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Aspects of the large-scale tropical atmospheric circulation

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ABSTRACT

An account is given of a number of recent studies with idealised models whose aim is to further understanding of the large-scale tropical atmospheric circulation. Initial-value integrations with a model with imposed heating are used to discuss aspects of the Asian summer monsoon, including constraints on cross-equatorial flow into the monsoon. The summer descent in the Mediterranean region and on the eastern sides of the summer subtropical anticyclones are seen to be associated with the monsoons to their east. An aqua-planet GCM is used to investigate the relationship between simple SST distributions and tropical convection and circulation. The existence of strong equatorial convection and Hadley cells is found to depend sensitively on the curvature of the meridional profile in SST. Zonally confined SST maxima produce convective maxima centred to the west and suppression of convection elsewhere. Strong equatorial zonal flow changes are found in some experiments and three mechanisms for producing these are investigated in a model with imposed heating.

1. Introduction

The understanding of the large-scale atmospheric circulation in the tropics is in some senses easier and in other senses more difficult than in middle latitudes. As the Coriolis parameter goes to zero, the Rossby number even on large length-scales is no longer small, and so quasi-geostrophic theory is not valid within some neighbourhood of the equator. Latent heat release in deep convection is of order one importance whereas in middle latitudes for many purposes the adiabatic dynamics sets the scene. Finally, there are mean easterlies at some longitudes and heights and mean westerlies at others, as opposed to the mean westerlies in higher latitudes, the linearisation about which is basic to many theories of the circulation there.

On the other hand, the relationship between the transient convective motions and the large scale circulations in which they are embedded may be simpler than that between the middle latitude synoptic systems and the larger space and longer time-scales in those latitudes. There is much more of a real scale separation in the tropics. The tropical responses to both sea surface temperature (SST) variations and continentality are clearly strong and direct, as seen in for example ENSO and the Asian monsoon (Charney and Shukla, 1981).

An estimate of the vertically integrated heating in June–August (JJA) is shown in Fig. 1a. Whereas in December–February the major heating is centred over the warm waters of the tropical west Pacific, in JJA this maximum is over the summer continental region of India and SE Asia. The American maximum shifts from Brazil to Central America, and the African heating follows the sun across the equator. The quasi-linear ITCZ regions are in general associated with latitudinal maxima in SST. There is cooling in the subtropics of the winter, southern hemisphere and localised cooling in the eastern N. Pacific and N. Atlantic.

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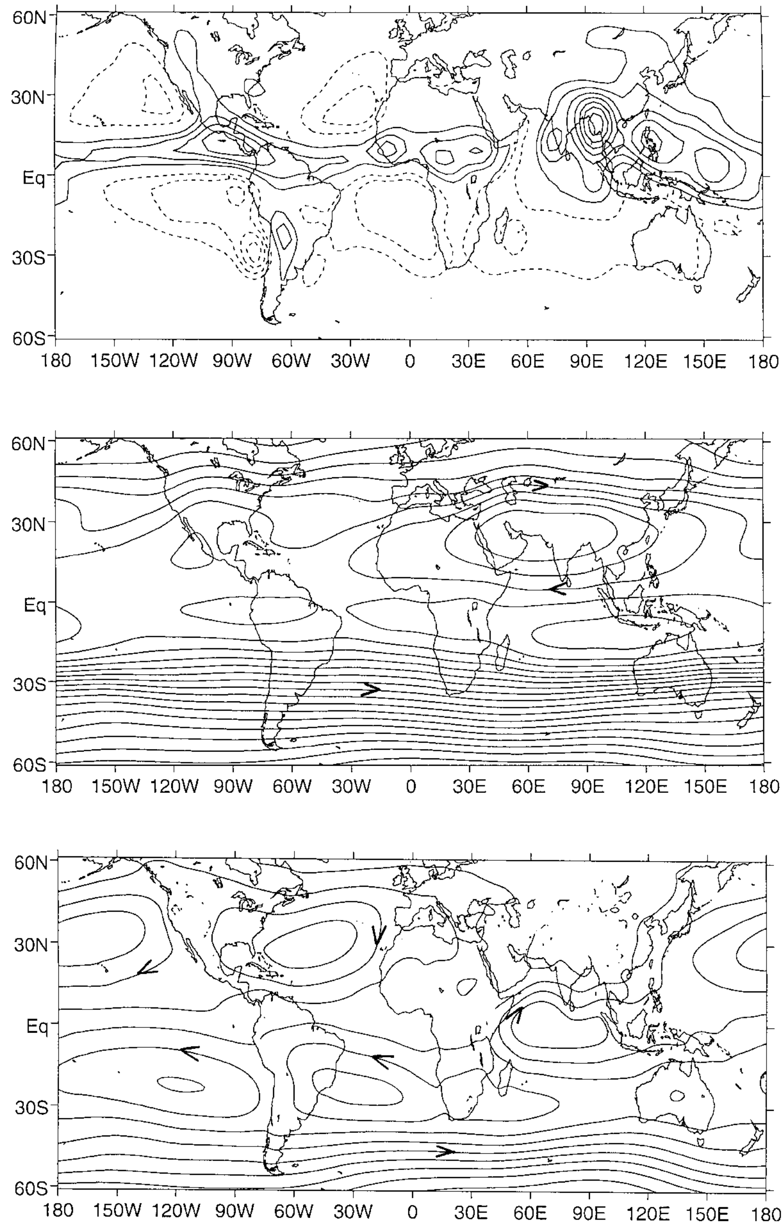

Fig. 1. Aspects of the mean climate for June–August as given by ECMWF Reanalysis data for 1979–94. (a) Vertically integrated heating, contour interval 0.5 K day^{-1}. The zero contour is not drawn. This field has been spectrally smoothed. (b) Streamfunction at 200 hPa, contour interval $10 \times 10^6\ m^2\ s^{-1}$. (c) Steamfunction at 850 hPa, contour interval $5 \times 10^6\ m^2\ s^{-1}$. In each case only the region 60°S–60°N is shown.

The upper and lower tropospheric circulations in JJA are summarised in Figs. 1b, c, respectively. In the upper troposphere the dominant feature is the anticyclone centred over southern Asia. In the southern hemisphere there is an anticyclone to the south and rather east of the Asian anticyclone. The lower tropospheric flow shows anticyclones in the northern ocean basins and a clockwise flow from the S. Indian Ocean region into the Asian monsoon heating region. There is a zonally elongated anticyclonic belt in the subtropics of the southern hemisphere.

The object of this paper is to summarise some recent studies in which the authors have been involved whose aim is to further understanding of large-scale circulations in the tropical atmosphere, including some of the features seen in Fig. 1. It is clear that the relationship between the heating and the flow field is an intimate one. However it can be argued that the existence of heating in those tropical regions, dominated by latent heat release in deep convection, is easily predictable given the nature of the lower boundary. Similar arguments can be given for some of the regions in which cooling, presumably dominated by long wave radiative loss occurs, though the strong localised nature of this, in the northern hemisphere in particular, is more surprising.

Therefore, one approach with potential to further understanding is to prescribe a heating field and determine the associated global flow field. There has been a range of such studies using the linearised primitive equations, e.g., Gill (1980), Hoskins and Karoly (1981), and more recently Jin and Hoskins (1995). Some recent research using this approach to study the Asian summer monsoon and summer subtropical descent and anticyclones will be discussed in Section 2.

Section 3 is devoted to discussing some recent studies which explore in more detail the relationship between SST and the heating, as well as between the heating and the circulation. The technique employed there is to use a full atmospheric GCM but in very idealised situations. Large variations in the tropical zonal flow are found in one of these experiments and are also found in real atmospheric data. Mechanisms for generating such zonal flow variations are discussed in Section 4, where the technique of specifying a heating field is again used.

2. Monsoons, sub-tropical anticyclones and deserts

Hoskins and Rodwell (1995) used an initial value technique to determine which parts of the flow field like that in Figs. 1b, c can be directly related to a heating field like that in Fig. 1a. They used a T31, 15-level spectral primitive equation model with only a ∇^6 hyper-diffusion, weak thermal damping towards a JJA zonal flow, and linear drag mimicking surface drag which was 4 times larger over land than sea. Over the first 5 days of the integrations, the mountains were raised and then the heating was switched on. Immediately the low-level flow over India started to accelerate. 3 days later, at day 8, the low-level cross equatorial flow began to accelerate and by day 12 it had reached very close to the observed value. For the next 8 days, this flow and all the other features mentioned earlier and seen in Figs. 1b, c were obtained as quasi-steady, quite accurately simulated, features. After this period middle latitude baroclinic instability produced unsteadiness in the tropics.

This experiment showed that the heating implied the circulation features. It is also clear that, for example, the simulated low-level flow into India and S.E. Asia would carry the moisture necessary for the latent heat release there. To some extent, the consistency of the observed climate has been verified. However, this experimental set-up enables understanding to be obtained. If now mountains are excluded and the heating is given a very small amplitude so that the calculation is linear, then the upper tropospheric circulation is still reasonably well simulated. The lower tropospheric sub-tropical anticyclones are present but not as well represented and the cross-equatorial Asian monsoon circulation is absent. Returning to finite amplitude heating improves the lower tropospheric features, but the clockwise Asian monsoon flow requires the presence of mountains in order to be well simulated.

In Rodwell and Hoskins (1995), this cross-equatorial flow into the Asian summer monsoon was studied further. This is the only season and place that such a significant, sustained cross-equatorial flow occurs. Air that crosses from one hemisphere to the other conserving its potential vorticity (PV) will tend to accelerate and move on an anticyclonic trajectory back into its original hemisphere. The only way that the air can stay a significant distance from the equator in the opposite hemisphere is for its PV to be modified to have the sign of the ambient air in that hemisphere. The impression that is present, even in the mean picture in Fig. 1c, of the air tending to curve to the south of India is a reflection of the fact that the PV of the air is still less than that of the ambient air at that latitude. The acceleration of the air into the "Findlater jet" off the Horn of Africa is associated with it still having negative PV but being in the northern hemisphere. Recently Tomas and Webster (1997) have also discussed the possible importance for the nature of the convection of cross-equatorial pressure gradients which lead to significant cross-equatorial flows carrying PV of the "wrong" sign.

The processes that modify the PV of the air from its negative value in the S. Indian Ocean region to a positive value as it crosses into Asia can be diagnosed. Shallow heating over the S. Indian Ocean starts the process. The strong drag associated with the E. African Highlands is vital and surface drag in the N. Hemisphere does the rest. It is clear that the presence of the E. African Highlands is vital to the existence of the low-level Asian summer monsoon circulation and to the nature and intensity of this monsoon.

The observed 477 hPa vertical velocity for JJA in the African and Asian region is shown in Fig. 2a. There is ascent over the Asian monsoon heating region and strong localised descent near 30°N off the coast of Africa, in the eastern Mediterranean and over the Kyzylkum desert (60°E). It can be seen in Fig. 1c that these are also regions of northerly wind in the lower troposphere.

Following Rodwell and Hoskins (1996), who studied the origin of these descent regions, it is shown in Fig. 2b that the model is able to capture these features, though there are quantitatives errors.

The ascent in the monsoon region is such that adiabatic cooling almost exactly balances the diabatic warming. However, this is not true near 35°N. There the horizontal advection of cold air from the north as well as the diabatic cooling are balancing the adiabatic warming in the descent regions. Equivalently, the air is descending at a steeper angle than, but in the same sense as, the mean isentropes.

Fig. 2c shows the vertical velocity from an identical model but with no diabatic cooling in the Mediterranean region. The descent is still there although weaker and less localised. Radiative cooling in the region is seen as providing a positive feedback on descent which is imposed from elsewhere. Further experimentation shows that it is the Asian monsoon heating that leads to this descent.

The results discussed here are consistent with

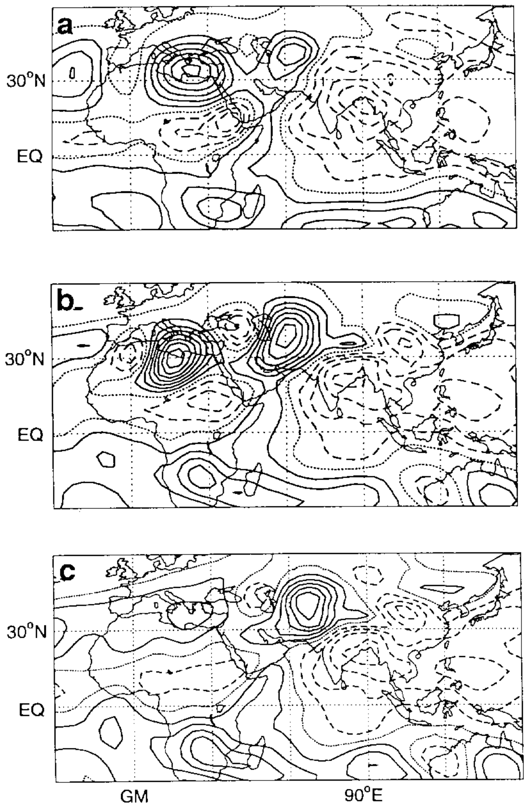

Fig. 2. Vertical velocity, ω, at 477 hPa in the region 30°W–150°E, 30°S–60°N. (a) From ECMWF analyses for June–August, (b) for day 16 of the model forced by the full heating field, and (c) as (b) but with no cooling imposed in the Mediterranean–North African region. The contour interval in each panel is 0.5 hPa hr^{-1} (adapted from Rodwell and Hoskins, 1996).

the mountain, no-mountain GCM study by Broccoli and Manabe (1992). Without the Himalayas, the Indian summer monsoon and the associated clockwise low-level circulation into it were not present. In addition the precipitation increased over the dessert region of Asia to the north and over eastern N. Africa.

There is observational support for the hypothesis of a link between the eastern N. Africa desert and the Asian monsoon from a number of sources. Data from variations of lake levels (Lézine and Casanova, 1991) and, therefore, moist events in the eastern Sahara appear to correlate well over the last 140 thousand years with expected weak Asian monsoon periods as deduced from northern hemisphere JJA insolation (Prell and Kutzback, 1987). GCM experiments by Claussen (1994) suggest that vegetation imposed over all of N. Africa would remain in the west but not in the east, consistent with non-local control of the climate in the latter region. Finally, each year the Mediterranean climate appears to settle into its summer mode at about the same time as the Asian summer monsoon becomes established.

Following this discussion, it is natural to propose (Hoskins, 1996) that the northerly winds and descent on the eastern sides of the summer subtropical anticyclones, and the anticyclones themselves, are also triggered by the monsoons on the continents to their west. The descent shown off the coast of Africa in Fig. 2c is supportive of this hypothesis. Further support is provided in Fig. 3 which shows the middle tropospheric vertical velocity and lower tropospheric stream-function when the only heating imposed is that associated with the Central and N. American monsoon, in the region 0–60°N, 110–40W. Comparison of Fig. 3b with Fig. 1c suggests that the essence of the North Pacific anticyclone structure is captured. As seen in Fig. 3a, there is descent off the west coast of N. America. As in the Mediterranean region, it can be expected that this will lead to a suppression of latent heat release and a dominance of radiative cooling in this region, consistent with the observed cooling seen in Fig. 1a. If this positive feedback cooling is added in the descent region, then the north Pacific subtropical anticyclone structure becomes more realistic. The ITCZ region heating is also helpful in increasing the realism.

3. Aqua-planet response to sea surface temperature distributions

Aqua-planet atmospheric GCMs are full GCMs but with the lower boundary simplified to that of an ocean covered earth with specified SST.

Aqua-planet experiments for prescribed simple SST distributions have two major uses:

I: They provide a means to develop hypotheses and to test and extend ideas developed in the context of simpler models with prescribed heating.

II: They provide a test-bed for studying the interaction of the dynamics and the parametrisations, particularly those for boundary layer fluxes, deep and shallow convection, and radiation.

The limitation in I is that the answers obtained may be very dependent on the GCM. However, this weakness is exploited as a strength in II, though in both uses it must be remembered that the “right” answer is not known. In use II, a standard suite of experiments, such as those described here, appear to provide a natural test-bed for new parametrizations, sitting between single column and full GCM tests, and complementing the dynamical core tests of Held and Suarez (1994) and Boer and Dennis (1997).

The focus here is on use I and results obtained with a single model, an aqua-planet version of the UK Meteorological Office Unified Model (version 2b). This model has been used for many climate change studies, e.g., Murphy and Mitchell (1995).

The response of aqua-planet models to SSTs

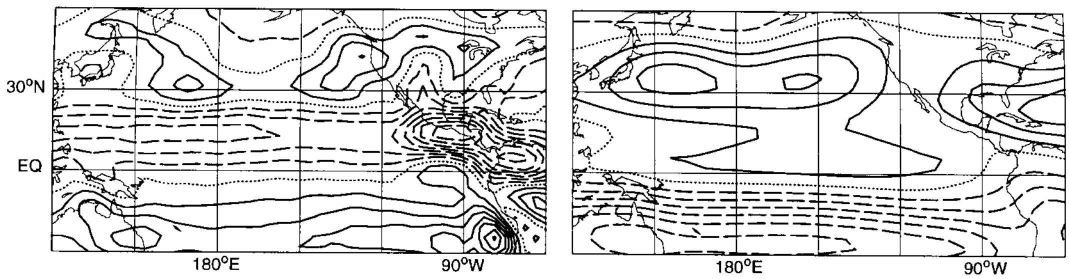

Fig. 3. Day 12 results from forcing the primitive equation model with only the heating in the American monsoon region 110°W–40°W, 0–60°N. (a) Vertical velocity,, with contour interval 0.5 hPa hr^{-1}, and (b) $\sigma = 0.89$ streamfunction with contour interval 5×10^6 m^2 s^{-1}. In each case only the region 120°E–60°W, 30°S–60°N is shown.

which are a function of latitude only has been studied by Hayashi and Sumi (1986) who were interested in the organisation of tropical convection into propagating regions, some of which had MJO-like space and time-scales. Hess et al. (1993) using an aqua-planet model with zonal SSTs showed how, for a particular meridional profile different convection schemes lead to the convective maximum occurring on or off the equator. They also showed the sensitivity to the meridional profile of SST.

The aqua-planet UM has been run for a range of zonally symmetric SST distributions and equinoctial insolation to study the nature and sensitivity of the response of a particular model. Hemispherically symmetric SST runs have been made with a range of SST profiles from 27°C at the equator to 0°C at 60° north and south. All of them have an SST of 0°C between 60° and the pole in both hemispheres. A linear profile of SST between the equator and 60° of latitude gives a very strong but narrow convective maximum on the equator. For an SST profile with a smoother equatorial behaviour, proportional to the square of latitude, the equatorial convective maximum is somewhat weaker and broader. However, for an even flatter equatorial SST profile, shaped as the fourth power of latitude, there is a broad region of weak convection in the tropics. This change in behaviour is accompanied by a disappearance of the tropical Hadley cells.

The behaviour in these experiments is consistent with the ideas of Held and Hou (1980). If it is assumed that a local radiative–convective balance occurs in the tropics, then the meridonal SST gradient will be communicated to the atmosphere. From thermal wind balance, assuming a near zero surface wind, this implies an upper tropospheric wind. It is hypothesised that this local radiative–convective balance will occur if the implied upper tropospheric winds are not too large to be attainable for an angular momentum conserving trajectory from zero velocity at the equator. If the equatorial SST gradient is not flat enough for the implied winds to be attainable then Hadley cells and concentrated equatorial convection are expected to occur.

The curvature in the equatorial SST profile at which the transition from a Hadley cell equatorial convective peak to a broad tropical region of almost local radiative–convective equilibrium occurs is likely to be sensitive to the parametrizations, particularly that of convection, used in the model. (In this model a mass flux convection scheme, Gregory and Rowntree, 1990 is used.) This test could form part of the GCM benchmark (use II).

The aqua-planet model can be used to look at the impact of a local anomaly in tropical SST in the simplest situation of an otherwise zonal SST distribution. A tropical anomaly of maximum value 3°C when added to the intermediate meridional gradient, zonal SST field yields the pattern shown in Fig. 4a. The convective precipitation for a GCM run with this SST field is shown in Fig. 4b. The difference of this precipitation from that in the control run without the local SST anomaly is shown in Fig. 4c. Here the finer contour interval for negative values should be noted. There is a region of significantly enhanced equatorial precipitation with its maximum slightly to the west of the SST maximum. However, there is also a widespread weaker reduction in precipitation, particularly to the west and off the equator and to the east near the equator. These minima may be associated with Rossby and Kelvin wave responses. However, detailed diagnosis shows that the response to the SST anomaly even in this very idealised situation is far from the simple first baroclinic mode in the vertical, Rossby and Kelvin wave response suggested by a Gill-type model. The very simple models of atmospheric response to SST designed for use in conjunction with tropical ocean models, such as that of Lindzen and Nigam (1987), can be tested against the aqua-planet GCM results for such idealised situations. In general, qualitative agreement can be obtained but the parameters required for quantitative agreement appear to vary with the position of the source and with location with respect to the source.

A zonal wavenumber 1 in equatorial SST of amplitude 3°C when added to the same zonal profile produces an equatorial convection with magnitude ranging along the equator from near zero to about double that in the control. Again the convective maximum is to the west of the SST maximum. The response of the upper tropospheric rotational flow to the strong zonal asymmetry in convection is very large. The upper tropospheric equatorial zonal wind varies from a westerly of 45 ms^{-1} to weak easterlies which occur in the

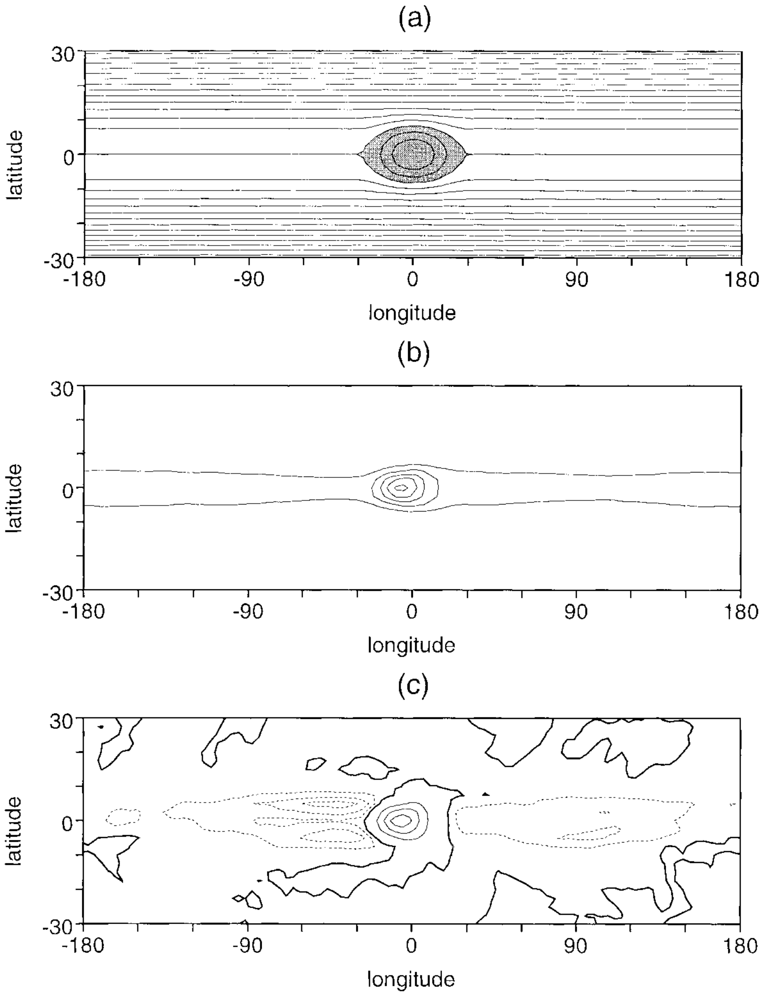

Fig. 4. An aqua-planet experiment with a localised equatorial SST anomaly. (a) The SST field with contour interval 2°C. The maximum SST is 30°C and the shaded region has SST greater than 27°C. (b) The convective precipitation with contour interval 10 mm day^{-1} (c) The anomaly in the convective precipitation from the control with zonally uniform SSTs. Here the contour interval for positive values is still 10 mm day^{-1}. The zero contour is heavy. Negative contours, shown as dashed, are drawn every 2 mm day^{-1}. In each panel only the region between 30°N and 30°S is shown.

region of the convective maximum. The zonally averaged wind for the control and this wavenumber one experiment are shown in Fig. 5. The lower tropospheric flows are similar. However in the upper troposphere weak equatorial easterlies have been replaced by 22.5 ms^{-1} westerlies, and the sub-tropical westerly jets are significantly weaker.

As in the zonal SST cases, the dependence of the response to local and wavenumber one tropical SST anomalies on the parametrizations employed in the model would be of great interest and could form part of a standard test.

4. Tropical zonal flow

One of the interesting aspects of the aqua-planet experiments was the generation of large super-

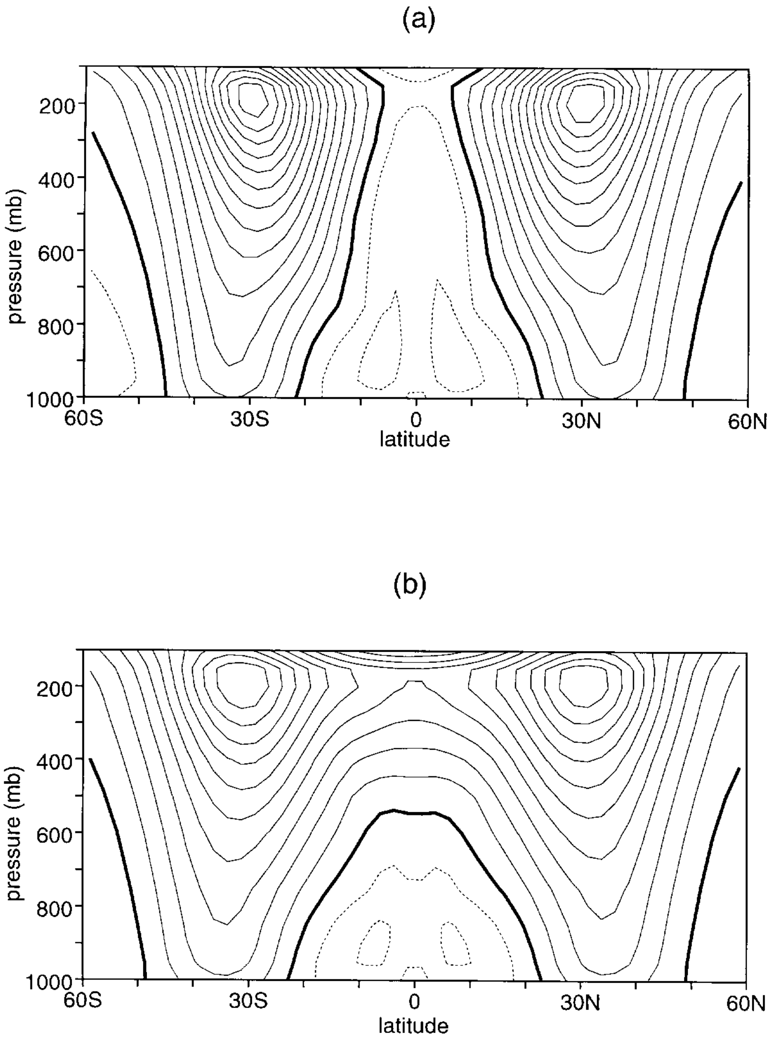

Fig. 5. Zonally averaged zonal winds for two aqua-planet experiments: (a) the control with zonally uniform SSTs, and (b) an experiment with a 3°C amplitude zonal wavenumber are added to the equatorial SSTs. The contour interval in each case is 5 ms^{-1} with the zero contour thickened and negative contours dashed.

rotation in the upper troposphere on the equator. Saravanan (1993) has also discussed a sudden transition to super-rotation in 2-level models such as the GCM of Suarez and Duffy (1992). However, such a transition has not been found in models with higher vertical resolution (Saravanan, 1993; Woolnough, 1997). Fluctuation in the upper tropospheric equatorial wind has been found to be an integral and robust part of the Madden Julian Oscillation (MJO). This behaviour has been used by Slingo et al. (1996) to diagnose the MJO simulations in atmospheric GCMs. Consequently it is of interest to consider mechanisms for the generation of tropical zonal wind anomalies.

The equation for the zonal wind perturbation from a 3-D equilibrium flow with x and y components U and V may be written with some approximation:

$$\frac{\partial[u]}{\partial t} = \underbrace{\left(f - \frac{\partial[U]}{\partial y}\right)[v]}_{1} - \underbrace{\frac{\partial}{\partial y}[u^*v^*]}_{2} - \underbrace{\frac{\partial}{\partial y}[U^*v^* + V^*u^*]}_{3}, \qquad (1)$$

where u and v are the perturbation horizontal wind components, [ ] denotes a zonal average and ( )* the departure from this.

Term 1 is the Hadley cell term. If there is a perturbation to the zonally averaged meridional wind, this will lead to a tendency in the zonal wind. To illustrate the nature of this, a very small amplitude deep near equatorial heating has been applied to a 15-level, T31 spectral model in which there is a basic state composed of a Dec–Feb climatological zonal flow. Fig. 6a shows the zonal flow perturbation at day 6. Consistent with the dominance of term 1, there are sub-tropical westerlies in the upper tropospheric outflow levels and easterlies in the lower tropospheric inflow levels. The signs would be reversed for a zonally averaged cooling anomaly.

Term 2 is the non-linear momentum flux convergence associated with the perturbation. To give an example of this the same model has been run with a finite amplitude, deep heating centred on the equator but in a restricted longitudinal region, with uniform cooling elsewhere in longitude to give zero zonal mean. The day 6 zonal flow perturbation for this experiment is given in Fig. 6b. This time there are upper tropospheric westerlies on the equator and easterlies in the sub-tropics. Almost the same picture is obtained for localised cooling balanced in longitude by weak heating. An interpretation of this is that stationary Rossby waves propagate polewards in both hemispheres in the upper tropospheric westerlies that reach almost to the equator. As they propagate polewards their phase-lines are tilted, north-west to south-east in the northern hemisphere, and westerly momentum is transported equatorwards. For a finite amplitude local heating without compensating cooling, a mixture of Figs. 6a, b is produced. In particular there are broad westerlies in the upper troposphere.

Analysis shows that this is the dominant mechanism operating to produce the equatorial super-

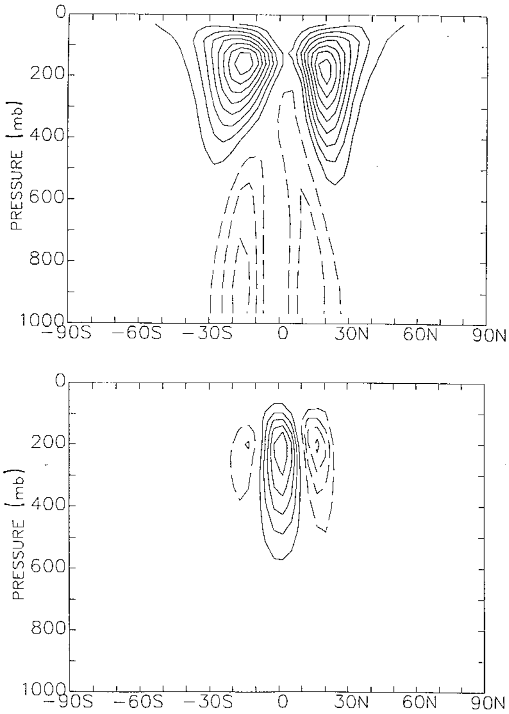

Fig. 6. Zonally averaged perturbation zonal winds for day 6 of a primitive equation model with a zonal Dec–Feb climatological basic flow, and a linear low-level drag on perturbations to it. (a) The linear response to zonal heating in the equatorial region. For the zonal component of a localised heating with maximum vertically averaged heating of 2.5 K day^{-1} in a region 60° of longitude and 30° latitude, the contour interval is 0.4 ms^{-1}. (b) The non-linear response to the heating mentioned in (a) but this time for the zonally asymmetric part, i.e. with the zonal average removed. The contour interval is 0.4 ms^{-1}.

rotation in the wavenumber 1 SST aqua-planet experiment discussed in Section 3. Saravanan (1993) found in his 2-layer model that as the equatorial easterlies decreased and became westerly the sub-tropical structure of the mid-latitude transients responded. Consistent with refractive index ideas there was a reduction in their transport of westerly momentum from the tropical to mid-latitude regions. This provided a strong positive feedback on the westerly tendency in the equatorial region. However, initial analysis suggests that this feedback does not occur in the aqua-planet GCM runs; the mid-latitude transients do not appear to reach deep enough into the tropics to be significantly affected by the equatorial wind changes.

The final term is linear in the perturbation and is dependent on the longitude of the heating with respect to the zonally asymmetric basic state. For example, at some longitudes a perturbation with equatorward winds would flux westerly momentum towards the equator, whereas at other longitudes it would flux easterly momentum. Examples are now shown for small amplitude, local equatorial heating with compensating cooling, perturbing a full 3-D Dec–Feb climatological flow. The model includes a forcing to make this basic state a steady solution. The day 6 zonal flows for heating centred at 150°W and 90°E are given in Fig. 7a, b, respectively. For the former case westerly momentum is transported from the sub-tropics to the equator and for the latter it is transported in the reverse direction.

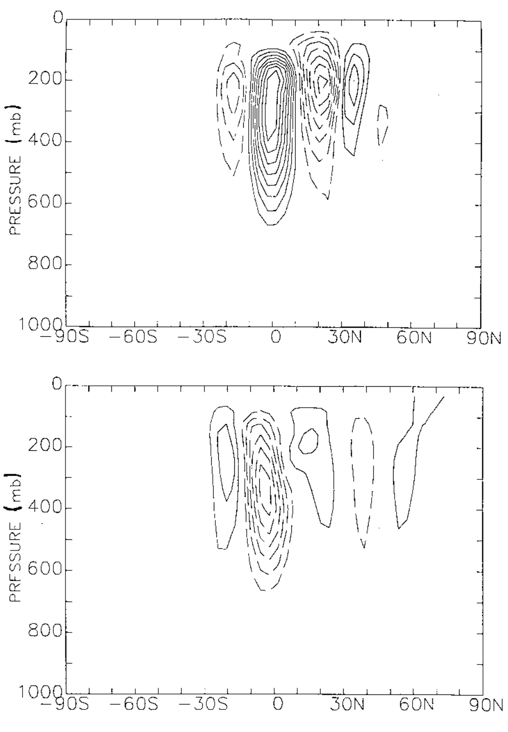

Fig. 7. Zonally averaged zonal winds for day 6 of a primitive equation model with a 3-D Dec–Feb climatological basic flow, a linear low-level drag on the perturbation, and a localised heating as used for the result shown in Fig. 6b, but in a linear calculation. The heating is centred at (a) 150°W and (b) 90°E. The contour interval is 0.4 ms^{-1}.

For a moving heating, such as that in the MJO, the equatorial zonal flow tendency will therefore change in time. Consequently this mechanism can be proposed to explain the equatorial zonal flow oscillation associated with the MJO.

5. Concluding comments

An account has been given of a number of recent studies with idealised models whose aim has been to probe aspects of the relationship in the tropics between surface conditions, the release of latent heat in large-scale convective regions, the suppression of convection elsewhere, and large-scale horizontal circulations.

Prescribed seasonal mean heating in a simple primitive equation model with specified zonal flow and mountains allowed discussion of a number of aspects of the atmosphere in the tropics and sub-tropics of the northern hemisphere. The potential vorticity constraint that limits cross-equatorial flow is broken for the clockwise low-level monsoon flow into India through a number of processes, but in particular by friction in the region of the east African highlands. Equatorward and descending flow leading to the dry summer conditions in the eastern Mediterranean/North African region were related directly to the Asian summer monsoon heating. Similarly it is proposed that the basic nature of the summer subtropical anticyclones in the ocean basins is set by the monsoons to their east.

An aqua-planet version of a Meteorological Office/Hadley Centre model was used to show the sensitivity of the distribution of tropical convection to the meridional gradient in SST. For a sufficiently flat equatorial distribution no equatorial convective maximum or Hadley cells occur. Idealised positive equatorial anomalies in SST, in this model, produce strong convective maxima which tend to be centred to the west of the SST maxima. There is suppression of convection elsewhere in the tropics in a manner suggestive of simple equatorial wave responses, though the flow suggests that the response is, in fact, more complicated than this.

It is proposed that aqua-planet experiments such as these would provide an excellent test-bed for the interaction of the dynamics and the physical parametrisations in atmospheric GCMs.

The aqua-planet experiments with SST anomalies showed interesting tropical zonal flow variation depending on the nature of the anomaly. The mechanisms for generating tropical zonal flows were analysed with the aid of prescribed tropical heating experiments. Examples were presented of zonal flows generated by Hadley cells, by poleward propagating Rossby waves and by the momentum fluxes associated with heating induced circulations at different longitudes with respect to a zonally asymmetric basic flow. The second, non-linear mechanism was considered to be responsible for the tropical zonal flows in the aqua-planet SST anomaly experiments. The third, linear mechanism was proposed as being operative in the tropical zonal flow fluctuations in the MJO.

More details of the unpublished studies summarised here will be given elsewhere. However, it is hoped that the presentation here of a range of experimentation with idealised models of differing complexity has shown that they play an important rôle in the production and development of hypotheses. These should then be tested using observed data and more complete models. Experimentation such as that described helps to build the necessary framework for diagnosing the results from GCMs and for improving their performance.

6. Acknowledgements

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