NOTES AND CORRESPONDENCE

Precipitation and Circulation Covariability in the Extratropics

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ABSTRACT

Extratropical precipitation is primarily produced by cold and warm fronts associated with surface cyclones and upper-level troughs. The growth of these midlatitude storms is partially controlled by the dry baroclinicity of the troposphere, which in turn can be roughly quantified by the intensity of the upper-level zonal flow. Orographic rainfall, an important component of the precipitation in several midlatitude regions, is also partially determined by the intensity of the cross-mountain midlevel winds. Thus, at monthly and longer time scales, variations of precipitation and zonal flow aloft (as well as wind shear) at a given location should exhibit some degree of coherence. In this work the local covariability of these variables is documented over intermonthly and interannual time scales, using global precipitation products and atmospheric reanalysis from 1979 to 2004. The spatial correspondence between the precipitation and two indices of synoptic activity in the extratropics is also documented.

The local correlation \( r \) between monthly anomalies of precipitation and upper-level (300 hPa) zonal flow varies in space, from moderately and even highly significant values \( r \approx 0.3 \) to 0.7 over the midlatitude oceans to near zero over the interior of continental areas. Broadly similar results are found when considering the monthly variance of the high-pass-filtered meridional wind (an index of eddy activity) or the midlevel Eady growth rate. The local correlation map between precipitation and low-level (850 hPa) zonal flow is similar to its upper-level counterpart, but the correlations over open ocean are somewhat weaker, while orographic effects show up more clearly. The correlations are positive and large upstream of the major north–south-oriented mountain ranges, as strong westerlies promote upslope rain in addition to storm-related precipitation. In contrast, the correlation tends to be negative downstream of the ranges, as strong westerlies enhance the rain shadow effects over the lee side.

1. Introduction

Tropical and extratropical precipitation are readily differentiated on the basis of the zonal average of annual mean rainfall (Fig. 1b) because of a marked subtropical minimum found in both hemispheres. When displayed on a map (Fig. 1a), the distinction between tropical and extratropical precipitation regimes is less evident since the bands of strong precipitation over the midlatitude oceans are connected with tropical areas of rainfall. Tropical precipitation (within 20° of the equator) accounts for nearly half of the global precipitation and it is mostly produced by convective clouds (deep and shallow) and their associated areas of stratiform clouds. Extratropical precipitation reaches a maximum in midlatitudes (40°–50°), primarily produced by deep stratiform clouds (nimbostratus) and embedded convective cells that develop along cold and warm fronts. The frontal systems are in turn associated with surface cyclones, an integral part of the baroclinic waves that populate the midlatitudes.

Although each midlatitude storm (or transient eddy) exhibits a unique evolution, they tend to move along rather narrow latitudinal bands known as storm tracks (e.g., Hoskins and Valdes 1990), whose identification and diagnosis has been performed using several methods and variables [see Paciorek et al. (2002) for a review]. An updated, detailed view of the storm tracks in the Northern and Southern Hemispheres is provided by

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Hoskins and Hodges (2002, 2005, respectively), based on the application of filtered variances and feature-tracking techniques. They describe the seasonal march of the hemispheric patterns and its association with the mean flow, and identify regions of preferent genesis/lysis of baroclinic systems associated with prominent topography and oceanic fronts. In the NH the prominent storm tracks are located over the midlatitude Atlantic and Pacific Oceans, attaining maximum intensity in early winter and fall/spring, respectively (e.g., Nakamura 1992; see also Fig. 1a). A secondary NH storm track is found during summer over high-latitude land masses (e.g., Yoon and Chen 2006). In the SH there is a single, subpolar storm track during the austral summer, most intense over the eastern Atlantic and Indian Oceans (e.g., Trenberth 1991; Nakamura and Shimpo 2004; see also Fig. 1a). During austral winter strong upper-level wave activity is also found at subtropical latitudes stretching from the Indian Ocean into the eastern Pacific.

According to the linear theory, the growth of individual eddies is highly dependent on the dry baroclinicity of the large-scale flow in which they evolve (e.g., Frederiksen and Frederiksen 1993). Nevertheless, nonlinear effects and diabatic heating complicate the eddy–mean flow relationship. Furthermore, there are several indices of baroclinic eddy activity between which the correlation is only low to moderate outside the core of the storm tracks (e.g., Paciorek et al. 2002). Thus, when considering monthly or annual means, the correspondence between the position and intensity of, say, upper-level jet streams and storm tracks is not particularly strong (e.g., Chang et al. 2002; Nakamura and Shimpo 2004).

Because of the lack of a direct relation between cyclone intensity and rain formation, the correspondence between circulation and precipitation on monthly and longer time scales should be further complicated. Such relation is particularly difficult to establish over subtropical regions, where most of the rainfall is caused by upper-level disturbances and the subtropical end of cold fronts arching equatorward from storms moving at higher latitudes (e.g., Eichler and Higgins 2006). Furthermore, most of the precipitation from midlatitude storms originates in warm conveyor belts (WCBs; an airstream originating in the cyclone warm sector and ascending slantwise along the cold front), but only about half of the cyclones are associated with WCBs, according to a recent climatology of these bands presented by Eckhardt et al. (2004). On the other hand, some of the rainiest areas in midlatitudes are found on the windward side of major mountain ranges (e.g., the Andes, the Rockies, and the southern Alps) where the amount of orographic precipitation is partially explained by the intensity of the cross-mountain flow at low and mid levels (e.g., Roe 2005).

Thus, it is not obvious whether precipitation and zonal flow aloft/eddy activity exhibit coherence in their
temporal variations at monthly and longer time scales and how the degree of coupling varies in space.\(^1\) In this short contribution we aim to answer the previous questions by mapping the local covariability between precipitation and several circulation variables over intraseasonal and interannual time scales. Among other aspects, these questions are relevant because inspection of the changes in model-derived large-scale circulation and storm track position between present-day and future/past simulations are being used to make inferences on precipitation changes (e.g., Bengtsson et al. 2006). The global precipitation products and atmospheric reanalysis used in our analysis are described in section 2. In section 3 we describe the covariability on intraseasonal (month to month) and interannual scales between precipitation–mean flow (section 3a) and precipitation–eddy activity (section 3b), as well as the association between large-scale circulation variability and regional-scale precipitation anomalies (section 3c). Section 4 presents our concluding remarks.

2. Data and methods

Two global datasets have been primarily used in this work: the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis (NNR; Kalnay et al. 1996) and the Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP; Xie and Arkin 1997). Our analysis is based on monthly mean values and extends over 26 years from January 1979 to December 2004. CMAP is a widely used precipitation product\(^2\) obtained by merging gauge observations, several satellite estimates, and blended NNR precipitation.

To document the precipitation–circulation covariability, we calculated the Pearson (ordinary) correlation between pairs of time series [precipitation (\(P\)) and circulation] at collocated grid points. The circulation variables include monthly means of zonal and meridional flow at 300 and 850 hPa (\(U_{300}\), \(V_{300}\), \(U_{850}\), and \(V_{850}\) respectively) as well as two indices of the eddy activity. The first is the monthly variance of the 300-hPa meridional wind (\(\Sigma V_{300}\)) calculated from daily mean values of \(V_{300}\) [the values were previously bandpass filtered (3–15 days)]. Our second index is the midtropospheric Eady growth rate \(\sigma = 0.31 (f \cdot N^{-1}) \partial \overline{V}/\partial z\), where \(f\) is the Coriolis parameter, \(N\) is static stability, \(\overline{V}\) is total wind, and \(z\) is height. The wind shear was evaluated between 500 and 700 hPa, and \(N\) is the average value for that layer; \(\sigma\) was calculated daily and then time averaged to form the monthly means.

Although CMAP and NNR are both on a 2.5° × 2.5° latitude–longitude grid, CMAP data correspond to the area-mean precipitation at each box (the first box centered at 88.75°S, 1.25°E) while NNR represents variables at the grid nodes (the first node at 90°S, 0°). Therefore, NNR variables were translated to the center of the boxes by averaging the values at the four surrounding nodes. The local correlation (\(r_l\)) was obtained for each of the 10 368 (144 × 72) grid boxes over the globe and displayed on a map if statistically significant. The significance was assessed locally using a two-tailed Student’s \(t\) test at the 95% confidence level. Results from spatially lagged correlation are also discussed in section 3c.

3. Results

a. Precipitation–mean flow covariability

The local correlation map between precipitation and 300-hPa zonal wind using monthly anomalies is shown in Fig. 2a. (At each grid box, the mean annual cycle was previously removed using the anomaly filter). Over most of the extratropical oceans, both variables are positively correlated with anomalies of upper-level zonal flow explaining from 30% to 60% of the precipitation variance on intraseasonal and interannual time scales. In the NH, the local correlations tend to increase eastward over the Atlantic and Pacific Oceans. In the SH, there is a belt of large, positive correlations at \(\sim 45°\)S, interrupted over eastern South America. High correlations (\(r_l > 0.6\)) are also found along the subtropical portions of the South Pacific and South Atlantic convergence zones and upstream off the west coast of southern South America. Over the equatorial central Pacific (and to some extent over the equatorial Atlantic) there are negative correlations between \(P\) and \(U_{300}\)
mostly produced by the ENSO signal in that region. Over the continental landmasses, variations of precipitation and zonal flow aloft are mostly uncorrelated. Nevertheless, at least part of the low correlation might stem from the low signal-to-noise ratio over dry continental areas.

The basic interpretation of the positive correlations between precipitation and 300-hPa zonal flow over the midlatitude oceans is that $U_{300}$ is an indicator of the dry baroclinicity of the tropospheric column. Thus, at monthly and longer time scales, strong westerlies aloft are conducive of a rapid growth and fast succession of baroclinic disturbances, leading to an increase of cyclone/frontal precipitation. As discussed in the introduction, however, the previous conceptual sequence is weakly linked, so the significant correlation between $P$ and $U_{300}$ is noteworthy. If the local correlations are calculated using seasonal or annual means (not shown),

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For instance, during El Niño years abnormally wet conditions over the central equatorial Pacific are associated with easterly anomalies at upper levels and westerly anomalies at low levels, thus producing negative $P-U$ correlations at 300 hPa (Fig. 2a) and positive correlations at 850 hPa (Fig. 2b).
the overall pattern remains similar to that in Fig. 2a, but the absolute values are slightly lower (yet statistically significant) than their counterparts using monthly anomalies. This indicates the relative short time (a month or so) in which zonal flow, eddy activity, and precipitation processes are actually coupled.

The increase in local correlation between $P$ and $U_{300}$ near most of the continental western seaboard, relative to the midocean conditions, is likely associated with orographic precipitation processes acting either in concert with or independent of baroclinic storms. Since these processes are more directly controlled by the flow near mountain ridge levels (e.g., Roe 2005), we also constructed the local correlation map between $P$ and $U_{850}$ (Fig. 2b). This map is similar to its upper-level counterpart (cf. Figs. 2a,b), but for slightly weaker correlations over the oceanic storm tracks and a sign reversal of $r_0$ over the tropical belt. The orographic effects, however, show up more clearly in Fig. 2b. The largest values of $r_0 (>0.7)$ are found off the western seaboard of South America and North America as well as upstream of the Iberian and the Scandinavian peninsulas, the British Isles, and the Siberian Plateau, while bands of significant, but negative, correlation are found on the eastern seaboard of Australia, Africa, and Asia and the interior of North and South America.

For a north–south oriented mountain range, stronger than normal low-level westerlies increases precipitation on the windward side by enhancing upslope rain. On the lee side there are competing factors. In one hand, stronger than normal westerlies at the mountain top produces more intense downslope flow, favoring dry conditions to the east of the ridge (rain shadow effect); on the other hand, enhanced westerlies might increase rainfall by advecting hydrometeors toward the lee slope and increasing the number of baroclinic disturbances moving over the region. At the scale of the present analysis, the enhanced rain shadow effect is dominant, as signaled by the pronounced west-to-east decrease of $r_0$ (even the change of its sign) observed across the southern Andes, the Rockies, the Australian Great Dividing range, and the southern Alps, as well as the South Africa Plateau and Greenland. For more east–west oriented topographic features, such as the Antarctica periphery and the Gulf of Alaska, the best local correlations between precipitation and circulation are found when considering $V_{850}$ (not shown).

Since the amount of precipitation also depends on the available water supply, several authors have included a moisture parameter at low levels in their statistical downscaling models of precipitation (e.g., Pandey et al. 2000). Here we consider this effect using relative humidity at 850 hPa ($RH_{850}$). The alternative use of specific humidity or moisture flux yields very similar results. Figure 3 shows the increase in the variance ($\Delta r^2$) of precipitation explained when using $U_{850}$ and $RH_{850}$ as “predictors” on a multivariable linear fit, relative to the case when only $U_{850}$ is used (i.e., $r_0^2$). The increase in local correlation over oceanic midlatitudes is modest ($\Delta r^2 \leq 0.1$). In contrast, large values of $\Delta r^2$ (>0.3) are found over continental landmasses (central Europe and the Siberian plains, central North America, Australia, and South Africa), where the correlation $RH_{850} - P$ is positive and large, and the correlation $U_{850} - P$ is very small. Whether low-level moisture variability over those continental areas actually con-

![Fig. 3. Increase in variance of precipitation ($P$) explained by a linear fit using 850-hPa zonal wind and relative humidity as predictors of $P$, relative to the case in which only the zonal wind is used. See text for further details.](image)
trols precipitation, or vice versa, is open to question and requires further analysis.

b. Precipitation–eddy activity covariability

We now document the covariability between precipitation and eddy activity, by means of the correlation maps $P - \Sigma V_{300}$ (Fig. 4a) and $P - \sigma$ (Fig. 4b). As in the case of precipitation and mean flow, the largest correlations are found when using monthly anomalies, although the overall patterns are also evident when considering seasonal or annual means.

The local correlation map constructed using $\Sigma V_{300}$ as the circulation variable (Fig. 4a) shows broad areas of positive, significant local correlation over the midlatitude oceans extending into the subtropics and encompassing several continental areas (e.g., the eastern seaboard of North America). Nevertheless, the largest $P - \Sigma V_{300}$ correlations tend to be located at $60^\circ$ latitude and downstream of the major jet stream (e.g., over the eastern North Pacific). The local correlation map between $P$ and $\sigma$ (Fig. 4b) shares many features with the previous one (as well as with those in Fig. 2), but in this case...
The largest values are found at the entrance of the major jet streams, especially over the western North Pacific and North Atlantic as well as over the South Atlantic convergence zone. Most of the covariability observed in this map stems from the month-to-month changes in the wind shear component of the Eady growth rate [i.e., \( \frac{\partial \sigma}{\partial t} = \frac{\partial (\sigma V/\partial z)}{\partial t} \)] and not from variations in the static stability term. Notably, the local correlations in Fig. 4 are, generally, lower than their counterparts when using \( U_{300} \) despite the closer conceptual link between precipitation and eddy activity. That is probably a reflection of the more “noisy” (in time and space) character of the eddy activity fields. Furthermore, the differential positioning of the correlation maxima with respect to the major jet streams, reflects, at least partially, the “preference” of the eddy activity indices to capture baroclinic storms in their developing (\( \sigma \)) or mature stages (\( \partial \sigma / \partial t \); e.g., Paciorek et al. 2002).

**c. Spatial patterns**

One may speculate that an increase of \( U_{300} \) over one particular grid box should produce not only a local but also a downstream increase in eddy activity and precipitation. To test this, precipitation at each grid box was correlated with \( U_{300} \) everywhere; an example of such a 1-point correlation map is shown in Fig. 5a for a grid box centered at 40°N, 160°W (central North Pa-
cific). In this case, the maximum correlation \( r_{\text{max}} \) is obtained for a \( U_{300} \) grid box offset 10° westward and 5° equatorward (about 1400 km away) from the precipitation grid box, but \( |r_{\text{max}}| - |r_d| \sim 0.1 \), a difference below statistical significance. Similar results are found elsewhere: the correlation between precipitation and upper zonal wind does maximize when considering a \( U_{300} \) grid box 5°–10° to the west of the \( P \) grid box, but the increase in correlation is, at best, marginal. In the case of \( P - \Sigma V_{300} \) and \( P - \sigma \), introduction of a spatial offset produces no significant changes in the correlation analysis.

The 1-point correlation map presented in Fig. 5a also indicates that rainier-than-normal conditions over the central North Pacific are associated with an elongated area of strong cyclonic anomalies at higher latitudes and weaker anticyclonic anomalies centered at roughly the same longitude but farther south. Such north–south oriented dipoles of geopotential anomalies enhance (relax) the meridional pressure gradient at midlatitudes, thus increasing (reducing) the zonal flow at 40°N, where \( P \) and \( U_{300} \) are highly correlated.

To investigate the generality of this pattern, precipitation at each grid box was correlated with 300-hPa geopotential height elsewhere. For reference, grid boxes in midlatitudes and where zonal flow and precipitation are (locally) tightly coupled, we obtained similar results to those found for the central North Pacific. Figure 5b encapsulates this result by showing a composite of all 1-point correlation maps based on grid boxes where \( r_0 > 0.3 \). The composite\(^4\) was constructed by averaging the correlation fields for a region of 90° latitude and 180° longitude centered in the reference grid box. The individual correlation field was shifted upside-down for reference grid boxes in the SH. With the exception of very persistent blocking events, actual (observed) geopotential anomalies lasting a month or longer are generally associated with global or hemispheric scale phenomena (e.g., ENSO, annular modes) and tend to exhibit a spatial extent larger than those shown in Fig. 5b. Nevertheless, global/hemispheric anomalies might be able to produce significant rainfall anomalies, provided that the largest zonal flow perturbations take place over regions where the local correlation between \( P \) and \( U_{300} \) is high.

4 The compositing analysis used here can be regarded as a basic tool for finding the large-scale circulation patterns that are best correlated with precipitation. More sophisticated techniques should include a multilevel, time-lagged analysis (e.g., lagged maximum covariance analysis as illustrated in Bielli and Hartmann 2004), but are beyond the scope of this note.

4. Concluding remarks

In this short contribution we have documented the covariability of the extratropical precipitation and circulation at intraseasonal (month to month) and interannual time scales, using upper- and lower-tropospheric mean flow and two indices of eddy activity. While the emphasis of this work has been on the local covariability, we also explored the large-scale circulation patterns that are best related with local precipitation anomalies. The key findings are as follows:

- The local correlation between monthly anomalies of precipitation and upper-level zonal flow is significant over most of the extratropical oceans. In the NH the largest values (\( r_0 > 0.5 \)) are found on the eastern side of the ocean basins; in the SH the largest values are found on a midlatitude circumglobal belt, as well as over the subtropical South Pacific and South Atlantic convergence zones. Over midlatitude continental areas precipitation is much better correlated with low-level moisture than with zonal flow aloft.

- Similar results are found when considering the monthly variance of the 300-hPa meridional wind or the monthly average of the midtropospheric Eady growth rate. The local correlations in these cases are, however, lower than their counterparts when using \( U_{300} \), likely a reflection of the more “noisy” (in time and space) character of the eddy activity fields.

- The significant, positive correlation between \( U_{300} \) and \( P \) over the midlatitude oceans and adjacent continental areas is consistent with the fact that strong westerlies aloft are conducive of a rapid growth and fast succession of baroclinic disturbances, leading to an increase of cyclone/frontal precipitation. The increase in local correlation near the western seaboard of continents and major mountain ranges seems associated with orographic effects superimposed on baroclinic storms.

- The orographic effects show up even more clearly in the local correlation map between precipitation and low-level (850 hPa) zonal flow. The correlations are positive and large upstream of the major meridionally oriented mountain ranges (e.g., the southern Andes), as stronger westerlies enhance upslope rain in addition to storm-related precipitation. In contrast, the correlation tends to be negative downstream of the ranges, as strong westerlies enhance the rain shadow effect over the lee side.

A variety of long-lasting, large-scale pressure patterns is able to disrupt the zonal flow aloft (e.g., ENSO, annular modes), leading to significant rainfall anomalies provided that the largest zonal flow perturbations
take place over regions where the local correlation between \( P \) and \( U_{300} \) is significant. Thus, the local covariability between precipitation and circulation described in this paper provides a simple, yet efficient, framework to connect large-scale circulation anomalies with regional-scale precipitation anomalies, suitable for studies of past, present, and future climate variability.

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