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12. Revised by

Kioshi Hada

9. Authorship

Prakki Satyamurti
Vadlamudi Brahmananda Rao
Yoshihiro Yamazaki

Responsible author

13. Authorized by

Marco Antonio Raupp
Diretor Geral

14. Abstract/Notes

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15. Remarks

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by

Prakki Satyamurti
Vadlamudi Brahmananda Rao
and
Yoshihiro Yamazaki

Instituto de Pesquisas Espaciais - INPE
12200 - São José dos Campos - SP

September, 1985

ABSTRACT

A brief review of the observed features of climate of Antarctica and the results of climatic simulation is given. Several interesting aspects of the climate of Antarctic region such as the role of topography and the effect of surface thermal inversions are discussed. The deficiencies of the General Circulation Models in simulating the observed features are mentioned.

1. INTRODUCTION

A comprehensive grasp of the climate and climatic theory demands a good formulation of the polar processes. The high latitude processes are as important as the convective processes in the tropics and the baroclinic processes in the middle latitudes. The ice covered polar regions act as sinks for the atmospheric and oceanic heat engines. It is generally agreed that among all the physical processes of the polar regions, radiative transfer near the surface stands out. The albedos of ice and snow surfaces are about five to eight times as large as those of sea and land surfaces. Such large differences in the surface reflectivity are likely to trigger climatic changes if the action remains sufficiently prolonged. The earth's climate is known to be sensitive to changes in CO_2 and dust concentrations and changes in solar constant due to variations in the planet's orbital parameters. A few authors conjectured the possibility of a stable completely ice-covered planet with the present value of solar constant (see Goody, 1980). It is, however, refreshing to note that such a state has never been realized in the geological history. Donn and Shaw (1966) and Budyko (1972) have speculated that an ice-free Arctic Ocean would remain free of ice once that condition is established by a climatic anomaly. Similar speculations are not advanced in respect of the Antarctic Continent.

There are evidences that global climate changed in the past and is changing in the present. It is of paramount interest to mankind to know if the climate is turning more hostile or more congenial. Economic planning requires the knowledge of the climate in coming years. The tools with which climatologists try to guess the course of the future climate are climatic models. For validating the climatic models, the past and present climates have to be successfully simulated in their totality. In the process of experimenting with those models, one may stumble upon the climates of obscure regions where the observational evidence has not yet established the regional climate. One such obscure region is Antarctica.

The south polar region differs from the north polar region mainly due to the existence of a huge land mass around the South Pole. As the physical features of the south polar region differ, the physical processes are bound to differ, both in quality and quantity, from those in the north. Thus the conclusions drawn from the studies of artic region may not be applicable to the antarctic region. The statistical properties of the general circulation of the atmosphere depend not only on the total amount of solar energy absorbed by the system but also on the vertical and horizontal distribution of absorption. Therefore large changes in surface properties such as albedo and thermal conductivity over the south polar region can influence the regional and even the hemispherical weather systems.

In the following sections we propose to present the observed climatic features of the antarctican region and the climates simulated by climatic models.

2. OBSERVED FEATURES

Earth's climate feels the presence of Antarctica for, essentially, two dynamic reasons. One is the topographic effect and the other is the albedo effect. The topography of Antarctica and the minimum and maximum pack ice extents around the continent are shown in Figure 1. The continent's elevation is more than 3000m in the interior with a peak of more than 4000m situated in the Eastern Antarctica. The slopes are rather gentle up to a few hundred kilometers from the coast. Very close to the coast the slopes are very steep. Table 1 gives the mean monthly extents of the sea ice. In September sea ice covered area around Antarctica is as much as 5% of the total global area, whereas it is less than 1% in March. The area of the continent is 14×10^6 km^2 which is almost completely covered by pack ice and snow throughout the year. The maximum ice covered area in summer is 16.4×10^6 km, whereas in winter it is twice as large.

The climatological charts of surface air temperature in the antarctic region are presented in Figure 2 for the two months, January and July as given by Schwertfeger (1970). The lowest temperatures are found in East Antarctica with values of - 30^{o}C and -70^{o}C in January and July respectively. The meridional surface temperature gradients are larger in winter than in summer. However, in the continental slopes the gradients are very strong in summer. Over the continent thermal inversions at the surface are very frequent, especially in winter, i.e., June through August. Figure 3 gives the distribution of the strength of the surface inversions in winter over Antarctica. We may note that in the interior of the continent the inversions are of the order of 20-25^{o}C. The depth of the inversion layer is about 1.2 km. The annual march of surface air temperatures at several Antarctic stations indicates a so-called "coreless winter". That is, the temperatures show a slight rise in July-August with falls before and after this period (see Van Loon, 1967a and Cavalieri and Parkinson, 1981).

The mean sea level pressure charts for January and July are shown in Figure 4. The summer and winter situations differ only in the strength of the four low pressure cells around the continent. In winter they are deeper. Average surface wind flow over the continent is given in Figure 5. We note that the air flows nearly down the slope up to the coast and there it goes round the continent anticyclonically. The surface winds are known to display a high degree of persistence (Parish, 1982). The wind regime at the coastal stations is katabatic. These winds are very cold and strong occasionally persisting for more than 24 hours.

The factor responsible for the surface wind regime in the interior of Antarctica seems to be the sloped-inversion. The vector difference between the geostrophic wind above the inversion layer and the thermal wind yields the surface geostrophic wind. Parish (1982) estimated the surface winds by considering a balance between the sloped-inversion, coriolis and surface friction terms in the following manner:

$$0 = -g \frac{\Delta T}{\bar{T}} \frac{\partial h}{\partial x} + fv - \frac{KVu}{H} \tag{1}$$

$$0 = -fu - \frac{KVv}{H} \tag{2}$$

$$V^2 = (u^2 + v^2),\ u = V \cos \beta,\ v = V \text{ sen } \beta \tag{3}$$

where x direction is taken down the slope of the terrain and y is taken positive to the right of x such that $\partial h/\partial y = 0$ in which h is the topography. u and v are the components of wind in the x and y directions respectively. ΔT is the strength of the inversion, $\bar{T}$ is the average temperature in the inversion layer, H is the thickness of the inversion layer, g is acceleration due to gravity and f is the coriolis parameter. Taking

$$F \equiv -g \frac{\Delta T}{\bar{T}} \frac{\partial h}{\partial x} \text{ and } J \equiv Hf^2/2KF \tag{4}$$

the solution of the Equations 1 and 2 is

$$V^2 = \frac{HF \cos \beta}{K} \text{ and } \cos \beta = [\sqrt{J^2 + 1} - J]. \tag{5}$$

The estimates of the surface winds given by Parish (1982) are shown in Figure 6. These winds agree fairly well with those shown in Figure 5. This analysis brings out the fact that the surface thermal inversions observed in Antarctica influence the general circulation through the surface winds.

The upper air climatology of antarctic region is not well determined because of lack of observations. However, van Loon (1967b) and Schwertfeger (1970) present gross features of antarctic climate. The winds at all levels above the inversion layer are fairly geostrophic. Figures 7 and 8 show the 500mb geopotential fields for

January and July respectively. The cyclonic center situated more or less near the South Pole is deeper in July than in January. The stationary low pressure centers in the Atlantic and Indian Ocean sectors around the continent are deeper in January. The circumpolar jet in the Southern Hemisphere (not shown here) is situated at 12 km height and is weaker in summer than in winter agreeing well with the thermal wind relation. The jet is stronger and closer to the equator in July. There is a second jet in higher latitudes at a height of 20 km in the Southern Hemisphere (S.H.).

The slope of the bottom topography perpendicular to the flow field has a stabilizing effect on the growing wave disturbances (Blumsack and Gierasch, 1972). However, the investigations of Mechoso (1980) show that the doubling times of the unstable perturbations in representative basic flows around East Antarctica remain less than two days. The combined action of finite amplitude migratory baroclinic waves from the mid-latitudes, topography of the Antarctic Continent and the meridional temperature gradient around the continent can generate a westerly jet in the zonal winds. The effect of the topography enters the dynamic equations to, essentially, change the "β-effect". This alters the speed of the Rossby waves in the slopes of the Antarctic Continent. The Rossby waves with differential phase speeds (i.e., phase speeds varying longitudinally) would create tilts. These tilts become important for the transport of momentum meridionally.

There are appreciable differences between antarctic and arctic climates. The thickness of the sea ice is more or less the same in both the Hemispheres which is about 2m. The maximum sea ice extent in the north is $\sim 15 \times 10^6$ km^2 as against 18.8×10^6 km^2 in the south. The minimum sea ice extents in the two Hemispheres differ appreciably with values of 8×10^6 km^2 in the north and only 2.6×10^6 km^2 in the south. Thus the seasonal variations in the two Hemispheres differ greatly. Surface temperatures are much colder and surface inversions are much more frequent over Antarctica than over

Arctic Ocean (Schwertfeger, 1970). The double jet structure observed in the S. H. does not show up in the N. H.

3. SNOW/ICE - ALBEDO - TEMPERATURE FEEDBACK

By all means the single most important effect of the polar ice caps in the ocean-atmosphere dynamics is the ice-albedo--temperature positive feedback. When by some chance the ice cover increases the albedo of the surface jumps from ~0.1 to ~0.7 and the solar radiation absorbed at the surface decreases drastically which causes a decrease in the surface temperature. The resulting lower surface temperatures, in turn, allow more ice to form in the adjacent areas. This mechanism of ice cover feeding itself to grow is a positive feedback. This feedback is taken into account in varying degrees of sophistication in many General Circulation Models thus far used to simulate global or hemispherical climates.

The change of phase of water is a rather sudden process. At a given point the phase change occurs almost instantaneously, depending upon the temperature of water, along with changes in the optical and thermal properties of the surface. These processes are highly difficult to be parameterized into the large-scale climatic models with exactness. For instance, if the error in the estimation of temperature is $1^{o}C$ in the model, some regions where the temperature is underestimated freeze to ice and in other regions where it is overestimated remain in the water phase.

Albedo for an ice covered surface is ~0.6 and for a snow covered surface it is ~0.8, whereas for land and ocean surface it varies between 0.06 to 0.20 (Goody, 1980). A 20% change in sea ice can affect a change of 1% in the total solar radiation absorbed by the Globe at the surface. The effect of such a change is no less important than, for instance, replacing the Amazon forest by a desert. Robock (1979, 1983) showed with a two dimensional energy balance model that the average global surface temperature would fall by more than $2^{o}C$ if the solar constant is decreased by 1%. Therefore large prolonged changes in sea ice cover and the attendant changes in absorbed solar radiation seem to be important for global climatology.

The formation of ice and snow depend on surface temperature, precipitation and other factors that determine the surface energy budget and hydrological cycle. New ice develops in high latitudes when the ocean surface is cooled to freezing temperatures followed by further surface heat loss. The rate of accretion of ice at the bottom of the ice sheet depends on the rate of heat conduction upwards through the ice at the top. Hence the rate of ice formation is sensitive to the thickness of existing ice. A useful review of polar atmosphere-ice-ocean processes is given by Polar Group (1980).

Thus a consistent climate model should determine the ice and snow cover explicitly. The albedo effect is included in a highly simplified and parameterized fashion in the energy balance models of Budyko (1969) and Sellers (1969, 1973) and in the GCMs beginning in the simulations of Wetherald and Manabe (1972). In the energy balance models the ice cover is related to temperature empirically so that surface albedo is temperature dependent. Budyko (1969) assumed that where surface temperature is less than the present mean temperature at 72^{o}N, the albedo of the earth-atmosphere system increases from 0.5 to 0.62. Sellers used a scheme in which the albedo increases at a rate of $0.09/1^{o}K$ if the temperatures drop below $10^{o}C$ until a maximum of 0.85. The GCMs calculate the snow and ice covers from the hydrologic cycle explicitly. Manabe and Wetherald (1980) distinguished rainfall from snowfall according to the temperature at a height of 350m by an empirical relationship. The accumulation of snow is calculated using

$$\frac{\partial s}{\partial t} = S_F - E - M_e, \qquad (6)$$

where s is the depth of snow, S_F is the rate of snowfall, E is the rate of evaporation and M_e is the rate of melting. If snow accumulated is older than ~15 days it is considered to form pack ice. In the GISS model (Somerville et al, 1974) the ice and snow effects are included very crudely. The latitude of the snow line in the N. H. is taken to be

$$\phi_s = 60 - 15 \cos 2\pi \left(\frac{t - 24.7}{365}\right), \tag{7}$$

where t is measured in days starting from 1 January. The snow line in the S. H. is always taken 120^0 south of ϕ_s. The albedos used in this model are 0.07 for sea, 0.14 for land and 0.7 for snow or ice. The change in boundary temperature for land, ice and snow locations is predicted from the net surface heating or cooling due to radiative, sensible heat and latent heat fluxes. Handerson-Sellers and Wilson (1983) summarized in a tabular form the albedo formulations considered in different GCMs around the world.

4. SIMULATION EXPERIMENTS AND RESULTS

In this section physics contained in and results obtained by various simulation experiments conducted using spectral and grid point GCMs are discussed giving attention to the polar regions.

Manabe et al (1979) presented an evaluation of a spectral GCM in simulating the earth's climate. They used a nine level GFDL (Geophysical Fluid Dynamics Laboratory) spectral GCM with three different truncations beyond 15, 21 and 30 wave numbers in the spherical harmonic representation of the fields. These three experiments are called M15, M21 and M30 with equivalent east-west grid spacings of 7.5^0, 5.625^0 and 3.75^0 respectively. The GCM determines surface temperatures over continents through the condition that the net fluxes of solar and terrestrial radiation and turbulent fluxes of sensible and latent heat locally add to zero. Over the oceans seasonal variation of surface temperature is prescribed. The surface albedos are prescribed as functions of latitude over oceans and pointwise over land. These values are replaced by higher values whenever ice or snow is simulated. The experiments are integrated for 2 to 3 years ending on 1 August or 1 July and the last year climatology is presented.

Figures 9 and 10 show the zonal mean surface pressure for January and July obtained in the three experiments M15, M21 and M30 and are compared with the observations. Except for the gross features in the tropics and subtropics, the observations differ very much from the simulations. The results of M15 are better than the results of M21 and M30. Especially over antarctic and subantarctic regions the sea level pressure in winter is superestimated by more than 15 mb. A comparison of these results with those obtained in grid point GCM experiments is worthwhile. Figure 11 presents the results of experiments with 250 km and 500 km resolutions for January. The agreement with the observations is worse here than in spectral model especially in polar regions. All these discrepancies may be largely attributable to lack of adequate physics of the polar processes.

Mc Aveney et al (1979) simulated the January global circulation using the GFDL nine-level spectral GCM with a resolution corresponding to M15 described above. Figure 12 shows the observed and simulated mean surface air temperature distributions for January. The GCM overestimates the temperatures in polar regions by at least 10^{o}C and underestimates in tropics by 5^{o}C, with the result the meridional temperature gradients are weaker in the simulated atmosphere. These discrepancies are due to the inability of the model to handle surface inversions. Figure 13 shows the zonally averaged precipitation rates as simulated by Mc Aveney et al and the observed rates. The model overestimates the precipitation rates in the equatorial and polar regions and elsewhere the agreement is good.

Washington et al (1979) conducted climate simulations with an eight-layer NCAR (National Center for Atmospheric Research) GCM. This GCM has a $5^{0} \times 5^{0}$ horizontal resolution. In this model surface temperature over the open oceans and the sea ice are prescribed from climatology. If the grid point falls over the sea ice or a nonvegetated continental area, then the surface temperature depends on the snow cover and soil moisture and is calculated from the surface energy balance. Figure 14 gives the zonal mean surface air temperatures obtained in the simulation along with observed values for January. Except in antarctic region the agreement is good. Over Antarctica the model atmosphere is warmer by about 10^{0}C as in the case of FGDL model. Figure 15 shows the observed and simulated mean zonal winds. The positions of the jets in the simulated atmosphere agree well with the observations. However, the jets are a little too strong in the model atmosphere. The midtropospheric equatorial easterlies are not reproduced in the simulations. The model generates easterlies over and around Antarctica throughout the troposphere and above whereas there are no easterlies in the observations.

The climate simulation experiments conducted with many other GCMs such as the AES GCM (Boer and Mc Fartane, 1979) and the GLAS GCM (Halem et al, 1979) produced results similar to the ones described above, i.e., the model atmospheres agreed qualitatively with the observed atmosphere in the tropics and midlatitudes but disagreed completely in the high latitude and polar regions. Curiously enough, results of the same quality are obtained with a two-layer model of the Oregon State University (Schlesinger and Gates, 1979). For instance, Figure 16 shows latitudinal variation of zonal average temperature as simulated by the two-level model and as observed for January. The agreement between the model temperatures and observations is as good as or better than in the multilevel models presented above.

Chervin (1979) performed experiments with a July version of the NCAR GCM to assess the models response to changes in surface albedo at selected places. A large-scale increase in surface albedo resulted in reduced precipitation over the prescribed change region and increased precipitation equatorwards. Herman and Johnson (1980) examined the response of GLAS GCM to sea ice boundary conditions around the arctic region. He concluded that the ice margin anomalies were capable of altering local climates in the high and mid-latitudes. Robock (1979) simulated the annual average surface temperatures with a modified version of the Sellers (1976) model with a 45 days time lag between temperatures and ice and snow to fit the observed conditions. The result is shown in Figure 17. The model temperatures agree with the observations well except near the South Pole. This may be because the model assumes the horizontal heat fluxes to extend throughout the 1000 mb deep atmosphere and in reality the high continent prevents this.

5. DISCUSSION AND CONCLUDING REMARKS

In the last two decades many climate models have envolved. They vary from one dimensional Energy Balance Models to very ambitious three dimensional GCMs. Many models take into account the polar processes in varying degrees of sophistication. The single most important variable polar process is the ice/snow-albedo-temperature feedback mechanism. Budd (1975) noted that a $1^{o}C$ change in annual mean atmospheric temperature corresponded to about 2.5^{o} change in the sea ice margin and a 70 day change in the duration of ice cover. The feedback mechanism in the models is parameterized because of the obvious complexities involved.

Using these models many authors have tried to simulate past and present climates as well as some hypothetical scenarios for the future (see International Council of Scientific Unions, 1979). We have tried to describe a few simulation results pertaining to Antarctic region in particular and the general circulation of the atmosphere in general with an emphasis to understand the role of polar processes in determining the climate.

There is little doubt that the Antarctic Continent and the sea ice around it influence the weather and climate of the S. H. The two important characteristics of Antarctica are its topography and the ice covered surface with large seasonal and interannual variability of sea ice around it. Whereas time derivatives in the weather variations are only one order of magnitude less than the measurable magnitudes of terms representing geostrophic balance, the time derivatives of ice mass are at least two orders of magnitude less than the measurable magnitudes of flux terms that produce them. These flux terms involve parameterizations, the forms of which are not even agreed upon (Saltzman, 1984).

The ice line margin provides a strong baroclinic zone where cyclogenesis and frontogenesis are favored. Although the study of Budd (1982) does not show a direct correlation between the position of the sea ice line and the cyclone frequency in the S. H., the preferred locations of the generation of cyclones and fronts may vary with the ice line. Schwertfeger and Kachelhofer (1973) observed that the latitudinal band of cyclone frequency in the transition seasons varies with the sea ice margin. There are indications from the study of Fletcher (1969) that an intensified zonal circulation in the S.H. coincides with above average Antarctic ice conditions. Weller (1982) and Fletcher et al (1982) visualize that the Antarctic Oceans influence both the longer term fluctuations of the global climate and the shorter term fluctuation in the S. H.

From the linear studies of Mechoso (1980, 1981) it is learnt that the Antarctican topography has a stabilizing effect on the growing baroclinic waves in the coastal slopes. The phase speeds of Rossby waves increase in this region due to the reduced β-effect, thereby creating horizontal tilts that favor transport of momentum which supports a jet like structure in the winds around the continent. Cavalieri and Parkinson (1981) and van Loon (1967a) have shown that around Antarctica the winter is 'coreless'. This suggests that when the temperatures fall below ~260°K around 60-70°S the baroclinity generates more wave disturbances which pump heat into the polar latitudes to raise the temperatures.

From the GCM simulation studies of Manabe et al (1979), Mc Aveney et al (1979), Washington et al (1979) and Boer and Mc Farlane (1979) it is noted that the grid point models fail to reproduce the antarctic climatology satisfactorily, in that the surface pressures and air temperatures over the antarctican region are overestimated by about 15mb and 10°C respectively. The model upper air winds present circumpolar jets that are a little too strong and false easterlies throughout the troposphere over Antarctica. The spectral GCMs do a little better than the grid point versions. The simulations produced by low vertical resolution models (ex. Schlesinger and Gates, 1979) are not inferior to those produced by high resolution models. It is evident from the experiments of Chervin (1979) and Herman and Johnson (1980) that the ice line and albedo anomalies are capable of producing changes in the climates of high and mid latitudes. Robock's (1979) study clearly points out that the polar processes have still to be refined to obtain better results in the high latitudes.

For a Brazilian meteorologist, the interest in Antarctica is to see the possible associations between antarctic climate and the climate over South America. More specifically, what changes in Antarctic climate precede changes in Brazilian climate, may it be over South Brazil where the frontal activity depends on the position of sea ice margin and antarctican climate or it be over Northeast Brazil where the large-scale hemispheric circulation changes may affect the drought and humid epochs. A cursory survey of the existing literature shows that there are few studies on this aspect.

Therefore, it is recommended that GCM experiments of the southern hemispheric general circulation, with the inclusion of the effects of sea ice margin and extent, be carried out to deduce the possible changes in Brazilian climate, one of the important questions not addressed earlier is how strong and how durable an anomaly in the sea ice conditions is necessary to bring appreciable changes in Brazilian climate. These can be carried out in the lines of desert--albedo-temperature feedback simulations. To start with simple two--level models such the Oregon State University Model (Schlesinger and Gates, 1979) may be used for such studies.

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FIGURE LEGENDS

1 - Topography of Antarctica and average extent of pack ice in March and September. Continuous lines drawn for 2000, 3000 and 4000m.

2 - Mean isotherms of surface air in a) January and b) July.

3 - Isolines of average strength of the surface inversion in winter (June-August).

4 - Mean sea level pressure in a) January and b) July.

5 - Average surface wind flow over Antarctica.

6 - Sloped-inversion winds over East Antarctica estimated by Parish (1982).

7 - Mean 500mb geopotential field in January.

8 - Same as Figure 7 except for July.

9 - Latitudinal distribution zonal mean sea level pressure simulated in M15, M21 and M30 experiments and as observed for January.

10 - Same as in Figure 9 except for July.

11 - Same as in Figure 9 except for a grid point GCM with horizontal resolutions of 500km and 250km.

12 - Surface air temperature for January simulated by Mc Aveney et al (1979) and as observed.

13 - Latitudinal distribution of precipitation simulated by Mc Aveney et al (1979) and as observed.

14 - Zonally averaged surface air temperature simulated by Washington et al (1979) and as observed.

15 - Latitude - heigth distribution of zonally averaged zonal wind component for January and July simulated by Washington et al (1979) and as observed.

16 - Zonal average of simulated and observed air temperature for January at the surface, 800mb and 400mb.

17 - Annual average zonal mean surface temperature.

TABLE 1

AREA OF THE REGION AROUND ANTARCTICA, COVERED WITH SEA ICE (10^6 KM^2)

MONTH	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC
AREA	6.8	4.3	2.6	6.4	9.2	11.0	13.9	15.7	18.8	17.8	15.2	11.4

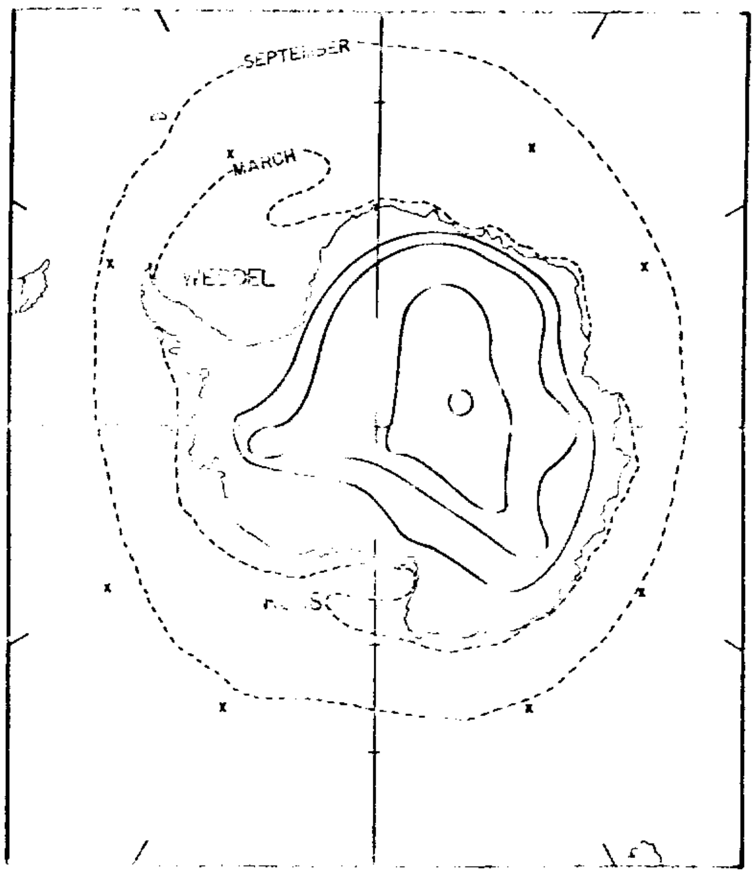


Fig. 1

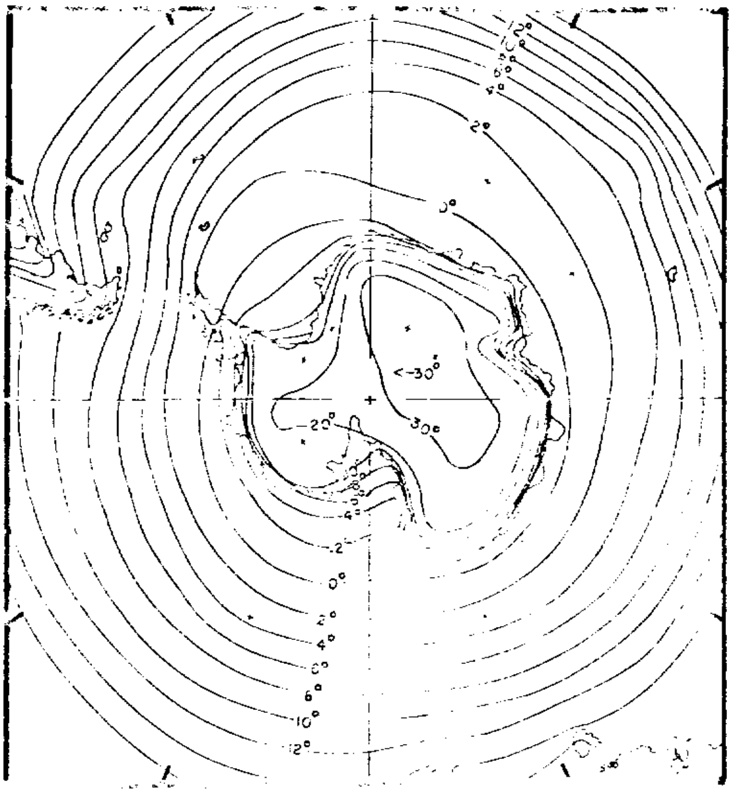


Fig. 2a

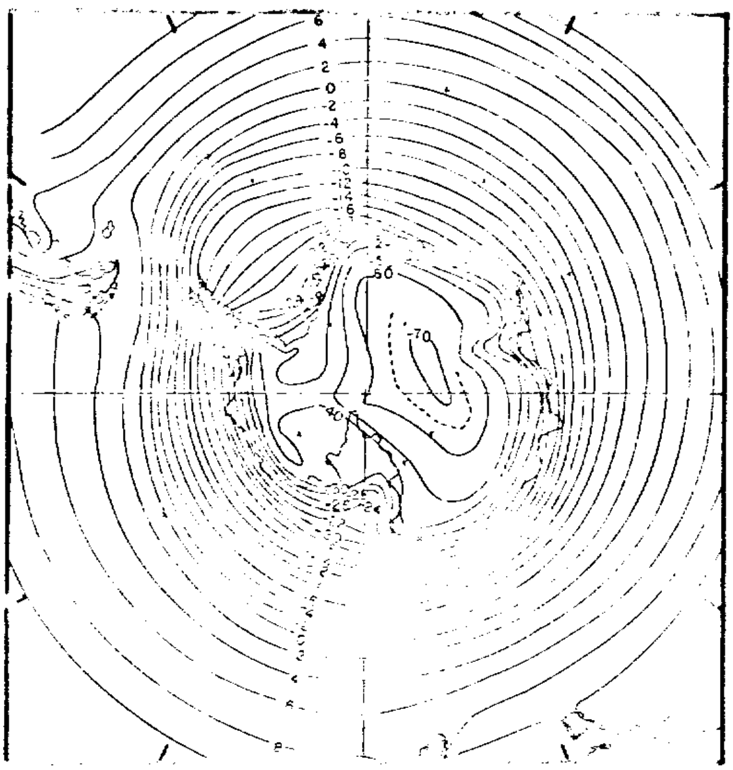


Fig. 2b

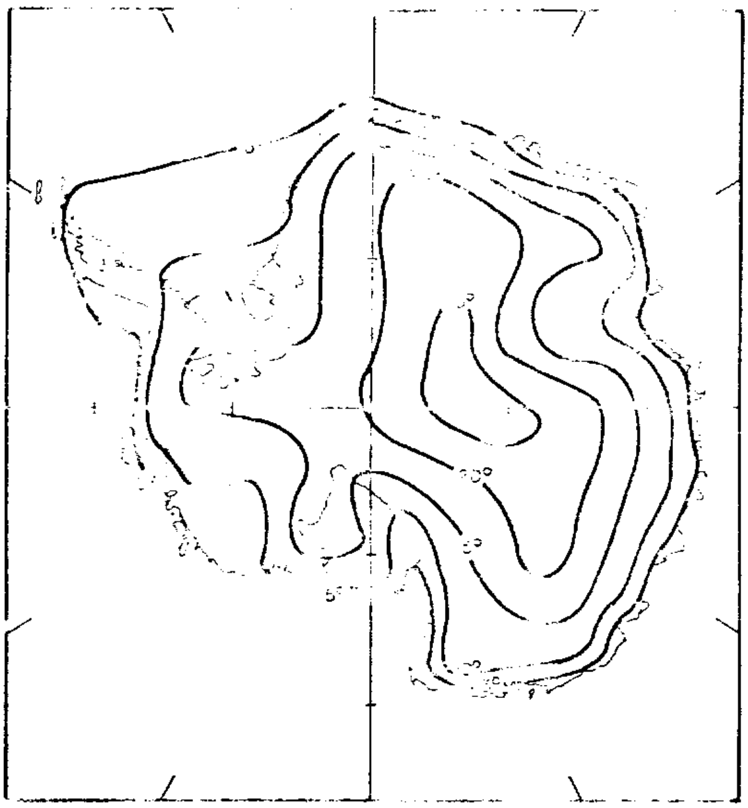


Fig. 3

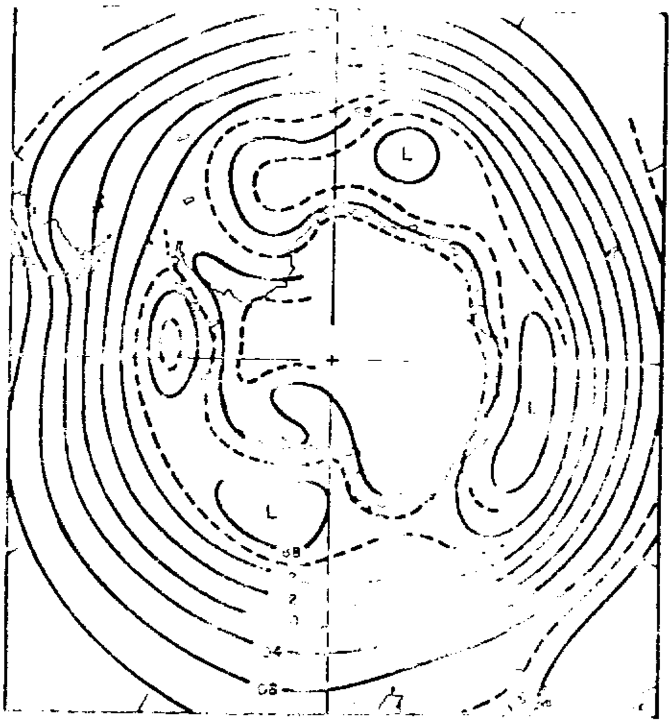


Fig. 4a

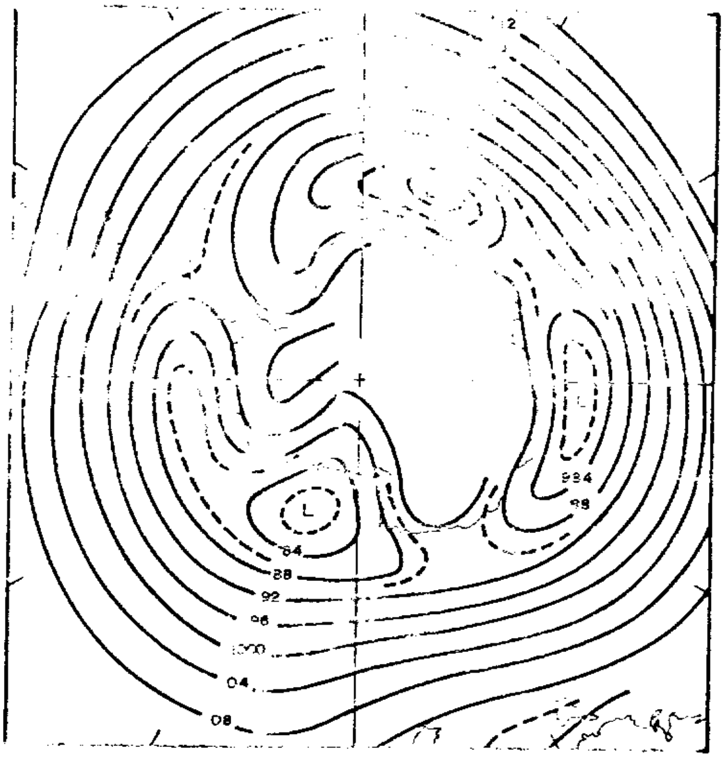


Fig. 4b

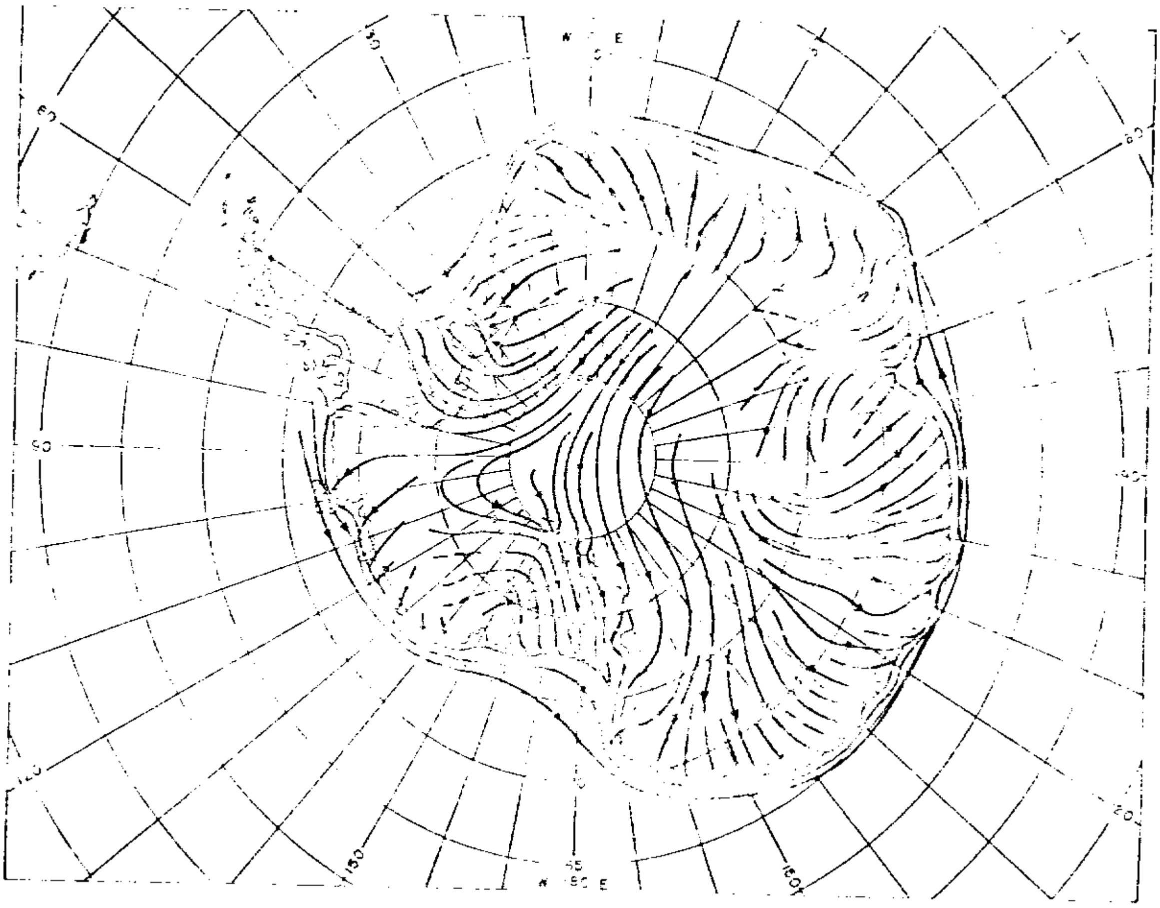
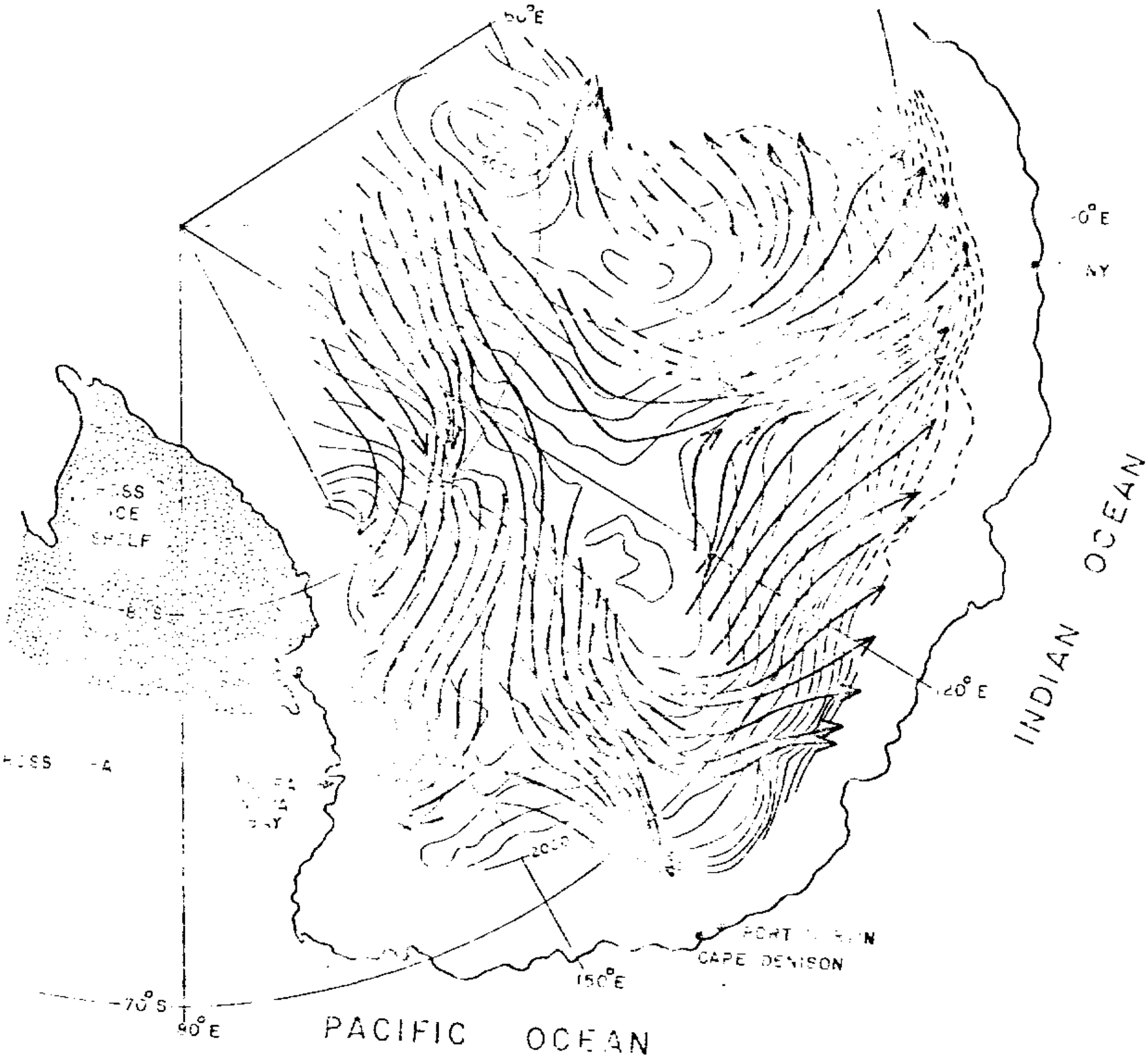


Fig. 5



PACIFIC OCEAN
INDIAN OCEAN
CAPE DENISON
120° E
150° E
90° E
70° S
2000

Fig. 6

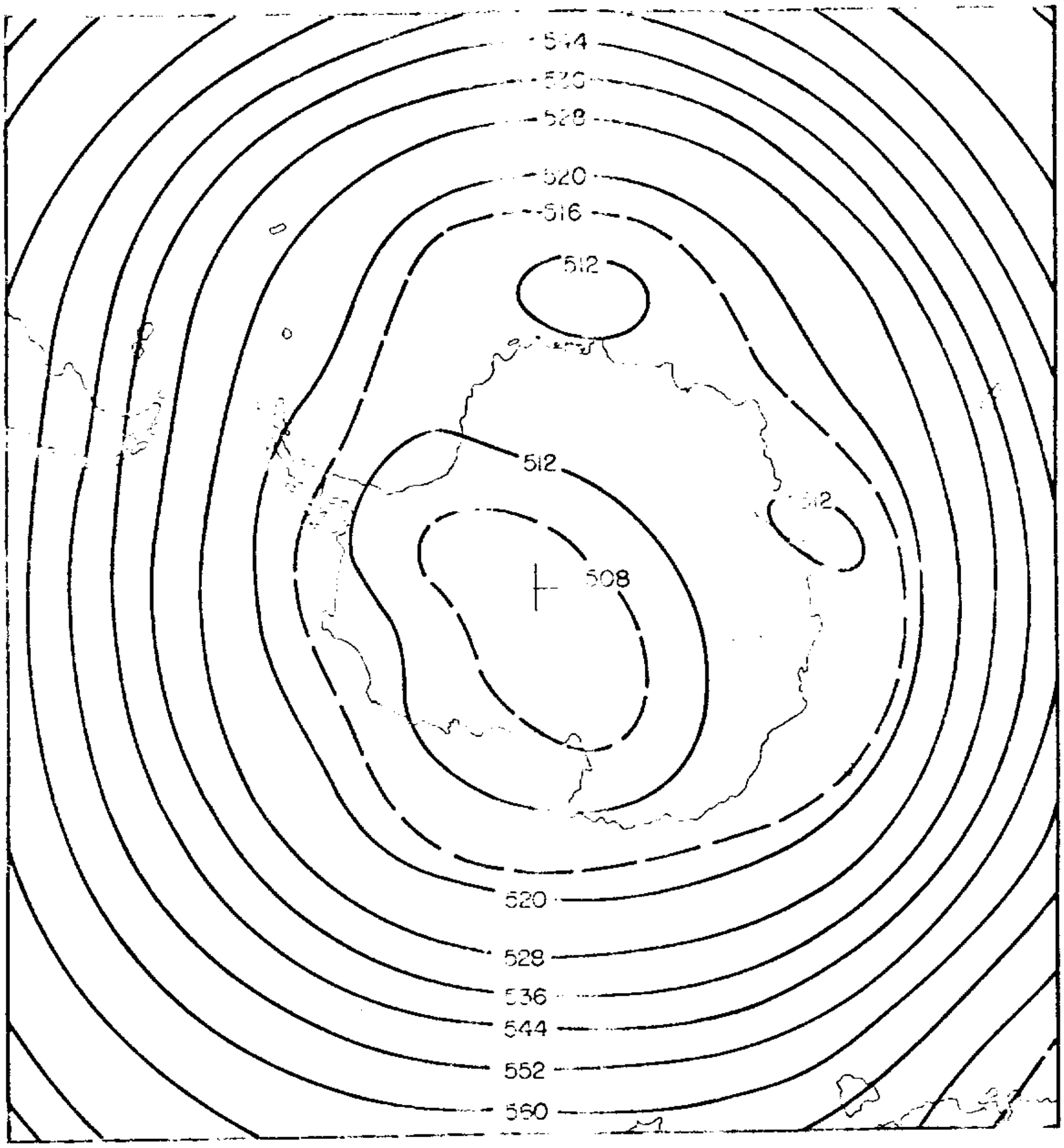


Fig. 7

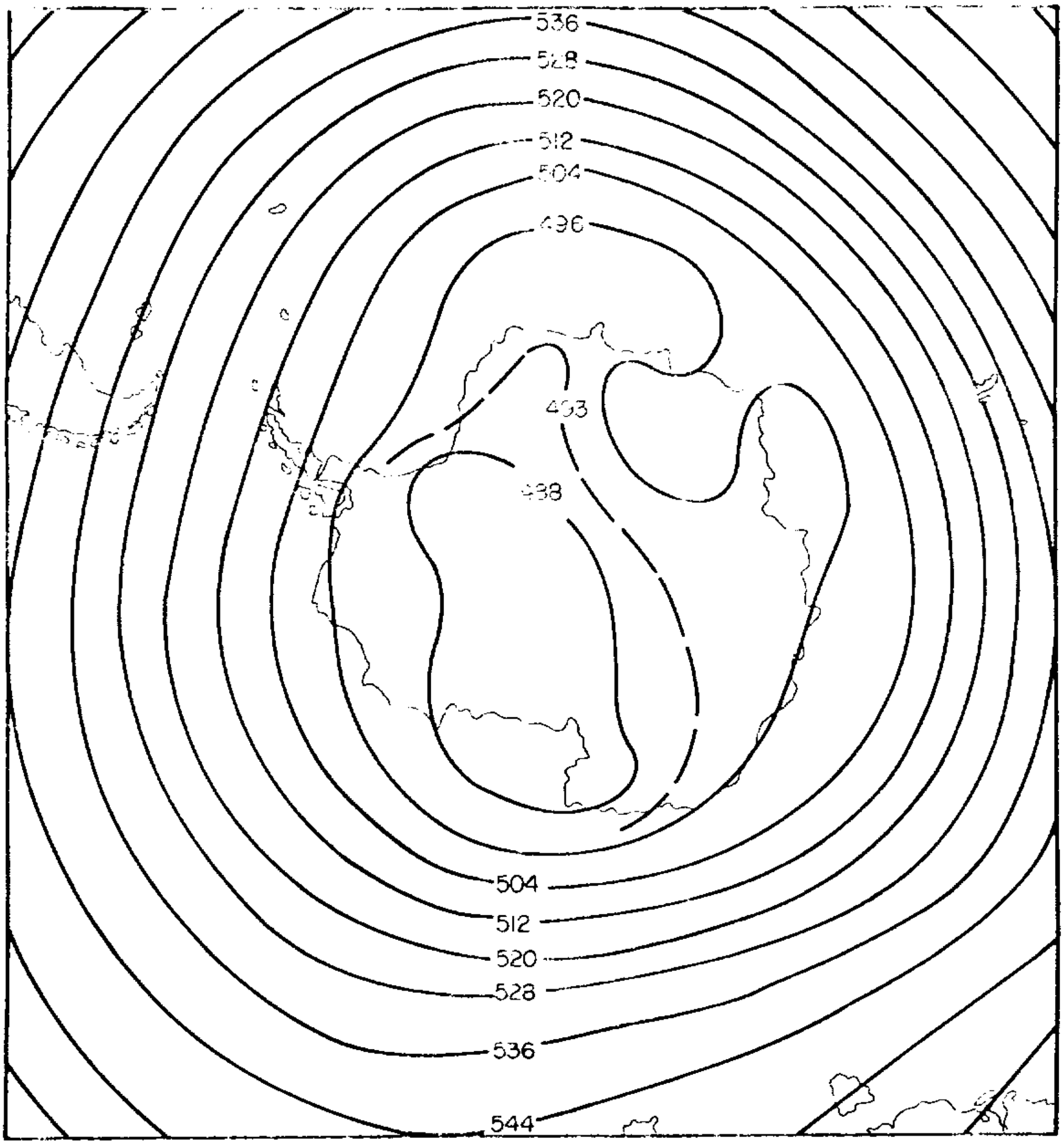


Fig. 8

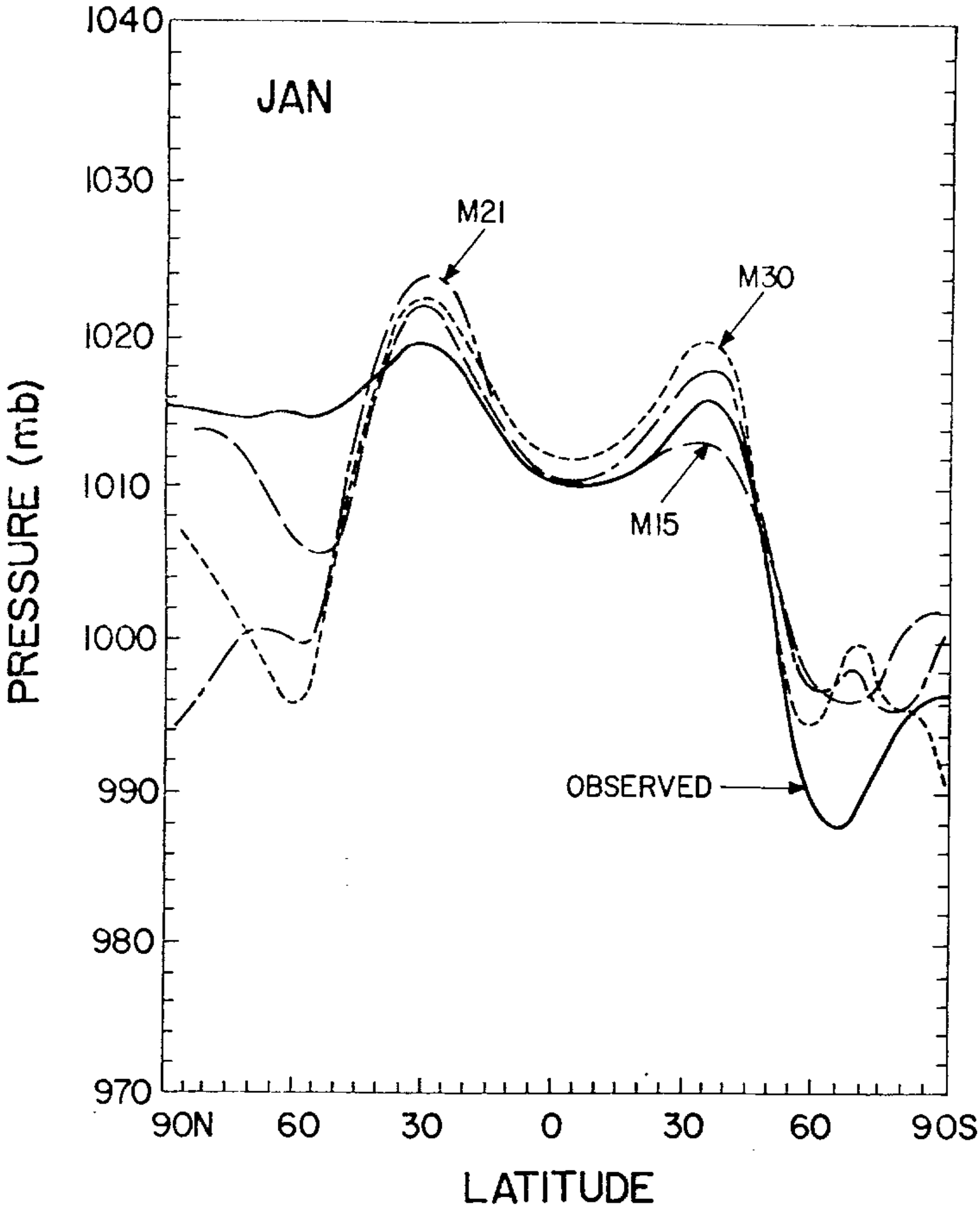

JAN
PRESSURE (mb)
1040
1030
1020
1010
1000
990
980
970
M21
M30
M15
OBSERVED
90N
60
30
0
30
60
90S
LATITUDE

Fig. 9

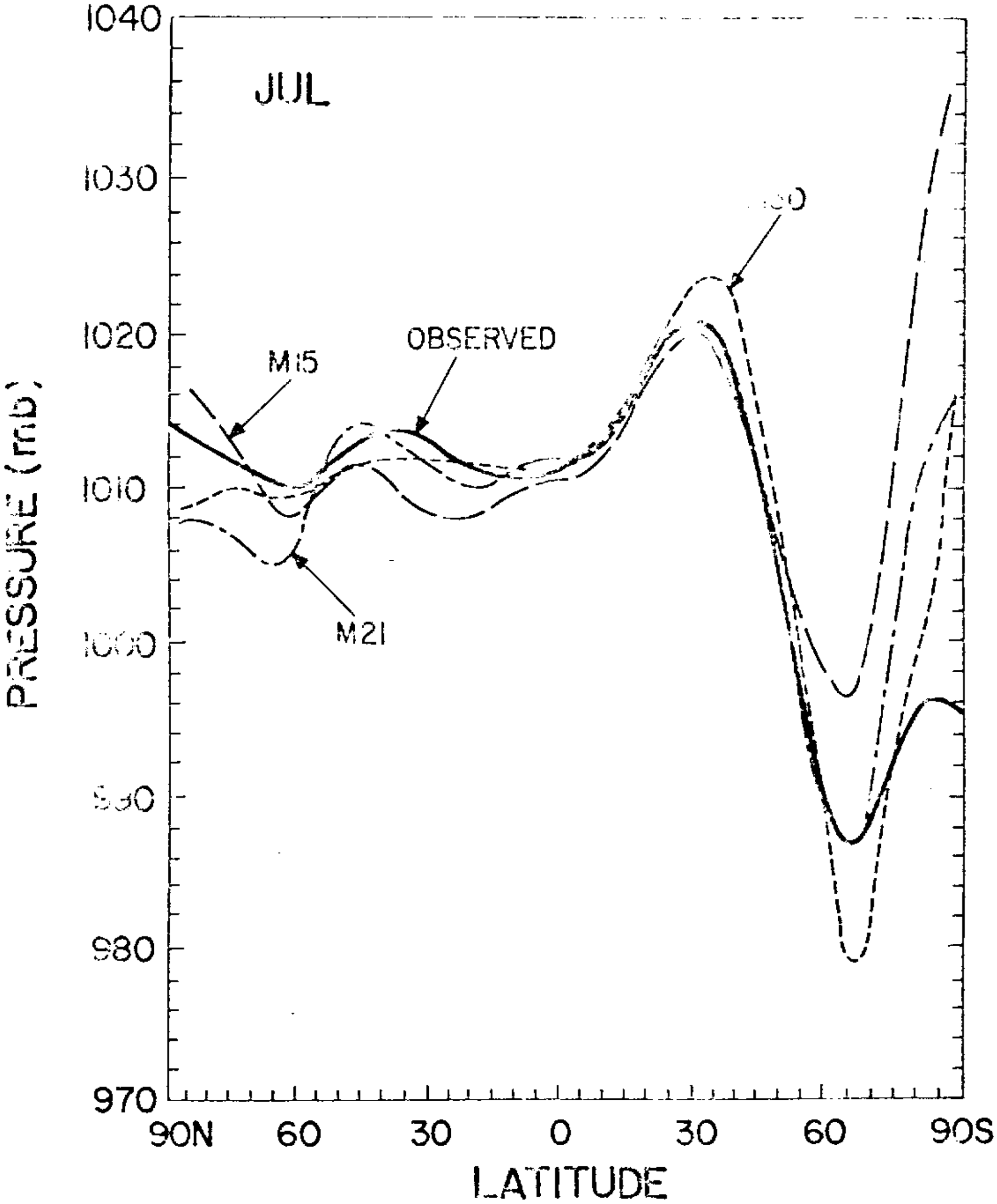

JUL
PRESSURE (mb)
LATITUDE
OBSERVED
M15
M21
1040
1030
1020
1010
980
970
90N
60
30
0
30
60
90S

Fig. 10

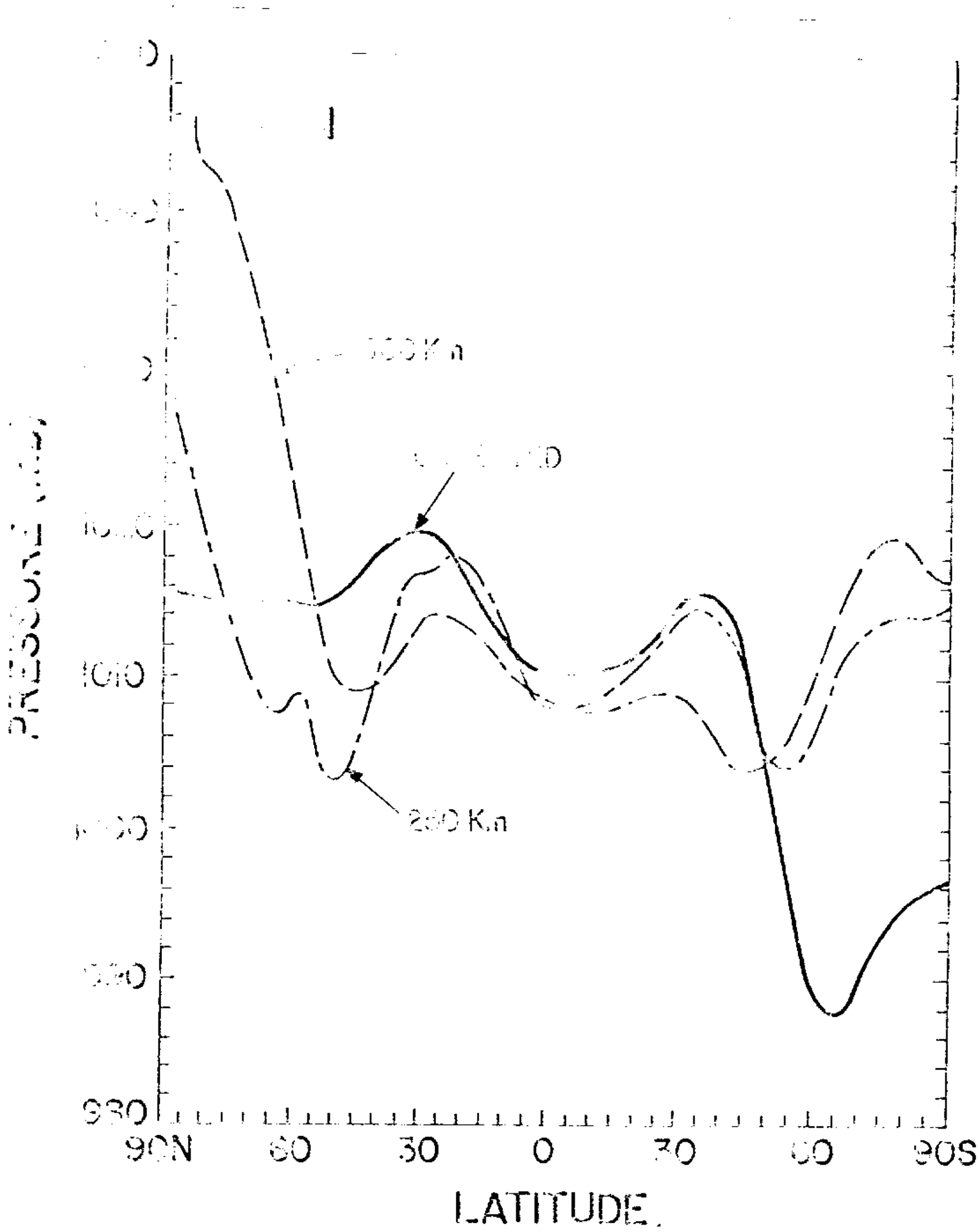


Fig. 11

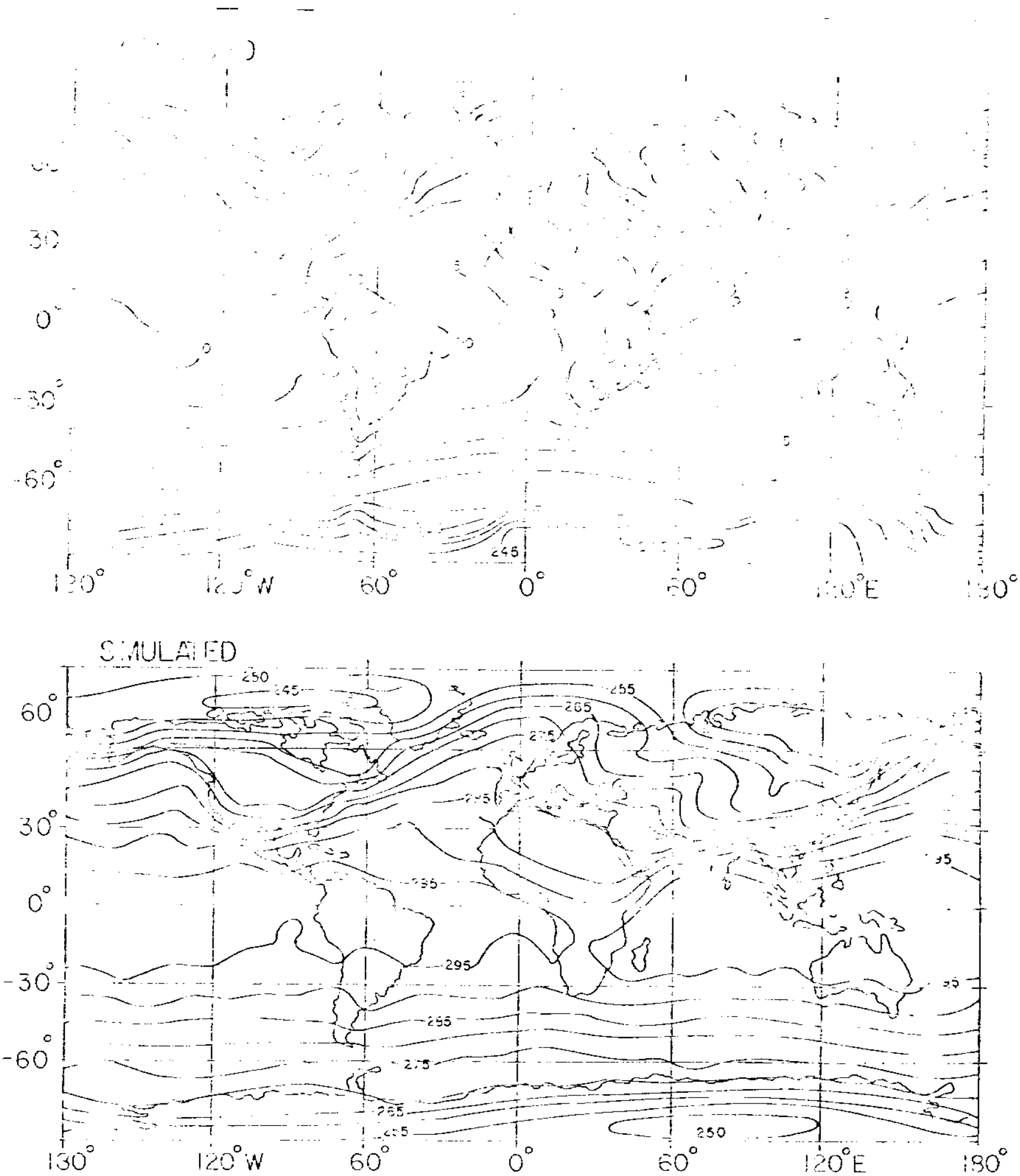

SIMULATED
250
245
255
265
275
285
295
60°
30°
0°
-30°
-60°
130°
120°W
60°
0°
60°
120°E
180°

Fig. 12

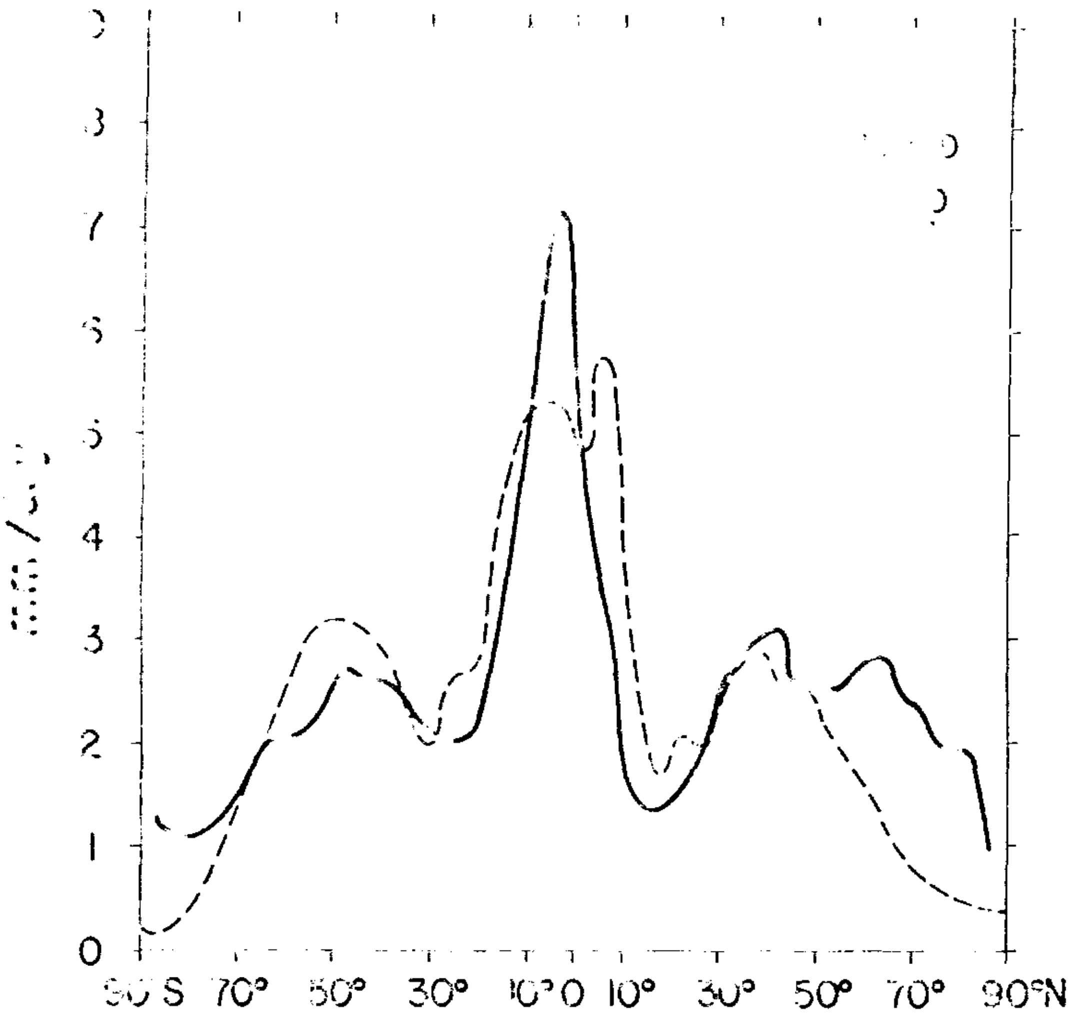


Fig. 13

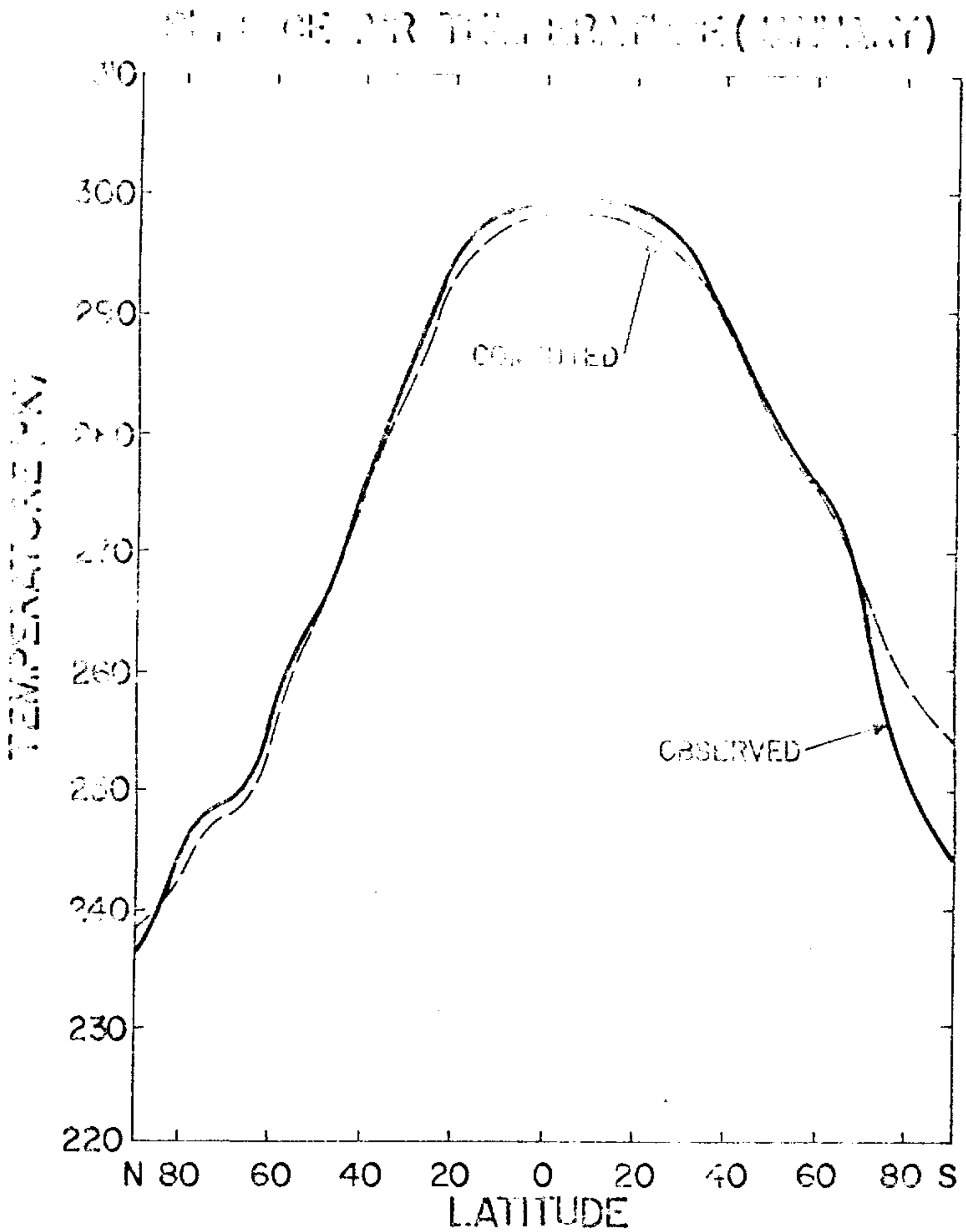


Fig. 14

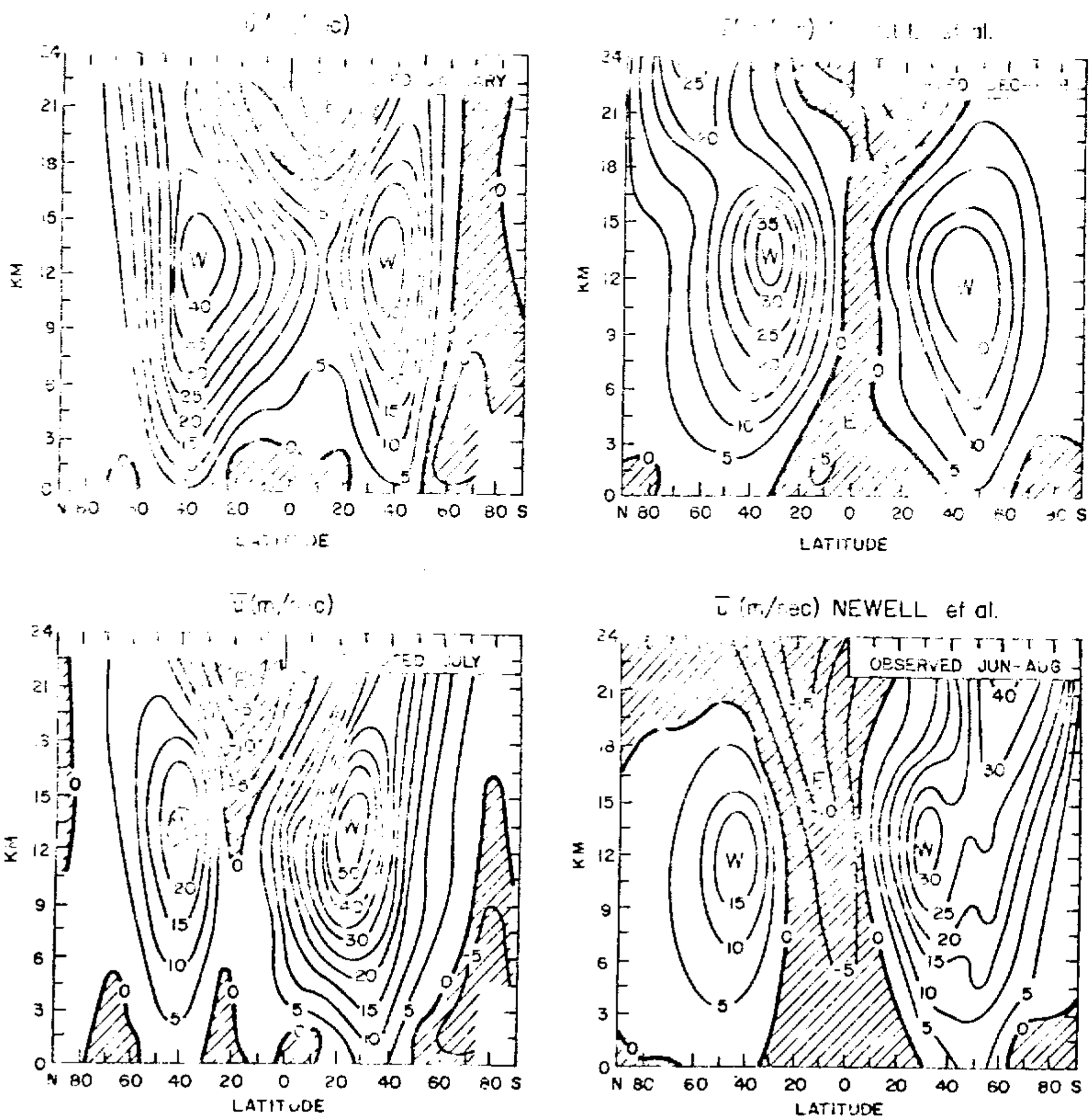


Fig. 15

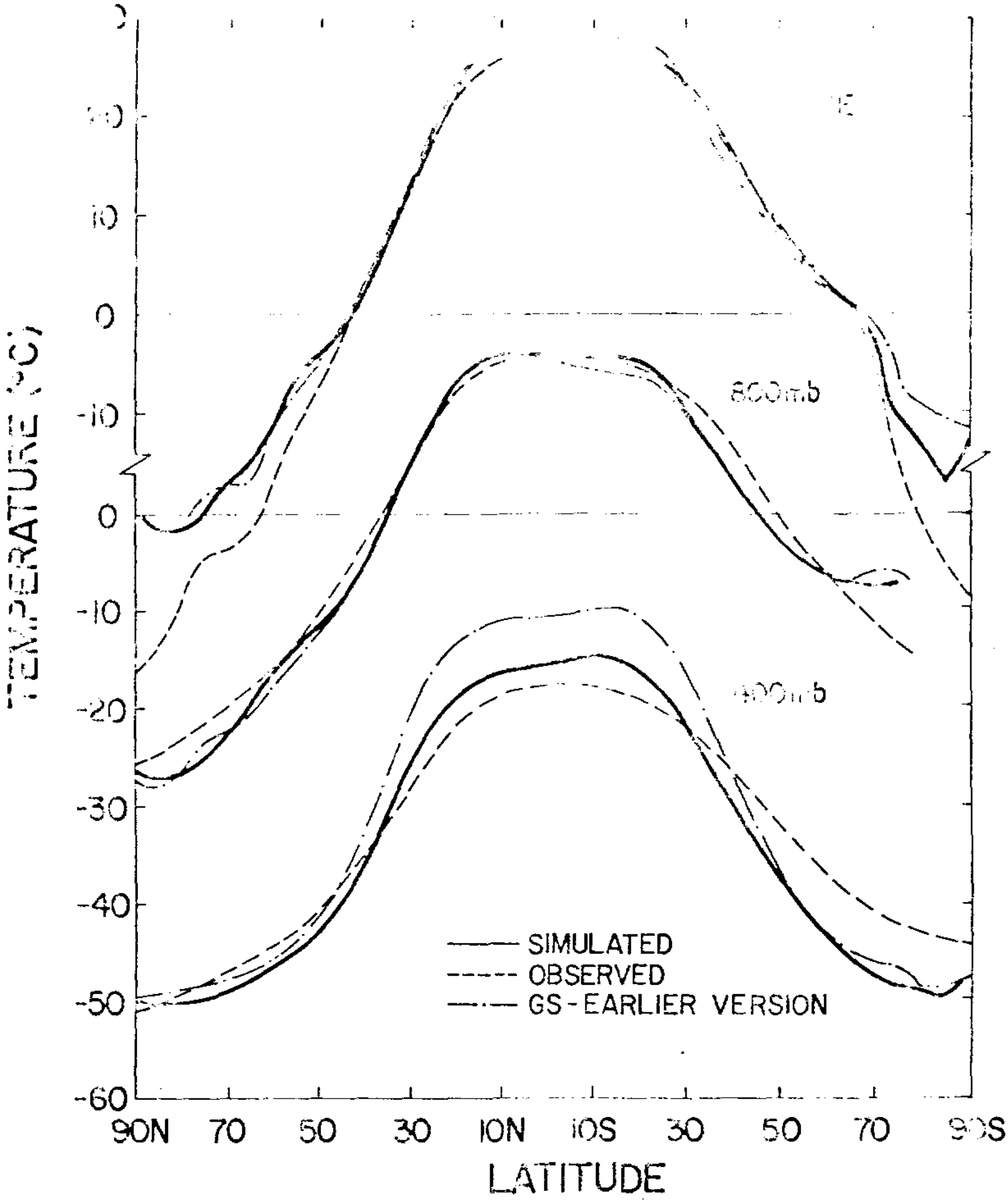

800mb
400mb
SIMULATED
OBSERVED
GS-EARLIER VERSION
TEMPERATURE (°C)
LATITUDE
90N
70
50
30
10N
10S
30
50
70
90S
10
0
-10
-20
-30
-40
-50
-60

Fig. 16

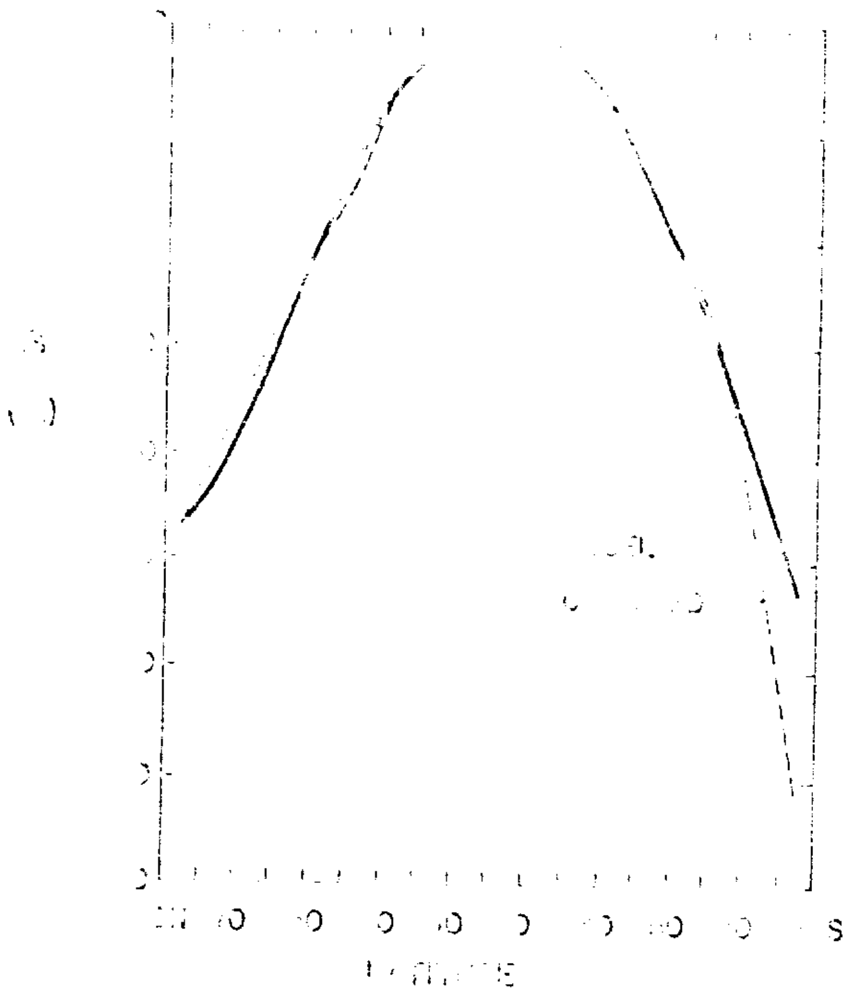


Fig. 17