Northern Winter Stationary Waves: Theory and Modeling

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ABSTRACT

A review is provided of stationary wave theory, the theory for the deviations from zonal symmetry of the climate. To help focus the discussion the authors concentrate exclusively on northern winter. Several theoretical issues, including the external Rossby wave dispersion relation and vertical structure, critical latitude absorption, the nonlinear response to orography, and the interaction of forced wave trains with preexisting zonal asymmetries, are chosen for discussion while simultaneously presenting a decomposition of the wintertime stationary wave field using a nonlinear steady-state model.

1. Introduction

The climate of the earth would be independent of longitude if the earth's surface provided a zonally symmetric boundary condition for the atmosphere. From the gross similarity between the two hemispheres, we are confident that the structure of the zonal mean flow within the troposphere is not dramatically dependent on the detailed asymmetries of the lower boundary. Therefore, one can visualize a theory for the tropospheric climate as being constructed in two parts: a theory for the zonally symmetric climate, and a theory for the deviations from this symmetry that assumes knowledge of the zonal mean. To the extent that the zonal asymmetries do modify the zonal mean, one can contemplate iterating and converging to a theory that encompasses this interaction as well.

Starting with the seminal work with quasigeostrophic models by Charney and Eliassen (1949) and Smagorinsky (1953) and evolving into studies of the primitive equations on the sphere (e.g., Egger 1976; Hoskins and Karoly 1981; Nigam et al. 1986, 1988; Chen and Trenberth 1988b; Valdes and Hoskins 1989; Ting 1994) history has shown that one can make some progress in modeling these zonal asymmetries using linear models in which the effects of transients are treated in a crude way or even omitted entirely. This contrasts with theories for the zonally averaged flow, which are very strongly dependent on one's models of transient eddy fluxes. It is this reduced need for accurate representation of transient eddy fluxes that hopefully allows one to separate the theory of *stationary waves* from the theory of the general circulation as a whole. We do not distinguish between the terms *stationary waves* and *stationary eddies.* The common usage of the former term testifies to the value of linear theory in analyses of the climatic zonal asymmetries.

A variety of issues arise as one moves from qualitative comparisons with idealized linear theories toward quantitative comparisons of linear and nonlinear steady-state models with observed climatic asymmetries, and toward the use of these models to diagnose the sources of interannual variability. We begin in section 2 with an introduction to some fundamental questions underlying the enterprise of stationary wave modeling. We then discuss aspects of the linear and nonlinear responses to orographically and thermally forced stationary waves, using as a backdrop a particular nonlinear steady-state simulation of the observed Northern Hemisphere stationary waves in January. The steady-state simulation is introduced in section 3. Sections 4 and 5 are devoted to the linear and nonlinear responses to orography. Section 6 is devoted to the response to thermal forcing, and also includes a brief discussion of the

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interaction between heating and orography. Most of the issues related to stationary wave propagation are discussed in the orographic sections although they are relevant for thermal forcing as well.

2. Some questions

a. Is the time mean flow dynamically significant?

The theory of stationary waves is an attempt to understand aspects of the time mean flow in the atmosphere. One is implicitly assuming that the mean (seasonally varying) flow is a quantity of both practical interest and dynamical significance. Yet there are dynamical systems, such as the three-component Lorenz model, for which the mean of a variable is located at a position in phase space that is so unrepresentative that the mean is not a statistic of prime interest. In addition, the linearized dynamics about this time mean may have little value for understanding the sensitivity of the climate to a change in a parameter. There are claims that the mean response of the atmosphere to perturbations in boundary conditions may, in part, be due to changes in the occupancy of different ''regimes'' that reside far from the mean (Hansen and Sutera 1986; Molteni and Tibaldi 1990; Kimoto and Ghil 1993; Palmer 1999).

There is controversy concerning our ability to reliably estimate probability distributions of the large-scale flow with the existing database (Cheng and Wallace 1993; Nitsche et al. 1994). We believe that it is useful to address this kind of critique pragmatically by examining the results of stationary wave theories. To the extent that one can create simple steady-state models that predict the mean response of the atmosphere or of GCMs to changes in boundary conditions, at a useful level of accuracy, one is effectively demonstrating that this mean flow does have dynamical significance. These theories should fail if the mean response is fundamentally a consequence of changes in occupancy of regimes that reside far from the mean. In addition, to the extent that one can create quantitative models of the midlatitude storm tracks, or of low-frequency variability, by linearizing about the time mean flow, one is also directly demonstrating the dynamical relevance of this mean flow (e.g., Whitaker and Sardeshmukh 1998).

The picture of stationary Rossby waves, forced by orography or heat sources, and propagating on a smooth background flow, is the starting point for theories of the climatological zonal asymmetries in midlatitudes. But one can question placing too much significance on the time mean flow as the appropriate background for stationary Rossby wave propagation. Stationary linear models are not linear in the basic state upon which the waves propagate. The mean response is, as a result, rather easily modified by adding random variations to the state about which one linearizes (e.g., Pandolfo and Sutera 1991). A problem for which low-frequency variability of the background for wave propagation is likely to be important is the extratropical response to tropical forcing (see Hall and Derome 2000). If only a small fraction of the states of a slowly evolving background are especially favorable for a large extratropical response, a steady model linearized about the mean background state will be unable to accurately capture the mean response. A diagnosis of the steady state would show ''transients'' playing a role.

A related concern of this type is discussed by Swanson (2001); see also Swanson (2002, this issue). If the potential vorticity (PV) distribution consists of two regions of homogenized PV separated by a sharp boundary, fluctuations of the boundary can be sufficient to create a smooth mean PV distribution. But is it then meaningful to compute the refraction of planetary waves on this smooth distribution, or do the waves feel the sharp discontinuity and propagate along it? See Ambrizzi and Hoskins (1997) for examples of stationary Rossby waves being ducted along the large PV gradients at westerly jet maxima. Once again, the success (or lack of success) of theories that assume that the smooth mean flow is an appropriate dynamical background on which to study stationary wave propagation is a good indicator of the extent to which this approach is or is not naive.

The path toward the construction of a linear stationary wave model need not always start with the linearization of the dynamics about some mean flow, followed by attempts to compensate for various effects that one has thereby omitted. One can directly obtain linear operators governing the evolution of deviations from the mean by empirically fitting atmospheric data, after selection of a suitably small number of degrees of freedom with which to work. Branstator and Haupt (1998) provide a good example, in which a model comparable in complexity to a linearized barotropic model is constructed empirically to simulate the evolution of the 500-mb flow and is then found capable of simulating the steady response of a GCM to tropical heating. The task then becomes one of trying to understand the structure of this effective linear operator. We do not discuss this path further, but it may very well be an efficient way of sidestepping complexities of the sort outlined above.

b. Is it meaningful to think in terms of stationary waves forced by specified heating distributions?

Historically, one of the central questions addressed by linear stationary wave theories has been the relative importance of orography and thermal forcing for the observed zonal asymmetries of the circulation. The problem of thermal forcing has often been thought of as decomposed further into two parts: determining the diabatic heating distribution generated from the lower boundary asymmetry, and then analyzing the response to this heat source. Often the first part of this problem is simply discarded and one examines the response to heating distributions obtained from observations or general circulation models.

One can imagine a number of scenarios in which the approach of studying the response to specified heat sources is problematic. If we are interested in the relative importance of orography and heating, there is, first of all, the complication that orographic forcing can modify the heat sources; we would need a model of this effect to discover the true effect of orography on the atmospheric state. Indeed, one would need a coupled atmosphere–ocean model, since the presence of the orography could modify SSTs as well, and these modified SSTs would further alter the heating field.

But there are other ways in which thinking in terms of the response to specified heat sources can potentially lead one astray. As a simple and relevant example, suppose that one's extratropical heat source over the oceans is a strongly increasing function of the difference between a prescribed ocean surface temperature T_s and the atmospheric temperature $T(0)$ near the surface. Think of the simplest case in which $Q = \gamma(z)[T_s - T(0)]$, where γ determines the vertical structure of the heating. A linear theory, $LT = Q$, forced by an estimate of Q could easily yield inaccurate temperatures if Q or the operator L are not exact. On the other hand, if the dependence of the heating on the surface temperature is incorporated into the model $L_* T \equiv LT + \gamma T(0) = \gamma T_s$ one is at least assured that the resulting temperatures near the surface will be close to T_s if γ is sufficiently large. See Shutts (1987) for a related discussion. In the Tropics, an analogous case can be made that parts of the thermal structure and circulation can be understood without thinking about latent heat sources directly, and that the heat sources are then constrained to be consistent with this circulation (e.g., Neelin and Held 1987; Emanuel et al. 1994).

Along the same lines, consider the eddy sensible heat fluxes in northern winter. The heating due to the convergence of these fluxes acts to dissipate the low-level stationary eddy temperature field (e.g., Lau and Wallace 1979; Kushner and Held 1998). Suppose that one's model of the stationary eddies, when forced by the full heating field and the observed transient eddy fluxes, is accurate. If one removes some part of the heating field so as to isolate its influence, thereby altering the low-level temperatures, but holds the transient eddy heat fluxes fixed, this damping effect will be distorted. A theory that incorporates the eddy heat fluxes into the operator L rather than as prescribed forcing is once again desirable.

One can make attempts along these lines at parameterizing eddy effects and diabatic heating. However, we do not feel that current theories are sufficiently credible to produce *quantitative* stationary wave models. More effort has gone into relating tropical heating to the lower boundary condition, motivated by the desire to create idealized models of ENSO. The problem of how the lower boundary conditions control extratropical heating is even less well established in our view (see Kushnir et al. 2002, this issue).

c. *Are nonlinear models of zonal asymmetries well posed?*

Despite the usefulness of the simplest linear theories, it is clear that we must move beyond linearization about the zonal mean to model the stationary eddies quantitatively. Several extensions of linear models have been examined. The most direct is the use of a model linearized about a zonally asymmetric flow to iterate toward steady nonlinear solutions (e.g., Valdes and Hoskins 1991). Unfortunately, one often finds that the linear operators obtained in this way are nearly singular. If such an iteration method is successful, then the last step in the iteration will involve linearization about a flow similar to that observed. As pointed out by Simmons et al. (1983), when one linearizes about observed mean flows, one typically finds nearly neutral low-frequency eigenmodes. Stationary forcing can resonantly excite these modes. A small change in the forcing can then lead to implausibly large responses (e.g., Ting and Sardeshmukh 1993; Ting and Yu 1998). The implication is that transients provide sufficient mixing or damping, in some generalized sense, to regularize the response.

One can add damping to remove these resonances, and then iterate to steady nonlinear solutions. It so happens that the damping required to make models linearized about observed asymmetric flows more robust is often comparable to that needed to stabilize the model completely, not only to low-frequency modes but to the dominant baroclinic instabilities as well. Therefore, a simple and efficient procedure suggests itself: add damping sufficient to stabilize the flow and then integrate forward to a steady state with the full nonlinear primitive equations on the sphere. Works along these lines typically incorporate some physical features in the damping prescription, such as enhanced damping of winds near the surface to represent surface friction, but they have generally not attempted to fully justify the damping as accurately mimicking the parts of the transient eddy fluxes or heating fields that react to changes in the mean state.

An alternative that avoids dependence on arbitrarily enhanced damping is that employed by Jin and Hoskins (1995) and Rodwell and Hoskins (1996) in which the integration of a time-dependent model is simply terminated before the dominant midlatitude instabilities develop. We prefer the steady-state model with added damping because it provides a framework within which one could, in principle, try to formulate physically based damping/mixing schemes.

3. An example of a steady-state model

While cognizant of these difficulties, we feel that the classic diagnostic decomposition of the stationary eddies into parts forced by orography and heating, in which the heating distribution is taken from data or models, as well as the more recent analyses of the non-

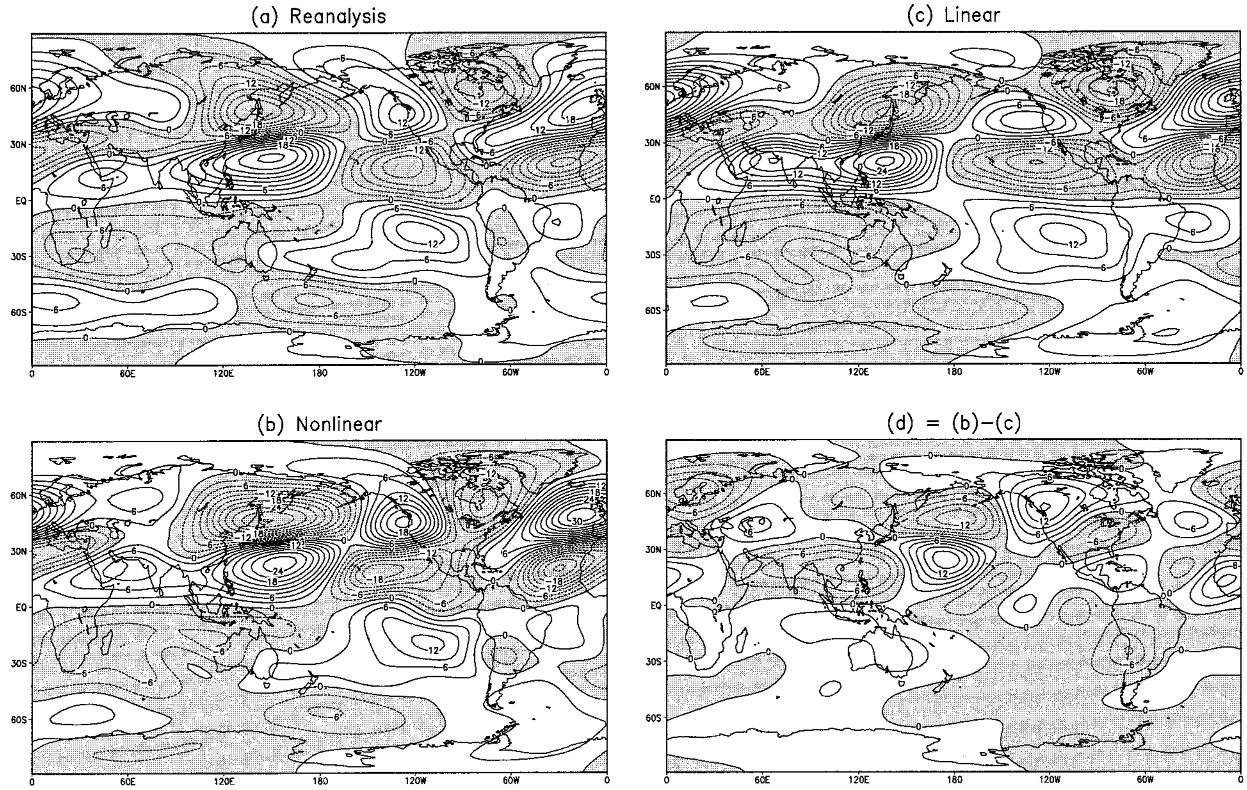

FIG. 1. (a) The observed stationary eddy 300-mb streamfunction from NCEP–NCAR reanalysis for Jan, (b) the steady nonlinear response of the 300-mb eddy streamfunction to global forcing by orography, heat sources, and transient eddy flux convergences, (c) the steady linear response to these same forcings, and (d) the nonlinear response minus the linear response. Contour interval is 3×10^6 m^2 s^{-1}.

linear interaction between these parts, remains an important stepping stone to a satisfactory understanding of the circulation. This kind of diagnosis continues to provide valuable information on the relative importance of different factors for maintaining the observed climate. For future work, it provides a backdrop from which one can attempt to construct steady-state models within which reliable theories for the heating field and transient eddy fluxes are embedded.

Figure 1a shows the 300-mb eddy streamfunction (the streamfunction with the zonal mean removed) in January from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis. Figure 1b is the response to the global distribution of heating, orographic forcing, and transient eddy flux convergences, as generated by a nonlinear but steady-state primitive equation model on the sphere, with prescribed zonal mean flow taken from the same NCEP–NCAR reanalysis for January. A description of the model, including the damping factors added to produce a steady (or very nearly steady) state can be found in the appendix. No attempt is made here to model the transient eddy fluxes or the heating distribution. The quality of the agreement is a measure of the distortion caused by the addition of the damping, the consistency of the reanalysis with the dynamical model utilized, and the internal consistency of the reanalysis itself. While the patterns agree reasonably well, the amplitude in this stationary wave model is significantly greater than that in the reanalysis in the extratropics, especially over the Atlantic. When using the same technique to model the climate of a GCM in which the dynamical equations are identical to those used by the steady-state model, the errors are smaller (Ting et al. 2001). Therefore, it appears that the introduction of the damping itself is not the dominant source of error. Rather we suspect the accuracy of the heating field is the issue, accentuated by the absence of feedback from the predicted low-level temperatures.

Figure 1c shows the *linear* response to the same combination of orography, heating, and transient eddy fluxes as used in the nonlinear model. The linear and nonlinear models contain the same damping terms. One can obtain the linear result by direct matrix inversion, linearizing about the prescribed zonal flow, or one can multipy all of the forcing functions by a small number, ϵ, generate the steady nonlinear solution, and then divide this solution by ϵ. Figure 1d shows the difference between the linear and nonlinear solutions. Several features, such as the high over western North America, are improved by the nonlinear simulation; one can also find features that seem to be degraded somewhat.

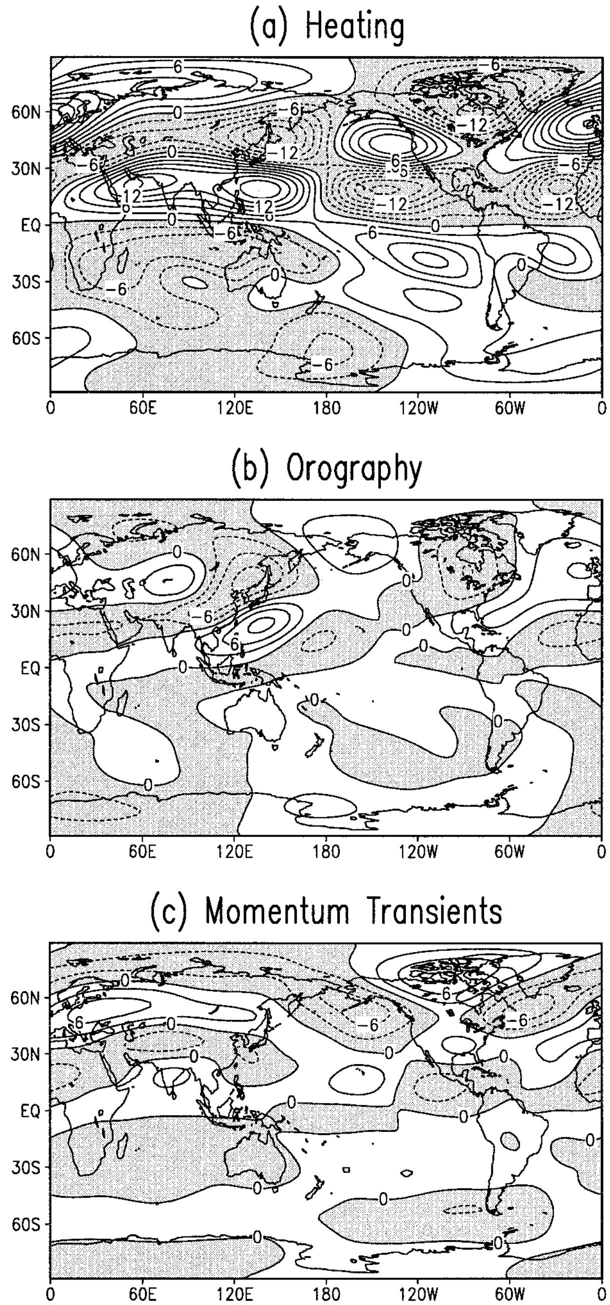

FIG. 2. The decomposition of the total linear response at 300 mb (Fig. 1c) into parts forced by (a) heating plus transient eddy fluxes in the temperature equation, (b) orography, and (c) transients in the momentum equations. Contour interval is 3×10^6 m^2 s^{-1}.

The decomposition of the linear solution into parts forced by heating, orography, and transients in the momentum equation is shown in Fig. 2. The transients in the temperature equation are lumped together with the heating field for the purpose of this decomposition. The sum of Figs. 2a,b,c is the linear solution shown in Fig. 1c.

The zonal asymmetries in the transient eddy vorticity fluxes responsible for Fig. 2c play only a modest role in maintaining the upper-level climatological stationary eddies, according to our steady-state model. There are cogent arguments for positive feedbacks between the eddy mixing associated with wave breaking and both the deformation of the flow along the jet core (Shutts 1983) and the deflection of the jet from its zonal orientation (Orlanski 1998). These works suggest that the role played by upper-level transients in shaping the zonal asymmetries of the flow might be greater than that indicated by diagnoses based on stationary wave theory. The fact that upper-level transient eddy fluxes can, at least in some models, play a central role in the response to extratropical SST anomalies (Kushnir et al. 2002, this issue, and references therein) also points in the same direction.

In this review, the decomposition in Figs. 1 and 2 is used to motivate a discussion of several issues in the theory of stationary waves. Theories of the heating field itself (and the transient eddy fluxes) as a function of the boundary conditions and the mean circulation are outside the scope of this review. We discuss northern winter exclusively, but the reader can find a discussion of similar calculations for other seasons in Wang and Ting (1999), Hoskins and Rodwell (1995), and Rodwell and Hoskins (2002). See also the study of the seasonal cycle of the stationary waves in a GCM by Ting et al. (2001). The reader is referred to Held (1983) for additional introductory material on stationary Rossby waves.

4. The linear response to orography

Figure 3a shows the linear response in winter to the orography of central Asia (primarily the Tibetan Plateau). Figure 3d is the analogous result for the orography of North America (primarily the Rockies). The sum of these two responses is close to the total linear orographic result in Fig. 2b. The shapes of the responses are qualitatively similar to those in other studies with multilevel primitive equation models on the sphere, such as Valdes and Hoskins (1989), Nigam et al. (1988), and Trenberth and Chen (1988), and also resemble the results of barotropic simulations (Grose and Hoskins 1979; Held 1983). In all cases we see wave trains emanating from the orography, with a part refracting strongly into the Tropics, and (most clearly in the Tibetan case) a part propagating poleward before arcing into the Tropics.

Our orographic responses are comparable in magnitude to those in Valdes and Hoskins, but weaker than those in Nigam et al., and Trenberth and Chen. The dominant low on the Asian coast forced by Tibet is only 30% of the magnitude of the observed climatological low. Because the orographic response has smaller zonal scales than the thermal component, its share of the total response increases if one examines the meridional wind or the deformation of the flow. As discussed in Held and Ting (1990), the linear response to orography is approximately proportional to the strength of the low-

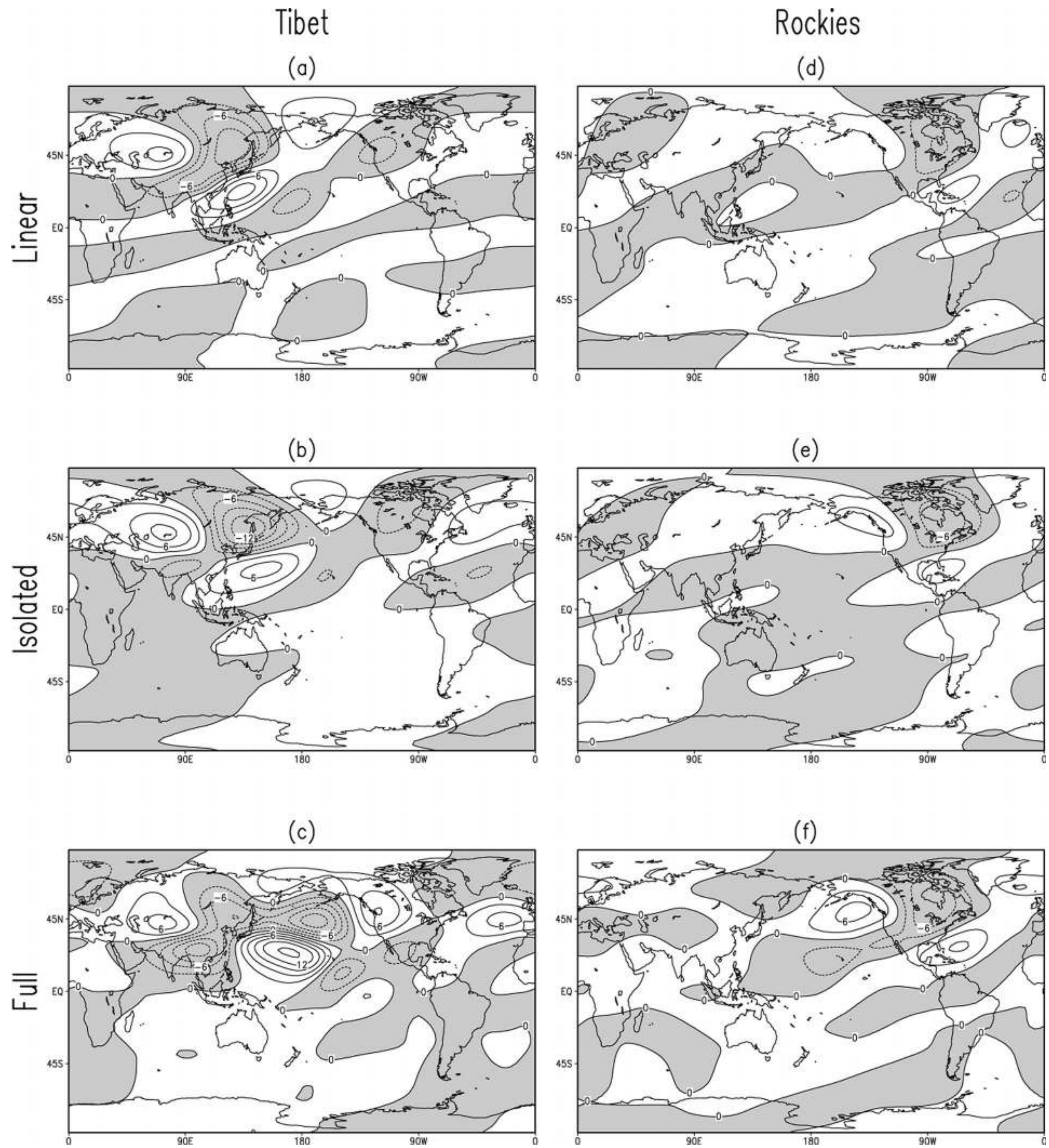

FIG. 3. The 300-mb eddy streamfunction for the (a) linear, (b) isolated nonlinear, and (c) full nonlinear responses to Tibet in Jan; the analogous (d) linear, (e) isolated nonlinear, and (f) full nonlinear responses to the Rockies. Contour interval is 3×10^6 m^2 s^{-1}.

level mean winds, and is inversely proportional to the strength of the low-level meridional temperature gradient. Differences in strength of the low-level winds appear to be the main reason for discrepancies between the different linear orographically forced models in the literature.

Figures 3b,c,e,f display what we refer to as the *isolated nonlinear* and *full nonlinear* responses to Tibet and the Rockies; we discuss these in section 5.

In this section, we first isolate some of the essential features of the linear solutions using more idealized models. We begin by focusing on the wavelength and the vertical structure of the external Rossby wave. We then turn to the paths followed by the radiating Rossby

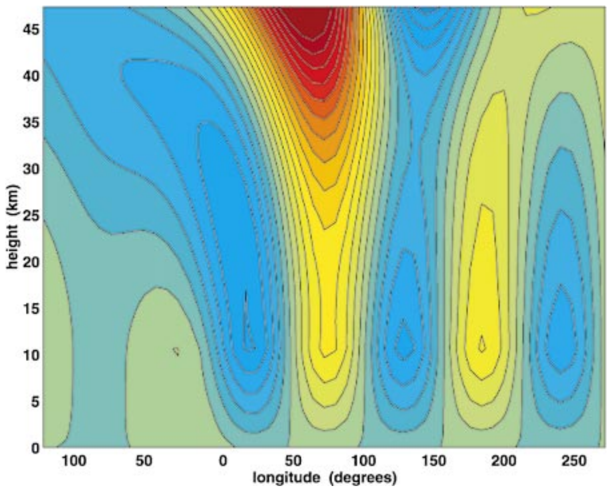

FIG. 4. Streamfunction response to orography in a QG model on a β plane with uniform Brunt–Väisälä frequency, in which the mean zonal flow is linear in height below the tropopause (at 10 km) and uniform above the tropopause. The orography and the solution are assumed to be independent of latitude. The orography is centered at 0° lon.

waves as they move eastward from their orographic sources. We then address the confusing issue of the potential for reflection from the Tropics. Idealized GCM results are also described that provide modeling evidence for the picture of near-perfect absorption within the Tropics, with little reflection. Finally, the possibility of the stratosphere modifying the tropospheric stationary wave field is briefly addressed.

a. The external mode

Figure 4 shows the two-dimensional (x–z) response to a localized topographic source with no y structure, in a quasigeostrophic (QG) model on a β plane. The simple basic state is described in the caption. A radiation condition is imposed at the top of the model that allows upward-propagating waves to pass through without reflection. One sees the distinction between the largest waves that escape to the middle atmosphere and the shorter waves captured within the troposphere. The latter part of the wave field organizes itself into a horizontally propagating wave train with the particular equivalent barotropic vertical structure of the *external Rossby wave.*

The external mode has maximum streamfunction amplitude in the upper troposphere. Below this maximum, regions of low pressure are cold and regions of high pressure are warm. These warm high/cold low wave trains are easily distinguished from the warm low/cold high signature of the local response to a shallow heat source. The stationary wave pattern in the extratropics is dominated by this structure (Wallace 1983), the most prominent exceptions being regions of monsoonal heating. For a detailed analysis of the external mode structure and dispersion relation, see Held et al. (1985, hereafter HPP). We review some of these results here.

In an idealized problem such as that in Fig. 4, one can solve for the eddy field by first performing a modal decomposition in the vertical and then solving for the horizontal structure of each mode. The resulting vertical modes can be divided into two classes: vertically trapped modes and modes that propagate vertically and escape to infinity. In the simplest case of a uniform flow U with no vertical shear and constant buoyancy frequency N, there is one and only one trapped mode. Its energy decays exponentially away from the surface, but its streamfunction is independent of height. The total horizontal wavenumber of this mode is given by Rossby's classic formula: $k = (\beta/U)^{1/2}$.

With the Charney basic state, a linear shear profile $U = U(0) + \Lambda z$ and constant N, and assuming that $\Lambda H \gg U(0)$, where H is the scale height, the number of trapped modes is the largest integer less than $1 + r^{-1}$, where $r \equiv h/H$ and $h \equiv f^2\Lambda/(\beta N^2)$, so there is one and only one trapped mode if $r > 1$. In the midlatitudes, we typically have $r \approx 1$–2. It is customary to define an *equivalent barotropic level*, z_e, so that one obtains the correct wavelength for the stationary wave by using $U(z_e)$ in the Rossby stationary wavenumber formula. In this case, one finds for the Charney basic state (HPP) that $z_e/H \approx 4/(2 + r^{-1})$. For typical values of r this expression predicts an equivalent barotropic level a bit higher than the scale height H. There is vertical motion in the external mode, but at z_e the vortex stretching due to this vertical motion is zero.

In more realistic flows with a jet maximum at the tropopause, the external mode streamfunction takes on a sharp maximum at the tropopause, with a shape that is similar to that of the zonal wind itself. In this more realistic case, the external model takes on some of the characteristics of an edge wave propagating on the tropopause (e.g., Rivest et al. 1992; Juckes 1994; Verkley 1994). An analytically tractable model can be constructed by using an idealized flow in which U is constant above the tropopause, while below the tropopause the vertical curvature of the flow is assumed to counteract β to produce homogeneous quasigeostrophic PV in the troposphere. For constant N, and ignoring compressibility and the presence of the lower boundary, the stationary wavenumber can be shown to be

$$K = \frac{1}{2}\left(\alpha + \frac{1}{\alpha}\right)K_R, \qquad (1)$$

where

$$\alpha \equiv \frac{K_D}{K_R}; \qquad K_D \equiv \frac{f^2\Lambda}{N^2U}; \qquad K_R \equiv \sqrt{\frac{\beta}{U}}. \qquad (2)$$

Here U is the wind at the tropopause and Λ is the shear immediately beneath the tropopause. This expression always gives $K \geq K_R$, consistent with the existence of an equivalent barotropic level in the troposphere.

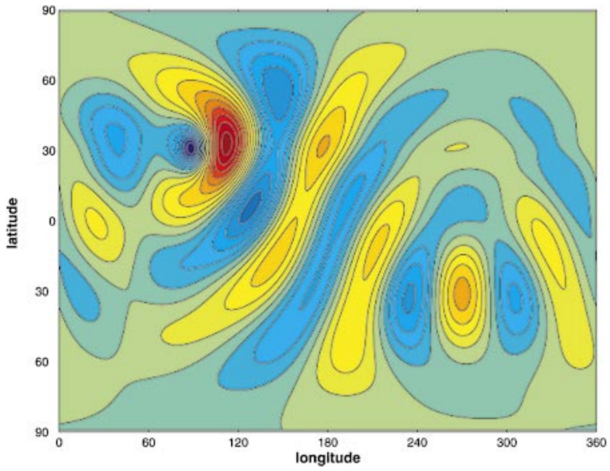

FIG. 5. Response to orography in a shallow water model, linearized about solid body superrotation. The topography is centered at 30° lat and 90° lon.

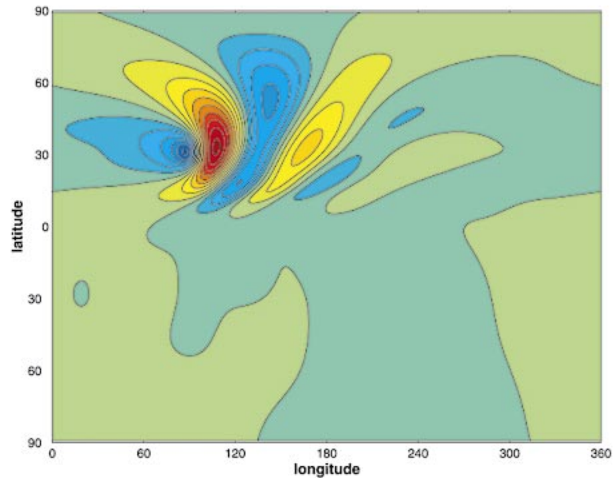

FIG. 6. As in Fig. 5, but linearized about a zonal mean flow that resembles that in the upper troposphere in winter. The contour interval and forcing are identical to those used in Fig. 5.

This analytic expression can be generalized to include compressibility, a jump in N at the tropopause, and a lower boundary. A closely related example is considered by Swanson and Pierrehumbert (1995), who show that the external mode wavenumber is hardly modified if one compares flows with smooth PV gradients with those in which the troposphere is homogenized by mixing the PV so as to create a tropopause. In this context one can think of the surface as perturbing the underlying tropopause edge wave structure of the external mode, providing one way of thinking about the upper-tropospheric maximum in the modal structure.

It is also worth noting that $\psi' \propto U(z)$ is the vertical structure of the eddy streamfunction that results in a balance between the zonal advection of the eddy temperature by the mean flow and the eddy meridional advection of the mean temperature (Hoskins and Karoly 1981; Held 1983). The poleward motion must be downstream of the cold air so that the cooling due to advection by the zonal wind can balance the meridional advection of warmer air from lower latitudes. This implies that low pressure must be in phase with the cold air, or, equivalently, that the eddy streamfunction must increase with height. Above the tropopause, the meridional temperature gradient is reversed and the same argument implies that the eddy streamfunction decreases with height. One is assuming here that adiabatic heating due to vertical motion can be ignored on the large scales of external Rossby waves, which can be a useful starting point but is not quantitatively accurate.

In a separable case such as that in Fig. 4, once one moves a bit downstream of the orographic source, the response at the surface is almost perfectly captured by a barotropic model designed by projecting the full equations onto the external mode (HPP). In the more general nonseparable case, with meridional as well as vertical structure in the mean flow, one can still think of the external mode as being trapped in a waveguide whose structure is a slowly varying function of latitude, and one can design a barotropic model by projecting onto the local structure of this mode. This systematic approach to the construction of linear barotropic models has not attracted much attention, possibly because of the ease with which one can compute the full linear baroclinic response, but it does help one understand how linear baroclinic and barotropic models are related.

b. Great circles

The wave trains in Fig. 2 do not propagate along latitude circles, as in Fig. 4. Instead, they are eventually refracted into the Tropics, where they are evidently absorbed. In a set of landmark papers, Hoskins et al. (1977) and Hoskins and Karoly (1981) emphasized the central importance of the fact that stationary Rossby waves on the sphere tend to propagate along great circles, not latitude circles. Ray tracing using the barotropic dispersion relation shows that the ray paths for stationary waves are exactly great circles when the zonal flow is a uniform superrotation, $U(\theta) \propto \cos(\theta)$, where θ is latitude. Figure 5 shows the classic eddy streamfunction response to a localized mountain in the uniform superrotation flow in a shallow water model, with a small amount of damping to prevent the wave train from passing undiminished around the earth.

The solution in Fig. 5 is dominated by rays that have an initial southward component, which quickly cross the equator, and another set of rays that point northward initially, but also follow a great circle to enter the Tropics farther east. The first downstream trough is due east of the source, and it is only for the next set of highs downstream that this splitting is observed.

Figure 6 is similar to Fig. 5 except that the mean state is now similar to the observed upper-tropospheric flow, with a transition from westerlies to easterlies in the deep Tropics. This response now bears a stronger resemblance to that in Fig. 3: one can still discern the two dominant bundles of rays near the source that we see in the superrotation solution, but all of these waves are absorbed in the Tropics with no signs of either transmission or reflection. Linear stationary Rossby wave theory is singular at the critical line in the latitude–height plane at which the zonal mean wind vanishes. The addition of some damping removes this singularity. The resulting dissipative linear model predicts essentially complete absorption of the incident wave (Dickinson 1968, 1970).

c. The possibility of reflection from the Tropics

Substantial literature has arisen on the question of whether this linear dissipative result is misleading, much of it implying at face value that one should expect reflection from the Tropics. Theories for the reflecting ''nonlinear critical layer'' are elaborate (Killworth and McIntyre 1985), but the essence of the underlying dynamics is easily understood. In the framework of nondivergent two-dimensional flow, consider a Rossby wave propagating from the midlatitudes into the Tropics. (One can argue that a nondivergent two-dimensional model captures the essence of the problem.) The lines of constant phase in an equatorward-propagating wave tilt northeast–southwest (NE–SW) (as in Fig. 4) so that the eddy zonal and meridional velocities are positively correlated and the eddy momentum flux $\overline{u'v'}$ is poleward. A reflected wave would possess the opposite tilt. If there is little reflection or transmission, the stationary wave is continually generating a momentum flux divergence in the Tropics, decelerating the mean flow, $\partial U/\partial t < 0$. In the limit that the dissipation is very weak, this deceleration is centered in a very narrow region around the critical latitude.

Stokes's theorem tells us that modifying the zonal mean flow is equivalent to changing the total vorticity integrated over the polar cap bounded by the latitude circle in question. To decelerate the zonal mean flow, the eddies must reduce the vorticity of the polar cap by creating an equatorward vorticity flux, down the mean (absolute) vorticity gradient. In the linear dissipative absorbing theory, the wave is continually transporting vorticity downgradient near the critical latitude. The mean flow vorticity gradient is not destroyed; in a linear theory it is simply prescribed.

In the nonlinear theories, the absolute vorticity in the vicinity of the critical layer is effectively homogenized by the breaking wave. Once the vorticity gradient is destroyed, there can no longer be any vorticity flux or mean flow deceleration, unless the region of wave breaking expands so that vorticity can be brought through the critical latitude from farther off. From this perspective, one of the principal results of nonlinear critical layer theory is that there are solutions in which this mixing layer does not expand in time.

The solution poleward of this critical layer must be consistent with the absence of momentum flux convergence in the vicinity of the layer. This is only possible if a reflected wave exists with the same amplitude as the incident wave. Nonlinear, inviscid critical layer theory predicts perfect reflection, no matter how weak the incident wave. The amplitude of the wave simply determines the width of the region that is effectively homogenized.

Much of the theory of nonlinear critical layers is limited to sources that are sinusoidal in longitude. Recent nonlinear simulations have addressed the problem, more realistic for the troposphere, of a localized wave train incident on the Tropics (Brunet and Haynes 1996; Magnusdottir and Haynes 1999). The dynamics is complex; reflection does seem somewhat harder to generate than in the case of sinusoidal forcing, evidently due in part to the reenforcement of vorticity gradients in the breaking region by zonal advection.

d. The momentum balance in the subtropical upper troposphere

Observations show that the northern winter stationary eddies transport angular momentum poleward, implying that these waves are preferentially propagating into rather than out of the Tropics (Peixoto and Oort 1992). Therefore, the incident waves must be at least partially absorbed. Since there are also tropical heat sources that generate poleward-propagating stationary waves, the fact that the stationary eddy momentum flux is still poleward, even in the presence of these tropical sources, becomes an even more compelling argument for the strong absorption of incoming waves. So how are we to think of the relevance of nonlinear critical layer theory?

One can think of the upper-tropospheric zonal mean flow equatorward of the subtropical jets as determined by the competition between the deceleration by mixing associated with breaking Rossby waves (primarily eastward-propagating baroclinic waves) on the one hand, and the Coriolis acceleration resulting from the Hadley cell on the other. This competition maintains the vorticity and potential vorticity gradients in the subtropical upper troposphere. When one forces a weak stationary wave in the midlatitudes that then propagates into the Tropics, this balance will be perturbed very little, and the wave can continue to be absorbed indefinitely. The inviscid nonlinear critical layer picture becomes relevant only if the eddy mixing by the stationary wave is strong enough to effectively homogenize the vorticity gradient despite the continuing presence of this underlying balance of forces. While there are several shallow water studies of the interaction between the Hadley cell and a stationary Rossby wave (Held and Phillipps 1990; Esler et al. 2000), they are not directly relevant to the

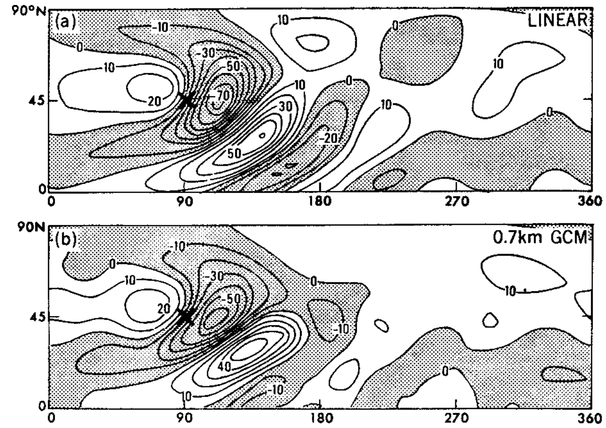

FIG. 7. Comparison of response to isolated orography in an idealized GCM with that predicted by a linear model, from Cook and Held (1992). The plots are of eddy geopotential in the upper troposphere, with the linear solution forced only by orography on the top and the GCM result on the bottom.

problem of how a preexisting balance between the Hadley cell and a spectrum of transient eddies is perturbed by the addition of a stationary wave.

Idealized general circulation models can be used to address this issue. Figure 7b is reproduced from Cook and Held (1992) and shows the stationary wave produced by a model in which an isolated midlatitude mountain is the only zonal asymmetry in the lower boundary condition. (The ''mountain'' here is the size of Tibet!) The height of the mountain is varied, and the figure shows the case with the smallest height ($\approx$0.7 km). The result conforms precisely to the picture of near-perfect absorption of stationary Rossby waves incident to the Tropics. Making this picture more persuasive is the linear response to this orographic feature, linearizing about the GCM control climate, shown in Fig. 7a, in which dissipation creates an absorbing Tropics. As the mountain height is increased in the GCM, the wave pattern changes but there is still no sign of reflection from the Tropics. This result was obtained with a low-resolution (R15) spectral model, and it remains to be seen if higher-resolution models change this picture qualitatively.[1]

In the absence of reflection and the resulting interference effects, it is difficult to imagine that the stationary response to orography is profoundly sensitive to the mean flow, as required, for example, to produce multiple equilibria in wave–mean flow interaction models of the Charney–Devore (1979) type. The remaining sensitivity is not insignificant, however. In particular, see the studies of Nigam and Lindzen (1989) on the sensitivity of the orographically forced waves to the position of the subtropical jet; Kang (1990) on the importance of the displacement of the polar turning latitude for poleward-propagating rays emanating from Tibet; and Ting et al. (1996) and DeWeaver and Nigam (2000) on the extent to which stationary wave models can explain observed correlations between zonal flow variations and stationary wave structure, potentially a key element in the dynamics of the North Atlantic Oscillation.

e. Refraction in the stratosphere

The larger scales that propagate into the stratosphere, as in Fig. 4, also propagate meridionally in ways that are dependent on the stratospheric zonal winds, and are then absorbed where these planetary waves break. Idealized models such as that shown in Fig. 4 can show strong sensitivity of the tropospheric response to the stratospheric winds, but this is a result of allowing only vertical and zonal propagation, which can grossly overestimate the significance of back reflection. On the other hand, observations in northern winter indicate that the phase tilt of the stratospheric stationary wave field is at times reduced to small values, suggestive of reflection (Perlwitz and Graf 2001).

Nearly all computations of the kind shown in Figs. 1–3, using the primitive equations on the sphere, retain very little resolution in the stratosphere and do not attempt to impose a radiation condition at the top of the model, so they are dependent on refraction and absorption in the stratosphere for the validity of the simulations. Ruosteenoja (1999) provides a careful study of the sensitivity of a stationary wave model on the sphere to a reflecting upper lid. The resulting sensitivity is quite modest.

This issue has recently come to the fore due to GCM experiments and observational analyses (Graf et al. 1993; Kirchner et al. 1999) that suggest that changes in stratospheric circulation caused by volcanic eruptions cause significant changes in wintertime tropospheric flow, and also due to the global warming simulations of Shindell et al. (1999) indicating that changes in the stratospheric winds due to CO_2 changes could have tropospheric consequences as well.

There is a pathway other than reflection through which stratospheric winds can affect the tropospheric circulation through the agency of the stationary (or low frequency) waves. This process is often referred to as ''downward control'' (Haynes et al. 1991). Unlike Rossby wave reflection, this process results in an essentially zonally symmetric tropospheric response to a change in the stratosphere: changes in the eddy driving of the zonal mean flow in the stratosphere are balanced by changes in the mean meridional circulation that, in turn, are balanced by oppositely directed return flows near the surface that modify the tropospheric zonal winds. Zonal

[1] *Note added in proof:* C. Walker and G. Magnusdottir (2002, personal communication) have recently found clear reflection in a GCM similar to that of Cook and Held (1992), but for winter rather than annual mean conditions, and at higher resolution. The weak tropical vorticity gradients created by the stronger winter Hadley cell, which are more easily saturated by the incident stationary wave, appear to be key.

asymmetries in the response can then be created as this zonal mean modification interacts with the preexisting asymmetric circulation.

5. The nonlinear response to orography

Can we expect a linear model to provide a good approximation for the climatic response to an orographic feature? By climatic response we mean the difference in climates with and without this feature present in the lower boundary condition, a response that one might estimate by comparing two integrations of a trustworthy general circulation model.

One complication is that the orographic feature will affect the way in which the atmosphere is heated, as well as the distribution of transient eddy fluxes. The strength of the midlatitude eddy field and the manner in which it organizes precipitation makes one skeptical that these interactions could ever be negligible. Yet the result of Cook and Held (1992) reproduced in Fig. 7 indicates that, in this idealized setting, and for an orographic feature of sufficiently small amplitude, stationary linear theory can be accurate, despite the fact that this flow is embedded in a sea of baroclinic, precipitating eddies. In addition, Nigam et al. (1988) compare the linear response about a zonally symmetric basic state with the difference in wintertime climates in GCMs with and without mountains, with encouraging results. The situation is likely to be very different in summer, as GCM experiments suggest that the latent heating in the Asian monsoon is dramatically altered by the presence of the Tibetan Plateau (e.g., Hahn and Manabe 1975).

Even if we can ignore the interaction between orography and the heating and transient eddy flux distributions, we must still understand the nonlinear response to an orographic feature in isolation, as well as the effect that the asymmetric circulation generated by other fixed sources of climatic asymmetries has on the response to this orographic feature. For this purpose, we distinguish between *isolated* and *full* nonlinear responses.

We denote the nonlinear response to some source of asymmetry A, obtained from our steady-state model, as $N(A)$. Let T represent the total forcing that produces the simulation in Fig. 1b, $N(T)$. We refer to $N(A)$ as the isolated nonlinear response to A and $N(T) - N(T - A)$ as the full nonlinear response to A. If we think of the different parts of the forcing as being added in sequence, the isolated nonlinear response to A is relevant when A is the first to be added, while the full nonlinear response is relevant when A is the last to be added, or the first to be removed. Figures 3b and 3e show the isolated nonlinear responses to Tibet and the Rockies. Figures 3c and 3f are the corresponding full nonlinear responses.

a. The isolated nonlinear response to orography

In these steady solutions the isolated nonlinear response is very similar to the linear response for Tibet as well as the Rockies. Valdes and Hoskins (1991) also find rather modest effects of isolated nonlinearity, while Trenberth and Chen (1988) find much larger effects. While there is a voluminous literature on the breakdown of linear theory in idealized atmospheric flows over obstacles on mesoscales, relatively little of this is directly relevant to the planetary scales on which we focus here. Additional work is required to isolate the key parameters that control the breakdown of linear theory.

In the simplest nonrotating problems, the key nondimensional parameter is a Froude number, Nh/U, where N is the buoyancy frequency, h the height of the mountain, and U the incident wind speed. When the Froude number is large, the flow tends to be blocked, rather than passing over the obstacle. In the presence of rotation, and assuming that the flow is balanced, the Froude number is replaced by the parameter Nh/fL as the most relevant measure of nonlinearity, and the Froude number loses its relevance for controlling whether the flow is blocked by the obstacle (Pierrehumbert 1985). Here L is the zonal extent of the obstacle. Comparing Nh/fL with unity, we might expect the isolated response to Tibet to be rather nonlinear, and the response to the Rockies, at least as they are represented at this resolution, to be relatively linear.

The theory of Pierrehumbert (1985) assumes that the incident flow has no vertical shear or, equivalently, no meridional temperature gradient. Yet increasing the meridional temperature gradient reduces the amplitude of the response and, presumably, extends the range of validity of the linear theory (Held and Ting 1990). One way of understanding the effect is to note that the flow at the surface over a large-scale mountain is anticyclonic, bringing air from lower latitudes up the slope; when the horizontal temperature gradients are strong, relatively small meridional displacements can then help balance the adiabatic cooling. Related discussions can be found in Valdes and Hoskins (1991), Cook and Held (1992), and Ringler and Cook (1997).

As an application of these ideas, Cook and Held (1988) argue that the large meridional temperature gradients in the ice age climate help to keep the response to the huge Laurentide ice sheet surprisingly linear. Despite the fact that inspection of the total flow in a GCM suggests that the air passes around rather than over the ice sheet to some extent, thereby splitting the jet, linear theory actually simulates this flow rather well.

(Many ice age theories revolve around the interaction between the North Atlantic Ocean and the Laurentide ice sheet. The orographically forced wave train is a central aspect of this interaction. While Cook and Held find that linear theory is a rather good qualitative approximation to the GCM's stationary wave, it is not able to simulate a GCMs response in the pattern of low-level winds in the North Atlantic at a level of accuracy that would be needed to force an ocean model. Whether steady nonlinear solutions would do a better job has not been determined, but if so they would be a valuable tool for studying this interaction.)

In QG linear theory about a zonal flow, one can think of an orographic wave as forced at the surface by the vertical motions created by flow up and down the orography $h(x, y)$: $w \approx w_L \equiv U\partial h/\partial x$. One intuitive measure of nonlinearity is the extent to which the near-surface vertical velocity departs from this value, or the extent to which $w_N = \mathbf{v} \cdot \boldsymbol{\nabla} h$ departs from w_L, where $\mathbf{v}$ is the total flow. By this measure, observations suggest that the flow is very nonlinear (Saltzman and Irsch 1972). Chen and Trenberth (1988a,b) argue that it is useful to retain linear dynamics in the interior of the atmosphere while using w_N rather than w_L at the lower boundary. The result is still a linear model, but the lower boundary condition couples different zonal wavenumbers even when linearizing about a zonal flow. Trenberth and Chen (1988) find that the response to Tibet is reduced in magnitude and substantially altered when they modify their model in this way.

This approach is problematic because it retains some terms quadratic in the amplitude of the forcing and not others. One can easily create a situation in which the use of the nonlinear condition on w in the context of a linear wave model causes the solution to be less rather than more accurate. In particular, in QG theory we have the remarkable result that if the mean flow U is dependent only on height and not on latitude, and if the flow is inviscid and adiabatic, then the linear response to topography is an exact solution of the nonlinear QG equations, even though w at the surface is not well approximated by w_L [see Tung (1983) and the discussion in Ringler and Cook (1997)]. When latitudinal variations are present in the basic state, further analysis suggests that an appropriate measure of nonlinearity is the ratio of the meridional displacement of a streamline to the scale over which the index of refraction (the mean potential vorticity gradient divided by the mean wind) changes by order unity.

Ringler and Cook (1997) provide an interesting comparison of linear and nonlinear responses to an isolated mountain in a QG model in which the flow is independent of latitude. This is precisely the model for which linear solutions satisfy the full nonlinear equations if the flow is inviscid and adiabatic. Yet these authors find very substantial changes in the flow as a function of mountain height in a model that includes Ekman pumping and small-scale thermal damping, emphasizing that these nonconservative terms can potentially exert considerable control over the character of the nonlinear modification. They find that the equatorward bundle of rays is favored over the poleward bundle as the amplitude of the topography is increased. In contrast, in the idealized GCM of Cook and Held (1992) the poleward-propagating rays appear to be enhanced at the larger forcing [as in Trenberth and Chen (1988) and Valdes and Hoskins (1991)]. A QG model imposes the boundary condition at $z = 0$ and therefore misses the physical ‘‘blocking effect’’ of a large obstacle, but it is unclear whether this is the key to this distinction. The latitudinal variation of the flow, which is particularly strong at low levels, may be the key difference between these models. There is clearly much yet to be learned by detailed analysis of the steady isolated nonlinear responses to orography in both QG and primitive equation models.

b. The full nonlinear response to orography

We turn now to the full nonlinear response to Tibet in Fig. 3c. Differences between Figs. 3c and 3b are primarily a consequence of the interaction between the response to heating and the Tibetan orography. Further decomposition shows that the interaction with extratropical and tropical heat sources are both important. The result is reminiscent of the large effect of preexisting zonal asymmetries on the response to tropical heating discussed in the following section. We are unsure of the robustness of this result. The result of removing all topography in Nigam et al. (1988) shows a pattern downstream of Tibet that resembles the linear prediction, or our isolated nonlinear solution, more closely than our full nonlinear solution.

We have also computed $[N(T) - N(T - \epsilon A)]/\epsilon$, where ϵ is a small number and A refers to Tibet. The solution is nearly unchanged from that in Fig. 3c, as is also the case for the computation $[N(T - A + \epsilon A) - N(T - A)]/\epsilon$. Therefore, one can generate the pattern in Fig. 3c by linearizing about the zonally asymmetric state generated by the steady-state model with or without Tibet present in the forcing. The height of Tibet is not the key ingredient.

Comparing the full and isolated responses to the Rockies in Figs. 3e and 3f, we find that, unlike Tibet, the pattern in not changed dramatically, although the full response is of larger magnitude.

6. Thermal forcing

Figure 8 shows the diabatic heating in January, averaged between the surface and 100 mb, as computed from the NCEP–NCAR reanalysis, using the method outlined in the appendix. This pattern suggests a natural division of the thermally forced stationary eddies into parts forced by tropical heating (south of 25°N) and by extratropical heating (north of 25°N). Figures 9a,d show the linear responses to tropical and extratropical heating at 300 mb. Our ‘‘extratropical heating’’ includes the eddy sensible heat flux convergence. Given the close relationship between low-level eddy sensible heat fluxes and the eddy latent heat fluxes that shape the heating in the storm tracks, we prefer not to separate the extratropical eddy flux convergence from the heating field. As described in section 2, ideally one would include a closure theory for these fluxes in one’s steady-state model.

We see Rossby wave propagation in both the tropically and extratropically forced eddy fields. Tropical

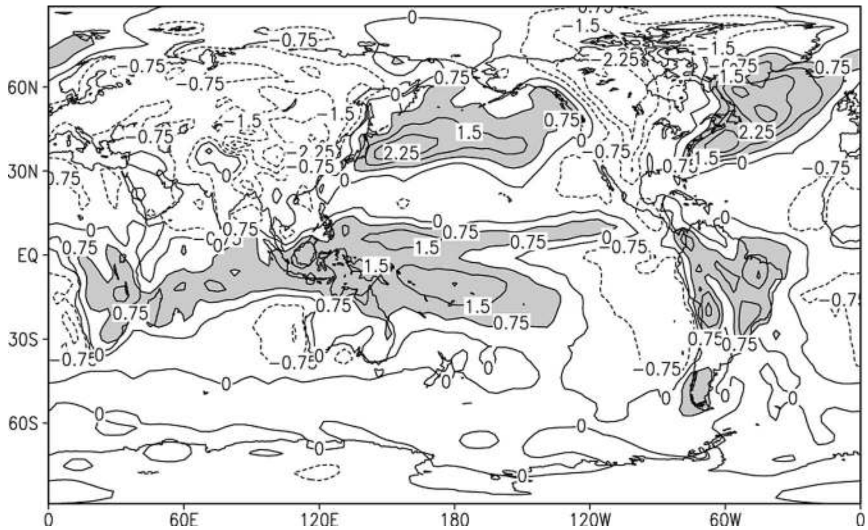

FIG. 8. The column-averaged diabatic heating field in Jan obtained from the NCEP–NCAR reanalysis as described in the appendix. The contour interval is 0.5 K day^{-1}.

heating produces waves arcing across approximate great circle paths into higher latitudes, familiar from the work of Hoskins and Karoly (1981). The response to extratropical heating has the NE–SW tilt that is the signature of equatorward propagation.

In this section, we first discuss the linear and nonlinear responses to tropical heating, focusing exclusively on the extratropical response. We then turn to the response to extratropical heating. Finally, we briefly describe some results that suggest that a key nonlinearity, according to our steady-state model, is that between the response to Tibetan orography and the response to tropical Pacific heating.

a. Extratropical response to tropical heating

The local temperature tendency due to a tropical heat source is balanced to an excellent approximation by adiabatic cooling. Given that the lapse rate itself responds only weakly to this local forcing, specification of the heating in the Tropics is essentially equivalent to the specification of the mean vertical motion and, therefore, of the divergence of the flow. The rotational part of the flow can be thought of as determined by the vorticity equation forced by this divergence at each level in the troposphere. The temperature response can then be diagnosed from the divergence equation through the requirement that the flow be balanced. See Sobel et al. (2001) for a discussion of these approximations. We do not discuss the *tropical* response to tropical forcing further in this review.

The *extratropical* response to tropical forcing is sensitive to the latitude of the source. As an example, Fig. 10 shows solutions to a shallow water model in which the same heat (or mass) source is in one case centered on the equator and in the other is centered at 10°N. The basic state is symmetric about the equator and has weak easterlies near the equator. Given this sensitivity, it is important to strive for a clear understanding of the factors that control the amplitude of the extratropical wave train.

Given the tropical balances outlined above, one is led to think in terms of a linearized two-dimensional barotropic model in which the vorticity source is the vortex stretching/compression $(f + \zeta)D$ in the tropical upper troposphere resulting from the prescribed divergence D. As the heating and divergence are moved poleward, the mean absolute vorticity, $f + \zeta$, and the associated stretching increase in magnitude, explaining the increase in amplitude of the wave train in Fig. 10. This argument has been refined somewhat, utilizing what is referred to as a ''Rossby wave source,'' the full vorticity tendency associated with the divergent flow $\nabla \cdot [\mathbf{v}_D(f + \zeta)]$, where $\mathbf{v}_D \equiv \nabla\xi$ and $\nabla^2\xi \equiv D$. This expression can displace the source farther poleward than the stretching term in isolation, to latitudes where the absolute vorticity is larger and where the mean westerlies favor stationary wave propagation (Sardeshmukh and Hoskins 1988).

This concept of a Rossby wave source is not fully satisfying, for it depends on the level at which the source is computed and does not tell one how the response depends on the vertical structure of the heating or of the zonal mean flow. If the zonal flow were, as an extreme example, independent of height, the heating would not force any external mode wave train. The fact that we can think of the wave train as forced only by the upper-level divergence is dependent on the fact that the low-level convergence is embedded in mean easterlies where it is ineffective as a wave source.

One is tempted to try to project the heating onto the external mode directly. But ray tracing in three dimensions informs us that rays tend to be nearly horizontal

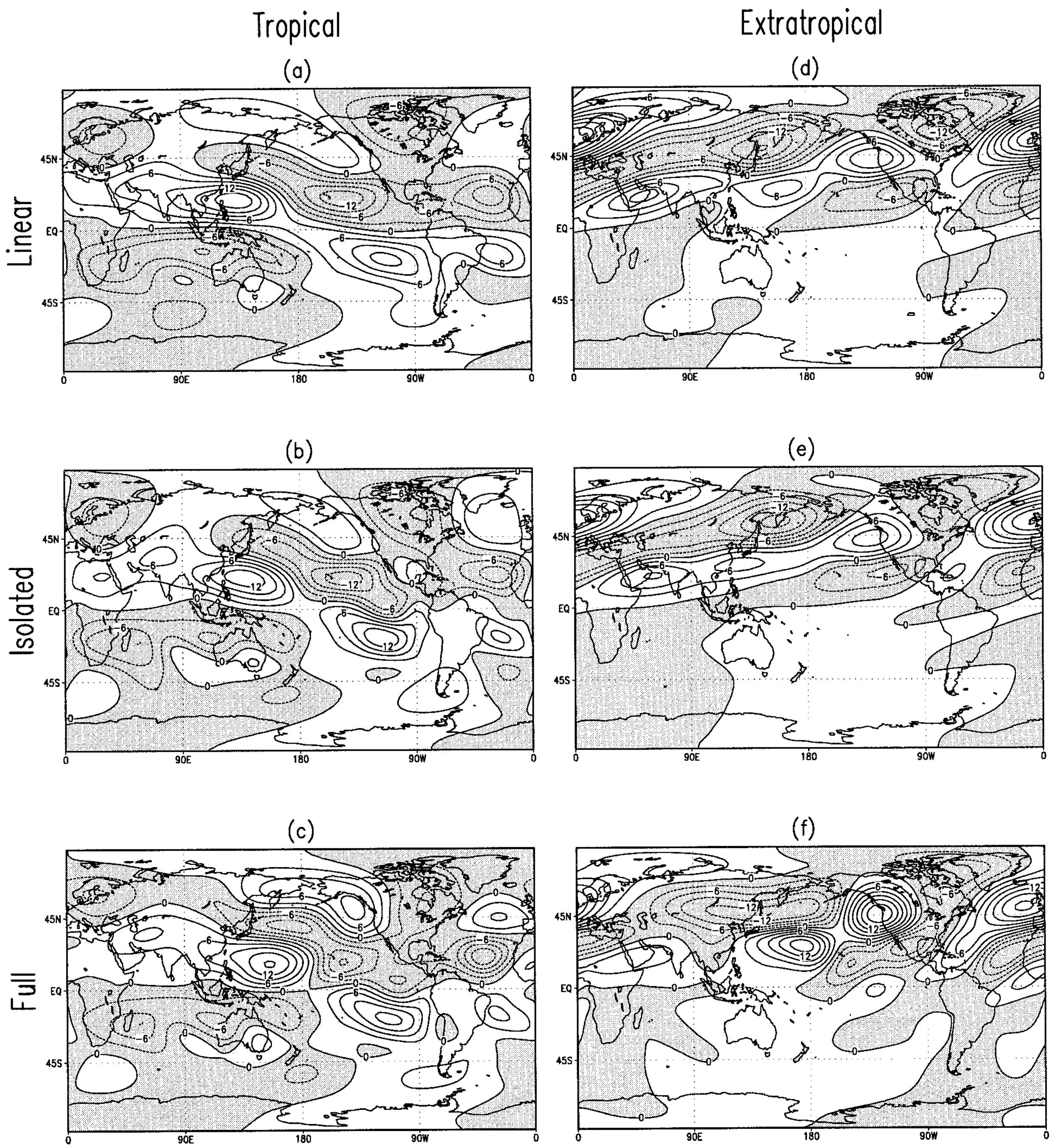

FIG. 9. The 300-mb eddy streamfunction for the (a) linear, (b) isolated nonlinear, and (c) full nonlinear responses to NH tropical heating in Jan; the analogous (d) linear, (e) isolated nonlinear, and (f) full nonlinear responses to extratropical heating. Contour interval is 3×10^6 m^2 s^{-1}.

in low latitudes (Hoskins and Karoly 1981). Focusing on modal structures in the vertical is only useful when rays propagate vertically across the troposphere before they propagate very far horizontally. For tropical forcing, it is more appropriate to think of the solution as being forced more or less independently at each level and then propagating horizontally away from the source until it can develop a significant vertical component in its group velocity. Consistent with this picture, Ting (1996), in a study of how to mimic a baroclinic model's extratropical response to tropical heating with a barotropic model, obtains the best fit by taking the divergence from the level of the tropical outflow, while linearizing about the zonal flow from the midtroposphere, close to the midlatitude equivalent barotropic level. Forcing a barotropic interpretation on the flow, with a

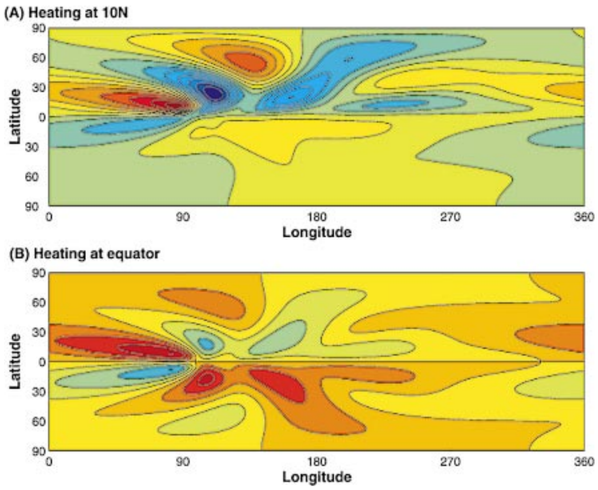

FIG. 10. The response to tropical heating in a linear shallow water model, with the heating (mass source) centered at (a) 10° lat and 90° lon and (b) on the equator. The size of the mass source and the contour interval are identical in both (a) and (b).

single equivalent barotropic level, can create the appearance of a Rossby wave source in the subtropics (cf. Held and Kang 1987). Further analysis of the transition in the subtropics from vertically decoupled vorticity dynamics to the equivalent barotropic extratropical wave train is needed for a more satisfying understanding of the amplitude of the extratropical response, even in this relatively simple case of a zonally symmetric basic state.

The isolated nonlinear response to tropical heating in Fig. 9b does not differ substantially from the linear response. Figure 9c shows the full nonlinear response to tropical heating, which is of substantially larger amplitude in the extratropics. This response is itself quite linear in the amplitude of tropical heating, just as the full nonlinear response to Tibet is fairly linear in the height of Tibet.

The result in Fig. 9c is consistent with the large body of work that shows strong effects of the asymmetric circulation generated by extratropical heating and orography on the structure of the wave trains generated by tropical heating. In particular, the response is enhanced in the North Pacific and over North America. This interaction was first analyzed in barotropic models by Simmons (1982) and Branstator (1985), and continuing work has been motivated by GCM studies, which often show that the extratropical wave train forced by SST anomalies is roughly fixed in longitude, independent of the longitude of the SST anomaly (see Hoerling and Kumar 2002, this issue, for an update). Care must be taken in interpreting these GCM results, since SST anomalies in different regions can produce heating (divergence) anomalies of very different amplitude; without further analysis, they do not imply that a fixed heating anomaly in the Tropics, when displaced in longitude, would favor a response in the Pacific–North American (PNA) sector. However, baroclinic stationary wave models (Ting and Yu 1998) do in fact show this longitudinal preference when a tropical heat source of fixed amplitude is displaced in longitude. These ENSO-related issues are relevant for the response to climatological forcing, as the Pacific mean state can be thought of having the flavor of La Niña when compared to the El Niño–like basic state with zonally symmetric heating.

The extratropical storm tracks are thought to play a significant role in the extratropical wave train forced by El Niño. The results of Held et al. (1989) using a model linearized about a zonally symmetric flow are suggestive, but linear studies with a zonally asymmetric basic state (Hoerling and Ting 1994) are more convincing. The impression from these studies is that the direct effect of tropical forcing must be large in the jet exit region, where the storm track eddy momentum fluxes are concentrated, if feedback from these fluxes is to be significant. This impression is reinforced by Ting and Held (1990), in which there is no hint of reenforcement of the extratropical wave train by midlatitude transients when a GCM with a zonally symmetric climate is perturbed by a tropical SST anomaly. However, it does not appear that this feedback is fundamental to the longitudinal localization of the response (Ting and Yu 1998; Hall and Derome 2000).

It is also unclear whether resonance with a modal structure underlies this longitudinal localization, given the substantial differences between the PNA pattern that is internally generated in midlatitudes and the pattern in the same region forced by tropical heating (see Straus and Shukla 2000, and references therein.) The passage of a wave train through a jet exit region, where $\partial U/\partial x < 0$, is often considered to be the key ingredient, based on analysis of local energetics. Ting and Yu (1998), on the other hand, suggest that one can understand longitudinal localization in linear models with asymmetric basic states by examining the Rossby wave source, implying that the zonal variations in the subtropical (absolute) vorticity may be the key ingredient.

b. Extratropical heating and the storm tracks

As is clear from Figs. 2 and 9, in our steady-state model extratropical heating forces a large fraction of the eddy streamfunction field in the extratropics, consistent with Hoskins and Valdes (1990). In some other linear diagnoses, such as Nigam et al. (1988), orographic and thermal components are more comparable. The primary causes of the difference between Nigam et al. and the present study seem to be that 1) the extratropical heating is weaker in the former, and 2) the near-surface mean winds are stronger, resulting in a weaker response to extratropical heating [because the air spends less time in the heated region—see Held and Ting (1990)] and a stronger orographic response.

One must be careful to avoid overinterpreting this diagnosis, since the midlatitude heating field can itself be influenced by tropical heating, orography, and the

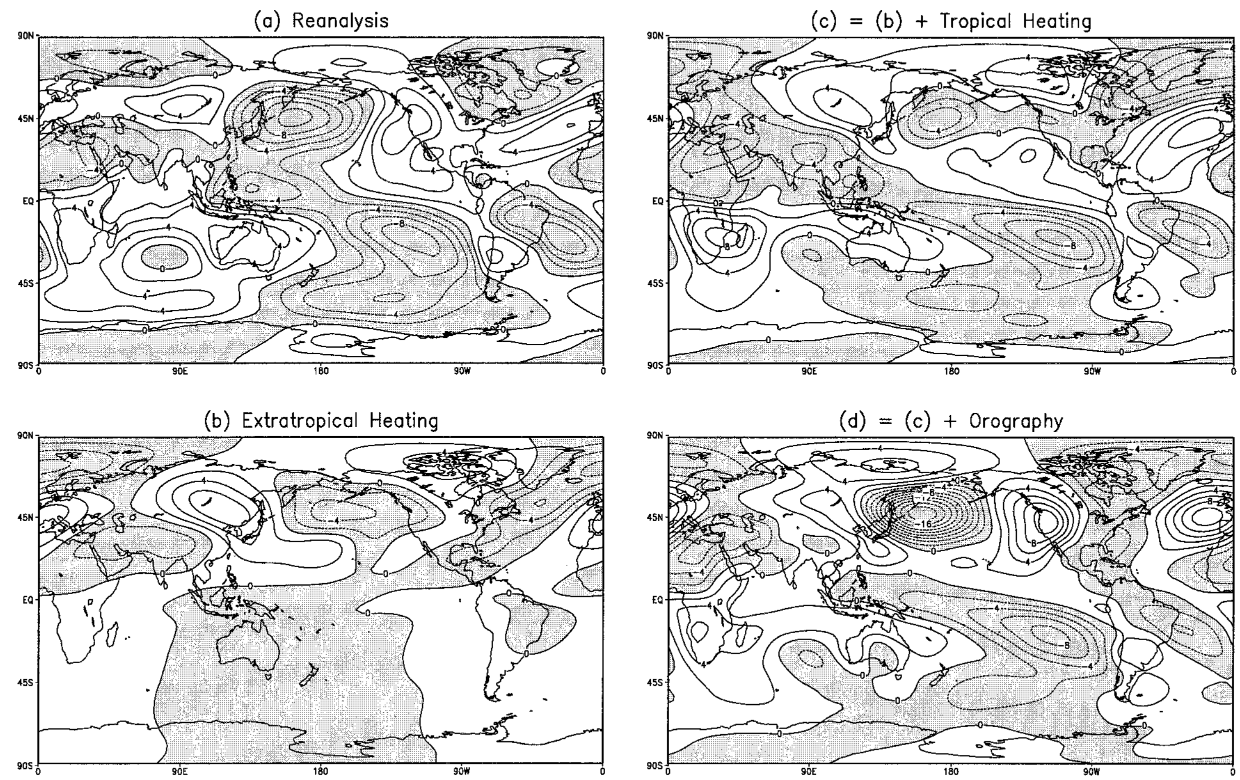

FIG. 11. The eddy streamfunction on the $\sigma = 0.866$ surface for (a) NCEP–NCAR reanalysis, (b) nonlinear response to NH extratropical heating, (c) nonlinear response to tropical heating and NH extratropical heating, and (d) nonlinear response to orography, tropical heating, and NH extratropical heating. Contour interval is 2×10^6 m^2 s^{-1}.

storm tracks. On the one hand, the excitation of the storm track eddies is controlled by the baroclinicity of the low-level flow, which should be most strongly tied to the local thermal forcing, as discussed by Hoskins and Valdes (1990). On the other hand, the idealized storm track model of Chang and Orlanski (1993) also illustrates clearly that localized low-level baroclinicity is not sufficient to create localized eddy activity.

Figure 11a shows the low-level stationary eddy field in the NCEP–NCAR reanalysis, using eddy streamfunction on the 0.866 sigma level. Figure 11b is the isolated nonlinear response to extratropical heating. (We continue to bundle extratropical thermal transients together with the extratropical heating throughout.) While many of the prominent features in the extratropical low-level flow are present, they are distorted to the extent that the resulting model would be of little value in studying regional climates—the oceanic lows are shifted too far eastward and the ridge over the western United States is hardly present. The linear response (not shown) is comparable, but slightly weaker in amplitude and further distorted. Figure 11c is the isolated nonlinear response to the total (tropical plus extratropical) heating, while in Fig. 11d topographic forcing is also included. With the addition of these other forcing factors, the pattern is now more accurate, although several features have too large an amplitude once again, suggesting that the reactive parts of the heating that tend to damp the response are not well represented (see section 2).

The difference between Figs. 11c and 11b is much larger than the response to tropical heating in isolation, comparable to the result at upper levels discussed above. Similarly, the nonlinear response to the addition of orography to these heating fields (the difference between Figs. 11d and 11c) is different in structure from the linear response to orography. See Wang and Ting (1999) for further discussion of these interactions.

In interpreting these results, it remains useful to start with linear theory, even though we see that this can only take us part way to the desired goal. The linear response to extratropical heating at low levels has a large component that is not a free stationary Rossby wave but rather can be thought of as a local particular solution. The bulk of the extratropical heating is rather shallow. In direct contrast with the tropical forcing problem, within this particular solution the temperature tendency within the source region is, to first approximation, balanced by horizontal advection rather than adiabatic cooling due to vertical motion. As a consequence, one can hope that a very simple advective solution can be used to estimate the low-level linear response. If zonal advection of eddy temperature by the mean flow is dominant, $U\partial T'/\partial x \approx Q$ then one can obtain T' by integrating Q/U in longitude and then computing the eddy streamfunction from hydrostatic balance, leading to cold highs and warm lows. Including meridional advection of the mean temperature field by the eddy meridional flow does not change this picture significantly, since the effects of zonal and meridional advection are in phase (see Held and Ting 1990).

One needs to add on a homogeneous solution to satisfy the lower boundary condition, and it is this homogeneous component that will dominate as one moves away from the forcing (e.g., Chen 2001). In particular, this homogeneous solution will contain an external Rossby wave component. One is tempted to assume that heating cannot excite the external mode, but this is only true in the unrealistic case of no vertical shear. In QG theory, assuming that $U(z) > 0$, the external mode response to a heat source $Q(z)$ is proportional to

$$\int \frac{Qw_e}{U^2}\, dz, \tag{3}$$

where w_e is the vertical velocity in the mode (HPP). Because of the factor of U^2 in the denominator, the linear far-field response to extratropical heating is sensitive to the vertical structure of the heat source near the surface, where U is relatively small.

Typically, the particular solution is expected to dominate within the source region, so one might hope to use the simple advective model as the starting point for more elaborate models in which one includes theories for the heating field. Indeed, diffusive energy balance models for the surface temperature often evolve into models of this type when one attempts to include the effects of horizontal advection. But, as indicated in Fig. 11, this is unlikely to result in practically useful models of the extratropical low-level flow without somehow also taking into account the much more nonlocal effects of tropical heating and orography.

The additional inclusion of the eddy vorticity fluxes has little impact on the low-level flow depicted in Fig. 11. While we tend to intuitively think of the wintertime oceanic lows as being the graveyard of extratropical low pressure systems, it is heating and orography, and not the transient eddy vorticity fluxes, that are responsible for these features in stationary wave decompositions. It is striking when one's synoptic intuition is so distinctly at variance with the results of stationary wave modeling.

7. Nonlinear interaction between heating and orography

The interaction between thermal and orographic forcing has been considered by several authors (Chen and Trenberth, 1988b; DeWeaver and Nigam 1995; Ringler and Cook 1999; Wang and Ting 1999) but much work remains before we can place all of these calculations in a consistent context. The large response to heating can alter the response to orography in several ways. It can modify the flow incident on the orography and alter the downstream wave trains at their source, or the changes in the local zonal wind structure can alter the propagation of the wave trains. Or the thermal forcing can create a flow that possesses a near resonance that is excited by the orography. We can make symmetrical statements about the effects of the response to orography on the thermally forced waves.

Whether we examine the total nonlinearity (Fig. 1d) by subtracting the linear solution from the nonlinear solution with all forcings present, or if we examine the difference between the full and the isolated responses to orography (Fig. 3) or the difference between the full and isolated responses to tropical heating (Fig. 9) the resulting pattern is always largest in the Pacific–North American sector. Therefore, in trying to isolate the important sources of nonlinearity in our steady-state model, we are finding it useful to focus on the interaction between Tibetan orography and the heating in the tropical Pacific.

Setting A = Tibet and B = heating in the tropical Pacific, we find that the difference $N(A + B) - N(A) - N(B)$ is similar in many respects to the total nonlinearity in our steady-state model (Fig. 1d), suggesting that this isolates an important part of the nonlinearity. We also find the analogous interaction between Tibet and extratropical heating to be significant, but to bear less resemblance to the total nonlinearity. We have also computed

$$\frac{N(\alpha A + \beta B) - N(\alpha A) - N(\beta B)}{\alpha\beta} \tag{4}$$

for values of α and β between 0 and 1 to see how this interaction evolves as the size of the two forcings increases. When normalized in this way, all of these difference maps are of the same amplitude. As α and β are varied continuously, we see a smooth evolution in pattern, with no suggestion of resonance. In fact, we see a rough similarity between the results with very small amplitudes to those obtained with full strength, indicating that an appropriate starting point for thinking about this nonlinearity could be a perturbation theory for the interaction between *infinitesimal* Tibetan and tropical Pacific sources.

8. Concluding remarks

We believe that the classic decomposition of the flow into zonal mean and stationary waves is a useful one, since the factors that maintain the zonal mean flow are often distinct from those controlling the structure of the stationary waves, and that the approach of modeling the stationary eddies while holding the zonal mean fixed continues to be fruitful. We have indicated in this review a few questions that have yet to be fully addressed with regard to theories and models of the stationary wave field. Many of these questions can be best attacked, in our view, with the *simultaneous* use of GCMs, both realistic and idealized; and stationary wave models, both linear and nonlinear. In particular, we feel that studies using GCMs with idealized boundary conditions are vital, given the complexity of the earth's boundary, but the value of such modeling would be greatly enhanced if it were more often coupled to attempts at steady-state modeling of the stationary waves. Work with steady models must focus more strongly on the prediction of heating rates, and relevant eddy fluxes, from the appropriate boundary conditions for the atmosphere, rather than being satisfied with diagnoses of the response to prescribed heating distributions. This is also a natural path toward constructing climate models intermediate in complexity between GCMs and the simplest energy balance models.

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APPENDIX

Description of Model

The nonlinear stationary wave model is based on the three-dimensional primitive equations in σ coordinates. All the basic variables are deviations from a prescribed zonal flow. The basic prognostic equations are those for perturbation vorticity, divergence, temperature, and log(surface pressure). Perturbation geopotential height and $\dot{\sigma}$ vertical velocity are calculated from the diagnostic hydrostatic balance and mass continuity equations. A semi-implicit time integration scheme is employed with a time step of 30 min. The stationary wave solution in this model is obtained by integrating the model to a quasi-steady state after a short period of time. The top of the model is formally at zero pressure but still effectively acts as a rigid lid. The model has rhomboidal wavenumber-30 truncation in the horizontal and 14 unevenly spaced σ levels in the vertical. The basic state employed in this study is the zonal mean climatological (1948–99) basic flow in January taken from the NCEP–NCAR reanalysis. The forcings for the nonlinear model include orography, diabatic heating, and transient vorticity and heat flux convergences. During model integration, the zonal mean of the basic variables is relaxed very strongly, with a timescale of 3 days, to the observed zonal mean. (We have found that relaxing the zonal mean with a short timescale is often preferable to simply specifying it at the observed value, perhaps because it allows small adjustments to a flow preferred by the model truncation; we have repeated the calculations with a prescribed zonal flow, and none of our results are altered significantly.) Linear simulations are simply obtained by reducing the strength of the forcing by a factor of 100. More details about the model equations can be found in Ting and Yu (1998).

The damping used in the nonlinear model includes Rayleigh friction, Newtonian cooling, and biharmonic diffusion. The Rayleigh friction damping times for both the vorticity and divergence equations are 0.3, 0.5, 1.0, and 8.0 days for the lowest four σ levels (0.997, 0.979, 0.935, and 0.866), and 25 days throughout the rest of the model. The timescale of the Newtonian cooling is 15 days at all levels. The biharmonic diffusion coefficient, identical for vorticity, divergence, and temperature, is chosen to be 1×10^{17} m^4 s^{-1}. This is significantly stronger than the values typically used in GCMs of this resolution, and helps to suppress model-generated transients.

The diabatic heating was computed from reanalysis data using the thermodynamic equation in pressure coordinates and then spatially interpolating onto the model resolution. Our experience with the stationary wave model is that this residually derived heating is more consistent dynamically than the heating directly provided by the reanalysis.

When subjected to the prescribed zonal mean climatological basic state from NCEP–NCAR reanalysis, the fixed stationary wave forcings, and the specified dampings, the nonlinear model reaches either a true steady state or a quasi-steady state after being integrated for around 20 days. In the latter cases, weak transient eddies are produced, but inspection shows that they are too weak to modify the mean flow significantly. The strongest transients are generated in models forced by extratropical wintertime heating in isolation, for reasons that are unclear. These transients are typically nearly periodic and have no significant low-frequency variability. The nonlinear model solutions shown in the text are averaged over days 31–50, a period adequate to generate results that are not sensitive to the averaging period.

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