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A linear study of the effects of heating and orography on westerly airstreams with particular reference to monsoonal flow over northern Australia

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A linear, steady-state model for the forcing of unsheared westerly air-flow by orography and surface heating has been produced. In a westerly air-flow, topography produces an upstream effect which extends further than in an easterly. Downstream, the topography forces a mixture of Rossby waves as well as shorter internal gravity waves modified by the Coriolis force. Long Rossby waves are not excited by easterly flow, so that upper air perturbations in a westerly have larger amplitudes than in easterly flow.

A bounded, steady-state solution to the problem of westerly winds over a heated surface cannot be found. A bounded solution can be obtained if the appropriate amount of compensatory cooling is introduced on the eastern side of the heating. The heating results in negative perturbation pressures, and an overall equatorwards displacement of the flow, while the cooling produces positive perturbation pressures and a polewards displacement of the flow.

Introduction

Adams (1993), hereafter A93, examined the effects of orography and heating on an unsheared easterly flow. The object of that study was to model the easterly trade-wind flow, which is prevalent over northern Australia throughout the year. In summer the easterlies over northern Australia are interrupted, from time-to-time, by bursts of westerly, monsoonal flow. Well-developed, westerly, monsoonal flow can persist for several days over much of northern Australia. The object of the present study is to develop a model for the perturbations induced by orography and land/sea temperature contrast upon summertime monsoonal flow over northern Australia. Although there is a vast literature concerned with the effects of orography on westerly flow, few analytical studies have included the effects of internal gravity waves, the Coriolis force, and the variation of the Coriolis parameter with latitude, as is done here. Furthermore, few studies have concentrated on tropical westerly flow.

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The development follows from the work of A93, where the general, steady-state, linearised equations for unsheared, zonal flow of an hydrostatic, Boussinesq fluid over low orography and a surface heat source are derived. The Coriolis force and the beta effect are included. The unsheared flow is assumed to have a constant base-state velocity and a fixed Brunt-Väisälä frequency.

The equations derived in A93 are noted in the next section, and applied specifically to westerly flow thereafter. The paper concludes with a summary.

Governing equations

A93 obtained an equation for the Fourier transform of the vertical velocity, w, for steady-state, unsheared, zonal flow (velocity U) with fixed Brunt-Väisälä frequency, N, over low orography, h(x), and including a surface-based diabatic heating function, q(x,z). Here x is zonal distance, positive to the east, measured from some origin, and z is the vertical coordinate. The Coriolis parameter, f, and its meridional gradient, β, assumed constant, are included. The equation obtained in A93 is

$$\tilde{w}_{zz} + m^2 \tilde{w} = m^2 \tilde{q}/N^2 \qquad \ldots 1$$

where $\tilde{w}(k,z)$ is the Fourier transform of w(x,z), $\tilde{q}(k,z)$ is the Fourier transform of q, k is zonal wave number, m is vertical wave number, and the subscript z means differentiation with respect to z. It is found that

$$m^2 = \frac{N^2 (k^2 - K^2)}{U^2 (k^2 - L^2)} \qquad \ldots 2$$

where $K^2 = \beta/U$ and $L^2 = f_0^2/U^2 + \beta/U$, f_0 being the value of f at a central latitude. The convention is adopted that m is always positive.

The form of the solution to Eqn 1 depends on the sign of m^2. When $m^2 < 0$, it is useful to work in terms of the real quantity μ, defined so that $\mu^2 = -m^2$ and $\mu > 0$. A93 showed that when $m^2 > 0$, the general solution to Eqn 1 is given by

$$\tilde{w} = A_1(k) \exp(imz) + A_2(k) \exp(-imz) + \frac{im}{2N^2}\left[\exp(-imz) \int \tilde{q} \exp(imz)\, dz - \exp(imz) \int \tilde{q} \exp(-imz)\, dz\right] \qquad \ldots 3(a)$$

When $m^2 = 0$,

$$\tilde{w}_{zz} = 0 \qquad \ldots 3(b)$$

and when $m^2 < 0$,

$$\tilde{w} = A_1(k) \exp(\mu z) + A_2(k) \exp(-\mu z) + \frac{\mu}{2N^2}\left[\exp(-\mu z) \int \tilde{q} \exp(\mu z)\, dz - \exp(\mu z) \int \tilde{q} \exp(-\mu z)\, dz\right] \qquad \ldots 3(c)$$

where A_1 and A_2 have to be determined from the boundary conditions. The linearised lower boundary condition is

$$w(x,0) = U\, dh/dx \qquad \ldots 4$$

The perturbation quantities must be bounded as z approaches infinity, and there is a radiation requirement that the vertical component of the group velocity should be directed upwards.

Expressions have been derived in A93 for the vertical and zonal components of the group velocity:

$$c_{gz} = \frac{2kU (k^2 - K^2) (k^2 - L^2)}{(\pm m) [2(k^2 - K^2)^2 + f_0^2K^2/U^2]} \qquad \ldots 5$$

and

$$c_{gx} = \frac{2f_0^2k^2/U}{2(k^2 - K^2)^2 + f_0^2K^2/U^2} \qquad \ldots 6$$

It is helpful to work in terms of a perturbation streamfunction Ψ related to w by

$$w = -\partial\psi/\partial x \qquad \ldots 7$$

As discussed in A93, it is straightforward to obtain expressions for the perturbation pressure, temperature, zonal velocity and meridional velocity from the perturbation streamfunction. If the meridional velocity tends to zero far upstream, it is possible to determine the flow trajectory.

Solutions

Solution for the streamline

When the basic flow is from the west, a complication arises which does not occur in easterly flow. The difficulty is due to the inclusion of β in the equations. Since $U>0$ in westerly flow, the factor (k^2-K^2) can be negative. Equation 2 shows that there are three regions in k-space to be considered. If $k^2<K^2$, then $m^2>0$, and the solutions to Eqn 1 are wave-like. This wave-like contribution from the long waves does not occur in easterly flow, and this region of k-space will be called region 1.

If $K^2<k^2<L^2$, then $\mu^2>0$ and the solutions are evanescent. This region of k-space will be called region 2. For the monsoonal base state, the same values of the constants will be used as in A93 ($N=10^{-2}$ s^{-1} and a central latitude of 20°S), except that U is given a value $+10$ m s^{-1} instead of -10 m s^{-1}. For these values of the parameters, the demarcation wavelength $2\pi/K$ (the barotropic Rossby wavelength) between the wave-like solution of region 1 and the evanescent solution of region 2 is 4285 km.

When $k^2>L^2$, the solutions are again wave-like, and we shall call this region of k-space region 3. For the values of N, U and the central latitude given, the demarcation wavelength $2\pi/L$ between regions 2 and 3 is 1210 km. The radiation condition that $c_{gz}>0$ must be applied in regions 1 and 3, while the boundedness condition for w as z tends to infinity imposes a restriction in region 2.

When $k = K$, then $\mu = m = 0$ and Eqn 1 becomes $\tilde{w}_{zz} = 0$ with solution $\tilde{w} = Az + B$, where A and B may be functions of k. As w must be bounded as z approaches infinity, $A = 0$. A similar consideration applies to ψ and shows that $B = 0$ also. Thus, there are no contributions to the integrals from $k = K$. When $k = L$, μ and m approach infinity and Eqn 1 degenerates into $\tilde{w} = \tilde{q}/N^2$.

In order to apply the condition that $c_{gz} > 0$ in regions 1 and 3, we need to examine the expression for c_{gz}, Eqn 5. In region 1, the terms (k^2-K^2) and (k^2-L^2) are both negative, so that their product is positive. As U and k are positive, and since we have adopted the convention that m (and μ) should be positive, the positive sign of $(\pm m)$ in Eqn 5 is required to ensure positive c_{gz}. In region 3, the terms (k^2-K^2) and (k^2-L^2) are both positive, and the positive sign with m is again appropriate. Equation 6 shows $c_{gx}>0$ when $U>0$, showing that the flow of wave energy is downstream.

Application of the lower boundary condition allows an expression for the Fourier transform of the vertical velocity to be determined. From Eqn 3, for $0 \leqslant k < K$, and for $k > L$:

$$\tilde{w} = \left[U\tilde{h}_x(k) + \frac{m}{N^2}\int_{(z=0)} \tilde{q}\sin(mz)dz\right]\exp(imz) +$$

$$\frac{im}{2N^2}\left[\exp(-imz)\int \tilde{q}\exp(imz)dz - \exp(imz)\int \tilde{q}\exp(-imz)dz\right] \quad \ldots 8(a)$$

while for $k = L$,

$$\tilde{w} = \tilde{q}/N^2 \quad \ldots 8(b)$$

and for $K < k < L$,

$$\tilde{w} = \left[U\tilde{h}_x(k) - \frac{\mu}{N^2}\int_{(z=0)} \tilde{q}\sinh(\mu z)dz\right]\exp(-\mu z) +$$

$$\frac{\mu}{2N^2}\left[\exp(-\mu z)\int \tilde{q}\exp(\mu z)dz - \exp(\mu z)\int \tilde{q}\exp(-\mu z)dz\right] \quad \ldots 8(c)$$

where $\tilde{h}_x(k)$ is the Fourier transform of the orographic gradient dh/dx. The $(z = 0)$ tag means that the integrals are evaluated at $z = 0$. The vertical velocity is obtained from Eqn 8 by application of the inverse Fourier transform. The solution is the arithmetic sum of the orographic and heating effects. This allows for separate examinations of the two forcings.

The orographic effect

The Great Dividing Range is assumed to have bell-shaped orography of form $h(x) = h_0/(1 + x^2/a^2)$, where h_0 is the maximum height, and a may be considered as one-quarter of the orographic width. The formula for the perturbation streamfunction is given by

$$\psi = -Uh_0 a\left\{\int_0^K \exp(-ak)\cos(kx + mz)\,dk + \int_K^L \exp[-(ak + \mu z)]\cos(kx)\,dk + \int_L^\infty \exp(-ak)\cos(kx + mz)\,dk\right\} \quad \ldots 9$$

The wave-like part of the flow consists of short waves and long waves. The short waves occur also in easterly flow over orography; they are found downstream of the orography, and are modified internal gravity waves. In the limit as k tends to infinity, these waves become pure gravity waves. These waves display the upstream slope with height which is characteristic of flow over orography, and which arises from the imposition of the radiation boundary condition.

The long waves are slightly modified stationary Rossby waves, their westward phase speed balanced by the westerly base-state flow. In the limit as k tends to zero, these waves become unmodified Rossby waves. The influence of these waves is felt mainly in the perturbation pressure and meridional velocity fields; the zonal and vertical velocities associated with the long waves are very small. Pierrehumbert (1984) has shown that, provided the length-scale of the terrain is considerably less than the stationary Rossby wavelength (as here), the forcing of Rossby waves is proportional to the mountain volume.

The evanescent portion of the solution is similar to its counterpart in easterly flow. It is symmetric about the origin for symmetric orography, has maximum amplitude at the orographic peak, and becomes small far from the orography.

Figures 1(a) and 1(b) show the perturbation streamfunction and total streamfunction in westerly flow over bell-shaped orography centred at 20°S with a maximum height of 800 metres and a width of 400 km. These are the same values that were used in A93. Comparison with the results of A93 shows the upstream influence extends further for westerly flow than easterly flow. The perturbation pressure in the vicinity of the orography is illustrated in Fig. 1(c). There is an area of high pressure upstream of the orography with the maximum value of perturbation pressure the same as in easterly flow. However, downstream, the perturbation pressure does not decrease steadily as in the easterly case but displays a marked wave-like structure with a well-defined wavelength $2\pi/K$. This is the barotropic Rossby wavelength which depends only upon the magnitude of the basic flow and the latitude. The waves tilt slightly to the west with height, and display no amplitude variation with height. They decrease gradually in amplitude with increasing distance downstream. This Rossby wave structure is not evident in the form of the perturbation zonal and meridional velocity fields which are similar in form to their counterparts in easterly flow.

Figure 1(d) shows flows in the x-y plane at $z = 0$. The corresponding flow for the easterly case is superimposed. Surface meridional displacements begin roughly twice as far upstream (about 4400 km) as in easterly flow. The maximum meridional deviation at the surface is about 150 km, comparable in magnitude to the value in easterly flow. However, there is a major difference between easterly and westerly flows downstream of the orography. The well-defined wave-like structure with a wavelength of 1320 km in an east-

Fig. 1(a) Perturbation streamfunctions in the x-z plane for westerly flow over bell-shaped orography. Values in $m^2 s^{-1}$.

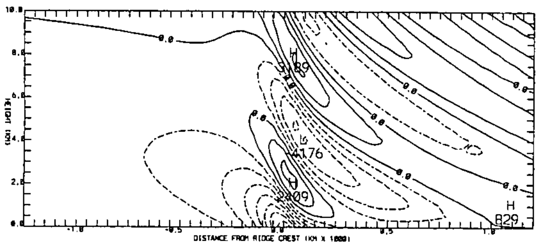


Fig. 1(b) Same as Fig. 1(a) except total streamfunctions. Values in units of $10^2 m^2 s^{-1}$.

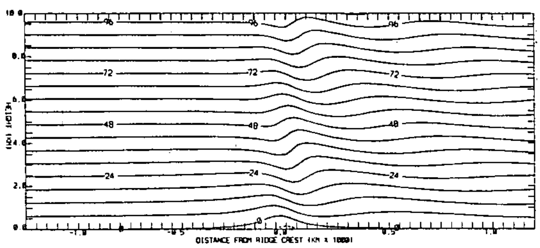


Fig. 1(c) Same as Fig. 1(a) except perturbation pressure. Values in hPa.

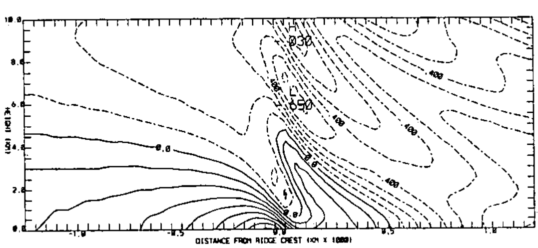


Fig. 1(d) Horizontal projection of a streamline at $z = 0$ for westerly flow over bell-shaped orography centred at $x = 0$. Zonal extent of major orography indicated by shaded region. x-axis labelled in units of 1000 km. y-values in km. Dashed line shows corresponding result for easterly flow.

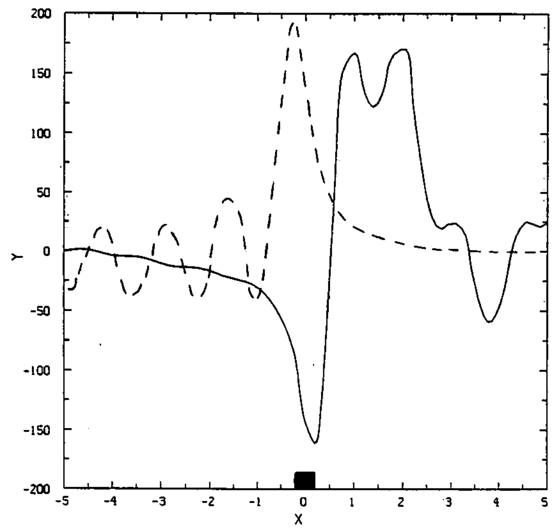
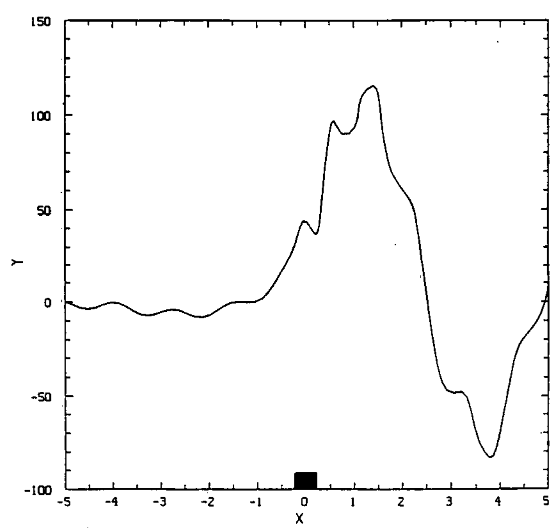


Fig. 1(e) Same as Fig. 1(d) except at $z = 5$ km.



erly regime is replaced in the westerly case by a complex interference pattern between Rossby waves and internal gravity waves modified by inertial effects. As the Rossby waves do not decay with height, upper-level westerly flow over orography displays larger meridional displacements than are evident in easterly flow. Figure 1(e) displays the flow at a height of 5 km.

As in the easterly case, different orographic profiles give broadly similar results to bell-shaped orography, although there are considerable differences in position between the major meteorological features close to the peak of the orography.

It is of interest to compare the present results for westerly flow with those of Eliassen and Rekustad (1971) who examined flow over the Scandinavian mountain range, Queney (1973), Pierrehumbert (1984) and Zehnder and Gall (1991) whose quasi-geostrophic analysis was directed at the influence of Central American orography. Queney's (1973) analysis for a barrier 300 km wide does not include consideration of the beta effect. It shows the westerly flow beginning to deviate polewards at an upstream distance of the same order of magnitude as the orographic width, the surface pressure maximum lying upstream of the streamline ridge (as in the present work), and a lee streamline trough lying about 300 km from the highest orography. The downstream streamlines and isobars do not display the wave-like features of the present study, though Queney's earlier (1947) work shows waves on the downstream streamlines.

Eliassen and Rekustad (1971) used a numerical technique to compute the time evolution of a vertically sheared initial state until it no longer changed appreciably. The beta effect was not included. Also, the Scandinavian mountains are at a latitude of 65°N where the Coriolis parameter is 2.7 times as large as in the present study. Eliassen and Rekustad's results for flow in the x-z plane are very similar to those of the present analysis, though the use of artificial friction in their model reduces the amplitude of upper-level disturbances. In the x-y plane, the present work gives results which are quite different from those of Eliassen and Rekustad. The main differences are that the Eliassen and Rekustad model does not admit Rossby waves and the upstream influence in their model extends only about 200 km from the highest orography. Close to and upstream of the orography, their flow deviates polewards as it does in the current work; just downstream of the orography, their flow turns equatorwards as it does here. However, the magnitude of the meridional deviations is larger in the present work; there is no evidence in the Eliassen and Rekustad (1971) model of wave-like perturbations in the x-y plane, and the flow results given here display a far more detailed structure.

Pierrehumbert (1984) expressed the Queney (1947) solutions for Boussinesq, hydrostatic, incompressible flow on an f-plane over orography in terms of Green's functions. Results for the horizontal perturbation velocities at z = 0 are very similar to those of the present study except that the upstream perturbation velocity values are somewhat smaller than those of Pierrehumbert.

Zehnder and Gall's (1991) results for westerly flow display an upstream equatorwards deviation beginning about 1000 km from the orographic crest. A ridge is formed close to the crest, but there are no lee waves. The lack of lee waves in the Zehnder and Gall study appears to be related partly to the width of the orography, and partly to the inclusion of a base-state meridional velocity in the analysis. The inclusion of these factors means that the preferred waves in a westerly air-flow have wavelengths which are too long to be evident, even on a planetary scale. Furthermore, Zehnder and Gall (1991) state that gravity waves, which have been filtered out in their work, may have some importance at the length scales under consideration.

The heating effect

We shall discuss forcing by the heating function given by

$$q = A\exp(-\alpha z)[H(x+b) - H(x-b)] \ldots 10$$

where H is the Heaviside step function, and A, α, and b are constants. This function was used in A93 for the study of easterly flow over a heated continent. It represents fixed surface heating between x = −b and x = b, the heating having a vertical e-folding decay scale of α^{-1}. Monsoonal flow over northern Australia is frequently accompanied by precipitation. The proposed heating function of Eqn 10 does not take account of factors related to precipitation, such as latent heat release, or the effects of surface moisture.

The solution for the perturbation streamfunction can be obtained from Eqn 8 in the same manner as for easterly flow (see Appendix). It contains the integral

$$\int_0^K \frac{m^2 \sin(mz)\{\sin[k(b+x)] + \sin[k(b-x)]\}\, dk}{k^2(\alpha^2 + m^2)}$$

Although m is a function of k, the behaviour of the integral is similar to that of the integral

$\int_0^K \sin[k(b \pm x)]dk/k^2$ which is unbounded. The integral's failure to converge (except at z = 0) is due to the factor k^2 in the denominator of the integrand. If k^2 is replaced by k, the integrand does converge. The offending k in the denominator can be traced to the imposition of the lower boundary condition. In the absence of orography, the condition is that w(x,0) = 0. In the easterly situation discussed in A93, the condition that the heated air had to maintain surface contact implied an upstream zonal perturbation velocity into the heated region. This holds also in the westerly situation, but the added complication of forcing the Rossby/gravity waves to maintain contact with the heated surface cannot be met.

Therefore, it is not possible to determine a steady-state, bounded solution to the heating problem in a westerly regime. This conclusion is not a peculiarity of the heating distribution chosen. If a bell-shaped heating distribution is investigated, the solution is again unbounded. Smith and Lin (1982), Lin and Smith (1986), and Lin (1987) encountered similar problems in their studies of the effects of heating on air-flows, and had to restrict their heating functions q so that

$\int_{-\infty}^{\infty} q\,dx = 0$, to obtain a bounded solution. This condition means that the net heating applied at any level must be zero, implying the application of compensatory cooling.

The introduction of a cooled region on the eastern side of the heated region to simulate cool ocean waters can be achieved by means of the heating distribution

$$q = A_0 \exp(-\alpha_0 z)[H(x+2b)-H(x)] + A_1 \exp(-\alpha_1 z)[H(x)-H(x-2b)] \quad \ldots 11$$

where $A_0>0$, and $A_1<0$. This heating distribution is intended to represent constant heating over a distance 2b followed by cooling over the same distance. The different vertical e-folding decay scales, α_0^{-1} and α_1^{-1}, are to take account of the different stability properties of cooled and heated air respectively.

It is possible to find a bounded solution for the combined heating and cooling distribution of Eqn 11 in a westerly wind regime only if the condition

$$\frac{A_0}{\alpha_0^2+N^2K^2/U^2L^2}+\frac{A_1}{\alpha_1^2+N^2K^2/U^2L^2}=0 \quad \ldots 12$$

is imposed. This condition is not equivalent to Smith and Lin's (1982) condition above unless $\alpha_0=\alpha_1$. The expression for the perturbation streamfunction pertaining to the heating distribution of Eqn 11, subject to Eqn 12, is given in Eqn A2 in the Appendix.

Use of L'Hôpital's rule shows that the wave-like part of the first integral of Eqn A2 will diverge unless Eqn 12 is obeyed. A second application of L'Hôpital's rule will verify that, provided Eqn 12 holds, the wave-like part of the first integrand of Eqn A2 equals zero when $k=0$.

Unless $\alpha_0=\alpha_1$, the contribution to the streamfunction from the evanescent part of the first integral of Eqn A2 at $k=0$ contains a term in x which becomes large as $|x|$ becomes large. Therefore, just as for easterly flow over a heated surface, there is an upstream effect, except when z equals zero.

Simulation of diabatic effects on a monsoon flow

To model the effects on a westerly airstream of diabatic continental heating and maritime cooling in the Coral Sea, parameters have been made equal to their values in the corresponding easterly situation. These values are: $b=1500$ km, $\alpha_0=5\times10^{-4}$ m^{-1}, corresponding to an e-folding decay scale of two kilometres over the hot land; and $\alpha_1=2\times10^{-3}$ m^{-1} corresponding to an e-folding decay scale of 500 metres over the cool ocean. A_0 has been chosen to produce the same value of maximum temperature difference (4.5°C) as in A93 for easterly flow over hot land followed by cool water. Once A_0 is fixed, A_1 is determined from Eqn 12. The mean latitude is 20°S, as in A93.

The perturbation flow in the x-z plane is shown in Fig. 2(a). Flow over the heated land is dominated by a single circulation with the hottest air close to the east coast rising. A secondary maximum in upwards motion is found nearly 1000 km east of the east coast. Flow over the cooled ocean, however, consists of a series of smaller-scale circulations. The perturbation pressure and perturbation temperature have simpler structures. Negative perturbation pressures are induced by the heating, and positive perturbation pressures by the cooling. Horizontal pressure gradient forces are partly balanced by the Coriolis force, implying equatorwards flow over the heated regions and polewards flow over the cooled area. Perturbation velocities and pressures are of the same order of magnitude as those obtained in the corresponding easterly situation. Figure 2(b) illustrates flow in the x-y plane at $z=0$. The flow is similar to its easterly counterpart; the maximum meridional deviation is slightly smaller in westerly than in easterly flow, but the far-downstream waves in the westerly case have a large amplitude.

Fig. 2(a) Perturbation streamfunctions in the x-z plane for westerly flow over a heat source extending from x = −3000 km to x = 0 km and cooling from x = 0 km to 3000 km. Values in m² s⁻¹.

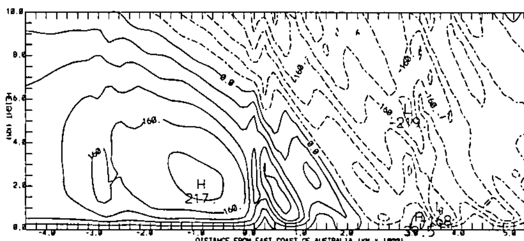


Fig. 2(b) Horizontal projection of a streamline at z = 0 for westerly flow over a heat source extending from x = −3000 km to x = 0 km (shaded region) and cooling from x = 0 km to 3000 km. x-axis labelled in units of 1000 km. y-values in km.

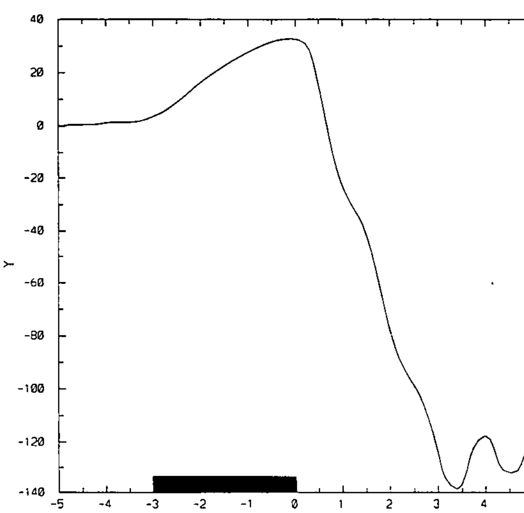


Fig. 3(a) **Perturbation pressure at z = 0 for westerly flow over orography centred at x = −200 km and a heat source extending from x = −3000 km to x = 0 km (shaded region) and cooling from x = 0 km to 3000 km. x-axis labelled in thousands of km. Values in hPa.**

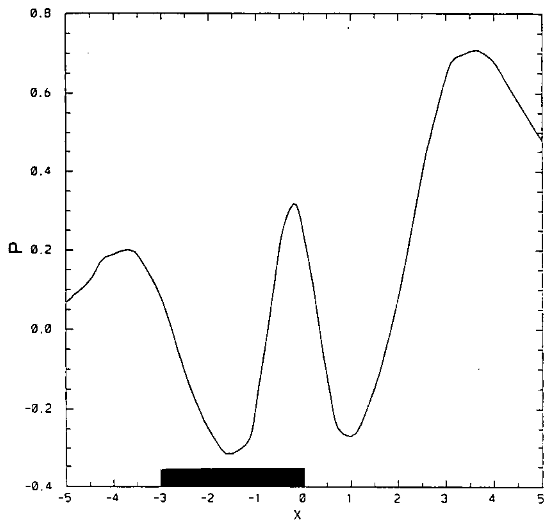
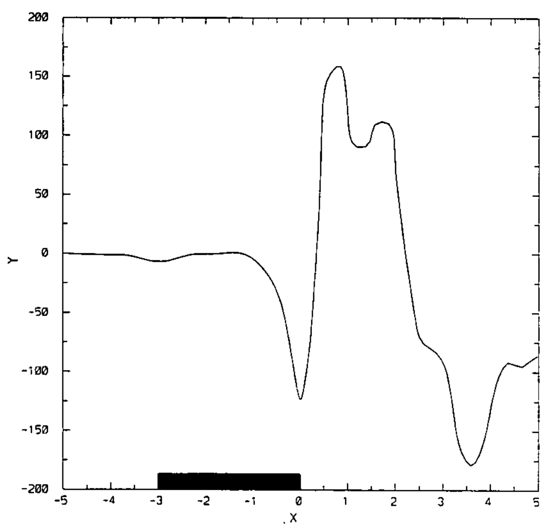


Fig. 3(b) **Same as Fig. 3(a) except horizontal projection of a streamline a z = 0. y values in km.**



Simulation of the combined and orographic effects on a monsoon flow

A major difference between the easterly and westerly flows over the continent is that the easterly flow crosses the mountains before traversing the heated continent whereas the westerly has already passed over most of the hot continent before it reaches the mountains. The perturbation streamfunction for the combined effect of warmed land, cooled sea, and orography is dominated by the orographic effect. The perturbation temperature, on the other hand, is influenced most by cooling over the water. Nonetheless, descent on the lee slopes of the orography is sufficient to produce a warm anomaly there. The perturbation pressure at z = 0 is shown in Fig. 3(a). The heating and the lee trough effect augment each other most strongly about 1000 km downstream of the continent to produce a well-defined pressure minimum. The flow at z = 0 for the combined heating/cooling/orographic effect is shown in Fig. 3(b). The displacements are approximately the arithmetic sum of the orographic and heating/cooling effects. A ridge occurs just inland of the east coast and there are two, marked troughs about 400 km and 1500 km east of the eastern boundary of the heating. These troughs arise from a combination of the lee trough effect and the flow of air from warm land to cool ocean.

Westerly, monsoonal flow over the whole of northern Australia is unusual. For example, during the period of intensive observations of the Australian monsoon during AMEX (10 January 1987) to 15 February 1987), there were only three days (22 January to 24 January) when there was substantially westerly flow over most of northern Australia. It is of interest to note that the streamlines over the Coral Sea did develop cyclonic curvature on 23 January, and that a week low-pressure system formed just off the Queensland coast in the Coral Sea at about the same time. The cyclonic streamline curvature and the weak low were maintained on 24 January, after which the westerly flow began to break down. These observations lend weight to the results outlined above.

Summary and conclusions

In a westerly wind, upstream effects due to the Great Dividing Range extend roughly twice as far from the orography as in an easterly, with the flow beginning to deviate polewards approximately 4000 km upstream.

Downstream of the orography, westerly flow consists of a mixture of internal gravity waves, modified by inertial effects, and Rossby waves. The downstream flow in the x-y plane is a complex superposition of waves without any clearly dominant wavelength. Maximum values of perturbation velocities and pressure, and maximum meridional displacement at the surface are very similar in a westerly to their easterly counterparts. Above the surface, these quantities tend to be larger in a westerly air flow than in an easterly.

It is not possible to find a bounded solution to the problem of westerly flow over a heated surface unless a specified amount of compensatory cooling is introduced. In the present study, compensatory cooling is introduced on the eastern side of the heated region. Cooling induces positive low-level perturbation pressures and polewards flow, while heating produces a decrease in the pressure, and an equatorwards deviation of the flow.

In a simulation of monsoonal flow over the whole of northern Australia, the addition of the orographic effect to a heated land surface and cooled ocean produces a well-defined pressure trough centred about 1000 km east of the east coast and sharp streamline troughs about 400 km and 1500 km east of the east coast. While westerly flow over much of the land surface of northern Australia is unusual, cyclonic streamline curvature and low pressure in the Coral Sea, just off the Queensland coast, were observed during a period of about three days of westerly flow over the continent during AMEX. The results obtained for heating should also be applicable to westerly flow over a region of elevated sea-surface temperature.

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Appendix

Equations for the perturbation function due to diabatic heating

For the heating function given by Eqn 10, the perturbation streamfunction is given by

$$\Psi = \frac{2A}{\pi N^2}\left\{\int_0^K \frac{m^2 \sin(kb)\,[\sin(kx+mx) - \sin(kx)\exp(-\alpha z)]}{k^2\,(\alpha^2+m^2)}\,dk + \int_K^L \frac{\mu^2 \sin(kb)\sin(kx)\,[\exp(-\alpha z) - \exp(-\mu z)]}{k^2\,(\alpha^2-\mu^2)}\,dk + \int_L^\infty \frac{m^2 \sin(kb)\,[\sin(kx+mz) - \sin(kx)\exp(-\alpha z)]}{k^2\,(\alpha^2+m^2)}\,dk\right\}, \quad \text{...A1}$$

while the perturbation streamfunction for the heating function of Eqn 11, subject to the constraint of Eqn 12 is given by

$$\Psi = \frac{2}{\pi N^2}\left\{\int_0^K \frac{m^2\sin(kb)}{k^2}\left[A_0\frac{\sin[k(x+b)+mz]}{\alpha_0^2+m^2} + A_1\frac{\sin[k(x-b)+mz]}{\alpha_1^2+m^2}\right]dk\right.$$

$$-\int_0^K \frac{m^2\sin(kb)}{k^2}\left[A_0\frac{\sin[k(x+b)]\,e^{-\alpha_0 z}}{\alpha_0^2+m^2} + A_1\frac{\sin[k(x-b)]\,e^{-\alpha_1 z}}{\alpha_1^2+m^2}\right]dk$$

$$+\int_K^L \frac{\mu^2\sin(kb)}{k^2}\left[A_0\frac{\sin[k(x+b)]\,e^{-\alpha_0 z}}{\alpha_0^2-\mu^2} + A_1\frac{\sin[k(x-b)]\,e^{-\alpha_1 z}}{\alpha_1^2-\mu^2}\right]dk$$

$$-\int_{K}^{L} \frac{\mu^2 \sin(kb)\, e^{-\mu z}}{k^2}\left[A_0 \frac{\sin[k(x+b)]}{\alpha_0{}^2-\mu^2} + A_1 \frac{\sin[k(x-b)]}{\alpha_1{}^2-\mu^2}\right] dk,$$

$$+\int_{L}^{\infty} \frac{m^2 \sin(kb)}{k^2}\left[A_0 \frac{\sin[k(x+b)+mz]}{\alpha_0{}^2+m^2}\right] + \left[A_1 \frac{\sin[k(x-b)+mz]}{\alpha_1{}^2+m^2}\right] dk$$

$$-\int_{L}^{\infty} \frac{m^2 \sin(kb)}{k^2}\left[A_0 \frac{\sin[k(x+b)]\, e^{-\alpha_0 z}}{\alpha_0{}^2+m^2} + A_1 \frac{\sin[k(x-b)] e^{-\alpha_1 z}}{\alpha_1{}^2+m^2}\right] dk\Big\}$$

. . .A2