An Empirical Study on the Parameterization of Precipitation in a Model of the Time Mean Atmosphere

H. M. VAN DEN DOOL

Department of Meteorology, University of Maryland, College Park, MD 20742

(Manuscript received 14 April 1986, in final form 18 August 1986)

ABSTRACT

An empirical study based on three years (1981–83) of monthly mean data revealed that colocated anomalies in precipitation ($\widehat{PP}$) and vertical motion at 500 mb ($\hat{\omega}$) are moderately well correlated over the United States, especially in winter. The $\widehat{PP}$ data are spatial averages in 344 Climate Divisions while the $\hat{\omega}$ are derived from initialized fields of the ECMWF and NMC NWP models. To a first-order approximation the deficit of rain associated with anomalous downward motion is just as large as the surplus of rain associated with anomalous upward motion. Therefore, it should be possible to generate some of the latent heat of condensation in a linear model for the time mean atmosphere by expressing $\widehat{PP}$ linearly in $\hat{\omega}$. Monthly mean $\hat{\omega}$ of the ECMWF and NMC are highly correlated with each other and relate about equally well to rainfall over the United States. The empirical constant a in the relation $\widehat{PP} = a\hat{\omega}$ turns out to be about -0.6 mm day^{-1}/(10^{-2} N m^{-2} s^{-1}), which is on the same order of magnitude as the theoretical amount of precipitation produced by adiabatic ascent of magnitude 10^{-2} N m^{-2} s^{-1}. Attempts to empirically extract the role of atmospheric moisture in the relation between $\widehat{PP}$ and $\hat{\omega}$ were made by comparing summer to winter, higher latitudes to lower latitudes and the United States to India but the results are at best modest.

Implementation of parameterized latent heat sources and sinks in a linear steady state anomaly model for the time mean atmosphere is equivalent to reducing its static stability by a sizable amount. This leads to increased response to a prescribed forcing.

1. Introduction

In the last ten years many studies were published in which the atmospheric response to prescribed heat forcing was studied in the framework of linear steady state models (Opsteegh and Van den Dool, 1980, for just one example). In the real world the heating is not constant, of course, but interacts with and is determined by the flow itself. Therefore, it would be interesting to see whether the heating can be expressed (at least partially) in terms of model variables. One such attempt was made by Webster (1981) who assumed that in areas of upward motion latent heat of condensation is released. This procedure is based on the way we think the *instantaneous* atmosphere works, namely, precipitation is associated with rising motion. However, this creates a non-linearity because downward motion can not be associated with negative precipitation. Non-linearity cannot be handled in a linear model and Webster had to resort to an iterative technique to find steady state solutions. But there is another possibility. One can assume that *anomalously* upward (downward) *time mean* motion is associated with more (less) precipitation than normal. The aim of the present paper is to check whether the latter assumption has any validity. This is done by performing an empirical study with monthly mean precipitation and vertical velocity data over the United States for the years 1981–1983. As we shall see, the outcome of the empirical study implies that some fraction of the release of latent heat can linearly be expressed in model variables. More specifically this means that, indeed, an anomalously positive (negative) latent heat source can be associated with anomalously upward (downward) motion.

Comparing precipitation and vertical motion at any time and space scale is interesting for its own sake as it is a mutual consistency check on the reliability of two poorly known quantities. The underlying assumption of course is that vertical motion and precipitation *should* somehow be correlated to each other and if they are not in a given study, we tend to conclude that something is wrong with the estimates/measurements of one or both. There are numerous classical studies dealing with the empirical relationship between calculated vertical motion and weather (Miller and Panofsky, 1958; Curtis and Panofsky, 1958; Hansen and Thompson, 1965; to mention only a few). Most of these studies were concerned with synoptic scale events in the midlatitudes. A moderate agreement between synoptic scale calculated vertical motion and station precipitation was indeed found. Bannon (1948) followed the opposite approach; he estimated vertical motion

Meteorological Society

from observed rainfall rates and concluded that in the midlatitudes the upward motion could be as strong as 0.1 to 0.2 m s^{-1}. After the introduction of the satellite the expected relationship between clouds (as observed from satellite) and vertical motion (as calculated) was confirmed for synoptic scale motion both in tropical (Saha et al., 1969) and midlatitude areas (Shenk, 1963; Hansen and Thompson, 1965). But again the correlation was at best fair. Major improvements can be expected soon, now that liquid water in clouds and rain over the ocean are being derived from satellite measurements (Prabhakara et al., 1985), and our knowledge of the divergent wind improves steadily.

The empirical relation between *time mean* upper air flow and precipitation has also been a popular topic, especially among researchers interested in 5, 30 or 90 day time-mean forecasts (Stidd, 1954; Klein, 1963; Weare and Hoeschele, 1983; Cayan and Roads, 1984; Englehart and Douglas, 1985; Harnack and Lanzante, 1985). This interest exists primarily because forecasts of rainfall are traditionally arrived at in a two-step-process. First the time mean circulation aloft is forecast and then the simultaneous time mean precipitation is inferred from the circulation. It suffices here to say that 10%–60% of the variance of precipitation was reported to be explained from simultaneous time mean height (wind or vorticity) fields.

The vertical motion used here is taken from (twice) daily post-initialized initial states of operational numerical weather prediction models. Since the introduction of nonlinear normal mode initialization (in the late 1970s/early 1980s) one may perhaps trust the vertical motion present in the initial state. Lau (1979) used 6 hour forecasts of ω as proxies for observations, but 6 hours may not be enough to achieve such a mass/windfall balance. We used data only from 1981 onward. Of course the adiabatic nonlinear normal mode initialization may reduce some of the physically relevant vertical motion (Rosen and Salstein, 1985), especially in low latitudes in areas of large diabatic heating. This problem could be of some importance over the United States as well, especially since we are interested in the relationship between vertical motion and latent heating.

We will compare precipitation over the United States with vertical motion obtained from both the European Centre for Medium Range Weather Forecasts (EC) in Reading, England and from the National Meteorological Center (NM), Washington, DC. In September 1982 EC introduced diabatic heating in the initialization process and it will be interesting to see whether that had any impact on the relationship between vertical motion and precipitation.

The data, analysis and a discussion concerning climatology will be presented in section 2. Results of the empirical study pertaining to anomalies are given in section 3. In section 4 we compare the empirical findings with theory regarding adiabatic ascent in the atmosphere. This includes a number of calculations in which the explicit role of moisture in the atmosphere is empirically considered. In section 5 we discuss how the latent heat source can be expressed in terms of vertical motion in a multilayer linear steady state model. A comparison of NM and EC vertical motion is presented in section 6. Finally, the conclusions and discussion can be found in section 7.

2. Data, analysis and climatologies

All data used in this study are monthly means over the period from January 1981 through December 1983 for an area equal to or slightly larger than the 48 contiguous United States. We use three datasets

(i) Spatially averaged precipitation (PP) for 344 Climate Divisions (CD) obtained from the Climate Analysis Center in Washington, DC. Within each CD there are many individual raingauges and the purpose of averaging is to reduce the noise inherent in all rainfall data. Some states have many divisions, others only a few. A map of the CDs can be found in Janowiak et al. (1986). In general, CDs tend to get larger towards the west.

(ii) Vertical motion (ω) and mixing ratio (q) at 850 and 500 mb, derived from daily (EC, 12Z, UTC time in two-digit hours) and twice daily (NM, 0 and 12Z) operational analyses (post initialized) on a $5° \times 5°$ (EC) and $2.5° \times 2.5°$ (NM) grid respectively. The ω is derived from horizontal motion analyses via mass continuity. The analysis and initialization scheme for both the NM and EC model are described in Hollingsworth et al. (1985). We will consider the initial states as *observations* even though the data are initialized. Model *forecasts* of ω are not considered here.

(iii) Rainfall in 29 Climate Divisions in India over the years 1981–83. A map of the CDs and relevant references can be found in Mooley et al. (1986).

Since ω and PP are not colocated we have to interpolate. The choice is either to interpolate the large-scale ω to the 344 centers of the CDs, or to interpolate the irregularly spaced small-scale PP data to a grid. The former is more easily done. Hence, we have 344 pairs of PP and ω data points in each month.

There is probably an incompatibility in spatial scales present in the PP, ω datasets. The ω fields contain mostly large scales (about 500 km and up). Therefore, all variance in PP contained in smaller spatial scales will reduce the correlation between PP and ω. This is not a major problem in the present context as we are trying to relate, via linear regression ($\widehat{PP} = a\hat{\omega}$), anomalies in PP and ω. Smoothing of PP by, for example, combining CDs increases the correlation but does not change a. Since, moreover, smoothing of PP data is not a trivial matter, we accepted the PP data as they are.

Even though the ω field is large scale, the smallest scales contained in the ω field may be noise. Applying a two-dimensional 1-2-1 filter three times on the gridded ω data before interpolation did indeed improve the correlation between PP and ω from -0.35 to -0.45 (yearly mean) the following results are based on smoothed ω fields.

In a given year j and month m we can express the similarity of anomalies in the fields of vertical motion ($\hat{\omega}(m, j)$) and precipitation ($\widehat{\mathrm{PP}}(m, j)$) by employing a pattern correlation coefficient

$$\mathrm{PC}(m,j) = \frac{\frac{1}{344}\sum_{c=1}^{344} \hat{\omega}(m,j,c)\widehat{\mathrm{PP}}(m,j,c) - [\hat{\omega}(m,j,c)][\widehat{\mathrm{PP}}(m,j,c)]}{\mathrm{sd}_{\omega}(m,j)\mathrm{sd}_{\mathrm{PP}}(m,j)} \quad (1)$$

where c is an index for the climate divisions, brackets represent a continental wide average, the caret represents the departure from a local, 3-year mean for month m, and the standard deviation (sd) in the denominator of (1) is given by

$$\mathrm{sd}_{\omega}(m,j) = \{\frac{1}{344}\sum_{c=1}^{344} (\hat{\omega}(m,j,c) - [\hat{\omega}(m,j,c)])^2\}^{1/2} \quad (2)$$

and similarly for $\mathrm{sd}_{\mathrm{PP}}$. In (1) all CDs are given equal weight, even though we realize that their sizes are different. When all terms in (1) are summed through over the three years a simpler pattern correlation emerges:

$$\mathrm{PC}^*(m) = \frac{\sum_{c=1}^{344}\sum_{j=1}^{3} \hat{\omega}(m,j,c)\widehat{\mathrm{PP}}(m,j,c)}{\{\sum_{c=1}^{344}\sum_{j=1}^{3} \hat{\omega}(m,j,c)^2 \sum_{c=1}^{344}\sum_{j=1}^{3} (\widehat{\mathrm{PP}}(m,j,c))^2\}^{1/2}} \quad (3)$$

which is only a function of the month m. In (3) all continental averages have vanished. Hereafter we will mostly use (3).

By summing over fewer than all 344 stations (3) may be used to study geographical variations. However, with only three years of data it is useless to present too much detail. So we will restrict ourselves to dividing the United States into four quadrants separated by the 40°N latitude circle and 95°W parallel.

Most of the discussion will be concerned with pattern correlations between precipitation and vertical motion *at 500 mb*. This level was found to give the best correlation although the difference with vertical motion at other tropospheric levels is negligible. At a preliminary stage of this study (only 22 months of data) we evaluated (1) with $\hat{\omega}$ at 500 mb replaced by 1) monthly mean height, 2) wind and vorticity fields at various levels, 3) variance (within the month) of vertical velocity and height fields at various levels, 4) monthly mean albedo and outgoing longwave radiation. Since the results of all these attempts were far inferior to using vertical motion we will not discuss them here. The only variant to be discussed in this paper is to involve the mixing ratio q. This is not done because we found improvement in the statistics but because it seems fundamentally correct to involve moisture availability in an expression for rainfall.

Before presenting the results concerning anomalies (in section 3) it is useful to display a few of the input data and climatologies (=3 year means) used in this study. Figure 1 shows (from top to bottom) the distribution of precipitation and vertical motion (EC middle, NM lower panel) over the United States in February 1982. In order to emphasize the large scale aspects we analyzed only the 50 mm per month line in Fig. 1a. The heavier (H, >50 mm) precipitation fell over the Southeastern and Northwestern United States while lighter (M, <50 mm) amounts fell over much of the

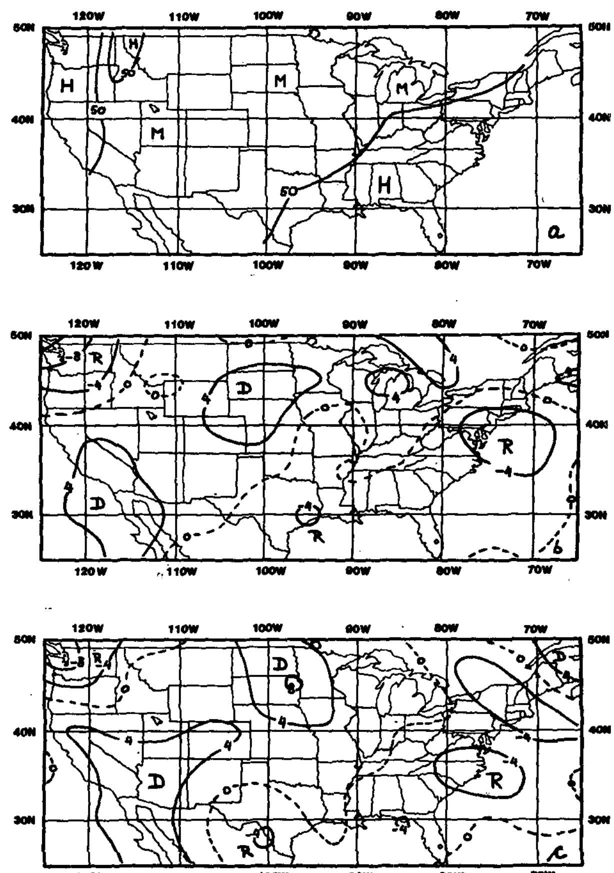

FIG. 1. A sample of the data used in this study. In the top panel (a) the precipitation collected in gauges over the United States during February 1982, averaged over Climate Divisions and analyzed for large-scale features only. In the lower two panels, the monthly mean vertical motion at 500 mb during February 1982 as derived from the initial state of the NWP models at ECMWF (b) and NMC (c). The units are mm month^{-1} and 10^{-2} N m^{-2} s^{-1}, respectively. The letters H and M stand for heavy and moderate precipitation, D and R for descending and rising motion, respectively.

interior. Within these large areas denoted by either M or H there is tremendous detail which has a lowering effect on the correlations to be discussed. There is a good deal of similarity between the PP pattern in Fig. 1a and the two ω-patterns in Figs. 1b and 1c, in the sense that most of the heavy precipitation fell in areas of rising motion (R), and the areas of moderate precipitation coincide with mostly descending motion (D). February 1982 serves as a typical example; it is not an exceptionally good case. After taking out the 3 year mean (for February) the PC between anomalies in PP and ω_{EC} (ω_{NM}) is -0.27 (-0.37), while the PC between $\hat{\omega}_{EC}$ and $\hat{\omega}_{NM}$ amounts to 0.66. It is encouraging that the two estimates of vertical motion used here are reasonably similar.

Figure 2 shows the climatology (1981–83) of seasonal mean vertical motion at 500 mb (based on EC data) over the United States. Each map is an average over 9 months and the seasons are defined as DJF, MAM, JJA and SON. The four maps have much in common and show (i) a dipole of rising (Northwest) and descending (Southwest) motion at about 120°W, (ii) broadscale descending motion east of the Northern Rockies extending northeastward over the Great Lakes, and (iii) a dipole consisting of rising motion over the Southeastern States and adjacent low latitude Atlantic Ocean, and descending motion over New England and adjacent high latitude Atlantic. The seasonality of this pattern is relatively small. The west–east oriented dipole of rising motion over the Northwest and sinking air over the North Central plains can be interpreted as being orographically induced. It is tempting however, to think of the north–south oriented dipoles at the west- and east coast as local meridional overturnings that act to decelerate and accelerate respectively the time mean flow at middle to high tropospheric levels.

Figure 2a (winter) is very similar to Lau's (1979) 11-year winter-mean vertical motion at 500 mb, especially over the western half of the United States. His data were based on 6 hour forecasts of various NM models used during 1965–76. Hence, there is some reason to believe that it is no longer necessary to use 6 hour forecasts of vertical motion as proxies for observations. It should also be pointed out that the four seasonal maps based on NM analyses (not shown) are very similar to the ones displayed in Fig. 2 (based on EC). A comparison of Fig. 2c (summer) with the 5 year summer mean vertical motion at 500 mb (based on 6 hr forecasts with the NMC model (1970–72; 1976–77)) as given by White (1983) is not quite as favorable. The major difference is that White finds very strong upward motion over Baja California, but as he points out this is not in agreement with the normal rainfall distribution in that area. In that respect Fig. 2c is more credible, and it may actually be that initial state vertical velocities during 1981–83 are better than the 6 hour forecasts used prior to 1980. A difference with both Lau and White is that the magnitude of our vertical motion seems to be smaller by a factor 1–3, which may be due to the initialization.

The monthly mean vertical motions used in this paper are derived from daily ω data rather than being derived from monthly mean data, the difference being primarily the explicit impact of transient eddies on $\bar{\omega}$. Using ECMWF's FGGE IIIb analyses for July 1979 data we calculated the vorticity advections

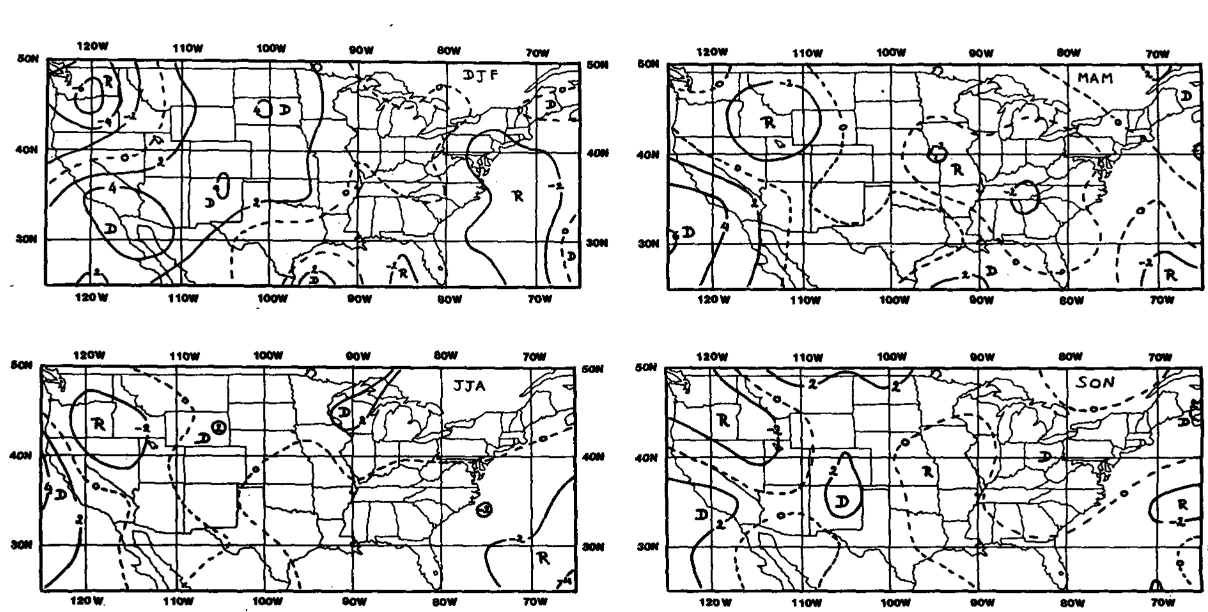

FIG. 2. A 3 year climatology (1981–83) of seasonal mean vertical velocity at 500 mb over the United States based on daily initial states of the ECMWF operational model. Winter is denoted as DJF, etc. The units are 10^{-2} N m^{-2} s^{-1}.

$$F_M = \bar{u}\frac{\partial \bar{\zeta}}{\partial x} + \bar{v}\frac{\partial \bar{\zeta}}{\partial y} + \beta \bar{v},$$

$$F_{\mathrm{TOT}} = \overline{u\frac{\partial \zeta}{\partial x}} + \overline{v\frac{\partial \zeta}{\partial y}} + \beta \bar{v},$$

where the bar stands for time averaging, and the other symbols have their usual meaning. Assuming $F = (\bar{\zeta} + f)\partial\bar{\omega}/\partial p$ we can now integrate downward to find $\bar{\omega}$ at 500 mb. The levels at which F_M and F_{TOT} were calculated from analysed wind data are 100, 200, 250, 300, 400 and 500 mb. Here $\bar{\omega}(0)$ was assumed to be 0. The results for $\bar{\omega}$ at 500 mb based on F_{TOT} and F_M are shown in Figs. 3a and 3b. As can be seen (from this one example) the impact of eddies is to change the magnitude of time mean vertical motion by 10%–30%, but to leave the pattern practically unchanged. We also note that Figs. 3a and 3b are very similar in pattern to the summer climatology (in Fig. 2), thereby giving credence to both methods of deriving $\bar{\omega}$. This conclusion was arrived at also by White (1983) who compared $\bar{\omega}$ based on 6 h forecasts by the NMC model to the vorticity equation method (F_{TOT} only). The most important point to be made here is that the transient eddy term in the vorticity balance has only a small (not negligible) impact on the derived vertical motion over the United States. This explains at least in part why Cayan and Roads (1984), Harnack and Lanzante (1985) and many others were able to use just time-mean heights in inferring precipitation from concurrent circulation. None of them used $\bar{\omega}$ directly, but combinations of Z, and geostrophically derived $\bar{u}$, $\bar{v}$ should produce a reliable proxy for $\bar{\omega}$. A more systematic approach to assessing the impact of transient eddies on $\hat{\omega}$ will be undertaken in the near future when data over 1984–86 become available.

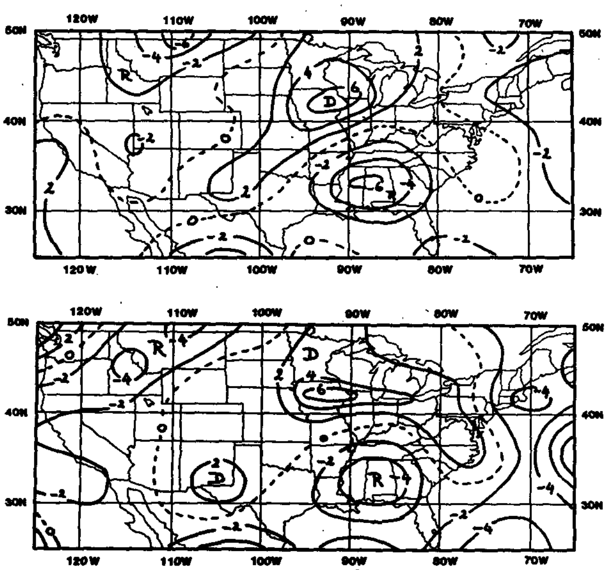

FIG. 3. The monthly mean vertical motion at 500 mb during July 1979 as derived from the vorticity budget. The top panel is based on monthly mean fields only while in the lower panel the effect of transient eddies is included. The unit is 10^{-2} N m^{-2} s^{-1}.

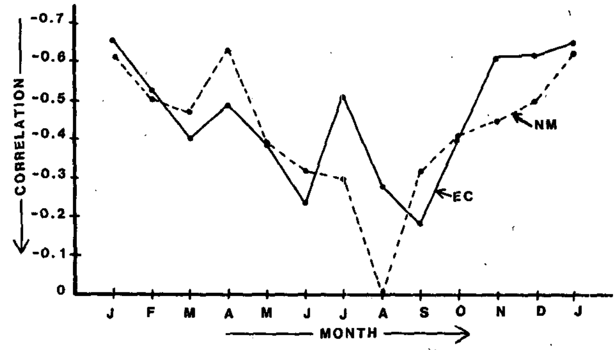

FIG. 4. The pattern correlation [see (3) for a definition] of anomalies in monthly mean precipitation with anomalies in monthly mean vertical velocity (500 mb) over the United States during 1981–83 as a function of the month. The curve labelled EC (NM) refers to correlation of precipitation with vertical motion of the European Center (National Meteorological Center).

3. Results

Figure 4 shows the PC* between the anomalous precipitation and vertical motion fields as a function of the month evaluated according to (3). Fortunately, the correlation is negative in all 12 months, in agreement with the expectation that anomalously upward motion ($\hat{\omega} < 0$) should be associated with more than normal precipitation ($\widehat{PP} > 0$) and vice versa. The correlation is better in winter (about −0.65) than in summer (about −0.20) which is probably due to the large amounts of precipitation associated with showers in summer. The occurrence of convective activity cannot always easily be linked to the large-scale dynamics for which $\hat{\omega}$ is representative. Regarding Fig. 4, we note that the EC and NM vertical motion correlate nearly equally well with precipitation over the United States. Figure 5 shows that, indeed, EC and NM monthly mean vertical motion anomalies are highly correlated, especially in winter (0.85). Whether this means that we know $\hat{\omega}$ fairly well or that ECMWF and NMC have the same biases in their analysis/initialization schemes, we cannot tell at this point.

a. Spatial degrees of freedom and statistical significance

In order to discuss the statistical significance of the foregoing results, we will ignore (just here) all seasonality and consider the 3-year mean correlation

$$\overline{\mathrm{PC}} = \frac{1}{36}\sum_{m=1}^{12}\sum_{j=1}^{3}\mathrm{PC}(m,j) \tag{4}$$

and the standard deviation

$$\mathrm{sd}_{\mathrm{PC}} = \left[\frac{1}{36}\sum_{m=1}^{12}\sum_{j=1}^{3}(\mathrm{PC}(m,j) - \overline{\mathrm{PC}})^2\right]^{1/2}. \tag{5}$$

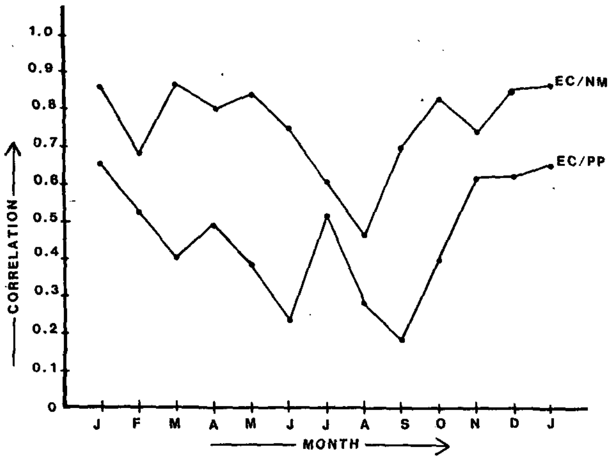

FIG. 5. The pattern correlation of anomalies in monthly mean vertical motion from EC with anomalies in monthly mean vertical motion from NM over the United States during 1981–83 as a function of the month. The curve EC/PP is repeated from Fig. 4 (sign reversed though) for comparison.

Similarly we define $\overline{PC^*}$ and sd^*_{PC} based on 12 cases.

For a normal distribution with zero mean the standard deviation of a linear correlation theoretically obeys (Panofsky and Brier, 1968)

$$sd = (N_e - 2)^{-1/2} \tag{6}$$

where N_e is the number of degrees of freedom. In order to assess statistical confidence limits of results such as those in Fig. 4, it is essential to know how large N_e (or sd) is. Besides, it is interesting for its own sake to know the effective sample size of the rainfall/vertical velocity data. The maximum possible N_e would be 344 (number of $\hat{\omega}/\widehat{PP}$ datapoints in space) but spatial dependence of the data will reduce this number considerably. In order to determine N_e we will resort to a poor man's Monte Carlo methodology.

In Table 1 we give $\overline{PC}$, $\overline{PC^*}$ and their associated N_e based on 36 (12) cases for a number of experiments where we have shifted the vertical motion data to the west, east, south and north. As expected the correlation is highest when the vertical motions are colocated with CDs (no shift, experiment 1). By shifting ω over 5° (10°) latitudinally (longitudinally), the correlation goes down considerably. The extreme shifts by 160, 180 and 200 degrees longitudinally serve here as Monte Carlo experiments. What we actually do is to correlate precipitation over the United States with vertical motion over South Central Asia. For all these cases $\overline{PC}$ or $\overline{PC^*}$ is very small and therefore the assumptions underlying (6) are best met in Experiments 8, 9 and 10 (see Table 1). Therefore, when using PC $N_e \approx 21$ seems to be our best estimate for the spatial degrees of freedom for the 344 data points of monthly mean precipitation/vertical motion. When using PC* $N_e \approx 57$ is appropriate.

Three final remarks about spatial degrees of freedom:

(i) $N_e = 21$ is a season independent number. In reality N_e is unquestionably a function of season but we do not have enough data (for each month) to address the seasonality.

(ii) $N_e = 21$ is the number of degrees of freedom for the *combined* precipitation and vertical motion anomaly field. From the 1 month lag *auto*correlations, see Table 2, we find N_e to be about 24 for precipitation, and 8 (EC) and 16 (NM) for vertical motion separately. As was expected precipitation has more spatial degrees of freedom than vertical motion. Spatial smoothing of rainfall data (or designing larger CDs) may decrease N_e and thereby increase the PC between PP and ω, but that is not the issue here.

(iii) Because PC* taken 3 times more data into account than PC it is not surprising that N_e associated with PC* is much higher than 21.

Since we had expected $\hat{\omega}$ and $\widehat{PP}$ to be negatively correlated we can assess the statistical significance of Fig. 4 by applying a one-sided test using $N_e \approx 57$ (sd = 0.135). Therefore all values of PC* < -0.22 are considered significantly different from zero at the 95% level (and negative), a criterion passed by all months in winter, spring and fall. For PC the criterion is more stringent: PC < -0.38.

b. Geographical distribution

With just 3 years of data one cannot expect to resolve all geographical variations in the relation between vertical motion and precipitation. Here we restrict our-

TABLE 1. Simultaneous correlations (%) between monthly mean precipitation and EC 500 mb vertical motion anomalies over the United States: 3 year mean (first line) and the spatial degrees of freedom (second line), evaluated according to Eq. (6). The shifts for experiment 1 to 10 denote by how much the vertical motion data were translated east- and northward relative to the rainfall data. PC and PC* refer to two different measures of correlation as defined by Eq. (1) and (3).

Experiment shift (°E, °N)	1 (0, 0)	2 (5, 0)	3 (10, 0)	4 (−5, 0)	5 (−10, 0)	6 (0, 5)	7 (0, −5)	8 (160, 0)	9 (180, 0)	10 (200, 0)	8, 9 and 10 combined
$\overline{PC}$	−38	−35	−19	−28	−17	−27	−28	−8	2	3	−1
N_e (PC)	34	35	31	32	28	18	36	23	20	20	21
$\overline{PC^*}$	−45	−43	−29	−35	−22	−36	−34	−9	3	2	−1
N_e (PC*)	48	49	50	54	54	27	40	63	28	63	57

TABLE 2. Yearly mean autocorrelation (%) of monthly mean precipitation and vertical motion anomalies over the United States at a lag of 1 month based on data over 1981–83. Second and third line give the standard deviation around the mean and the implied spatial degrees of freedom for each of the fields.

		Vertical motion	
	Precipitation	EC	NM
$\overline{PC}$	3	10	6
sd_{PC}	22	41	26
N_e	24	8	16

selves to considering four quadrants defined by the 40°N and 95°W latitude/longitude circles. Table 3 gives the yearly mean correlation between monthly vertical motion anomalies (EC) at 500 mb and precipitation. The correlations are poorest (although still negative!) over the southwestern quadrant where, probably, desert-like conditions (negative rainfall anomalies are rare) interfere with the desired linear relationship between precipitation and vertical motion.

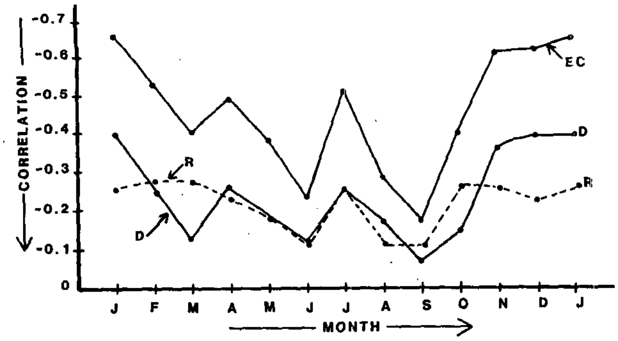

FIG. 6. The decomposition of the pattern correlation between anomalies in EC vertical motion and precipitation into contributions from areas and months with descending (D) and rising (R) motion. Area: the United States; years 1981–83.

c. *Linearity of the relationship*

A central issue here is whether the relationship between vertical motion and precipitation anomalies is linear. To that end we decompose (3) into

$$\mathrm{PC}^*(m) = \frac{\sum\sum \hat{\omega}^+(m,j,c)\widehat{\mathrm{PP}}(m,j,c) + \sum\sum \hat{\omega}^-(m,j,c)\widehat{\mathrm{PP}}(m,j,c)}{\text{denominator}} \quad (7)$$

that is, we sum separately over descending motion ($\hat{\omega}^+$) and rising motion ($\hat{\omega}^-$) cases while keeping the denominator as in (3). By doing so, we end up having two terms which, in themselves, are not a correlation, but their sum is PC*(m). Figure 6 shows how the correlation between $\widehat{\mathrm{PP}}$ and $\hat{\omega}$ is decomposed into contributions from areas of anomalously descending (D) and rising (R) motion. Since the D and R curves contribute roughly equal to the total (curve labeled EC) it appears that anomalies in precipitation and vertical motion over the United States are, to a first-order approximation, linearly related. In other words: the deficit in precipitation associated with anomalously downward motion is just as large as the surplus in precipitation associated with an equally strong anomalously upward motion. It is this linear anomaly property that enables us to apply the results of this empirical study to a model of the monthly mean atmosphere. A similar linear relation between precipitation and height anomalies was suggested for some areas by Klein and Bloom (1987).

TABLE 3. Yearly mean pattern correlation between vertical motion (EC, at 500 mb) and precipitation anomalies over the United States during 1981–83. The four quadrants are defined by the 40°N and 95°W axes and count 82 (Northeast), 86 (Northwest), 59 (Southwest) and 119 (Southeast) Climate Divisions, respectively.

	95°W		
	−47	−54	40°N
	−31	−42	

4. Comparing theory and observations

With only large scale ω data at hand one can expect to parameterize only that part of PP associated with large-scale vertical motion.[1] The parameterization of large scale precipitation goes back to Fulks (1935) and can be expressed concisely as (Haltiner and Williams, 1980)

$$\Delta\mathrm{PP} = -\frac{\delta F\omega}{g}\Delta p \quad (8)$$

where ΔPP is the amount of instantaneous precipitation generated in a second in a layer of thickness Δp as a result of adiabatic ascent or descent, g is acceleration of gravity, $\delta = 0$ when $\omega > 0$ or when q (mixing ratio) is less than the saturation mixing ratio (q_s) and $\delta = 1$ when both $\omega < 0$ and $q \geqslant q_s$. Finally, F is given by

$$F = \frac{q_s T}{p}\left(\frac{LR - c_p R_v T}{c_p R_v T^2 + q_s L^2}\right) \quad (9)$$

where p is pressure, R (R_v) is the gas constant for air (water vapor), c_p is specific heat at constant pressure

[1] By this we do not mean to say that convective precipitation is not at all related to vertical motion. However, the physical relationship between large scale vertical motion and convective precipitation is more indirect and not expected to be as simple as PP $\propto$ ω.

and L is the latent heat of vaporation. The F describes how efficient the adiabatic precipitation process will be.

For comparison with the empirical study we have tabulated in Table 4 some values of F and the associated rainfall according to (8) for some choice of ω. In these examples, the rain precipitated in a day by the 14 50-mb layers between 825 and 175 mb is tabulated for a cold, moderately warm and warm atmosphere. The assumptions made here are that the atmosphere is saturated from 825 to 175 mb, the temperature follows the wet adiabat, the 850 mb temperature is specified, and ω follows a parabolic profile with maximum upward motion of -10^{-2} N m^{-2} K^{-1} at 500 mb and zeros at 0 and 1000 mb. The most important features for our subsequent discussion are (i) vertical motion of -10^{-2} N m^{-2} s^{-1} at 500 mb is theoretically associated with a rainfall on the order of 1 mm day^{-1}, and (ii) at constant pressure, F (and ΔPP) increase with T, that is F increases with moisture availability. In other words: the same vertical motion will generate more precipitation in a warm wet atmosphere than in a cold dry atmosphere, at least as long as the assumptions underlying Table 4 are valid in different climates and seasons.

TABLE 4. Precipitation generated by a cold (upper), moderately warm (middle) and warm (lower panel) saturated atmosphere. The pressure at the center of the 14 50-mb-layers is denoted by p (mb), the temperature by T(K), the precipitation efficiency F (10^{-7} m^2/K), precipitation (mm day^{-1} or kg m^{-2} day^{-1}) from the layer (ΔPP) and accumulated precip ($\Sigma\ \Delta$PP). The precipitation amount is calculated for a parabolically sloped vertical motion profile with $\omega = -10^{-2}$ N m^{-2} s^{-1} at 500 mb and $\omega = 0$ at 0 and 1000 mb. The temperatures in bold at 850 mb are specified and define a cold (265), moderate (280) and warm (295) atmosphere.

p	T	F	ΔPP	$\Sigma\ \Delta$PP
		Cold		
200	184.0	0.00	0.000	0.000
250	195.1	0.01	0.000	0.000
300	204.8	0.03	0.001	0.002
350	213.4	0.07	0.003	0.004
400	221.2	0.13	0.005	0.010
450	228.2	0.20	0.009	0.019
500	234.5	0.30	0.013	0.032
550	240.2	0.41	0.018	0.050
600	245.4	0.52	0.022	0.072
650	250.1	0.63	0.025	0.097
700	254.3	0.73	0.027	0.124
750	258.2	0.82	0.027	0.151
800	261.7	0.90	0.025	0.177
850	**265.0**	0.96	0.021	0.198
		Moderate		
200	204.9	0.07	0.002	0.002
250	217.0	0.20	0.007	0.009
300	227.3	0.42	0.015	0.024
350	236.1	0.69	0.027	0.052
400	243.6	0.97	0.041	0.092
450	249.9	1.21	0.053	0.145
500	255.5	1.41	0.062	0.208
550	260.3	1.56	0.068	0.275
600	264.5	1.65	0.070	0.345
650	268.2	1.71	0.068	0.414
700	271.6	1.74	0.064	0.478
750	274.6	1.75	0.058	0.536
800	277.4	1.74	0.049	0.585
850	**280.0**	1.73	0.039	0.624
		Warm		
200	240.3	2.66	0.075	0.075
250	250.3	3.38	0.112	0.187
300	257.9	3.67	0.136	0.323
350	264.0	3.72	0.149	0.472
400	269.1	3.64	0.154	0.625
450	273.4	3.51	0.153	0.778
500	277.1	3.35	0.148	0.926
550	280.4	3.20	0.140	1.065
600	283.4	3.05	0.129	1.194
650	286.1	2.91	0.116	1.311
700	288.6	2.77	0.103	1.414
750	290.9	2.65	0.088	1.501
800	293.0	2.54	0.071	1.573
850	**295.0**	2.43	0.055	1.627

The precipitation generated by adiabatic ascent is linear in ω at any level. Therefore, Table 4 and the above discussion apply also to anomalies in vertical motion and precipitation, at least as long as $\omega < 0$. If the atmosphere were constant over a month, Table 4 would even apply to monthly mean anomalies.

In section 3 we discussed the relationship between $\widehat{\mathrm{PP}}$ and $\hat{\omega}$ in terms of a linear correlation. The correlation can easily be converted to a regression constant a

$$\widehat{\mathrm{PP}} = a\hat{\omega} \tag{10}$$

where $a = \mathrm{PC}^*\mathrm{sd}_{pp}/\mathrm{sd}_{\omega}$. Figure 7 shows the regression coefficient a based on EC data as a function of the time of the year for the entire domain (center) and for the four quadrants (defined earlier). By expressing $\widehat{\mathrm{PP}}$ in mm day^{-1} and $\hat{\omega}$ in units of 10^{-2} N m^{-2} s^{-1} we find a (year-round) to be about -0.6. This is close to the values in Table 4 where a moderately warm atmosphere was calculated to produce $\sim$0.6 mm day^{-1} when lifted continuously by -10^{-2} N m^{-2} s^{-1} at 500 mb.

The calculations shown in Table 4 also indicate that a should vary with the moisture content of the atmosphere. That is, a should be a function of season (high in summer, low in winter presumably) and a should be high in a warm moist climate and low in a dry cold climate. There is some evidence in Fig. 7 that such variations indeed occur. Supportive evidence is that a is lowest in the desert Southwest ($a \sim -0.3$) and also that a is sharply higher during the period of "monsoon" in the Southwest. Otherwise, we cannot conclude that a in the Southeast is higher than, say, in the Northeast even though the atmospheric moisture content in the Southeast is undoubtedly larger. Also, there is no clear evidence for a to be high in summer and low in winter. This may be due to the fact that the correlation in summer is insignificant to start with ($\mathrm{PC}^* \sim -0.2$) so that any further physical interpretation is impossible.

Another check on the numerical value of a was performed with Climate Division rainfall data over India.

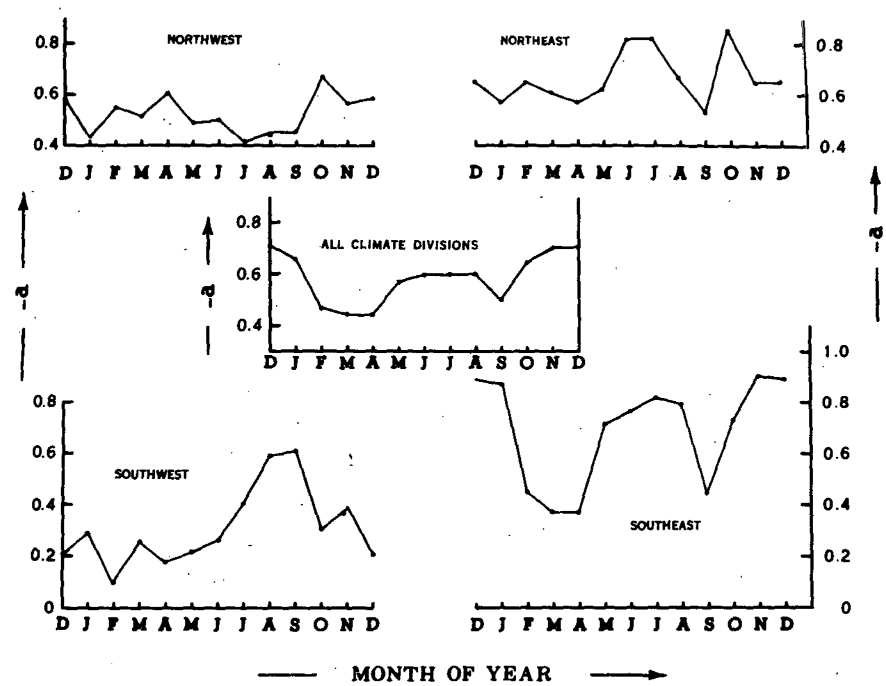

FIG. 7. The regression parameter a as a function of season for the United States as a whole (central portion) and for the four quadrants. The a is determined from a least-squares regression fit between monthly mean precipitation and EC vertical motion anomalies at 500 mb observed over the United States during 1981–83.

Vertical motion data at 500 mb for 1981–83 derived from EC were averaged over the four months June, July, August and September (summer monsoon season), and the anomalies were then correlated with June–September mean anomalies in rainfall in 29 Climate Divisions in India. The results can concisely be given as

$$\widehat{\mathrm{PP}} = -0.4\hat{\omega}$$

where $\widehat{\mathrm{PP}}$ and $\hat{\omega}$ have the same units as before. It is encouraging to see that also over India rainfall and vertical motion anomalies are negatively correlated (PC* = −0.45), but the value of $a = -0.4$ is lower than we had expected for a warm moist climate. Situated at low latitudes ($8°\mathrm{N} < \phi < 30°\mathrm{N}$) and in an area where diabatic processes are important the EC (and NM) vertical motions are perhaps less reliable. Moreover the convective rainfall may not have a simple relationship with large scale vertical motion.

At this point, we can also make a comparison with Nitta's (1972) work concerning the vertical motion (calculated from divergence of observed winds) and rainfall at the tropical Marshall Islands in 1956 and 1958. Dividing the precipitation collected at the Islands during March–July by the time mean 500 mb vertical velocity during that period yields $a = -0.8$ and $a = -1.4$ for 1956 and 1958, respectively. This can be taken as empirical evidence that a is larger negative in a moist, warm atmosphere.

We will consider now variations of the regression equation so as to allow for variations in atmospheric moisture availability represented by the mixing ratio at 850 mb (q). Specifically, we will investigate the performance of

$$\widehat{\mathrm{PP}} = b\widehat{q\omega}, \qquad (11a)$$

$$\widehat{\mathrm{PP}} = cq_n\hat{\omega}, \qquad (11b)$$

The motivation is the following. In (11a) we weigh the vertical motion at 500 mb with the simultaneous (monthly mean) mixing ratio at 850 mb since it seems essentially correct to expect more (less) precipitation if q is high (low). We may not always know q. Therefore, in (11b) we have degraded (11a) by using the climatological space dependent q_n, that is the 1981–83 average.

In contrast to most previous results the exercises (11a)–(11b) are entirely based on NM data (because the January and February 1981 q fields were missing in the EC dataset). Multiplying ω by q does improve PC* in 10 out of 12 months: the yearly mean PC* increases from −0.414 to −0.434. This is modest empirical evidence that a wet atmosphere over the United States produces more rain (for the same $\hat{\omega}$ at 500 mb) than a dry atmosphere. In (11b) we consider $\hat{\omega}q_n$, thereby ignoring the interannual variability of q; and indeed PC* (yearly mean) drops back to −0.423.

Finally we consider

$$\widehat{\mathrm{PP}} = d\widehat{\left(-u\frac{\partial q}{\partial x} - v\frac{\partial q}{\partial y}\right)}_{850}. \quad (11c)$$

That is, we assume that horizontal moisture advection in a steady state situation should be related to precipitation, without any reference to vertical motion. The yearly mean PC* for (11c) is +0.210, a rather disappointing result.

In summary, our attempt to extract the role of moisture in the relationship between $\widehat{\mathrm{PP}}$ and $\hat{\omega}$ from empirical data gives at best modest results. It follows that given upward motion either moisture is not a limiting factor for precipitation at these time-scales, or that the moisture analyses are of a low quality.

5. Potential application in linear models

The empirical study has indicated that, on the monthly time scale, precipitation and vertical motion anomalies are linearly related. Therefore it makes sense to consider application of this finding in linear anomaly steady state models. Since the seasonality and geographical dependence of a, in (10), are hard to interpret we will assume for the sake of this discussion

$$\widehat{\mathrm{PP}} \approx -0.5\hat{\omega} \quad (12)$$

at all times and places, where the units for $\widehat{\mathrm{PP}}$ and $\hat{\omega}$ are mm day^{-1} and 10^{-2} N m^{-2} s^{-1}, respectively. Assuming, for simplicity, a layer of 500 mb over which $\hat{\omega}$ varies little with height and assuming further that precipitation is formed uniformly within that layer, (12) translates into a heating rate uniform with height given by

$$\frac{\partial \hat{T}}{\partial t} = \frac{\hat{Q}_{\mathrm{LH}}}{c_p} \approx a^*\hat{\omega} = -2.5\times10^{-4}\hat{\omega} \quad [\text{units: K s}^{-1}] \quad (13)$$

The context in which we want to apply (13) is a linear stationary anomaly model of the type described in Opsteegh and Van den Dool (1980). The linearized thermodynamic equation reads typically:

$$U_n\frac{\partial \hat{T}}{\partial x} + \hat{v}\frac{\partial T_n}{\partial y} - \sigma_n\hat{\omega} = \hat{Q}/c_p \quad (14)$$

where U_n, T_n and σ_n are basic state zonal wind, temperature and static stability, and $\hat{T}$, $\hat{v}$, $\hat{\omega}$ are the anomalies in temperature, meridional wind and vertical velocity in response to anomalous heat forcing $\hat{Q}$. Here $\hat{Q}$ consists in principle of latent, sensible and radiational heating, i.e.

$$\hat{Q} = \hat{Q}_{\mathrm{LH}} + \hat{Q}_{\mathrm{SH}} + \hat{Q}_{\mathrm{RAD}} \quad (15)$$

Since $Q_{\mathrm{LH}}/c_p = a^*\hat{\omega}$ holds approximately [see (13)], it is possible to rewrite the $-\sigma_n\hat{\omega}$ term in (14) as $-(\sigma_n + a^*)\hat{\omega}$. In other words: expressing precipitation anomalies in terms of vertical velocity anomalies is equivalent to a redefinition of the static stability of the basic state. Since σ_n is typically on the order of 5×10^{-4} K/Pa and $a^* \approx -2.5 \times 10^{-4}$ K/Pa the change of σ_n to $(\sigma_n + a^*)$ is very large and has important consequences in a model.

There are reasons to believe that the empirical a^* is too large. In some areas the climatological mid-tropospheric σ_n can be as low as 3.10^{-4} K Pa^{-1} thereby making $\sigma_n + a^*$ almost zero. The reason for a^* being too large is that in all likelihood the empirical ω data are too small in magnitude especially in areas of precipitation.

Some preliminary experiments with linear stationary models indicate that with decreased static stability the response to a prescribed heat source becomes more vigorous. This was to be expected, especially in the tropics, since a larger $\hat{\omega}$ is needed to balance the forcing. However, once large divergence/convergence is established in the upper level tropics the effect will be visible also in the midlatitudes through wave propagation.

Although attempts to include moisture were not very successful from an empirical point of view, see section 4, we feel it should be correct to write

$$\mathrm{PP} = cq_n\hat{\omega} \quad (16)$$

In this case the constant of proportionality (cq_n) becomes a function of season and location. A linear model with a basic state varying in three-dimensions is needed to effectively apply (16).

The idea to reduce the static stability as an easy parameterization of rainfall is not new. But it is only in the framework of anomalies and sufficiently long time averages (allowing negative precipitation) that the idea is powerful. In this study we have tried moreover to estimate quantitatively the reduction of σ_n from empirical data.

The reduction of σ_n seems to be at variance with the climatological effect of the release of latent heat, which is to warm the middle troposphere and to stabilize the lower troposphere. Of course we are not concerned here with a maintenance equation for σ_n, but rather one for $\hat{T}$. Nevertheless it is good to reflect on what σ_n (based on measurement) really stands for. The basic question is: Does the observed σ_n already reflect the relation between $\widehat{\mathrm{PP}}$ and $\hat{\omega}$ and is it therefore redundant to reduce σ_n to $\sigma_n + a^*$? This is a philosophical problem encountered in all studies where a prescription of the basic state is necessary, and will not be solved here.

6. Comparison of NM and EC vertical motion

Starting from the assumption that vertical motion and precipitation *should* be correlated, we have an opportunity here to closely compare the quality of NM and EC 500 mb vertical motion anomalies over the United States during 1981–83. Using (1) the pattern correlation between $\hat{\omega}$ and $\widehat{\mathrm{PP}}$ has been calculated for all 36 months, for both EC and NM vertical motion. The results are plotted in Fig. 8. Individual values of

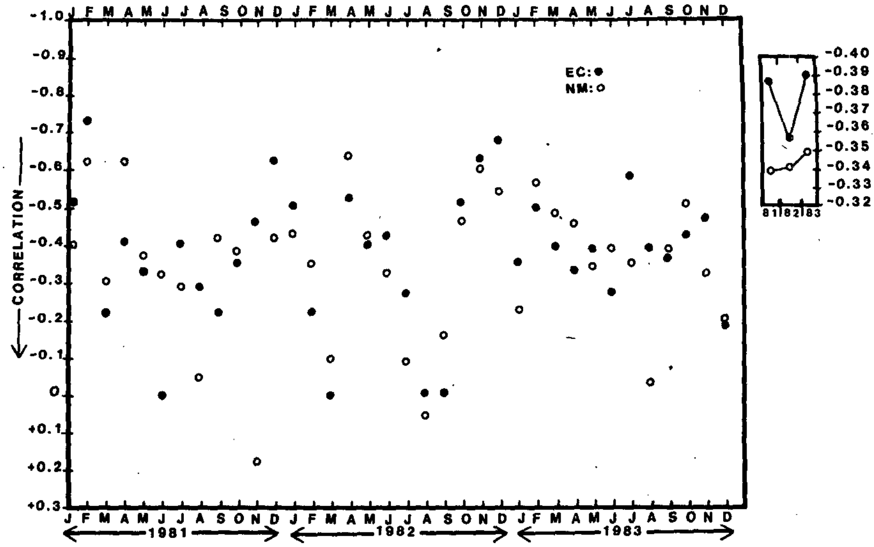

FIG. 8. The pattern correlation [see (1) for a definition] of anomalies in monthly mean precipitation with monthly mean vertical velocity at 500 mb over the United States for all months during 1981–83. The circles (full dots) refer to NM (EC) vertical motion. The inset on the right represents yearly averages.

−0.38 (or larger negative) pass the test of statistical significance based on $N_e \approx 21$, see Table 1. There is incredible scatter from month to month, but nevertheless it is clear that the EC- and NM-based correlations are generally close in a given month. This implies that most often $\hat{\omega}$ is similar in the analyses of these two centers. A notable exception is November 1981.

It is only upon averaging these values over the whole year, see inset at the right in Fig. 8, that it becomes clear that the EC vertical motions score higher than those of NMC every year by a small margin, typically $\overline{PC} = -0.37$ versus $\overline{PC} = -0.34$. Some of this difference may be due to the systematically smaller scales found in NM vertical motion fields (see Table 2). In addition it is worth pointing out that the magnitude of NM $- \hat{\omega}$ is generally larger then EC $- \hat{\omega}$. However additional smoothing of NM $- \hat{\omega}$ does not improve the $\overline{PC}$ any further.

Three years are not enough to see the improvement in the analysis of the divergent motion that most likely is achieved at both centers. In September 1982 a diabatic initialization scheme was introduced operationally at the EC. It is therefore of some interest to compare correlations before and after that month. We fail to detect in Fig. 8 any change in PC due to this potential improvement of 'observations' of ω.

7. Conclusions

An empirical study using monthly mean vertical motion and precipitation data over the United States during 1981–83 was undertaken. The results are:

(i) Anomalies in precipitation ($\widehat{PP}$) and vertical motion at 500 mb ($\hat{\omega}$) are correlated moderately well. The sign of the correlation is negative so that upward motion implies precipitation. The correlation is best in winter (approximately −0.65) and worst in summer (approximately −0.20), most probably reflecting the seasonality in the relative importance of large-scale and convective rains.

(ii) To a first order approximation $\widehat{PP}$ and $\hat{\omega}$ are linearly related implying that the deficit of rain associated with anomalous downward motion is just as large as the surplus of rain associated with anomalous upward motion of the same magnitude.

(iii) The empirical parameter a (in $\widehat{PP} = a\hat{\omega}$) has an order of magnitude comparable to what one can expect from adiabatic ascent in a saturated atmosphere. The seasonality and geographical dependence of a, however, are harder to interpret. There is only weak empirical evidence for a to be large in a moist atmosphere.

(iv) Attempts to explicitly include atmosphere moisture in the relation between $\widehat{PP}$ and $\hat{\omega}$ did not lead much improvement of the empirical results. Parameterization of $\widehat{PP}$ in other variables leads to results inferior to the simple $\widehat{PP}$, $\hat{\omega}$ correlation.

(v) The linear relation $\widehat{PP} = a\hat{\omega}$ can easily be implemented in a linear steady state anomaly model. A parameterized latent heat source implies a significant reduction of the static stability.

(vi) Vertical velocities taken from the post-initialized initial states of Numerical Weather Prediction models during 1981–83 appear credible. Their patterns are reasonable but the magnitude of ω is probably too small.

Most calculations reported in this paper were based on monthly anomalies. It suffices to say that correlations and regression constants for seasonal mean anomalies were very similar and need not be discussed here.

In this study we compared vertical motion as present in the initial state of NWP models to independently observed rainfall. For one month, February 1982, we compared also with the rainfall calculated by the NMC model during the first 24 hours of the 28 forecasts. While the pattern seems alright the precipitation amount produced by the model is only half of what is observed. This can perhaps be taken as another indication that, at least initially, vertical motion is too small.

Knowing $\hat{\omega}$ at 500 mb explains some 35–40% of the interannual variance of monthly precipitation amounts in winter over the United States, a percentage very close to Klein and Bloom's (1987) explained variance obtained from screening the 700 mb height field. It seems therefore that, as of the early 1980s, the application of the continuity equation to (u, v) fields derived from state of the art global data assimilation systems yields a reasonable measure of vertical motion from which precipitation can be inferred.

Acknowledgments. We thank Klaus Arpe, Robert Mureau, Philip Arkin, John Janowiak and Chet Ropelewski for providing us with the vertical motion and USA rainfall data. Thanks are due to ECMWF for letting us use the ECMWF/WMO analyses set. The Indian rainfall data was communicated to us by D. A. Mooley and originally compiled by B. Parthasarathy. We also acknowledge the help of Muthuvel Chelliah, Suranjana Saha, Åke Johansson and Erland Källen in some of the discussions and calculations. This research was supported by the Climate Dynamics Program, Division of Atmospheric Sciences, National Science Foundation, under Grant ATM-8314431.

REFERENCES

Bannon, J. K., 1948: The estimation of large-scale vertical currents from the rate of rainfall. *Quart. J. Roy. Meteor. Soc.*, **74**, 57–66.

Cayan, D. R., and J. O. Roads, 1984: Local relationships between United States West Coast precipitation and monthly mean circulation parameters. *Mon. Wea. Rev.*, **112**, 1276–1282.

Curtis, R. C., and H. A. Panofsky, 1958: The relation between large-scale vertical motion and weather in summer. *Bull. Amer. Meteor. Soc.*, **39**, 521–531.

Englehart, P. J., and A. V. Douglas, 1985: A statistical analysis of precipitation frequency in the conterminous United States, including comparisons with precipitation totals. *J. Climate Appl. Meteor.*, **24**, 350–361.

Fulks, J. R., 1935: Rate of precipitation from adiabatically ascending air. *Mon. Wea. Rev.*, **63**, 291–294.

Haltiner, G. J., and R. J. Williams, 1980: Numerical prediction and dynamic meteorology. Wiley & Sons, 497 pp.

Hansen, J., and A. H. Thompson, 1965: Vertical motion calculations and satellite cloud observations over the western and central United States. *J. Appl. Meteor.*, **4**, 18–30.

Harnack, R. P., and J. R. Lanzante, 1985: Specification of United States seasonal precipitation. *Mon. Wea. Rev.*, **113**, 319–325.

Hollingsworth, A., A. C. Lorenc, M. S. Tracton, K. Arpe, G. Cats, S. Uppala, and P. Kållberg, 1985: The response of numerical weather prediction systems to FGGE level IIb data. Part I: Analyses. *Quart. J. Roy. Meteor. Soc.*, **111**, 1–66.

Janowiak, J. E., C. F. Ropelewski, and M. S. Halpert, 1986: The precipitation anomaly classification: A method for monitoring regional precipitation deficiency and excess on a global scale. *J. Climate Appl. Meteor.*, **25**, 565–574.

Klein, W. H., 1963: Specification of precipitation from the 700-millibar circulation. *Mon. Wea. Rev.*, **91**, 527–536.

Klein, W. H., and H. J. Bloom, 1987: Specification of monthly precipitation over the United States from the surrounding 700 mb height field in winter. Submitted to *Mon. Wea. Rev.*

Lau, N-C., 1979: The structure and energetics of transient disturbances in the Northern Hemisphere wintertime circulation. *J. Atmos. Sci.*, **36**, 982–995.

Miller, A., and H. A. Panofsky, 1958: Large-scale vertical motion and weather in January 1953. *Bull. Amer. Meteor. Soc.*, **39**, 8–13.

Mooley, D. A., B. Parthasarathy and G. B. Pant, 1986: Relationship between Indian summer monsoon rainfall and location of the ridge at the 500 mb level along 75°E. *J. Climate Appl. Meteor.*, **25**, 633–640.

Nitta, T., 1972: Energy budget of wave disturbances over the Marshall Islands during the years of 1956 and 1958. *J. Meteor. Soc. Japan*, **50**, 71–84.

Opsteegh, J. D., and H. M. van den Dool, 1980: Seasonal differences in the stationary response of a linearized primitive equation model: Prospects for long-range weather forecasting? *J. Atmos. Sci.*, **37**, 2169–2185.

Panofsky, H. A., and G. W. Brier, 1968: Some applications of statistics to meteorology. The Pennsylvania State University, 224 pp.

Prabhakara, C., D. A. Short and B. E. Volmer, 1985: El Niño and atmospheric water vapor: Observations from Nimbus 7 SMMR. *J. Climate Appl. Meteor.*, **24**, 1311–1324.

Rosen, D. R., and D. A. Salstein, 1985: Effect of initialization on diagnoses of NMC large-scale circulation statistics. *Mon. Wea. Rev.*, **113**, 1321–1337.

Saha, K. R., D. R. Dikka and P. Krishna Rao, 1969: Some flow features of the Indian Summer monsoon deduced from Nimbus-II radiation data. *Tellus*, **6**, 806–813.

Shenk, W. E., 1963: Tiros II Window radiation and large scale vertical motion. *J. Appl. Meteor.*, **2**, 720–775.

Stidd, C. K., 1954: The use of correlation fields in relating precipitation to circulation. *J. Meteor.*, **11**, 202–213.

Weare, B. C., and M. A. Hoeschele, 1983: Specification of monthly precipitation in the western United States from monthly mean circulation. *J. Climate Appl. Meteor.*, **22**, 1000–1007.

Webster, P. J., 1981: Mechanisms determining the atmospheric response to sea surface temperature anomalies. *J. Atmos. Sci.*, **38**, 554–571.

White, G. H., 1983: Estimates of the seasonal mean vertical velocity fields of the extratropical Northern Hemisphere. *Mon. Wea. Rev.*, **111**, 1418–1433.