). The large-scale features are successfully simulated, with major troughs over the east coasts of North America and Asia and a weaker trough over Europe. The main deficiency in the simulation is that the 500 mb surface is systematically too low, consistent with the colder-than-observed tropospheric temperatures previously mentioned.

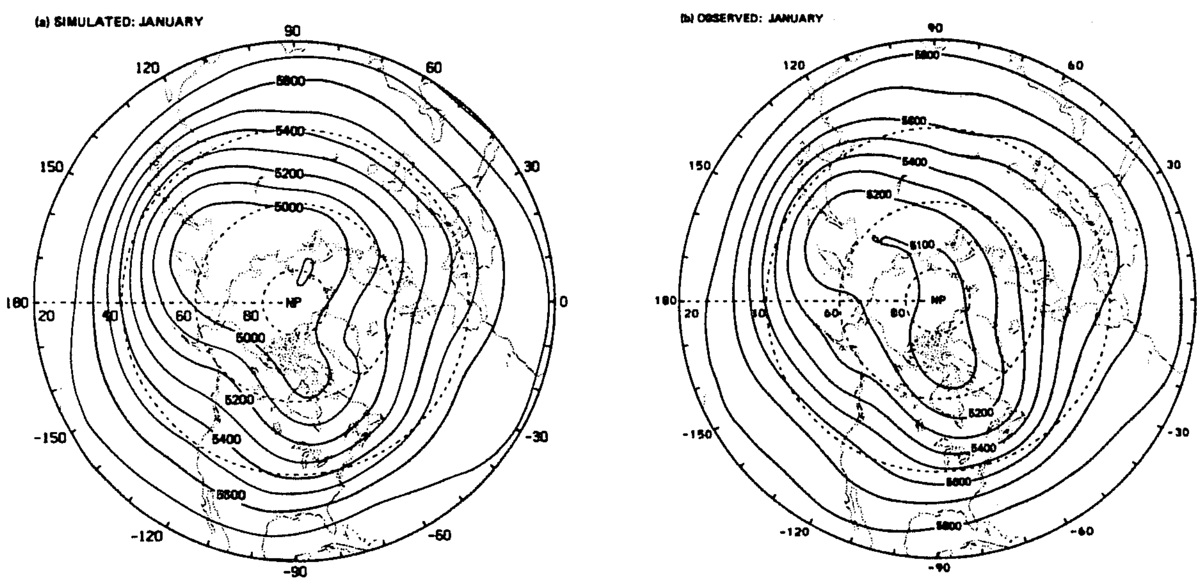

Fig. 6. 500 mb geopotential heights (m) for January, simulated and observed.

Fig. 7 compares the 300 mb mean zonal winds for Northern Hemisphere winter, model and observations. Agreement between model and observations for both the position and magnitude of the jets is quite good. Lau (1978) has shown that the dominant term in the local maintenance of the mean zonal wind is the mean meridional ageostrophic wind. Comparison of this quantity (not shown) also reveals good agreement between model and observations.

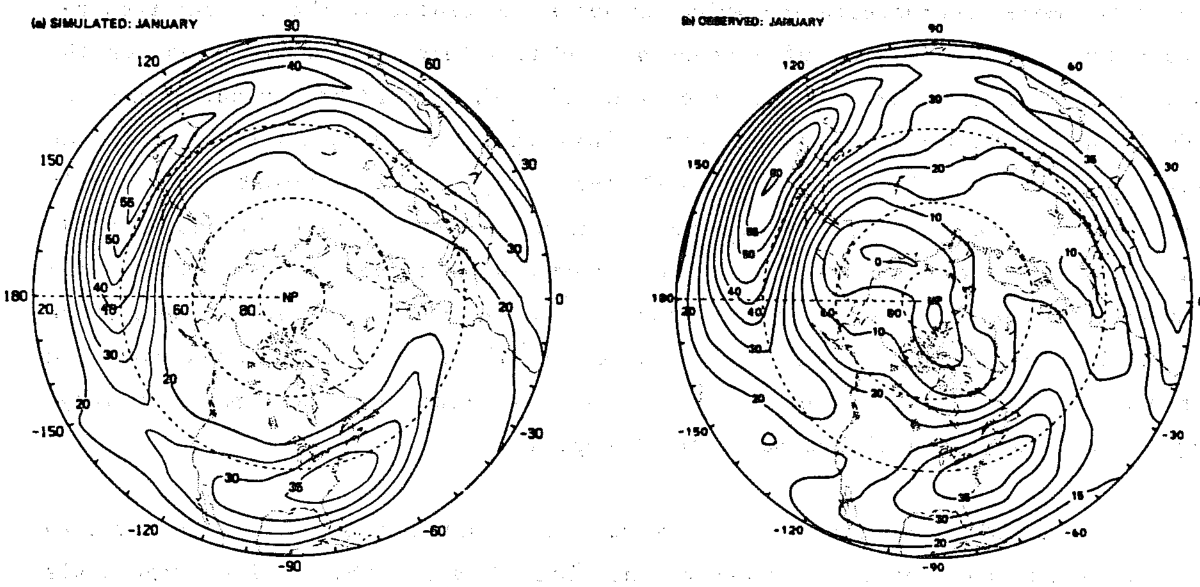

Fig. 7. 300 mb mean zonal wind (m s^{-1}) for January, simulation and observations.

Since the zonal average simulations of u and T are quite good in the model stratosphere, Figs. 2 and 3, we will show three comparisons of regional fields to confirm the quality of the model simulation. Fig. 8 shows the mean 100 mb temperature for January over the Northern Hemisphere. The simulated temperature fields in the troposphere generally show the (approximately) correct wave structure, but are too cold. At 100 mb, however, the model temperatures are now much closer to observed (Lau et al., 1981). The model also does a creditable job in simulating the midlatitude warm belt and the near polar minimum.

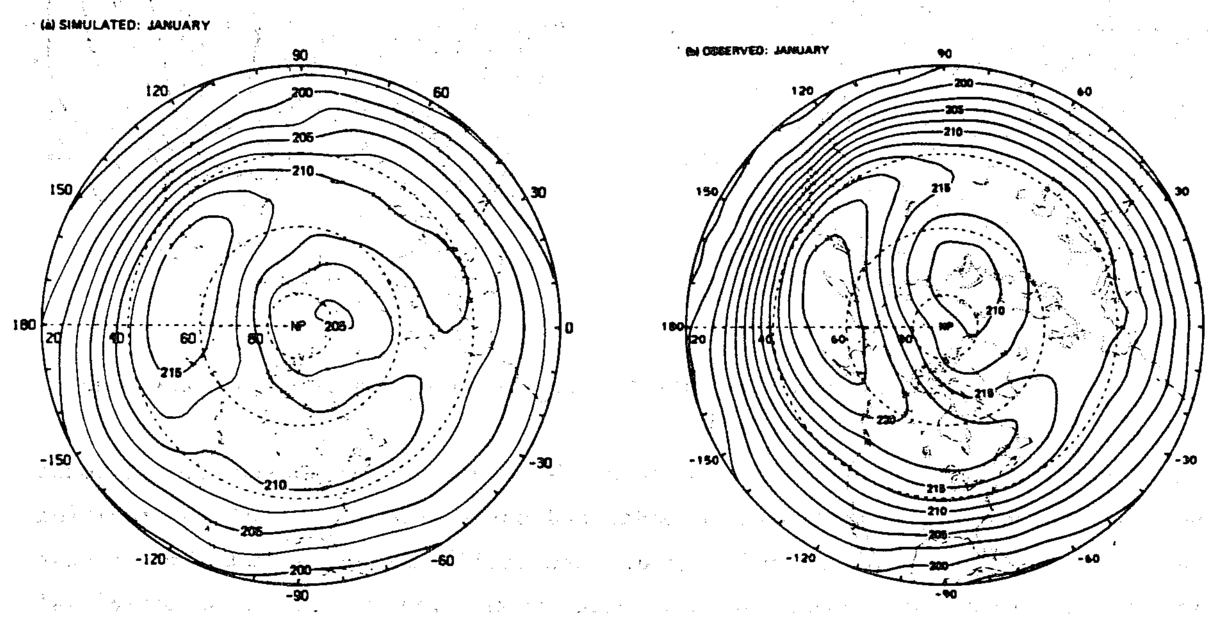

Fig. 8. 100 mb mean temperatures (K) for January, simulation and observations.

We show in Figs. 9 and 10 the mean temperature and mean resultant wind at 10 mb for January. The observations are from Labitzke et al. (1972). This pressure level is at the top of the model, so one makes this comparison somewhat skeptically, but also, perhaps, somewhat charitably. The comparison is, indeed, surprisingly good. The phase of the mean temperature waves, Fig. 9, are quite good, with a minimum temperature near 0°E and warmer temperatures around 110-120°E. The mean wind is in close agreement with observations, both in magnitude and position of the major features. Examination of these fields at other pressure levels shows that the model does not get the transition from the tropospheric jet structure at 200 mb to the polar night jet structure at 10 mb at exactly the right altitude. Given the limited resolution in the model stratosphere (see Table 1), this is not surprising. However, the success of the simulation at 10 mb suggests that the limited vertical resolution in the stratosphere is sufficient to allow simulation of the gross features of the stratospheric circulation.

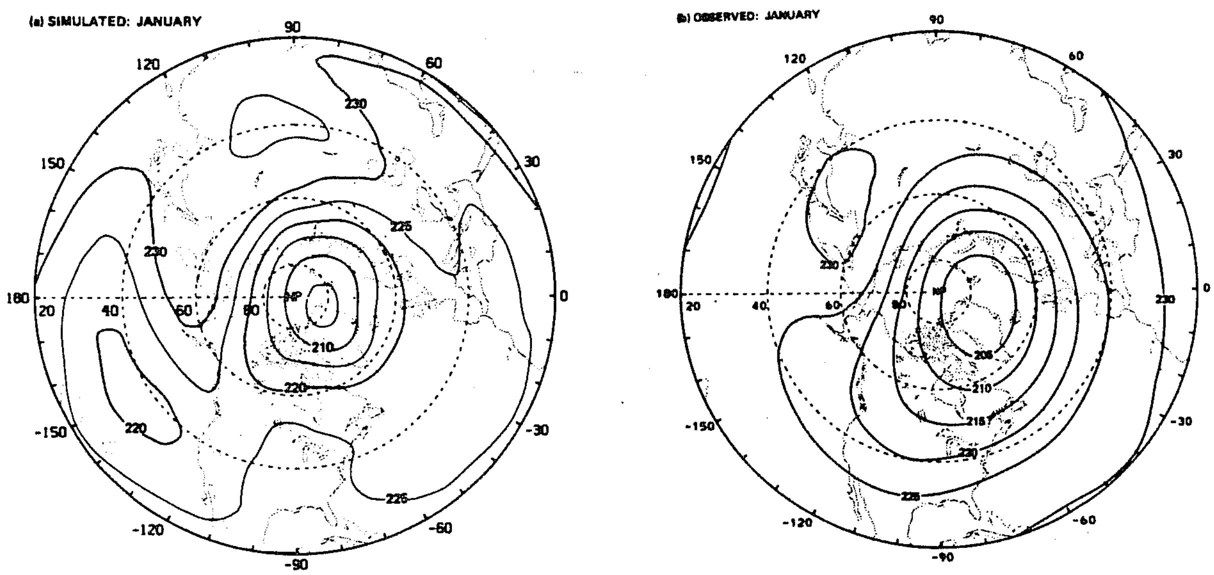

Fig. 9. 10 mb mean temperature (K) for January, simulation and observations.

3. SIMULATION OF STANDARD DEVIATION FIELDS

We turn now to the characteristics of the fluctuation of the model simulation. In Fig. 11 we show the latitude-height distribution of the standard deviation of geopotential height for January and July. Observations are taken from Oort (1982). In the January simulation we see that the wintertime midlatitude variability is quite close to observations although it is a bit weak in the lower stratosphere. The variability in the tropics is also somewhat too weak, by ≈5 m near 100 mb, although not as suppressed as that reported by Manabe and Hahn (1981). The variability in the summer midlatitudes is also weaker than that observed, by ≈30 m. In July, the same general characteristics are observed, with the largest discrepancy in variability again occurring in the summer midlatitudes.

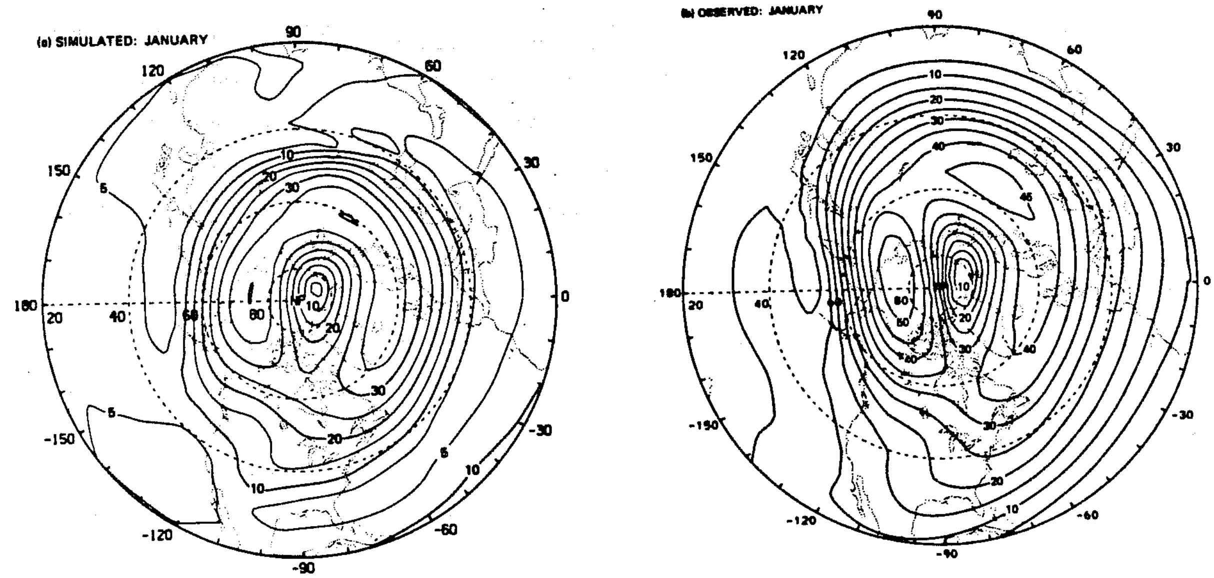

Fig. 10. 10 mb mean resultant wind (m s^{-1}) for January, simulation and observations.

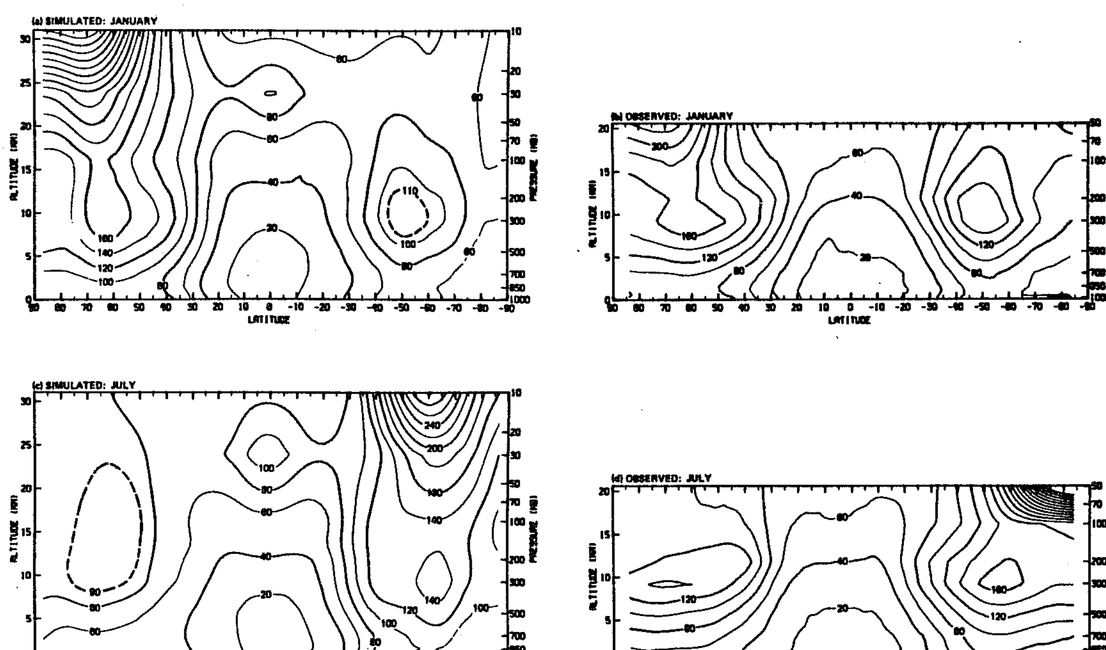

Fig. 11. Latitude-height distribution of the zonal average of the standard deviation of geopotential height (m).

In Fig. 12 we show standard deviation of the 500 mb height for January, simulated and observations. The model data used were a 1200-day series of 500 mb heights. Observations are from Lau et al. (1981). Several features are worth noting. The values of the maxima over the Pacific and the Atlantic are much closer to the observed values than values reported for other GCMs (see, e.g., Blackmon and Lau, 1981). The Pacific maximum is 10° north of its observed position, while the Atlantic maxima are on either side of the observed maximum. The maximum over Siberia (75°E,65°N) is completely absent in the model simulation. Although the local minima in the standard deviation near (120°E,50°N) and (120°W,70°N) are present in the model simulation, the one over North America is distorted in shape, shifted to the east and not low enough in magnitude.

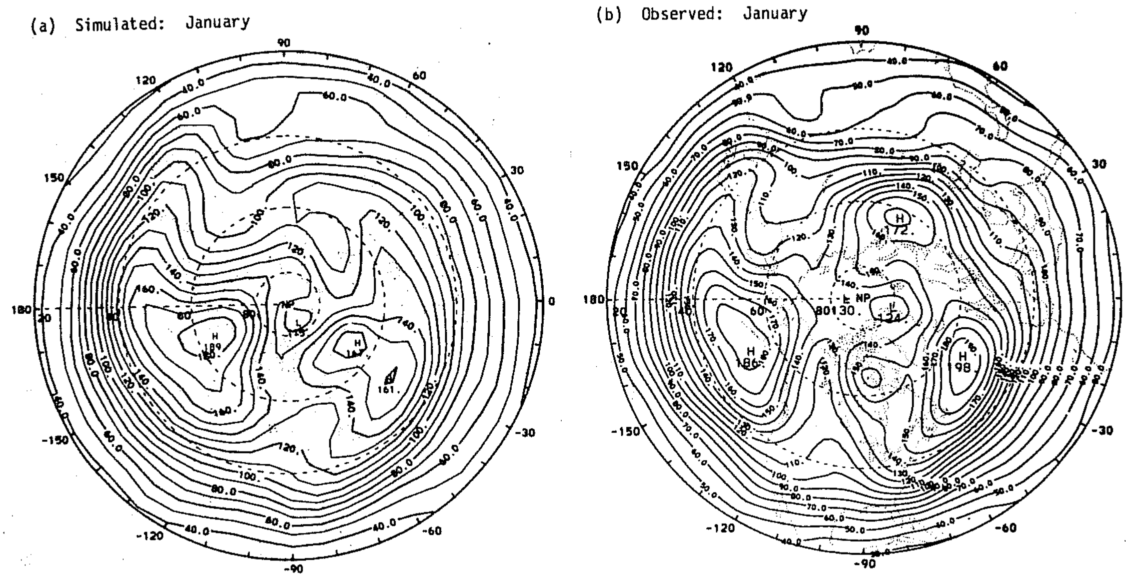

Fig. 12. Standard deviation of the 500 geopotential height (m) for January, simulation and observations.

Fig. 13 compares the bandpass filtered standard deviation of 500 mb geopotential height for January. The 1200-day time series used above was filtered as discussed in Blackmon (1976) to emphasize fluctuation with periods of 2.5 to 6 days. The maxima in the simulations are close in magnitude and position to those observed. The Pacific maximum is about 5° (or one model grid point) north of the observed position. The maximum over Siberia (50°N,70°E) is approximately 10° west of the observed location. The faithfulness of the high frequency simulation seems considerably better than that for the low frequency fluctuations.

4. SUMMARY AND DISCUSSION

Experience to date with climate simulations with the NCAR Communicty Climate Model indicates that some striking improvements have been made in the simulation of the stratosphere, not withstanding the limited vertical resolution of the model there. Preliminary analysis by B. Boville (private communication) indicates that increasing the model resolution in the stratosphere does not degrade the simulation of the mean fields, and indeed, improves it. The impact of additional model levels on the transients in the stratosphere is as yet unclear.

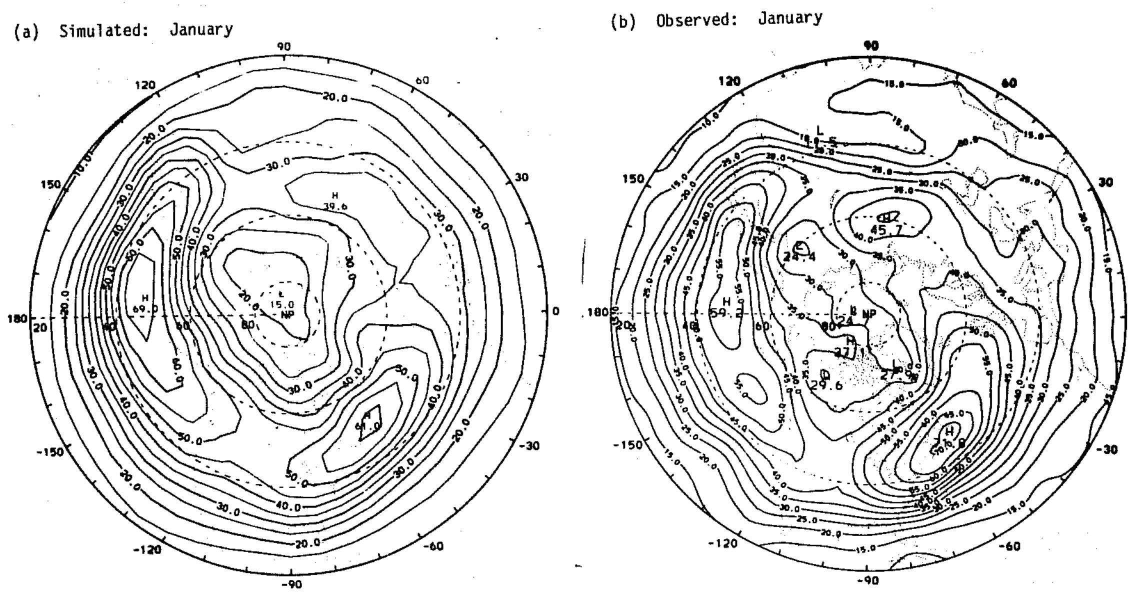

Fig. 13. Bandpass filtered standard deviation of the 500 mb geopotential height (m) for January, simulation and observations.

The simulations of the mean tropospheric fields show many interesting, reasonably accurate features. However, obvious discrepancies remain. Some of these seem common to those produced by other GCMs. Perhaps the most important systematic bias is that towards cold tropospheric temperatures. Mean jets in the summer hemisphere are also considerably too weak.

The variability of the model is quite strong, with the wintertime standard deviation of the height fields being nearly equal to observed values. This is in spite of the fixed boundary conditions in the model. A similar result was found by Manabe and Hahn (1981) for mid and high latitudes. The largest deficiency in model variability is in the summer hemisphere, where the standard deviation fields are subsantially weaker than observed.

The cold tropospheric temperatures, the weak summertime jets and the weak summertime variability are probably due, at least in part, to deficiencies in the present model's radiation and cloud parameterisations. Preliminary experiments incorporating cirrus clouds, with emissivity coupled to liquid water content, shows an increase in the summertime jets speeds. The treatment of radiative interaction with water vapor is presently being revised in the model with the expectation that the troposphere will be warmed by a few degrees, especially in the tropics.

5. REFERENCES

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