The Role of Transients in Weather Regimes and Transitions

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ABSTRACT

Transition of weather regimes is examined in a highly simplified model. Two completely distinct internal methods of transition are identified. The first is a synoptically triggered large-scale instability, while the second is an energy inconsistency between the large-scale and synoptic scales that does not allow the two scales to equilibrate. In the atmosphere, the first case appears as a sudden propagation and damping (or vice versa) of the large-scale pattern with no obvious warning, while the second is consistent with the synoptician's description of a regime being disrupted by a single catastrophic event such as explosive cyclogenesis. The first method is always fast (on a synoptic time scale), while the second does not have to be, though often is. By examining what causes the regimes to fail, one can better understand the role of the transients during all phases of weather regimes.

1. Introduction

Weather regimes (Reinhold and Pierrehumbert 1982, hereafter RP), persistent anomalies (Dole 1982), blocking and/or quasi-stationary waves are phenomena whose common feature is that *they persist for time scales long compared to not only baroclinic instability but the passing of mobile baroclinic systems* (Dole 1982). Perhaps the most fascinating aspect of these weather regimes and/or persistent anomalies is that they exist at all in an atmosphere otherwise characterized by continuously developing, *apparently* chaotic, synoptic-scale mobile disturbances. Furthermore, indications are that these regimes can develop and persist without any outside stimulus from changing boundary conditions. The internal nature of regimes is well documented in simple models (RP; Legras and Ghil 1985) and more complicated systems (Vautard and Legras 1988) as well as general circulation models (Blackmon et al. 1986; Lau 1981). In these latter two studies, the models show equally robust and intense regime phenomena with and without any time varying boundary conditions such as sea surface temperature anomalies. Somehow the planetary-scale waves are able to both equilibrate with the mobile systems as well as avoid self destruction through instability. On the other hand, their persistence is not indefinite, so there is some mechanism that allows them to become established and another mechanism (maybe the same one) that terminates them.

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In this study, we identify two distinct mechanisms of regime transition in the highly simplified model of RP, both of which appear to have analogs in the atmosphere. Though the model's severe truncation and simplicity have raised several concerns (Cehelsky and Tung 1987; O'Brien and Branscombe 1988, 1989, 1990), the study of Vautard and Legras (1988) has shown that the regime equilibrium process hypothesized by RP is still valid in their 21×21 wave model (see their Fig. 2). Thus, we feel that these transition mechanisms are also valid in more complicated systems that possess weather regimes. The interesting point is that by examining what causes the regimes to fail, we can better understand the total role of the transients during maintenance.

The model is the quasigeostrophic two-layer spectral channel model of RP, Cehelsky and Tung (1987), and O'Brien and Branscombe (1988, 1989, 1990). The nonconservative processes consist of a radiative thermal relaxation (which drives the model), Ekman dissipation (of different time scales) on the lower and upper layers as well as between the layers, and Newtonian cooling. The constant static stability, beta effect, and orographic height are also parameters that must be specified for each run. The channel width is 5000 km, which in combination with the parameter n (the aspect ratio between the channel width and zonal periodicity) gives the scale of the largest wave in the system. For further formalisms, the interested reader is referred to the preceding studies.

2. Approach

As a basis for our analysis we consider idealized cases of weather regime transitions. The parameter sets are chosen such that regime transition is rare and characterized by highly different flow patterns. The separation in phase space between the two regime states in these cases is large, which makes it easy to graphically display the two-regime behavior in two dimensions without any filtering. There is also no question about which regime state the model is in at any given time. In the cases examined, one regime tends to consist of a steady, high amplitude trough upstream of the orographic ridge with weak zonal flows, which we refer to as the trough regime; while the other consists of an erratically varying large-scale wave (whose time mean state is given by a ridge almost in phase with the orographic ridge) and strong zonal flows, which we refer to as the high-index regime. (The behavior of the large scale is so erratic in the high-index regime that one would be reluctant to classify it as a regime until one sees the model evolve into the trough regime. Then it is very clear that the model has two highly separated flow states.) We set the driving and dissipation such that each regime state is absurdly persistent by atmospheric standards, lasting several months to years (and is even indefinite, but still chaotic, at low enough driving or high enough dissipation). Transitions, nevertheless, when they finally do occur, happen in just a few days. Although all of these cases occur on an f plane (they had the cleanest separation in phase space), the addition of beta does not qualitatively alter the mechanisms.

To further simplify the analysis, we concentrate upon the breakdowns of the trough regime to an erratic, high-index regime. The steadiness and quiescence of the trough regime allow us to more easily identify the causes of the eventual breakdowns.

3. Synoptically triggered large-scale instability

For the first idealized case we use the same model truncation as in RP (2×2) with the following external parameters:

$$k = 0.04, \quad k' = 0.005, \quad j = 0, \quad \beta = 0, \quad h = 0.045,$$
$$\sigma = 0.15, \quad \theta^* = 0.13, \quad \hbar_{1,1} = 0.15, \quad n = 1.2.$$

By atmospheric standards, the dissipative parameters (surface friction k and Newtonian cooling h) are too large, and by terrestrial standards, the mountain height ($\hbar_{1,1}$) unrealistically high; but these values appear to be necessary not only to enhance the idealized behavior of the regimes, but to dampen the excessive activity of the baroclinic processes of the quasigeostrophic two-layer model. The static stability σ is a reasonable estimate of the static stability over the depth of the troposphere. With the above "n" value, the largest-scale wave corresponds to approximately global wavenumber 3, while the "synoptic wave" (which has half the space scale) is global wavenumber 6.

In Fig. 1 we show a phase–space picture of the climatological behavior of the largest-scale streamfunction wave in the model for 2×2 truncation. The dots are plotted each day for a period of 85 years. The ordinate is the sinnx component of the wave, and the abscissa is the cosnx component. Since the orography has the same structure as the largest-scale wave, all

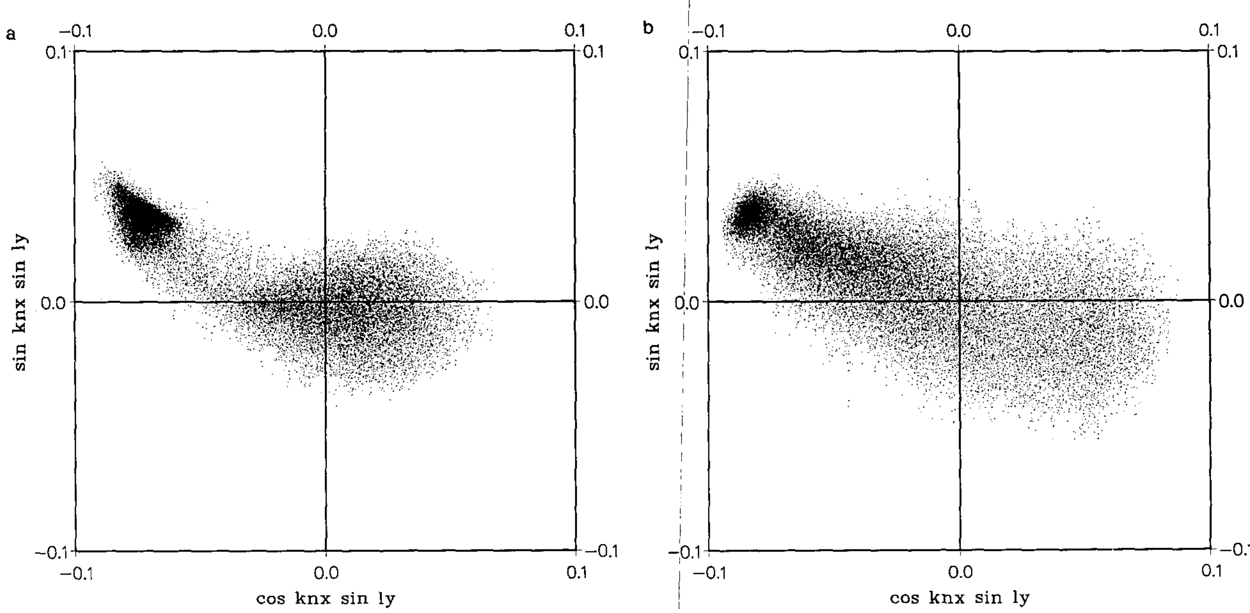

FIG. 1. Climatological plots of the large-scale streamfunction for (a) 2×2 and (b) 5×5 truncations. The abscissa is the cosnx component and the ordinate is the sinnx component of the streamfunction. Since the large-scale wave has the same structure as the topography, points along the positive abscissa correspond to a ridge over the mountain ridge. The points are plotted each day for a period of 85 years in both cases.

dots that lie along the positive abscissa correspond to a 500-mb ridge in phase with the mountain ridge.

The single most impressive feature of the diagrams is the difference between the two regime states, a difference that would be even more enhanced in a three-dimensional diagram with the third axis as the zonal flow. One regime is characterized by a relatively tightly knit clump of points in the upper left-hand corner of the diagram, while the second regime is characterized by a much wider scattering of points centered to the right of the origin. There are a few points outside of these two distinct regions. From the position of the two clumps we can see that the steady regime corresponds to a large amplitude trough about 40 degrees upstream of the mountain ridge, thus its name, while the remaining regime varies widely about a ridge centered over the mountain ridge. The rapidity of the transitions is clearly seen in Fig. 2, which shows a time series of the amplitude of the large-scale wave irrespective of phase for a 1200-day segment of the above run, where the time is given along the abscissa. Each dot represents 1.5 hours. It is clearly seen that the switch between the two regimes is accomplished in a matter of days.

An examination of several different variables (the energies, phase, structure, zonal flows, small-scale eddies, etc.) in a similar set of time series does not reveal any systematic precursor to the regime transitions. Instead, the transitions in all the aforementioned quantities (that differ between the two regime states) occur essentially simultaneously. This behavior suggests some type of large-scale instability.

To examine this hypothesis we compute traditional stability analyses of the large-scale state as a function of time, where the large-scale state is defined as the current values of those components having a nonzero time average over the composite regime. For the 2×2 model truncation the eigenfunctions that have nonzero time means are those with 2D wavenumber $(m, l) = (0, 1)$ and $(1, 1)$. There are only three possible structures of the unstable modes with such a basic state in a 2×2 truncation; a pure (0, 1), (1, 1) and/or (0, 1), (2, 2) structure and/or a coupled (0, 1), (0, 2), (2, 1), and (1, 2) structure. The coupled structure gives a crude approximation of a storm track. It consists of a train of disturbances whose centers follow the large-scale wave pattern with regionally enhanced amplitudes downstream of the ridge and trough axes. This mode is almost always observed to be unstable, though the growth rates are much larger during the chaotic regime with high zonal indices. The (0, 1), (1, 1) instability is a large-scale instability, and it is an orographically modified baroclinic instability that is often stationary (a so-called orographic instability). It is the large-scale instability that turns out to be of interest for the transition. The (0, 1), (2, 2) mode is always stable during the trough regime.

The results of this stability analysis are also depicted in the amplitude time series of Fig. 2. Large-scale stability is denoted by dots, while the occurrence of large-scale stationary instability is denoted by crosses. Instability is seen to occur frequently during the high-index regime (but the high variability of the so-called basic state in this regime makes the interpretation of the stability analysis dubious) but most importantly at the onset and collapse of the trough regime. Examination of several trough regime cases shows that the instability always occurs with transition, but transition does not always occur with the instability (as can be seen in Fig. 2 at day 235 shortly before the transition at day 265. On the other hand, the occurrence of the instability is accompanied by an enhanced variance in the large-scale amplitude even if transition does not occur, suggesting that one is on the threshold of transition.

It is of interest to examine which feature (amplitude, vertical structure, phase with respect to the mountain, zonal shear, etc.) of the time-mean large-scale weather regime state has to be altered to give rise to the instability, even though we anticipate that the exact details are parameter dependent. Thus, we start with the time-mean regime state and independently vary each of the variables holding the remaining variables constant. We find that decreasing the amplitude, *decreasing* the westward phase tilt with height (increasing equivalent barotropy), and moving the large-scale disturbance westward with respect to the orography, all contribute toward decreased stability, though the greatest sensitivity is seen in the structural rearrangement and amplitude. These properties are summarized in Fig. 3, which shows the marginal curve as a function of amplitude (ordinate) and phase difference between the streamfunction and potential temperature components, denoted by G (abscissa). The ratio between the temperature and streamfunction amplitudes is constant. The shaded area indicates those regions where the large scale is stable to large-scale perturbations, while the circle denotes the time-mean amplitude and phase difference of the trough regime state.

It is clear that as the wave decreases in amplitude and/or becomes more and more equivalent barotropic (G approaches $0°$), the stability of the wave radically decreases. The structural sensitivity is of interest because the vertical structure, unlike the amplitude, which has often been qualitatively recognized by forecasters as a critical element for weather regime persistence, is not so easily observed synoptically. Thus, significant changes in this quantity can easily go unnoticed, leading to surprise transitions. On the other hand, we do not mean to imply that it is equivalent barotropy per se that leads to greater weather regime instability in the atmosphere (such details are surely parameter and case dependent), but that the vertical structure, in general, may be an important element.

The critical mechanism in stationary (orographic) baroclinic instability is the zonal flow induced by the surface component of the primary wave interacting with the orography. Small changes in this induced zonal flow strongly influence the phase speeds and growth rates of the primary disturbance by altering the vertical structure. Consequently, weak surface flows can dramatically change the total flow without contributing significantly to the fundamental energetics of the changes, which are baroclinic (Reinhold 1990). The sensitivity of the instability to the induced zonal flows makes it no surprise that both the vertical structure and amplitude should be critical factors in determining the stability threshold of the weather regime as it is these two properties that determine the surface flows.

Regardless of the details of the instability threshold, be it amplitude and/or vertical structure, the important aspect of the above transition mechanism is the role of the mobile synoptic systems as a *trigger*. Here the major role of the mobile systems is to provide an environment in which the large scale is *stable to itself* by releasing the inherent baroclinicity of the system. In

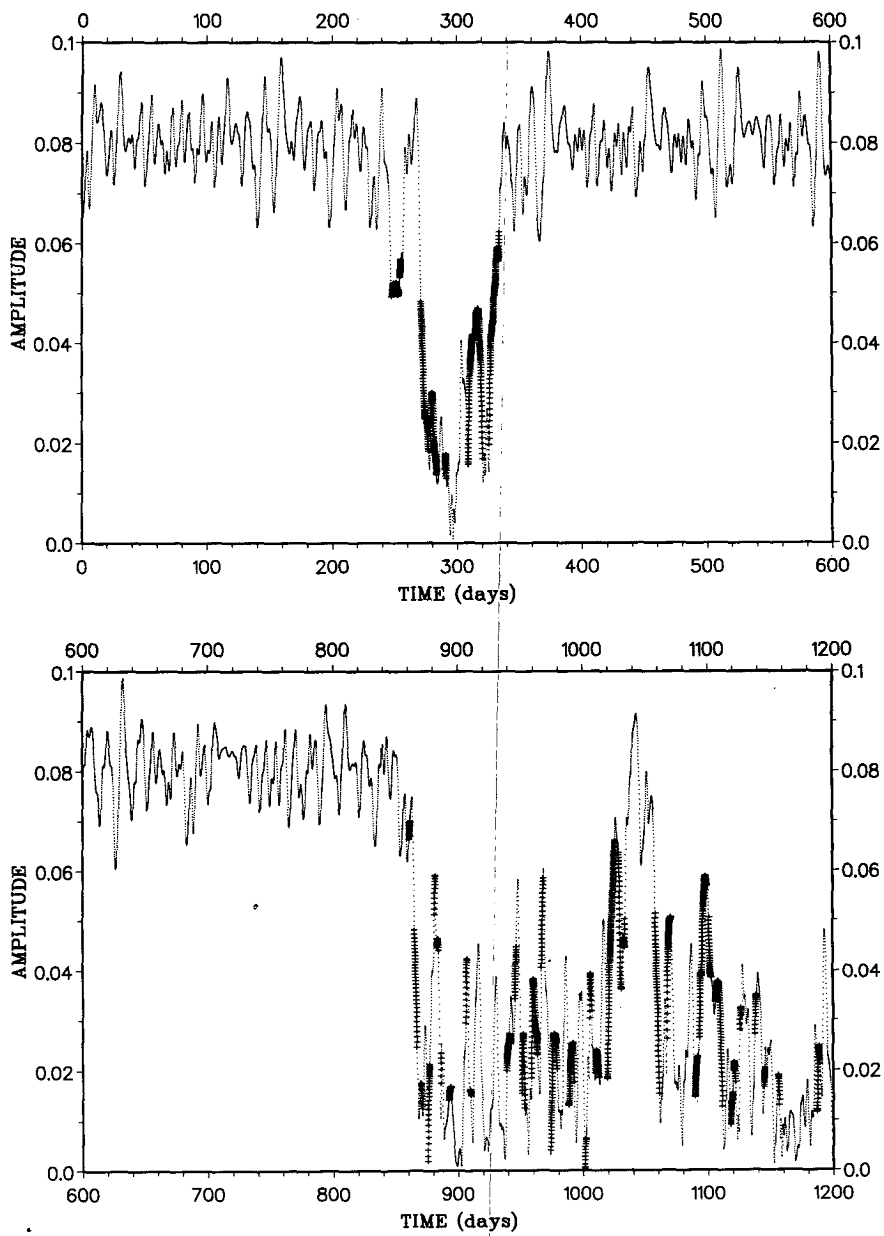

FIG. 2. Time series plot of the amplitude of the large-scale wave for a 1200-day segment of the case shown in Fig. 1. The points are plotted every 1.5 hours as dots if they are stable to large-scale perturbations and as crosses if they are "orographically" unstable. The figure shows both rapid transitions as well as the occurrence of the stationary instability associated with the trough-regime breakdown.

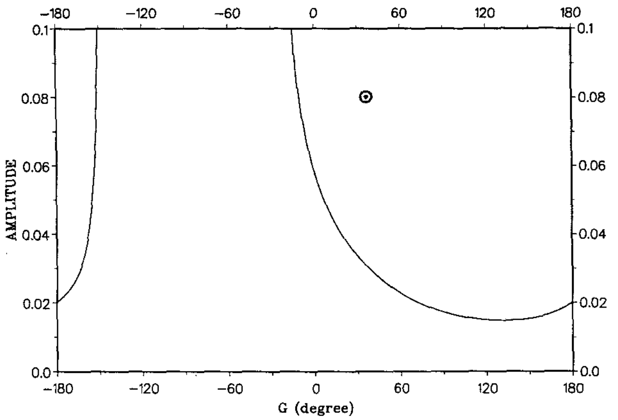

FIG. 3. Stability diagram for perturbations about the time-mean trough regime state. The two variable parameters are the amplitude (the ratio between the temperature and streamfunction amplitudes is held constant) and phase difference between the streamfunction and thermal components of the wave, G (abscissa). The point is the time-mean regime state and the shaded regions are orographically stable. We see both increased amplitude and increased westward phase tilt with height increase regime stability.

the absence of the small scales, this baroclinicity would have to be released through the large scale. When the transients fail to accomplish this goal, the regime collapses through a large-scale instability whose speed is determined by the time scale of the instability, which in this case is baroclinic and fast.

In the atmosphere, the release of baroclinicity through the smaller scales is probably much more effective than in the two-layer quasigeostrophic model. Shallow layers of low static stability favor the growth of very short shallow disturbances (Blumen 1979; Staley et al. 1977; Staley 1991; Whitaker and Barcilon 1992), which help erradicate the development of intense baroclinic zones upon the large scale. These phenomena are well recognized in the atmosphere (Bosart 1981; Uccellini et al. 1985). In fact, any process that reduces the baroclinicity and phase speed of the large-scale disturbances enhances the relative influence of orography and, thus, stationarity (Reinhold 1990). The dominance of baroclinic instability in the two-layer model is probably one reason why we must resort to high mountains and excessive dissipation to obtain even reasonably persistent weather regimes.

In spite of the net role of the transients described above, the kinematic picture of the transition is an instantaneous "push" of the large-scale wave over the stability threshold by an individual synoptic event. However, it need not be a violent disruption of the planetary scale as discussed by Sanders and Gykum (1980) or Colluci (1985), as the energetics of the transition are baroclinic on the large-scale and thus, the triggering event does not have to be spectacular. The effectiveness of the trigger depends more upon the state of the flow at the time of the individual event than anything else.

An interesting implication of this transition mechanism for atmospheric regimes is the fast speed and absence of any impressive precursor. A feature revealed by Dole's (1986, 1988) point-selection criteria analyses that has not been possible to elucidate by other techniques is the rapid onsets and collapses of persistent anomalies. The speed of transition and strong baroclinic character (Dole 1986) of the anomalies during the transition periods are at least consistent with the above ideas. Cessi and Speranza (1985) have also suggested that some type of large-scale orographic instability may be responsible for the development of Dole's persistent anomalies.

Furthermore it may be possible that the persistent Pacific negative event of 1976–77 (discussed by Namias 1978) collapsed through such a process. In that case no single spectacular event is identified which, on 20 February 1977, led to the collapse of a flow pattern (in a couple of days) that had persisted for months.

4. Regime-equilibrium instability

For the parameter set discussed above, all transitions appear to be associated with a synoptically triggered large-scale instability. However, if we simply drop the parameter n from 1.2 to 1.0 in the 2×2 model, the breakdown of the high-amplitude regime is no longer marked by large-scale instability, even though the climatology of the flow and regime evolution are strikingly similar. Presumably, the longer wavelength of the largest scale in the 2×2 model and the increased baroclinicity of the synoptic scale wave do not allow the large-scale state to come near enough to any stability threshold. In this case, the regime collapses because the energy in the synoptic scales becomes too large; thus, the equilibration process between the large scale and the transients or *regime equilibrium* (section 8 of RP) becomes unstable.

The key to the regime-equilibrium instability comes from the time series plot of Fig. 4, which shows both the amplitude of the large-scale wave (small dots at 1.5-h increments) and the synoptic-scale wave (large dots). While all characteristics of the large scale (amplitude, structure, phase, etc.) are nearly constant until the point of transition, the amplitude of the synoptic-scale wave systematically grows during the final stages of the regime until it reaches a certain threshold and the regime collapses. To demonstrate that the regime collapses from the high energy in the transients we compute the regime equilibria as a function of the energy in the synoptic scales.

The results are presented in Fig. 5, which shows the amplitude (heavy line) and phase (thin line), with respect to the mountain, of the large-scale component of the two regime equilibria as a function of the energy (in percent of the zonal flow energy) of the synoptic eddies. At zero energy in the synoptic scales, the regime equilibria reduce to the stationary equilibria of Charney

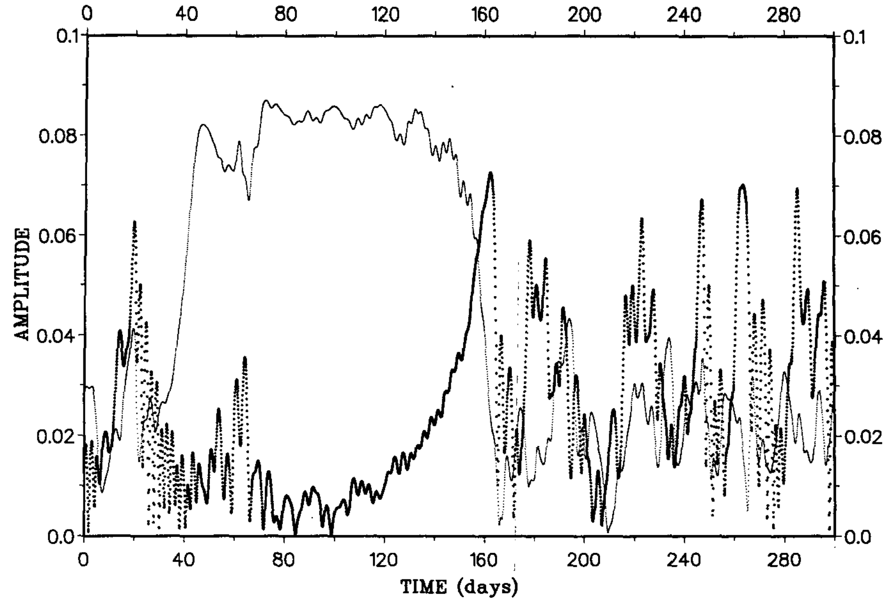

FIG. 4. Time series of the amplitude of the large-scale wave (small dots) and synoptic-scale wave (heavy dots) during the case used in Fig. 1a with $n = 1.0$. The dots are plotted every 1.5 hours for a period of 300 days. The phenominal rise in the synoptic-scale energy prior to the trough regime breakdown is clearly seen.

and Straus (1980), which have, of course, no associated storm track. As the energy in the synoptic scales is increased, the amplitude and phase of the large-scale component evolves from the stationary solutions, and there is an associated storm track. However, when the synoptic energy reaches a value of about 22%, the amplitude of the trough regime rapidly decreases and then ceases to exist. The high-index regime continues to exist to a value of about 24%, above which a jump is made

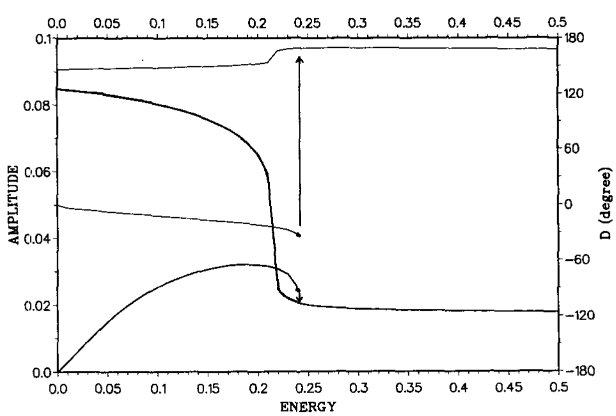

FIG. 5. The amplitude and phase (ordinate) of the regime equilibria for the case studied in Fig. 4 as a function of the energy (abscissa) of the transients as computed by the method of RP. The heavy lines are the amplitudes and the thin lines the phase with respect to the mountain. A clear break in the two solutions occurs about the value 0.25, but the most important feature is the rapid change of the trough regime solution at values exceeding 0.22.

to the one single solution branch remaining, which also has high zonal indices. However, one has to be careful in interpreting the details of this threshold, as the energy is given in percent of the zonal flow energy, and 22% in the trough regime represents much less transient energy than 22% in the high-index regime. Furthermore, the regime-equilibrium balance is not so concerned with the total energy of the synoptic disturbances per se, but with the intensity of the *transports* due to these eddies, which depend upon the *structure* as well as the energy of the disturbances.

We do not understand what causes the transients to systematically intensify in this case. However, the synoptic appearance of the transition is much more in accord with the mechanisms discussed by Sanders and Gykum (1980) and Colucci (1985), in the sense that a series of mobile events continues to pass until an event comes along that is so intense that a catastrophic change in the planetary flow results.

Though explosive cyclogenesis has been frequently suggested as a means of radically changing the planetary-scale flow, too little synoptic energy may lead to a similar result. The regime equilibrium calculations for several parameter sets tend to show instability for *low* synoptic-scale energies for one regime (usually the higher-index regime), while the reverse is true for the remaining regime.

A major difference between this transition mechanism and the triggered instability mechanism (which shows no increasing tendency in the synoptic-scale energies at all) is the time scale. The baroclinic processes of the instability mechanism for all intent and purpose force that mechanism to be fast. The regime-equilibrium instability, whether it occurs from too much or too little transient energy, need not be, even if it happens to be in the case examined here. The fast speed of the transition observed here is a consequence of the insensitivity of the phase position and amplitude of the large-scale component of the regime equilibrium to the equilibrated energy of the transients (all the way up to the point of nonexistence). It could very well be that the large-scale components of the flow change gradually as a function of the transient energy, and a steady distortion of the flow would ensue, leading to a much less spectacular shift.

5. Mixing the mechanisms

The preceding two case studies are highly idealized, and the breakdown of the respective trough regimes is always accomplished by the same mechanism. However, as we start increasing the driving, lowering the friction, increasing the resolution, or in general, start moving away from these highly idealized cases, both mechanisms are seen to operate, either intermittently or together. Even if we retain our basic parameter set and simply change the value of n to something more in the middle (1.1 or so), we can see the combination of both transition mechanisms, which is easily observed (as long as the model regimes continue to behave in a reasonably ideal manner) by computing time series of the large-scale stability and energy in the synoptic transients. Since the model is so idealized, we do not feel that there is any significance in the parametric dependence of the types of transition mechanisms, or for that matter, their statistics. The important aspect is the existence of the two different mechanisms that can operate.

In the real atmosphere there are a myriad of other possibilities to consider, including time-dependent variations of the boundary conditions, such as the annual cycle and sea surface temperature anomalies. Any one of these external variations could change the characteristics of the internal weather regimes such that the *current state is no longer compatible with the new boundary conditions.* The sensitivity of the regime *statistics* (regime frequency distribution, phase, amplitude, structure, etc.) in the model of RP to changes in the external parameters certainly suggest the potential importance of the time-varying boundary conditions for any forecast in the atmosphere. This type of sensitivity further implies that regime transition, even if caused by a slowly changing external forcing, could be catastrophically rapid, as in the regime-equilibrium instability considered above.

6. Conclusions and discussions

In the weather regime model of RP we find two completely distinct mechanisms that lead to weather regime collapse. Given the limited dynamics of the model, these two mechanisms probably account for the basic framework of all the model transitions. They can operate together, or independently, or in any random combination. The first is a *synoptically triggered large-scale instability.* Here the primary role of the transients is to maintain an environment in which the large-scale disturbance remains stable to *large-scale* perturbations. If, for example, the transients do not succeed in reducing the zonal shear sufficiently, or do not sufficiently confine the vertical structure of the large-scale wave within certain limits, the large-scale wave becomes unstable to perturbations of itself. The instability is energetically baroclinic and thus, occurs on that time scale and accounts for the rapidity of the transitions in these cases. Large-scale orographic instability has been hypothesized as a possible means for the onset and breakdown of Dole's (1982) persistent anomalies by Cessi and Speranza (1985) in the atmosphere; while in the RP model, Mukugawa (1987) claims it is the only mechanism that is sufficiently energetic to drive transition.

However, there is a second mechanism in the RP model that involves a *regime-equilibrium instability.* This process is highly nonlinear as it involves the equilibrated energy levels of the transients. If the energy of the transients becomes too high or too low, the regime equilibrium is unable to be obtained. The current large-scale flow is incompatible with the current transients and the regime collapses. Though one would anticipate that this transition mechanism could be slow, it does not have to be. If the characteristics of the large-scale component of the regime turn out to be rather insensitive to the synoptic-scale amplitudes right up to the point where the balance is no longer possible, the collapse in the large scale can occur suddenly. On the other hand, the regime transition is marked by a precursor in idealized cases: a gradually intensifying series of synoptic disturbances, though it may be difficult in real time to identify precisely the disturbance that is going to cause the transition without previous experience.

These two transition mechanisms allow us to view the role of the transients in weather regimes in a somewhat different light. The first and most important is so trivial (in some sense) that it is generally overlooked. This role is *to reduce the overall baroclinicity of the flow such that the large-scale disturbances are stable to large-scale perturbations.* Such a requirement implies that the bulk of the eddy statistics in a classic budget analysis should be acting against the time-mean state, which is what most analyses tend to show for both the traditional stationary waves (Lau 1979) and the persistent anomalies (Dole 1982), as well as the weather regimes in the RP model. That the barotropic component of the eddies may act with the mean flow (Holopainen and Fortelius 1988) may be the zonally inhomogeneous analog to classical negative viscosity, which is a secondary phenomenon driven by baroclinic instability, and in that sense these feedbacks are not responsible for the maintenance of the time-mean flow. The second role of the transients, which is much more involved, *is the regime-equilibrium process* in which the transients and the large-scale flow mutually influence and organize each other. Failure of the transients to accomplish the first goal leads to the synoptically induced large-scale instability, while the failure of the transients to accomplish the second leads to a regime-equilibrium instability.

In the atmosphere, Dole's (1986, 1988) analyses of the time evolution of persistent anomalies, in particular the rates of anomaly onset and collapse and the large-scale baroclinic structure at these times, are at least consistent with the above hypotheses, though they by no means confirm them. We suggest that the appearance of the synoptically triggered large-scale instability will manifest itself as a sudden change in the character of the planetary-scale flow, such as the rapid propagation and decay of a previously persistent ridge, with little or no precursor. The most obvious cause of a regime-equilibrium instability, on the other hand, is explosive cyclogenesis. In either case, it is evident that the detailed evolution of the individual synoptic events is critical for the prediction of weather-regime transition, and their capricious nature will probably be the limiting factor for weather forecasts. On the other hand, a better understanding of the threshold tolerances of regime balances would allow forecasters to identify those situations in which impending changes are more likely.

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