JOURNAL OF CLIMATOLOGY, VOL. 5, 417–421 (1985) 551.513.1:551.524.2(4/9–13)

ON THE ZONALLY AVERAGED TEMPERATURE IN THE SOUTHERN HEMISPHERE

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Received 19 June 1984
Revised 30 October 1984

ABSTRACT

The observed meridional temperature gradients in the middle and high latitudes of the southern hemisphere are much higher than the critical gradients necessary for the manifestation of baroclinic instability in a two-level quasi geostrophic model. This is in contrast to the 'baroclinic adjustment' observed in the northern hemisphere by Stone (1978).

KEY WORDS Baroclinic adjustment Southern hemisphere

INTRODUCTION

Recent observational studies (e.g. Oort and Peixoto, 1983) suggest that the mean atmospheric state in the southern hemisphere (S. H.) differs considerably from that of the northern hemisphere (N. H.). One aspect of the mean atmosphere which has interesting implications is the zonally averaged temperature. Flohn (1980) noted that the meridional, equator to pole, temperature difference during southern winter in the S. H. is much higher than that in the N. H. during northern winter. He discussed extensively the possible climatic consequences of a man-made global warming and suggested the implications of a possible coexistence of an ice free Arctic and a glaciated Antarctic.

Since the mean atmospheric state forms the so-called basic state for the instability studies, some interesting differences could be expected in the development of transient eddies and their associated processes in the S.H.

Stone (1978) noted that the zonal mean midtropospheric meridional temperature gradients in the middle latitudes of the N. H. agree closely with the local values of the critical gradient as given by the two-layer Phillips (1954) model. Stone interpreted this observation as the indication of a negative feedback between the meridional eddy flux of heat and the meridional temperature gradient. He termed the associated process as 'baroclinic adjustment'.

In view of the aforesaid Flohn's observation, it would be interesting to compare the observed and critical temperature gradients for the S. H. case, which is the objective of the present note.

COMPARISON OF THE OBSERVED AND CRITICAL TEMPERATURE GRADIENTS IN THE S. H.

The critical shear for a two-layer model (Holton, 1979) is given by

$$U_c = \frac{\beta \sigma_2 \Delta p^2}{2 f_0^2} \tag{1}$$

0196–1748/85/040417–05$01.00

where $U_c = (U_1 - U_3)/2$; σ_2, the static stability at level 2, is given by

$$\sigma_2 = -\frac{RT_2}{p_2}\left[\frac{1}{T_2}\frac{T_3 - T_1}{\Delta p} - \frac{R}{p_2 C_p}\right] \tag{2}$$

Subscripts 1, 2 and 3 refer, respectively, to the upper, middle and lower levels of a two layer model. $\Delta p = p_3 - p_1$. Using the thermal wind relation, the critical shear can be written in terms of potential temperature. The final expression for the critical gradient at level 2 for a 5° latitude interval is given by

$$\Delta T_c = \frac{\pi}{36}\cot(\phi)\frac{RT_2}{g}\frac{\Theta_1 - \Theta_3}{Z_1 - Z_3} \tag{3}$$

where ϕ is the latitude and g is the acceleration of gravity. In the above equations, U is the eastward component of wind, f is the Coriolis parameter, R is the gas constant for dry air, T is the temperature, Θ is the potential temperature and Z is the height above a reference level (1000 mb). All the meteorological quantities are the zonal averages.

In order to calculate ΔT_c, we used temperature data given by Taljaard *et al.* (1969). These are given at standard pressure levels. We have chosen the same pressure levels as did Stone (1978) to represent levels 1, 2 and 3, namely 400, 600 and 800 mb, respectively. Also, following Stone (1978), we have taken the mass weighted average temperature in the layer 200–600 mb as the temperature at 400 mb and the mass weighted average in the layer 600–1000 mb as the temperature at 800 mb level. The observed temperature gradients at the 600 mb level are compared with the critical gradients calculated by (3) and these are shown in Figures 1–5 for various seasons and for the annual mean. In these Figures continuous lines are the values of the critical gradients and the crosses give the observed gradients.

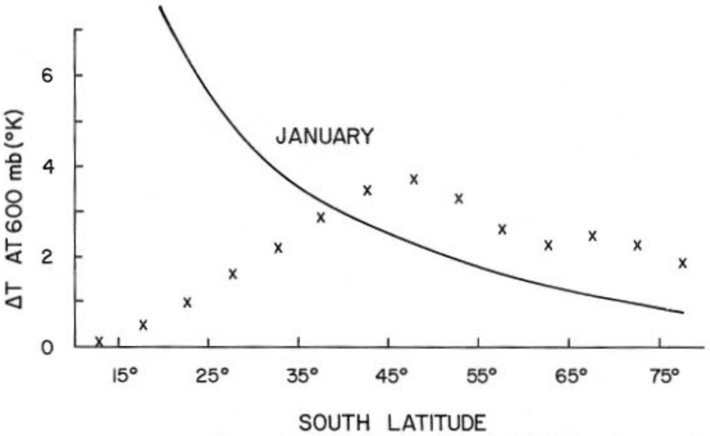

Figure 1. Temperature gradients, observed (crosses) and critical (continuous line) for January

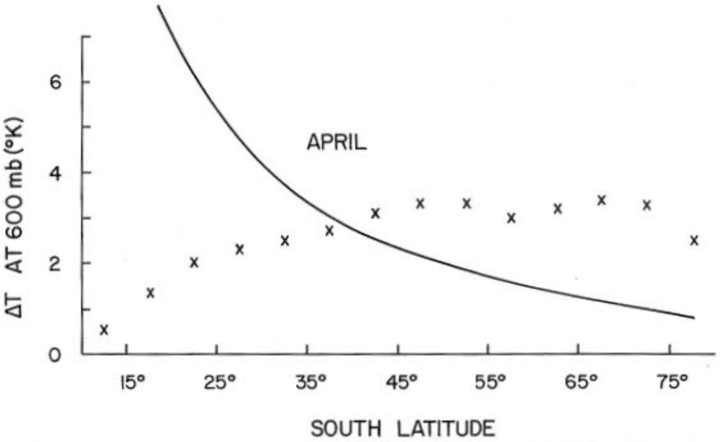

Figure 2. Temperature gradients, observed (crosses) and critical (continuous line) for April

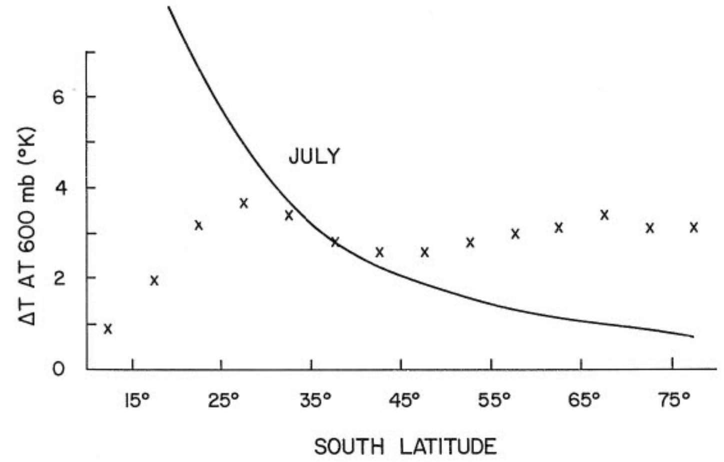

Figure 3. Temperature gradients, observed (crosses) and critical (continuous line) for July

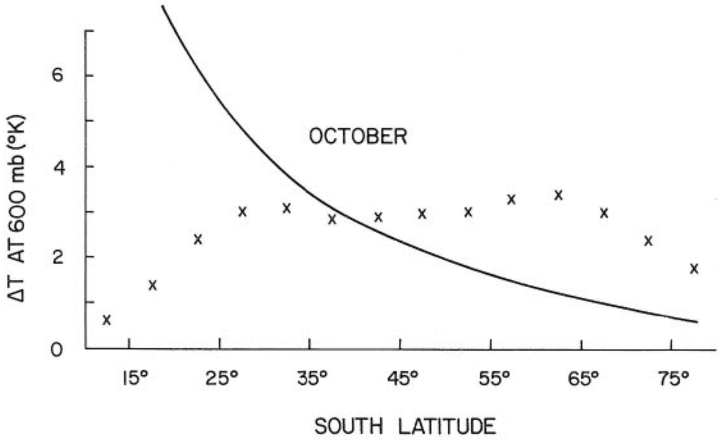

Figure 4. Temperature gradients, observed (crosses) and critical (continuous line) for October

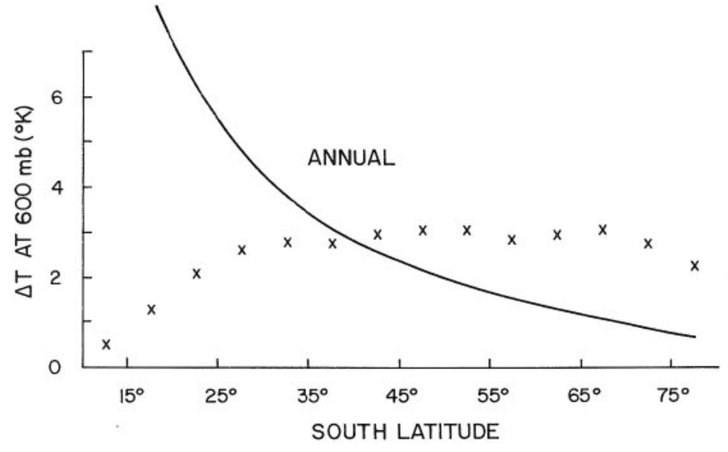

Figure 5. Temperature gradients, observed (crosses) and critical (continuous line) annual

DISCUSSION

In the month of July (winter), the observed values of temperature gradient around 37°S are approximately equal to the critical gradient values. But, in general the observed temperature gradients are much higher than the critical gradients south of about 45°S. This is different from what has been noted by Stone (1978) for the N. H.

The observed temperature gradient shows other interesting features. In all the four seasons the temperature gradient shows two maxima. This, in some way, is connected to the double maxima of the westerly jet observed by several authors and mentioned recently by Trenberth (1981). The subtropical maximum of temperature gradient shows a clear seasonal variation in position, occupying a more northerly latitude in July and a more southerly latitude in January. However, unlike what is found in the N. H., the change in the magnitude from January to July in the subtropical maximum is small. In the annual mean (Figure 5), the observed gradient has a more or less constant value over a broad belt.

The above-mentioned observations raise several questions: (1) what factors explain the supercritical nature of the temperature distribution in the S. H.?; (2) how well is a simple two layer model suited for understanding the baroclinic adjustment process in the S. H.?; (3) how far has the data inadequacy affected the results shown in Figures 1–5?

Apparently, there are no simple answers to these questions. The values of the critical gradient depend directly on the values of the static stability, which in turn depend on the zonally averaged temperatures in the atmosphere. Estimates of zonal mean temperatures in the S. H. might be biased towards continental values. In any case the large differences between the critical gradient and the observed gradients noted earlier cannot completely be ascribed to the lack of observations.

The observed thermal gradients south of 45°S are a little higher than those north of 45°N and the critical gradients south of 45°S are a little lower than their counterparts in the N. H. (Compare Figures 1–5 with those given by Stone, 1978). The latter difference is due to smaller static stability ($\theta_1 - \theta_3$ in (3)) in the S. H. There are indications in the observational study of Oort and Peixoto (1983) also that, at least in winter, the static stability in the S. H. is smaller than in the N. H.

It may be verified, by comparing the critical thermal gradients of the two hemispheres, that the static stability in S. H. in higher latitudes is about 0·7 times that in the N. H. Held (1978), using the classical baroclinic instability theory of Charney, showed that the poleward transport of sensible heat by long deep unstable baroclinic waves is proportional to the product of the square of meridional thermal gradient and the Brunt–Väisälä frequency (or square root of static stability). Oort and Peixoto (1983) noted that the poleward transport of sensible heat by transient waves in the S. H. is nearly the same as in the N. H. A simple estimation, based on the observation cited above and Held's expression for the transport of sensible heat, indicates that the meridional thermal gradient in the S. H. could be about 10 per cent higher than that observed in the N. H. This may explain in part the large difference between the two curves shown in Figures 1–5 in higher latitudes. An equally justifiable reverse argument may be offered for the lower values of static stability observed in the S. H.

Topographic forcing also seems to be important for the long baroclinic waves (Yao, 1980; Stone, 1977) which are more efficient in transporting heat polewards. The quasi-stationary waves in the middle latitudes of the N. H. transport nearly as much sensible heat as do transient waves. In the mainly oceanic S. H. long quasi-stationary waves transport negligible amounts of heat (Van Loon, 1983). Thus the total transport of sensible heat by waves in the S. H. is somewhat smaller than in the N. H. This may again explain partially the higher temperature gradients. Further, meridional variation of zonal wind in the model appears to be necessary to take into account the peculiarities of the S. H., such as the double jet.

The main utility of the baroclinic adjustment mechanism is in the climate energy balance models where the transport of sensible heat by baroclinic waves is parameterized so that the meridional temperature gradient in the model equilibrates to the critical gradient. However, the disagreement between the observed temperature gradients and the critical gradient suggests that the utility of a two-layer model without meridional variation for such purposes is less good in the S. H.

It would be worth while to repeat the calculations made here with new data sets, such as the one compiled by an Australian group (Le Marshall *et al.,* 1983). The new data sets, although not necessarily better than the one used here, might reveal the structure of the zonally averaged temperature for a different period. This is interesting in the S. H., in view of the interannual variations.

ACKNOWLEDGEMENTS

Thanks are due to Dr. A. D. Moura and Y. Yamazaki for discussions. This study is financed partially by Convênio CNPq-INPE/CIRM-Estudos teóricos e sinóticos da Circulação Atmosférica sobre a região Antártica-9503.

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