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Idealised simulations of the seasonal evolution of the middle atmosphere for a range of planetary-wave amplitudes in the lower stratosphere.

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1 INTRODUCTION

A first look at the seasonal evolution of the extra-tropical middle atmosphere shows that it is mainly driven by solar radiative heating. In the lower and middle stratosphere this results in an annual cycle which oscillates between a cold winter pole surrounded by a strong westerly vortex with a steep meridional gradient in strength and a warm summer pole surrounded by weaker easterlies. However, more detailed examination of the evolution reveals that this regular march of winds and temperatures is punctuated in the winter hemisphere by the development of large-scale disturbances. Some examples from the northern hemisphere include Canadian warmings, major midwinter warmings and final warmings; examples from the southern hemisphere include South Pacific warmings, the so-called travelling wave 2, and final warmings.

These disturbances can be divided into two broad classes: those involving the growth of a quasi-stationary anticyclone (major warmings and final warmings), and those involving the growth of eastward-travelling anticyclones (Canadian warmings, South Pacific warmings and the travelling wave 2). All involve highly nonlinear dynamics and are therefore difficult to understand. Previous workers have advanced various explanations based on particular combinations of the stratospheric state and the presence or absence of specific tropospheric phenomena - generally planetary waves.

In this work, we use a numerical model with a simplified lower boundary condition to study such planetary-scale disturbances. By carrying out a number of seasonal integrations using a range of (time independent) planetary-wave amplitudes at the lower boundary, one can "explore parameter space" under controlled conditions. This approach is essentially a numerical analogue of the sort of laboratory experiments which have already contributed a great deal to our understanding of flow regimes in planetary atmospheres (e.g. Hide and Mason, 1975). This study will concentrate on a synoptic view of the results obtained; other studies will present more conventional diagnostics.

It will be shown that phenomena reminiscent of many of those described above occur quite naturally in the simple numerical environment chosen. While it is accepted that the results obtained here cannot be compared directly to real atmospheric events, an inescapable conclusion of this work is that (at least synoptically) many of the observed disturbances could occur without prior conditioning of specific tropospheric events.

2 DESCRIPTION OF THE NUMERICAL ENVIRONMENT

The UKMO stratosphere-mesosphere model was run through a full annual cycle with the bottom boundary being held to a seasonally varying climatological zonal mean plus a constant amplitude wave one forcing. Simulations with a range of amplitudes were run, in accord with the aforementioned philosophy of exploring parameter space. At the time of writing, some further simulations with wave number two forcing have not yet been examined. In order to simplify our understanding of the results the description which follows is in terms of differences from the wave-free run (rather than from simple zonal or time means) for the weak forcing cases (up to and including 100m) and in terms of the total field for the larger perturbations.

For the range of amplitudes chosen, the model response ranged from linear (in the sense that the response initially scaled linearly with forcing) to highly-nonlinear. However, even in the weakest forcing cases, it is clear that simple ideas of waves propagating on the mean state completely fail to describe the evolution observed.

3 SYNOPTIC EVOLUTION

In all the cases of weak forcing (up to 100m geopotential amplitude) at the bottom boundary, the model produced basically the same seasonal evolution. As expected the wave one forcing did not alter the stratospheric mean state during the summer season. However, as autumn progressed the effect of the wave one forcing manifested itself initially with the advent of an anticyclone above and downstream of the maximum positive perturbation. As the circulation of the vortex strengthened, the anticyclone appeared to "lose contact" with the forcing below and was advected cyclonically around the vortex, generally weakening as it moved. As winter progressed, further anticyclones formed, strengthened and then were advected around the vortex. At times up to three such anticyclones were present in various stages of the life cycle. Analysed in terms of perturbations from the zonal mean, such events seemed very similar to the travelling

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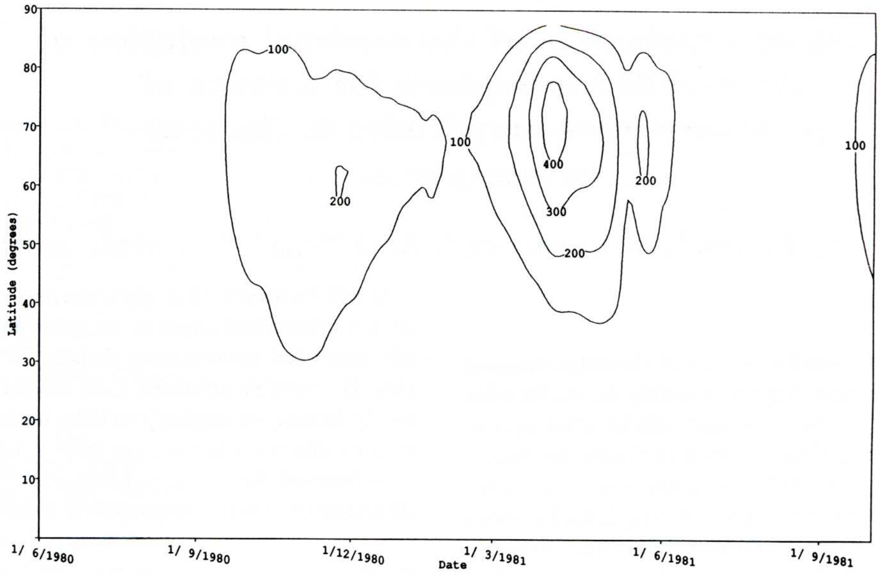

Fig. 1. Evidence for Plumb type variability with low amplitude (50m) forcing: Latitude-Time section of the amplitude (m) of zonal wave 1 at 1 mb.

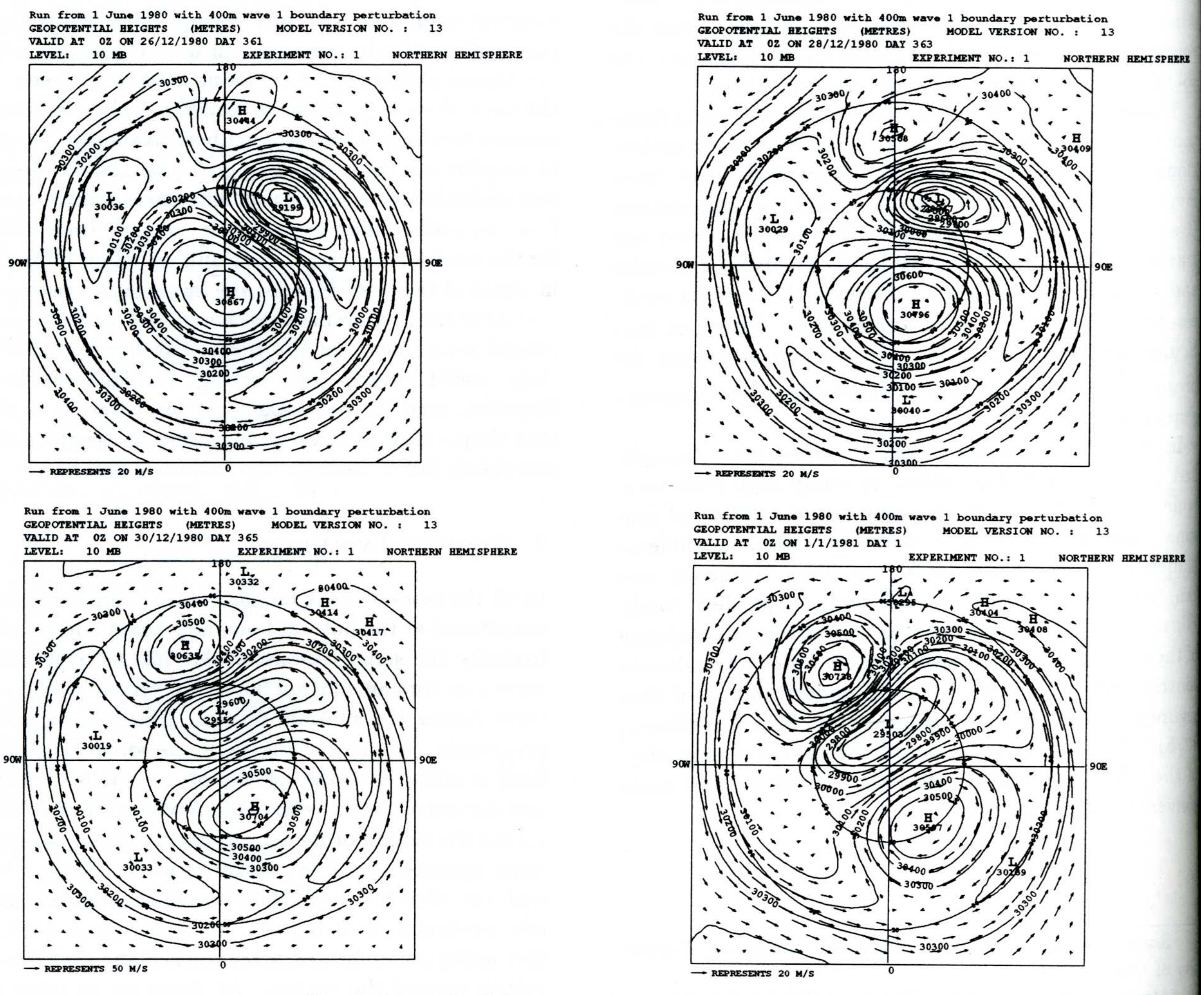

Fig. 2. Example of the onset of a travelling wave 2: 10 mb synoptic maps of the geopotential field and winds due to the 400mb forcing (late December).

wave two phenomena seen in the southern hemisphere. In later winter the general cyclonic advection of these vortices is arrested and they remain stationary or even move retrogressively against the cyclonic circulation. The final warming occurs as the growth of the stationary anticyclone in mid-latitudes becomes ubiquitous as it reverses the geopotential gradients around the polar vortex which then weakens and disappears. At these lower amplitudes, some evidence was found (figure 1) for the double peak phenomenon in seasonal planetary wave variance observed in the southern hemisphere (e.g. Plumb, 1989 and references therein). It should be stressed that in all of these cases there was no polar warming observable in the total zonal means.

With larger forcing on the bottom boundary more drastic dynamic events take place, all of which dominate the model fields. With 200m forcing, the model exhibits behaviour similar to that of weak Canadian or South Pacific (Farrara *et.al.*, 1992) warmings — although without the subsequent full stratospheric recovery. Once again, during the initial stages of winter, anticyclones grow and are advected around the polar vortex. However, with the stronger forcing the poleward and cyclonic advection result in the polar vortex being pushed off the pole. Although the anticyclone on the pole does weaken during spring, the final warming occurs as new anticyclones grow and coalesce while the (off) polar cyclonic vortex continues to weaken.

The run with the largest forcing (400m) exhibited a range of phenomena. The early growth in strength of the vortex did not occur centred on the pole and a succession of "minor" warmings occurred as anticyclones migrated around the vortex. At times a very strong travelling wave two was observed as the central vortex was squeezed back onto the pole. The sequence of warmings were interspersed with stratospheric recoveries before the final warming when the vortex was pushed off the pole. Essentially the difference between this and the lower forcing cases was that the anticylones which formed were much stronger: as one advected cyclonically, it moved poleward and displaced the vortex. As it weakened, the second began forming and the cyclonic vortex recovered its position on the pole — at which stage a strong wave-two was evident in the synoptic maps (figure 2). The resulting slowly travelling wave lasted for a brief period before the warming which pushed the cyclonic vortex off the pole and split it into two weaker surrounding vortices which were eventually washed out in a final warming.

4 Conclusions

As stated elsewhere (O'Neill and Pope, 1988) we find that the description of complex nonlinear phenomena is most usefully carried out in terms of vortex interactions. Certainly for the seasonal evolutions portrayed here it is difficult to find a more amenable framework for understanding the similarities and differences between the different runs. As in that previous work, the nonlinearity in these runs was manifested by the eastward advection of anticyclones coupled with concomitant weakening of the cyclonic vortex. However, it should be stressed that even with the weakest forcing, where we believe the dynamics to be linear, the perturbations from the control run also manifested themselves as eastward propagating anticyclones. Increasing the forcing seemed to result in a stronger poleward component of the propagation of the disturbances and produced warming-like events. Further studies of the dynamics of these events will be forthcoming.

Although the simulations described here were rather crude in terms of their exact correspondence to the observed atmosphere, they did show a range of phenomena which bear a strong resemblance to observed nonlinear disturbances in the middle atmosphere. In particular, disturbances with travelling anticyclones were produced, as well as disturbances with stationary anticyclones. It is thus plausible that such phenemona could arise in the real atmosphere without requiring transience in the tropospheric forcing.

5 References

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