THEORY OF LARGE-SCALE STATIONARY WAVES

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1. OBSERVATIONS

In general, stationary waves are defined with respect to an available data set and with respect to a season or a month. Nowadays typical data sets span 10-20 years of observations and January appears to be a favorite month for model-data intercomparison. The "observed" January stationary wave pattern is arrived at by averaging over all days in January on record. In Fig. 1 we show the stationary 300-mb height field in January as obtained from the DWD-data set for the years 1967-1981. By and large the pattern is fairly smooth with just two pronounced troughs. Typical amplitudes of the stationary waves are of the order 150 m or so.

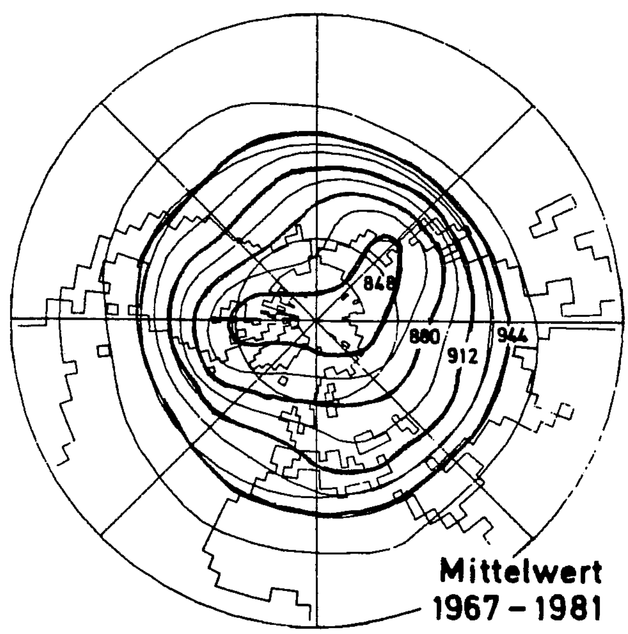

Fig. 1 January mean 300 hPa height (gdm) for the years 1967-1981. Analyses of the German Weather Service. Adapted from Fischer and Storch (1982).

As is well known there is considerable interannual variability of the average mean flow patterns for individual Januaries. Fischer and Storch (1982) pointed out that the stationary pattern shown in Fig. 1 is not typical of individual months. In Fig. 2 we show January mean patterns for years where the deviation from the long-term mean was particularly large. In all these years the flow pattern over Central Europe deviated considerably from the long-term mean field and there are remarkable shifts of the main troughs. January 1983 provided another example of a large deviation (Storch, 1984). On the other hand, the main troughs never disappear so that there is always a certain similarity of individual fields to the long-term mean.

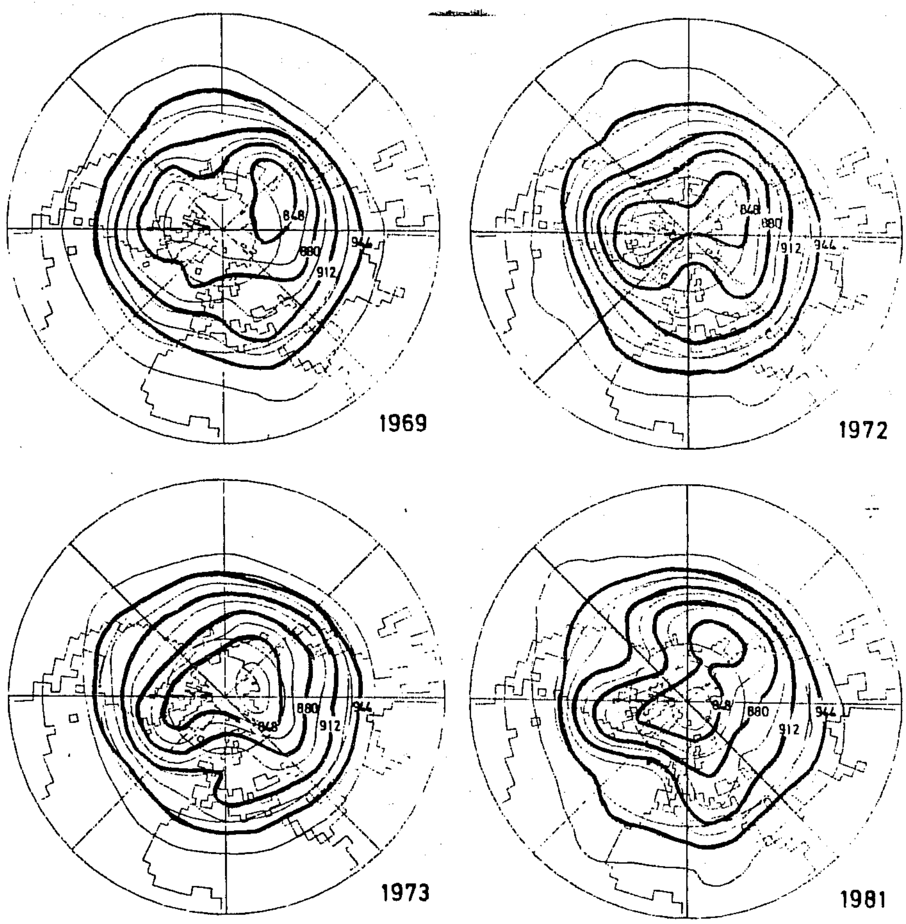

Fig. 2 January mean 300 hPa height (gdm) for years where deviations from the long-term mean in Fig. 1 were large. Analyses of the German Weather Service. Adapted from Fischer and Storch (1982).

In Fig. 3 we show the standard deviation of 30-day mean data in winter. It is seen that the standard deviation over the oceans is almost as large as the amplitude of the standing waves in Fig. 1.

All the theories of stationary waves which have been proposed so far aim at an explanation of the long-term mean pattern as depicted in Fig. 1. This is somewhat surprising since these mean fields are hardly ever observed in an individual month (Fischer and Storch, 1982). They are to some extent artifacts of the averaging procedure. One may wonder if such mean patterns can be explained by theories which neglect this aspect. What one would like to have is a theory of quasistationary waves which is capable of explaining also the interannual variability. Thus it is not only Fig. 1 which needs to be explained but also Fig. 3.

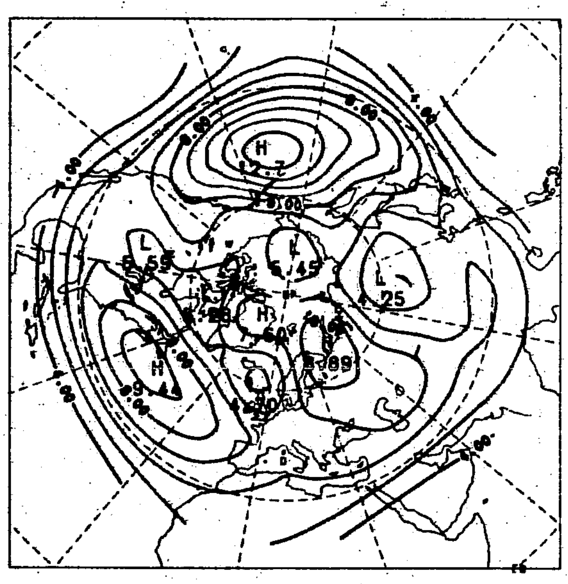

Fig. 3 Standard deviation of the 500 hPa 30-day mean stream function in winter. The map is based on 18 winter months. Analyses of ECMWF. After Metz (1986c). Planetary modes only. $10^6 m^2 s^{-1}$

This lecture is organized as follows. First (Section 2) we discuss the general problems to be solved by a theory of quasi-stationary waves. Next we describe the conventional approach (Section 3) and discuss related problems and results. Extensions of the conventional approach towards a more general theory of quasi-stationary waves are described in the following section. To restrict the scope of the paper we shall consider only quasi-stationary extra-tropical motions.

2. Theoretical approach

For illustrative purposes we assume that the atmospheric large-scale circulation is governed by a quasi-geostrophic potential vorticity equation

$$\left(\frac{\partial}{\partial t} + \underset{\sim}{v}\cdot\underset{\sim}{\nabla}\right) q = -d q + F \qquad (2.1)$$

where q is a potential vorticity, ψ the corresponding streamfunction, so that $\underset{\sim}{v} = \underset{\sim}{k} \times \underset{\sim}{\nabla}\psi$, d a damping constant and F some kind of forcing. Since we want to study slow motions it is appropriate to perform a Fourier transform in time

$$\hat{q}(x,y,z,\omega) = (2\pi)^{-1/2} \int_{-\infty}^{\infty} q(x,y,z,t)\exp(-i\omega t)\,dt \qquad (2.2)$$

Then (2.1) reads

$$i\omega\hat{q} + \int_{-\infty}^{+\infty} \hat{\underset{\sim}{v}}(x,y,z,\omega-\omega')\cdot\underset{\sim}{\nabla}\hat{q}(x,y,z,\omega+\omega')\,d\omega' = -d\hat{q} + \hat{F} \qquad (2.3)$$

Motions with frequencies $|\omega| < \omega_s$ will be called slow or quasi-stationary. In particular, the interannual variability of monthly means is slow if $\omega_s \gtrsim 2\pi/30$ days^{-1}. Motions with $|\omega| > \omega_s$ will be called transient. In what follows we shall consider slow motions only.

It is obvious from (2.3) that motions of all time-scales contribute to the dynamics of slow motions. The internal dynamics of slow motions contribute the term

$$I_1 = \int\limits_{\substack{|\omega-\omega'|<\omega_s \\ |\omega+\omega'|<\omega_s}} \hat{v}\cdot\underset{\sim}{\nabla}\hat{q}\,d\omega' \qquad (2.4)$$

to the interaction integral in (2.3). The interaction of slow and transient motions is described by

$$I_2 = \int\limits_{\substack{|\omega-\omega'|<\omega_s,\ |\omega+\omega'|>\omega_s \\ \text{or} \\ |\omega-\omega'|>\omega_s,\ |\omega+\omega'|<\omega_s}} \hat{v}\cdot\nabla\hat{q}\,d\omega' \qquad (2.5)$$

and, finally, there is the interaction of transient motions

$$I_3 = \int_{\substack{|\omega - \omega'| > \omega_s \\ |\omega + \omega'| > \omega_s}} \hat{v} \cdot \nabla \hat{q} \, d\omega' \tag{2.6}$$

Thus

$$i\omega \hat{q} + I_1 + I_2 + I_3 = -\alpha \hat{q} + \hat{F} \tag{2.7}$$

for $|\omega| < \omega_s$. A comprehensive theory of slow motions would have to come up with a solution to (2.3) for a given forcing. Such a theory does not exist, of course. In what follows we try to relate work on quasi-stationary waves to this general frame work.

3. Standard approach

There is a highly developed branch of the theory of slow motions where certain approximations are traditionally accepted. Since excellent reviews have been written on this topic (e.g. Held, 1983) it may be sufficient to list the main assumptions underlying this approach and to briefly discuss the main problems and results. Main assumptions are

i) It is sufficient to consider motions with $\omega = 0$.
ii) The interaction term I_3 is neglected as are wave-wave interactions in I_1.
iii) The basic equation (2.3) is linearized with respect to a zonal mean state $\bar{q}$.

The corresponding equation reads

$$\bar{U} \frac{\partial \hat{q}'}{\partial x} + \hat{v}' \frac{\partial \bar{q}}{\partial y} = -\alpha \hat{q}' + F' \tag{3.1}$$

where $\bar{U}$ is the zonal mean flow linked to $\bar{q}$. The primes denote perturbations. A solution of (3.1) can be achieved by standard methods. In particular, the dependence on x is separated out:

$$\hat{q}'(x,y,z,0) = \sum_m q_m(y,z) \exp(i k_m x) \tag{3.2}$$

where k_m is a zonal wave number.

To solve (3.2) one has to specify the boundary conditions. The orography enters at the lower boundary. In almost all theories worked out so far one relies on a linear lower boundary condition where the vertical perturbation velocity w' at the ground is assumed

$$w' = \overline{u}(y,0)\frac{\partial h}{\delta x} \qquad (3.3)$$

and h is orography. It is well know that (3.3) is highly problematic.

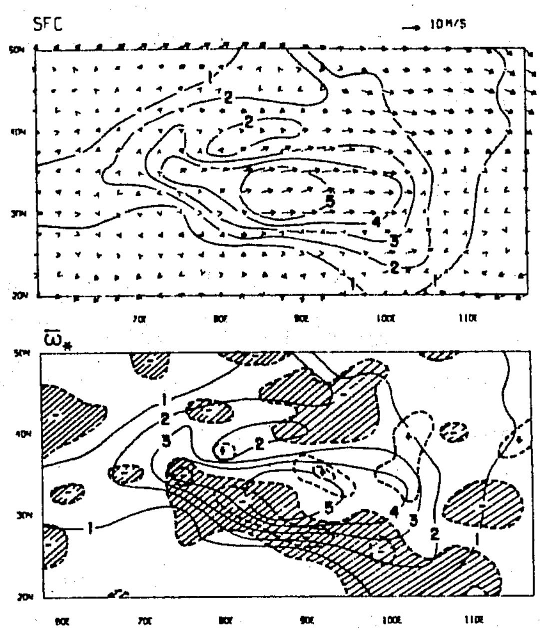

Fig. 4 Top: Winter mean surface winds (arrows) over smoothed Himalayan topography. Bottom: Winter mean pressure coordinate vertical motion in 10^{-3} mbs^{-1}. Hatching denotes areas of ascending motion. After Murakami (1981).

Fig. 4 gives the winter mean surface winds in the Himalayas as derived from observations (Murakami, 1981). Although even this pattern is highly idealized it is immediately obvious that the surface winds are not zonal as is assumed in (3.3). The "observed" vertical velocity field differs strongly from what is given by (3.3).

At the upper boundary one would prefer to have a radiation condition in order to avoid the spurious reflection of energy caused by the rigid lid condition used by many authors (see Chen and Trenberth, 1985, for a review). Another solution to this problem is to place the upper boundary at large heights. If some damping mechanism is built into the theory one can eliminate the spurious reflection this way (e.g. Jacqmin and Lindzen, 1985).

The specification of the forcing is also quite problematic. One would like to include heating due to fluxes of sensible heat at the ground, to release of latent heat and due to radiation. The corresponding distribution of heat sources is not well known. As has been pointed out by Webster (1981) the forcing is not independent of the stationary motion itself. It should be noted, however, that all the problems mentioned so far must be faced also by more general theories which do not rely on the specific assumption i)-iii). There is, however, a set of problems which is clearly linked to these assumptions. First of all, one does not really know to which observations the results ought to be compared. As has been pointed out above, the long-term average as obtained from the observations is atypical and cannot readily be identified with the solution of (3.1). Second the mean distribution of the potential vorticity may be unstable to small perturbations. This instability is suppressed since we require $\omega = \omega_s = 0$. One may argue that it is these instabilities which support through their transports of heat and momentum the mean state $\bar{q}$. Moreover, these instabilities are linked to transient phenomena. It is, however, questionable, if we can disregard this instability altogether. Next, there is considerable evidence that the interaction term I_3 and wave-wave interactions in I_1 are not negligible. We shall come back to this point in the next section. In a model equation like (3.1) there is exist critical lines. These are the lines $\bar{U}(y,z) = 0$. Reflections may occur at critical lines as well as absorptions. Moreover, nonlinear effects appear to be important. It is generally agreed that critical lines are of fundamental importance to the solution of (3.1) (e.g. Held, 1983). It is, however, doubtful how one ought to identify critical lines in the atmosphere. After all, the atmosphere is a turbulent medium with constantly changing winds. Planetary waves may experience the gradual shift from a westwind regime to a eastwind regime on the globe in a manner which may differ fundamentally from what is assumed in the steady-state equation (3.1).

Given all these problems and ambiguities one might think that the standard approach is fairly hopeless. This is, however, not correct. Quite to the contrary work based on the standard approach has produced results which bear relatively good similarity to the observations.

As an example we show results published quite recently by Jacqmin and Lindzen (1985). These authors solved the linearized primitive equations on the sphere for motions with $\omega = 0$ using a fairly good resolution in the latitude height grid (1° latitude x 1 km height). In Fig. 5 we show the perturbation height field for zonal wave number one as computed and as observed.

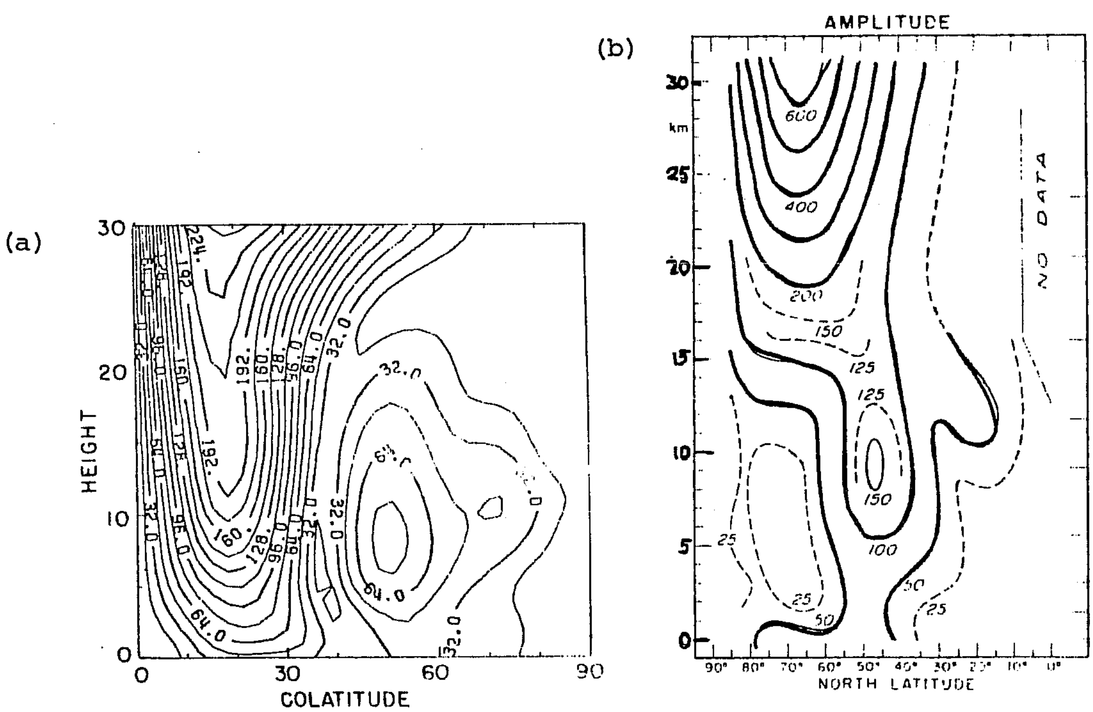

Fig. 5 Amplitude of the zonal wavenumber one height field (m) in January (a) as calculated by Jacqmin and Lindzen (1985) and (b) as observed (v. Loon et al. 1973). Northern hemisphere.

In the troposphere the model predicts maximum response at 70° N whereas the observed maximum is at 45° N. Calculated amplitudes are of the right order of magnitude but the tropospheric response south of 30° N and to the north of 60° N is overestimated. The calculations for wavenumbers two and three are even better. In Fig. 6 we show the calculated perturbation height field at 10 km to be compared to Fig. 1.

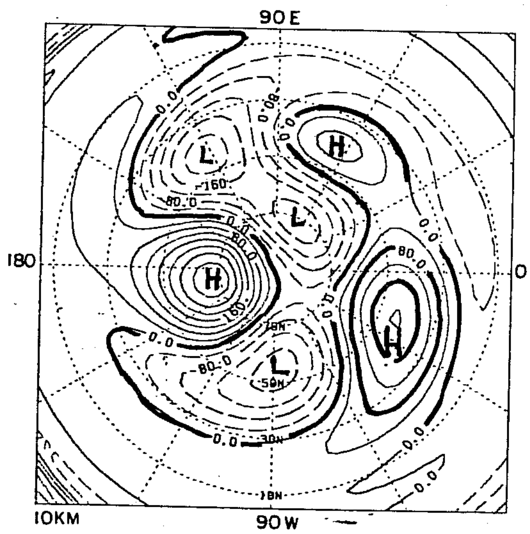

Fig. 6 Calculated January perturbation height field (m) at 10 km after Jacqmin and Lindzen (1985).

The model's primary deficiency is its intense high over the Ateutians. However, the main features of the observed distribution are captured by the linear model. The model output may be even more close to the January mean field in 1969 (Fig. 2) than to the longterm average. The overall satisfactory agreement is difficult to understand in view of the arguments presented above. It must be kept in mind, however, that the model is tested against one case only. Moreover the construction of the zonal basic state is in part by trial and error (Jacqmin and Lindzen, 1985).

4. Extensions to the standard theory

What one would like to have is a probabilistic theory of slow motions which predicts the probability to find, say, a certain January mean pattern. It may be possible to write down the corresponding equations for the probability distribution. It is, however, unlikely that these equations can be solved. Moreover, much of the information needed to specify these equations is lacking. However, there have been several attempts to relax some of the main assumptions underlying the standard theory. We shall try to briefly review some of this work.

a) "Forcing" by transients; $\omega_s = 0$.

It is difficult to include the interaction of transients with the stationary field in a theory of slow motions. Though West and Lindenberg (1984) presented an interesting analysis of the problem there is little hope that the term I_3 in (2.7) can be "parameterized" in terms of the quasi-stationary fields (see also Barnett and Roads, 1986).

Holopainen (1978), Lau (1979) and others evaluated the term I_3 using data. It appears that I_3 cannot be prescribed as a forcing term in (2.3). Holopainen (1983) computed the enstrophy budget of the stationary waves and showed that the transients remove potential enstrophy from the stationary wave field. Of course, a "forcing term" should not act as a sink of enstrophy. However, one need not reject the idea of a forcing by transients althogether. Most of the energy of the stationary waves resides in the planetary scales. At such large scales the barotropic part of the motion contributes little to the enstrophy. It is the temperature, i.e. the baroclinic part of the motion, which dominates potential enstrophy. Thus Holopainen's result means that the baroclinic part of the stationary waves is not forced by the transient eddies. This has been confirmed by Müller (1986). On the other hand Egger and Schilling (1984) provided evidence that slow barotropic motions are indeed forced by transients. Metz (1986 a) computed the steady-state planetary-scale response of a barotropic atmosphere to the forcing by synoptic-scale transients. The result is shown in Fig. 7 and may be compared to the observed barotropic stationary wave field. The amplitude of the forced waves is about the same as that of the observed ones and most of the observed features are seen in the computed fields as well. The model's primary deficiency is its intense low over Europe. This agreement is difficult to understand in view of the fact that mountain forcing and heating are neglected. It must be kept in mind, however, that the model is tested against one case only. Similar computations have been made by Youngblut and Sasamori (1980) and Opsteegh and Vernekar (1982) but these authors did not separate barotropic and baroclinic effects.

(a)

(b)

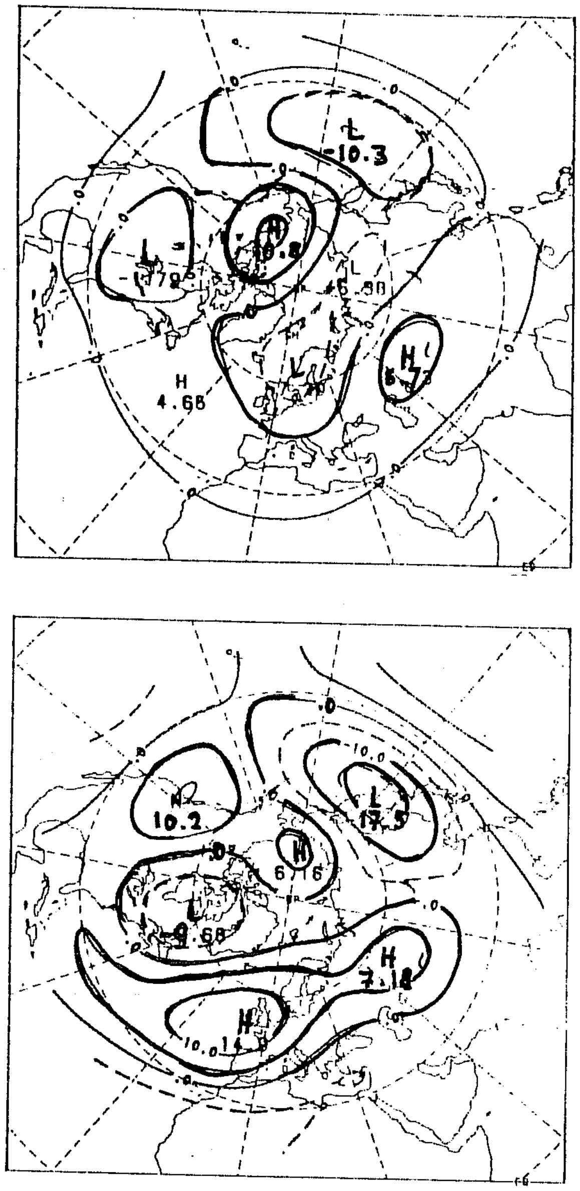

Fig. 7 Planetary scale stationary wave field in winter (a) as obtained from integrations of the linearized barotropic vorticity equation with observed forcing by synoptic-scale transients and (b) as observed (vertical mean). After Metz (1986 a). Stream-function in $10^6 m^2 s^{-1}$.

We conclude that the evidence as to the detailed effect of the transient motions on the long-term mean stationary fields is conflicting. There is, however, little doubt that the transients have a strong impact on the structure of the standing waves (see also Hendon, 1986, where this conclusion is supported by numerical experiments). There are indications that the transients may destabilise the stationary waves (Held et al. 1985).

b) Forcing by transients; $\omega_s > 0$.

Egger and Schilling (1983, 1984) and Metz (1986 b) suggested that a large fraction of the observed slow variability can be understood as a response of slow barotrope planetary modes to the forcing by transients. They solved the barotropic version of (2.3) whereby I_3 and I_2 where prescribed according to observations. Similar techniques have been used by Metz (1986 c) to study the variability of monthly means. In Fig. 8 we show the computed standard deviation of the barotropic monthly mean stream functions for the winter season. Except for the predicted center of variability at 120° W, 75° N the agreement of the theoretical result with the observations is quite good. In particular, the location of the maxima is simulated quite well. Obviously, the forcing by transient motions has a pronounced impact on the barotropic monthly mean flows. If similar techniques are used to study the forcing of the baroclinic part of slow motions by transient eddies the result is negative. The forced baroclinic motions are relatively weak and have little in common with the observed baroclinic flow component (Müller, 1986). As has been mentioned, this failure is due to the fact that there is a transfer of enstrophy from the slow planetary-scale baroclinic modes to the transients which cannot be modelled as a forcing of the planetary scales. It is the reverse for the barotropic modes (Egger and Schilling, 1984).

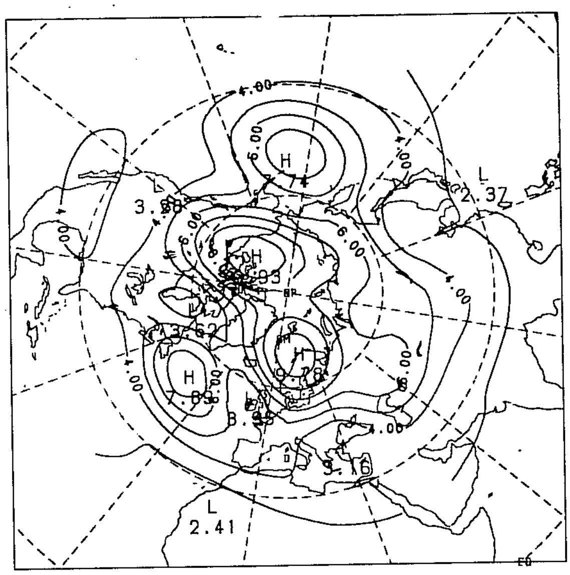

Fig. 8 Standard deviations of the barotropic stream function of monthly means as obtained from a linear barotropic vorticity equation where planetary modes are forced by synoptic-scale transients (in $10^6 m^2 s^{-2}$). After Metz (1986 c).

Kok and Opsteegh (1985) looked at an individual event in order to find out about the relative importance of mountain forcing, transient forcing and diabatic heating. They considered the 1982-1983 El Nino and solved the steady state form of (2.3) for six consecutive seasons. Again I_2 and I_3 were prescribed according to observations. In Fig. 9 we show the standard deviation of observed and simulated anomalies in the zonally asymmetric part of the seasonal mean zonal winds for each of the six seasons. It is clearly the response to the transient forcing which is most important. Mountains appear to have played a minor role during that event.

All this demonstrates that the mean pattern of individual month or seasons cannot be understood without taking the impact of transient motions into consideration.

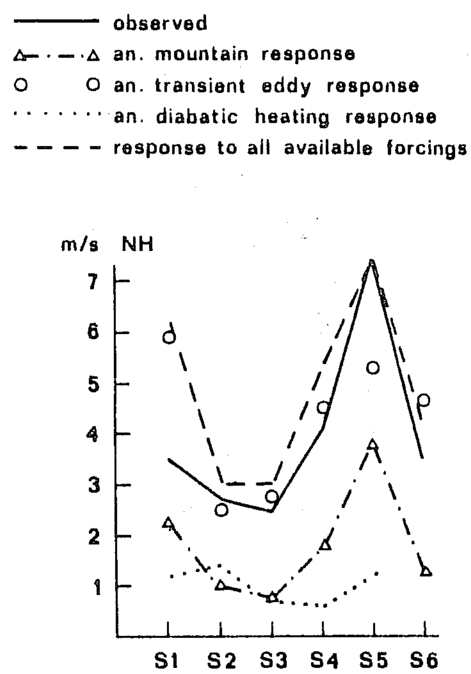

Fig. 9 Root mean square (rms) of observed and simulated anomalies in the zonally asymmetric part (m=1 to 6) of the seasonal mean zonal winds (ms^{-1}) for each of the six seasons of the 1982-1983 El Nino. After Kok and Opsteegh, 1985.

c) Transient mountain forcing; $\omega_s > 0$.

In the context of the linear theory transient mountain forcing can arise through a time variability of the zonal wind $\bar{U}$ in (3.3). The corresponding variations of w provide time dependent sources and sinks of vorticity at the mountain slopes. The power spectrum of the zonal wind is almost red (Schilling, 1984). Therefore most of the response to the variations of $\bar{U}$ will reside in the domain of slow motions with $\omega \lesssim \omega_s$. Monthly mean fields must be influenced by this mountain forcing. A first attempt to assess this effect has been made by Egger (1984). The linear barotropic ß-plane equation has been solved with transient mountain forcing where the variability of $\bar{U}$ has been chosen according to observations. It has been found that the Himalayas were the only obstacle where this kind of effect could be important. However, most of the computed variability has been found in the lee of the Himalayas, in an area where there is little observed variability on time scales of a month and more.

The computations have been repeated with atmospheric data instead of artificial data as in Egger (1984). First the zonally averaged flow $\overline{U}(y,t)$ is prescribed in the linearized barotropic vorticity equation

$$\frac{\partial}{\partial t}(\nabla^2 - \Lambda^{-2})\psi' + \overline{U}\frac{\partial}{\partial x}\nabla^2\psi' + J(\psi', f_0 h/H + f) + \overline{U} f_0 \frac{\partial h}{\partial x} H^{-1} = A\nabla^4\psi' \qquad (4.1)$$

on a day-by-day basis according to analyses by ECMWF. For the sake of convenience (4.1) is written in Cartesian coordinates but the actual computations have been made on the sphere. In (4.1) ψ' is the perturbation stream function in this linear problem and Λ^{-1} is a radius of deformation. The vorticity equation (4.1) has been integrated throughout the winter 1980/1981 with $\overline{U}(y,t)$ as the observed zonal mean flow at 300 hPa. In Fig. 10(a) we show the response for February 1981 data to a mountain of similar shape and location as the Rocky Mountains. We obtain the familiar picture of wave trains emanating from the "Rockies" towards the Tropics and also towards higher latitudes. The variability of this pattern from month to month appears to be weak. The trough in the lee of the obstacle, for example, had a minimum of $-0.21 \times 10^8 m^2 s^{-1}$ for December 1979 and of -0.22×10^8 for January 1981. This corroborates the earlier finding that the variability of monthly means as produced by variations of orographic forcing linked to the zonal mean winds are small. We can go one step beyond (4.1). Let $\psi_0(x,y,t)$ be the observed 300-hPa stream function in the winter 1980/81 and ψ' the perturbation induced by the mountain. Then the perturbation ψ' is obtained from

$$\frac{\partial}{\partial t}(\nabla^2 - \Lambda^{-2})\psi' + J(\psi_0 + \psi', \nabla^2\psi' + f_0 h/H) + J(\psi', \nabla^2\psi_0 + f) = A\nabla^4\psi' \qquad (4.2)$$

Note that we have a nonlinear equation for ψ'. The mountain term $J(\psi_0, f_0 h/H)$ induces waves as in (4.1) but these waves now have to move in the observed flow field. This allows us to estimate the impact of transient two-dimensional motions on the evolution of slow orographically induced motion.

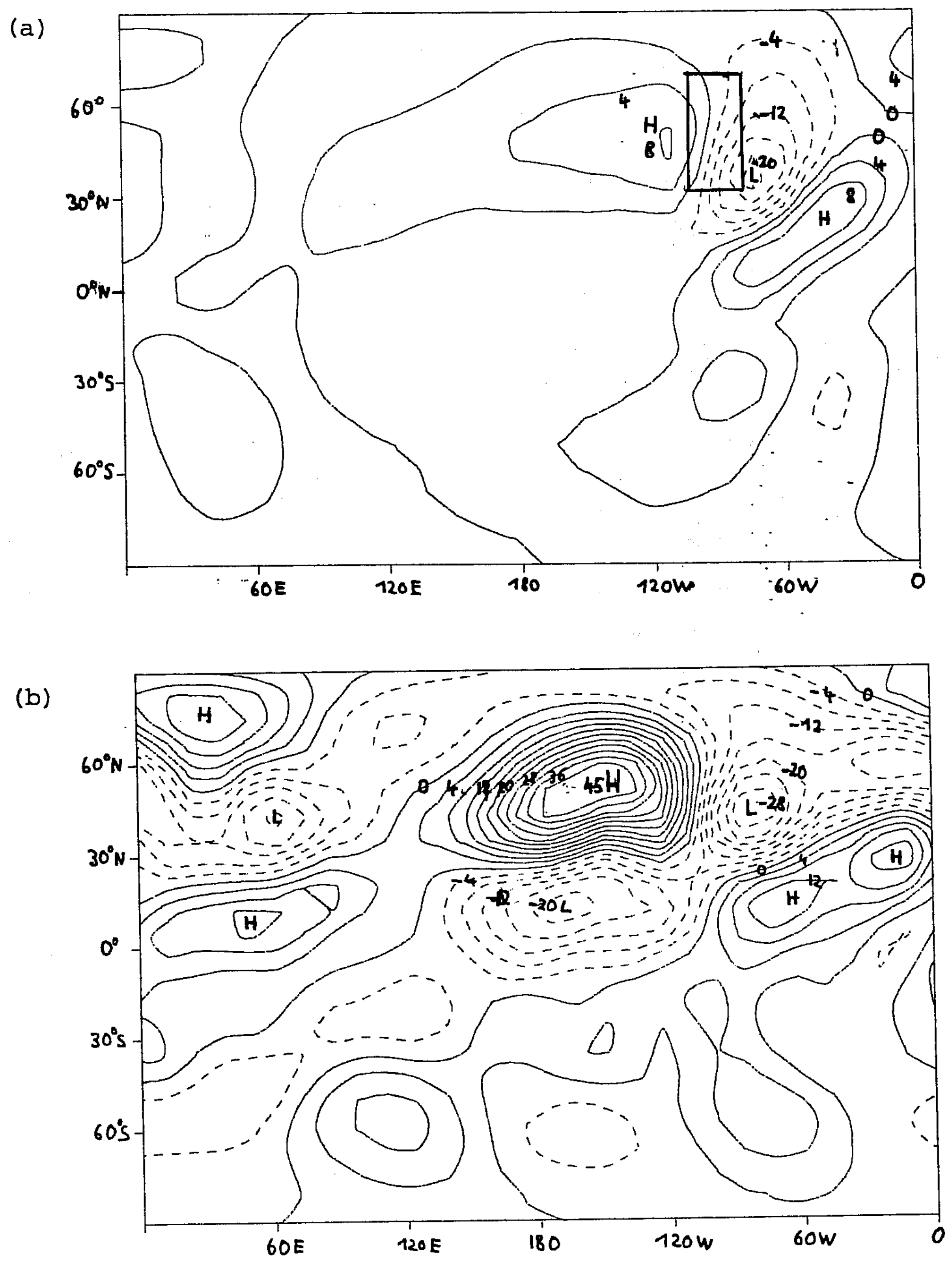

Fig. 10 Barotropic monthly mean stream function response to the "Rockies" (in $10^6 m^2 s^{-1}$) as obtained (a) from (4.1) for observed $\bar{U}$ and (b) from (4.2) for observed ψ_o. Data from January 1981, at 300 hPa; global domain of integration.

In Fig. 10(b) we show the mean stream function for January 1981 and for the same mountain as in Fig. 10(a) but we clearly now have an "instability". The monthly mean pattern of the response has enormous amplitudes. An inspection of the evolution of this pattern in time shows that this pattern has been growing from December to January. We have here another example of the instability of the observed 300 hPa flow detected first by Simmons et al. (1983). In such a situation there is little point in searching for the quasi-stationary response to a mountain. Moreover it is doubtful what conclusions to draw from Fig. 10(a). Is it really meaningful to study the wave response to a mountain in a zonally averaged mean flow when there is instability if we do not average the observations? On the other hand the relevance of this instability to atmospheric flows is not obvious.

5. CONCLUDING REMARKS

The theory of quasi-stationary waves has not yet reached a stage where there exist generally accepted explanations for most of the observed phenomena. Quite to the contrary more and more problems come up with ongoing research. For example, the interannual variability of the quasi-stationary waves and the impact of transients are aspects of the problem we are just beginning to explore.

Most of the results presented in this review are theoretical, i.e. they are derived by solving equations. It appears, however, that diagnostic methods became increasingly important in stationary wave research. For example, Plumb (1985) generalized the Eliassen-Palm flux to three dimensions and applied this concept to quasi-stationary three-dimensional patterns. His results appear to indicate that mountains play a relatively small role in estabilishing the quasi-stationary waves.

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