Oceanic response to midlatitude Rossby waves aloft and its feedback in the lower atmosphere in winter Northern Hemisphere

Kwang-Yul Kim, Hanna Na, and Jong-Gap Jhun

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[1] Midlatitude Rossby waves substantiated in the upper tropospheric (300–200 hPa) potential vorticity field exert significant influences on surface conditions. The physical changes associated with the propagation of Rossby waves are investigated from a climatological perspective, particularly over the oceans. Detailed analysis quantitatively shows that the passing of positive Rossby waves with cyclonic vortexes increases net shortwave and longwave radiation over an ocean’s surface by promoting a clearer sky condition. Further, the reduced humidity and temperature above the sea surface increases latent and sensible heat fluxes. The increased latent and sensible heat fluxes, in turn, induce secondary atmospheric circulation, characterized by anticyclonic and divergent motion in the lower atmospheric layer above the sea surface. Although the increased latent and sensible heat fluxes may change sea surface temperatures, the complexity of ocean dynamics with complicated land–ocean configurations and the mismatch of the temporal and spatial scales of atmospheric and oceanic motions smears the footprint of Rossby waves.


1. Introduction

[2] Rossby waves are prominent atmospheric features that exert a strong influence on local weather in the midlatitudes [Holton, 1992]. The passing of Rossby waves in the upper atmosphere (300–200 hPa) changes the temperatures and winds of the entire troposphere significantly. Theoretical studies and case studies have qualitatively investigated the surface response to upper atmospheric Rossby waves and their feedback [Hoskins et al., 1985; Davis and Emanuel, 1991; Davis, 1992; Romero, 2001; Ahmadi-Givi et al., 2004; Romero, 2008]. According to these studies, surface cold/warm advection induced by upper-atmospheric Rossby waves changes surface heat fluxes significantly. The propagation of Rossby waves plays an important role in developing cyclones in association with the changes in latent heat fluxes [Romero, 2001]. Further, Rossby wave breaking is known to organize turbulent heat flux into the spatial pattern of the Pacific decadal oscillation through moisture and temperature advection [Strong and Magnusdottir, 2009]. Considering that the effects of latent heat fluxes over land and over the ocean are not identical, the maritime surface response to Rossby waves may differ from the continental surface response.

[3] Near the surface of the ocean, air-sea interaction through turbulent heat flux relates changes in the atmospheric variables, including air temperature and wind speed, and sea surface temperature variations in the extratropics [Tanimoto et al., 2003]. It is not immediately obvious, however, whether there will be a noticeable oceanic response to these weather-scale disturbances aloft because of the vastly different spatial and temporal scales of motion in the ocean and the atmosphere. For this reason, many studies have focused on a resonant oceanic response to stochastic atmospheric forcing [Frankignoul and Müller, 1979; Willebrand et al., 1980; Müller and Frankignoul, 1981]. Other studies have focused on the oceanic response to hurricanes and typhoons based on observational case studies [D’Asaro et al., 2007; Sanford et al., 2007; Morozov and Velarde, 2008; Wada and Chan, 2008] or model experiments [Tsai et al., 2008; Zedler et al., 2009].

[4] In this study, we aim to quantitatively investigate the effects of weather-scale Rossby waves near the ocean surface. We do not focus on individual cases; rather, we approach the subject from a climatological perspective. A systematic and quantitative assessment may reveal the distinctive characteristics of the oceanic response to Rossby waves. Particular emphasis is placed on Rossby waves with time scales of approximately 6–8 days and spatial scales of approximately 3500–6500 km or equivalent wave numbers of 4–7. Rossby waves are one of the most significant sources of midlatitude weather disturbances [Hoskins et al., 1985].

[5] A careful investigation of surface conditions over the oceans reveals that there is a systematic and concerted oceanic response to passing Rossby waves nearly 10 km above
the surface. Additionally, the ocean exerts a significant influence on the lower atmosphere above the sea surface as feedback. Thus, a systematic and quantitative assessment of the oceanic response to Rossby waves and its feedback in the lower atmosphere will be discussed in this study in an effort to improve our understanding of Rossby wave response near the surface of the ocean. An accurate rendition of the climatological vertical structures of Rossby waves will also be addressed below.

2. Data and Methods

[6] Detailed physical mechanisms of the oceanic response to midlatitude Rossby waves and the resulting ocean-lower atmosphere feedback were derived from the daily NCEP/NCAR reanalysis data [Kalnay et al., 1996] and AVHRR sea surface temperatures [Reynolds et al., 2007] over the domain [170°–40°W × 20°–80°N]. Rossby waves were extracted from the potential vorticity (PV) field [Hoskins et al., 1985]. Because the role of Rossby waves is more important in the winter in terms of surface air temperature changes, we used a 120-day record from November 17 through March 16 each year from 1979 to 2008.

[7] Cyclostationary EOF (CSEOF) analysis [Kim et al., 1996; Kim and North, 1997] was conducted on PV between 250 hPa and 200 hPa:

\[
P V(r, t) = \sum_n PV_n(r, t)PC_n(t),
\]

where \( PV_n(r, t) \) are the CSEOF loading patterns and \( PC_n(t) \) are the corresponding principal component (PC) time series. The CSEOF loading patterns, \( PV_n(r, t) \), show the space-time physical evolution of Rossby waves and the PC time series measures the year-to-year fluctuations of the magnitude of propagating Rossby waves. Physically consistent evolutions of other physical variables (e.g., \( T(r, t) \)) are then found as follows:

\[
T(r, t) = \sum_n T_n(r, t)PC_n(t).
\]

This can be accomplished via CSEOF analysis of \( T(r, t) \) followed by a regression analysis between the principal components of \( PV(r, t) \) and \( T(r, t) \). That is,

\[
PC_n(t) = \sum_{m=1}^M \alpha_m^{(n)} T_m(t) + \epsilon_n(t),
\]

where the regression coefficients \( \alpha_m^{(n)} \) are determined so that regression error variance, \( \text{var}(\epsilon_n(t)) \), is minimized. The evolution patterns of the predictor variable, \( T_n(r, t) \), are then determined as follows:

\[
T_n(r, t) = \sum_{m=1}^M \alpha_m^{(n)} B_m(r, t),
\]

where \( B_m(r, t) \) are the CSEOF loading vectors of the predictor variable. Further details of the CSEOF analysis and regression analysis can be found in the Methods of Analysis section and the Appendices of Seo and Kim [2003]. Physical consistency between \( PV_n(r, t) \) and \( T_n(r, t) \) is ensured by their common evolution history, \( PC_n(t) \), whereas the two space-time patterns, \( PV_n(r, t) \) and \( T_n(r, t) \), may not be identical. The physical evolutions depicted in \( PV_n(r, t) \) and \( T_n(r, t) \) should, in fact, satisfy the governing equation of the physical processes they represent. In this way, we have produced space-time evolution patterns of oceanic and atmospheric variables that are physically consistent with those of PV in the 250–200 hPa layer. That is,

\[
Data(r, t) = \{PV_n(r, t), T_n(r, t), V_n(r, t), ..., LH_n(r, t)\}PC_n(t),
\]

where \{\( PV_n(r, t), T_n(r, t), V_n(r, t), ..., LH_n(r, t) \)\} denotes the physical evolution of the oceanic and atmospheric variables we investigated (Tables 1 and 2). These spatial and temporal patterns of physical evolutions, \( PV_n(r, t), T_n(r, t), V_n(r, t), ..., LH_n(r, t) \), describe the detailed nature of the atmospheric and oceanic response to Rossby waves in the 250–200 hPa layer (see, for example, Figures 4, 7, and 11).

### Table 1. Correlations Between the Upper-Level (500–200 hPa) Potential Vorticity and Various Physical Parameters

<table>
<thead>
<tr>
<th></th>
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</tr>
</thead>
<tbody>
<tr>
<td>Total Cloud Fraction</td>
<td>−0.51 ± 0.07 (−2)</td>
<td>−0.53 ± 0.10 (−2)</td>
</tr>
<tr>
<td>Divergence (400–200 hPa)</td>
<td>−0.66 ± 0.04 (−4)</td>
<td>−0.60 ± 0.05 (−5)</td>
</tr>
<tr>
<td>Geopotential Height (500–200 hPa)</td>
<td>−0.64 ± 0.06 (−1)</td>
<td>−0.73 ± 0.07 (0)</td>
</tr>
<tr>
<td>Vertical Velocity (500–200 hPa)</td>
<td>−0.54 ± 0.07 (−4)</td>
<td>−0.61 ± 0.05 (−4)</td>
</tr>
<tr>
<td>Vorticity (500–200 hPa)</td>
<td>0.80 ± 0.06 (0)</td>
<td>0.82 ± 0.05 (0)</td>
</tr>
<tr>
<td>Air Temperature (850–400 hPa)</td>
<td>−0.48 ± 0.14 (−3)</td>
<td>−0.69 ± 0.08 (−1)</td>
</tr>
<tr>
<td>Geopotential Height (1000–600 hPa)</td>
<td>−0.46 ± 0.08 (2)</td>
<td>−0.51 ± 0.06 (3)</td>
</tr>
<tr>
<td>Vertical Velocity (1000–600 hPa)</td>
<td>−0.44 ± 0.07 (−4)</td>
<td>−0.57 ± 0.06 (−4)</td>
</tr>
<tr>
<td>Divergence (1000–600 hPa)</td>
<td>0.52 ± 0.15 (−4)</td>
<td>0.61 ± 0.07 (−5)</td>
</tr>
<tr>
<td>Vorticity (1000–600 hPa)</td>
<td>0.63 ± 0.04 (1)</td>
<td>0.60 ± 0.05 (3)</td>
</tr>
<tr>
<td>Near-Surface Relative Humidity</td>
<td>−0.40 ± 0.08 (−4)</td>
<td>−0.44 ± 0.13 (−2)</td>
</tr>
<tr>
<td>Net Longwave Radiation at Surface</td>
<td>0.54 ± 0.06 (−2)</td>
<td>0.58 ± 0.07 (2)</td>
</tr>
<tr>
<td>Net Shortwave Radiation at Surface</td>
<td>0.34 ± 0.08 (−1)</td>
<td>0.30 ± 0.09 (2)</td>
</tr>
<tr>
<td>Latent Heat Flux</td>
<td>0.55 ± 0.12 (−2)</td>
<td>0.56 ± 0.25 (−1)</td>
</tr>
<tr>
<td>Sensible Heat Flux</td>
<td>0.59 ± 0.06 (−2)</td>
<td>0.58 ± 0.19 (−1)</td>
</tr>
</tbody>
</table>

*Each entry measures the closeness of the physical evolutions for the 120 winter days as represented by the CSEOF loading vectors between the upper-level (500–200 hPa) PV and other variables. The one-standard deviations were derived from a set of ten correlations based on the first 10 CSEOFs. The correlation values at each grid point were averaged over the specified domains. The numbers in parentheses are the spatial lags in units of \( \Delta x (=2.5^\circ) \) with positive values implying that the specific variables lead the upper-level PV and vice versa. Here, “near-surface” represents the 0.995σ level, which, on average, is \( \sim 50 \) m above the surface [Kalnay et al., 1996]. The direction of positive flux is upward.*
Table 2. Correlations Between the Turbulent Heat Flux and Various Physical Parameters$^a$

<table>
<thead>
<tr>
<th>Correlation With</th>
<th>Turbulent Heat Flux</th>
<th>NE Pacific</th>
<th>NW Atlantic</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sensible Heat Flux</td>
<td>0.95 ± 0.11 (0)</td>
<td>0.86 ± 0.21 (0)</td>
<td>0.91 ± 0.06 (0)</td>
</tr>
<tr>
<td>Sea Surface Temperature</td>
<td>−0.05 ± 0.04 (−5)</td>
<td>−0.13 ± 0.05 (4)</td>
<td>−0.22 ± 0.11 (−4)</td>
</tr>
<tr>
<td>2 m Air Temperature</td>
<td>0.53 ± 0.12 (−4)</td>
<td>0.67 ± 0.08 (−4)</td>
<td>0.49 ± 0.17 (−5)</td>
</tr>
<tr>
<td>10 m Wind Divergence</td>
<td>0.60 ± 0.10 (−4)</td>
<td>0.52 ± 0.10 (−1)</td>
<td>−0.40 ± 0.16 (−2)</td>
</tr>
<tr>
<td>10 m Wind Vorticity</td>
<td>0.53 ± 0.10 (−2)</td>
<td>0.61 ± 0.11 (−3)</td>
<td>0.53 ± 0.08 (4)</td>
</tr>
<tr>
<td>Air Temperature (1000–850 hPa)</td>
<td>−0.32 ± 0.11 (−1)</td>
<td>−0.54 ± 0.16 (−1)</td>
<td>−0.45 ± 0.09 (−3)</td>
</tr>
<tr>
<td>Vertical Velocity (1000–600 hPa)</td>
<td>−0.32 ± 0.11 (−1)</td>
<td>−0.54 ± 0.16 (−1)</td>
<td></td>
</tr>
</tbody>
</table>

$^a$Each entry measures the closeness of the physical evolutions for the 120 winter days as represented by the CSEOF loading vectors between the turbulent (sensible + latent) heat flux and other variables. The one-standard deviations were derived from a set of ten correlations based on the first 10 CSEOFs. The correlation values at each grid point were averaged over the specified domains. The numbers in parentheses are the spatial lags in unit of $\Delta x$ (=2.5') with positive values implying that the specific variables lead the turbulent heat flux and vice versa. Here, “near-surface” represents the 0.995σ level, which, on average, is ~50 m above the surface. The direction of positive flux is upward.

[8] Due to the irregular phase and wavelength of PV variations, the total variance of PV is split into many CSEOF modes. We used the first ten modes to explain approximately 50% of the total variability. It should be noted that the physical and dynamical mechanisms addressed below apply equally to all ten modes. Once Rossby waves were extracted from the PV field, physically consistent patterns of other variables could be identified from the surface to the tropopause, as described above. Next, high-pass filtering was conducted on all variables using the Parzen window [Newton, 1988] with a 10-day cut-off period so that ocean-atmospheric interactions with weather-scale Rossby waves could be addressed. The high-pass filtered data retain approximately 1/3 of the variance in the first ten CSEOF modes of PV; the explained variance (~17% of the total variability of 250–200 hPa PV) represents a significant fraction of the high-frequency variability of PV. The correlations in Tables 1 and 2 are derived from the filtered version of the physical evolutions in equation (5) for each mode $n$ up to the first ten modes.

[9] Once physically consistent patterns of evolution are derived from key physical variables as in equation (5), the horizontal and vertical structures of key variables can be seen in equation (5) because physical variables are analyzed at all standard levels. Because these horizontal and vertical structures vary significantly as a function of space and time, time-averaged structures are derived by calculating the correlations between the target variable (upper-level PV) and predictor variables (all other variables) at two vertically aligned points. This should be compared with the conventional method of one-point correlation or regression [e.g., Hoskins et al., 1985; Lim and Wallace, 1991; Chang, 1993]. Because Rossby waves are fairly variable in space and time with significantly different temporal, horizontal and vertical structures for various locations, the one-point regression approach does not necessarily yield accurate Rossby wave structures.

[10] For example, Figure 1 shows the vertical structures of geopotential height and temperature based on one-point regression with the 10-day high-pass filtered PV field in the first CSEOF loading vector. The vertical structure of geopotential height at ~80°N shows that the upper-level response leads the lower-level response when the base point is at (100°W, 45°N), whereas the lead-lag relationship reverses when the base point is at (50°W, 45°N). Although this discrepancy is not serious, a similar disagreement can be seen in the vertical structures of temperature. This indicates that the vertical structure is very sensitive to the base point in a conventional method. This sensitivity is easily understood considering that the speed and the wavelength of Rossby waves vary as a function of time and location; a low correlation compounded by contamination due to variable wavelength and speed makes it difficult to accurately extract the vertical structure of these waves.

[11] In the present study, we investigate the vertical structure of Rossby waves by computing correlations at two vertically aligned points. That is, the response of a certain variable (say, $T_n(r, z, t)$) to upper-level PV (say, $PV_t(r, z = z_{top}, t)$) at a specific point $r = r_1$ is determined by computing the correlation between the two variables at a given point $r_1$ as a function of elevation $z$. Here, $n$ is the CSEOF mode number. One complication of this approach is that the vertical structure of response is often tilted with the elevation; this tilt should be accounted for to accurately extract the vertical structure of the Rossby response from a given variable. It should be noted that the vertical tilt varies from one variable to another. This vertical tilt was taken into account by spatially lagging one variable against another in our study. The spatial lag between each target variable and a predictor variable was determined at each vertical level so that the maximum correlation is obtained at the given lag (Table 1).

3. Response to Rossby Waves

[12] Figure 2 shows the real part of the zonal Morlet wavelet coefficients of the 10-day high-pass filtered first CSEOF loading vector of the upper-level (250–200 hPa) PV at 50°N, where maximum PV variance is observed. A nearly continuous eastward propagation of waves is clearly seen at this latitude. Judging from the alternating signs of the wavelet coefficients, evolution in the loading vector represents primarily Rossby waves and perhaps partly cyclones. Despite high-pass filtering, both the spatial and the temporal compositions of waves appear variable. It should be noted that similar eastward propagations are observed in the first 10 CSEOFs of the 250–200 hPa PV field; the wave number-frequency spectrum shows wave energy residing mainly in the wave numbers between 3 and 7 and periods of 6–9 days for the first ten CSEOF loading vectors (Figure 3). These waves have the characteristic length and time scales of planetary Rossby waves in the upper atmosphere. The dispersion relationship of the linearized vorticity equation for free barotropic Rossby waves is provided as follows:

$$\omega = Uk - \beta k^2 \frac{1}{k^2 + \beta^2}.$$ 

\(\text{(6)}\)
where \( \omega \) is the frequency, \( k \) and \( l \) are the zonal and the meridional wave numbers, and \( U \) is the mean (background) westerly speed. Based on the maximum gradient of the wave number-frequency spectrum (Figure 3), the mean zonal speed is estimated to be approximately 15–21 m s\(^{-1}\) with the assumption of \( k = l \). This coincides closely with the winter mean barotropic zonal speed in the atmosphere, which is approximately 16 m s\(^{-1}\).

Figure 4 shows the 10-day high-pass filtered first CSEOF mode of PV in the atmosphere and the physically consistent responses of geopotential height, temperature and winds over the eastern Pacific and near the North American continent. As Rossby waves pass the selected location, a concerted response of the atmospheric column is clearly seen in the geopotential height and temperature anomalies. The sign of the geopotential height response is nearly uniform through the entire atmospheric column, although the column of response to passing Rossby waves, in general, is slightly tilted toward the west with height. It is remarkable that the physical response is fairly substantial near the surface, although Rossby waves are confined to the upper atmosphere between 400 hPa and 200 hPa.

During the passage of positive PV anomalies, the atmospheric column as a whole cools down, as reflected in the negative temperature and geopotential height anomalies in Figure 4. In fact, there are concerted changes in the physical variables, as summarized in Table 1. Convection is discouraged, as reflected in the negative correlations between PV and vertical velocity in the lower atmosphere and between PV and the cloud fraction. The increase in net longwave radiation from the surface is due to the decreased amount of moisture in the atmospheric column and cloudiness; longwave radiation leaving the surface is less completely trapped in the atmosphere, reducing the amount of downward longwave radiation from the atmosphere and cloud. During clear days, increased longwave radiation is often observed in terms of colder surface conditions. The reduction of convection during the passing of positive PV anomalies is also seen in the increased net shortwave radiation at the surface. The positive PV anomalies promote clear skies, increasing the amount of shortwave radiation reaching the surface.

The wind field also varies according to the sign of the PV (Figures 4c and 4d). As a positive vortex approaches a location (140°W, 45°N), positive meridional wind is observed, while its sign is reversed as the vortex leaves the location. The situation reverses when a negative PV approaches the location. The zonal wind has no clear relationship with the vortex because the sign of the zonal wind depends on the latitudinal location of the vortex with respect to the observing location.
to the location; when the center of a positive vortex is to the north of the location, positive zonal wind should be observed and vice versa.

Figure 5 shows the vertical structures of various physical variables over the eastern Pacific and the western Atlantic. The vertical structures were determined based on the maximum correlation between the upper-level PV and the other physical variables zonally within $-15^\circ$ to $15^\circ$ ($\pm 6$ grid points) with respect to the position of upper-level PV. As shown in Figure 5, geopotential height (dark blue) is tilted westward and temperature (red) is tilted eastward with height; the vertical structures of these parameters are similar to those of Lim and Wallace [1991] and Chang [1993]. Vorticity (light blue) is also tilted westward with height. In contrast, vertical velocity (gray) and divergence (green) appear $\sim 10^\circ$ upstream of the upper-level PV over the oceans. Correlation values indicate that vertical velocity is negatively correlated and divergence is positively correlated with the upper-level PV below about 600 hPa (Figure 6). Generally, correlation decreases near the ocean surface, although it remains to be significant, with the exception of the vertical velocity ($w$) and the temperature over the eastern Pacific.

Discontinuities in the vertical structure are seen in temperature at $\sim 300$ hPa and in divergence at $\sim 500$ hPa (Figures 5 and 6). These discontinuities in the vertical structure are reflected in the sign change of the correlation (Figure 6); in the transition zone at which the sign of correlation switches, the vertical structure is not well traced due to the insignificant correlation between the upper-level PV and these physical variables. A kink is seen in the vertical structure of divergence near the 850 hPa level for both of the oceans (Figure 5), while there is no abrupt change in the degree of correlation at the same level (Figure 6).

In addition to the changes in the atmospheric conditions, Rossby waves dramatically alter the surface conditions; notably, this surface response is most pronounced over the ocean. During the passage of positive PV anomalies, latent and sensible heat fluxes increase over the ocean (Figure 7 and Table 1). The increased sensible heat flux can

**Figure 2.** Plot of the real part of zonal Morlet wavelet coefficients of 10-day high-pass filtered 1st CSEOF loading vector of the 250–200 hPa potential vorticity at 50°N for zonal wave number (a) $k = 9$, (b) $k = 4$, and (c) $k = 3$. The abscissa is longitude and the ordinate is time (days). A similar plot is obtained for the imaginary part of zonal Morlet wavelet coefficients.

**Figure 3.** Log plot of wave number-frequency spectrum of the high-pass filtered CSEOF loading vectors of the 250–200 hPa potential vorticity at 50°N. An averaged spectrum for the first 10 CSEOF loading vectors is shown here.
Figure 4. Ten-day high-pass filtered potential vorticity (shade; red: positive, blue: negative) and (a) geopotential height anomalies (contour), (b) temperature anomalies (contour), (c) zonal wind anomalies (contour), and (d) meridional wind anomalies at (140°W, 45°N) over the eastern Pacific Ocean near the North American continent during the period of November 17-December 26 for the first CSEOF mode.
be explained in terms of the increased temperature gradient between the sea surface and the atmosphere above it (Table 1). The increased temperature gradient is the result of decreased air temperature above the sea surface and the negative anomaly of vertical velocity near the surface (Table 2). The latent heat flux also increases because the air above the sea surface is drier during the passing of positive PV anomalies. This is reflected in the decreased relative (and specific) humidity of air near the surface (Tables 1 and 2). The increased gradient of vapor pressure between the sea surface and the air above it explains the increased latent heat flux. The sensible and latent heat fluxes can also be dependent on wind speed. No significant correlation, however, is found between the PV aloft and the wind speed near the surface. While propagating PV anomalies produce consistent wind anomalies, wind speed, which is a nonlinear function, does not show any consistent changes. The magnitudes of latent and sensible heat fluxes induced by Rossby waves are large, often reaching an average of 50 W m$^{-2}$ in a day, particularly along the Atlantic seashore. Based on the first ten CSEOFs, typical one-standard deviations of area-averaged daily latent and sensible heat fluxes are, respectively, 7 W/m$^2$ and 4 W/m$^2$ for the northeastern Pacific Ocean (160°–130°W × 35°–55°N) and 13 W/m$^2$ and 12 W/m$^2$ for the northwestern Atlantic Ocean (70°–50°W × 35°–55°N). These numbers, however, should be interpreted as rough estimates in lieu of the significant systematic errors in turbulent flux particularly over the western boundary currents [Moore and Renfrew, 2002].

4. Secondary Circulation

[19] The release of latent and sensible heat fluxes from the surface of the ocean should be consumed somewhere within the atmospheric column. To understand how the turbulent (sensible + latent) heat flux is consumed within the atmospheric column, correlations between the turbulent heat flux induced by the upper-level PV and other physical variables

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**Figure 5.** The vertical structures of temperature (red), geopotential height (dark blue), vorticity (light blue), divergence (green) and $\omega$ (gray) for (a) the eastern Pacific Ocean (35°–55°N × 160°–130°W) and (b) the western Atlantic Ocean (35°–55°N × 60°–40°W). The abscissa denotes the distance in degrees from the center of the PV anomaly. The vertical structures are determined based on the maximum correlation $>0.3$ with the 500–200 hPa averaged PV for the first ten modes. The temperature profile is drawn only below the 300 hPa. The divergence profile is disconnected in the transition zone approximately 500 hPa. See Table 1 for correlation values.

**Figure 6.** Vertical structure of correlation between the upper-level (500–200 hPa) PV and temperature (red solid), geopotential height (blue dashed), $-\omega$ (black dotted), divergence (green dashed) and vorticity (blue dot-dashed) for (a) the eastern Pacific Ocean and (b) the western Atlantic Ocean.
were computed (Table 2). A comparison between Tables 1 and 2 shows that correlations with turbulent heat fluxes are nearly as high as those with the upper-level PV for the lower atmospheric variables. This implies that turbulent heat flux actively participates in the evolution of physical variables in the lower atmosphere. If turbulent heat flux, as a passive response to the upper-level PV, had not exerted an influence on the evolution of physical variables, its correlations with physical variables would have been much lower than those of the upper-level PV because the upper-level PV and turbulent heat flux are only moderately correlated (Table 1). It appears that the increased turbulent heat flux induces secondary atmospheric circulation above the sea surface.

[20] Figure 8 shows correlations of 1000-hPa divergence and vorticity with the upper-level PV (solid) and turbulent heat flux (dotted) for the Pacific and Atlantic Oceans. The positive 1000-hPa divergence is maximally correlated with the upper-level PV, which is $\sim10^\circ-15^\circ$ to the east (Figures 8a and 8b). This upper-level positive PV induces a positive turbulent heat flux slightly to its west (Table 1). This positive turbulent heat flux induces divergence, as shown in the increased correlation of divergence with the turbulent heat flux (red triangle versus black triangle in Figures 8a and 8b). Likewise, the 1000 hPa vorticity is maximally correlated with the positive upper-level PV, which is $\sim5^\circ-10^\circ$ to the west (Figures 8c and 8d). The positive upper-level PV induces positive turbulent heat flux slightly to the west. This positive turbulent heat flux, in turn, further increases the vorticity, as seen in the increased correlation of vorticity with the turbulent heat flux (red triangle versus black triangle in Figures 8c and 8d).

Figure 7. Correlations at spatial lag zero between the physical evolutions of 500–200 hPa averaged PV anomalies and that of (a) latent heat flux anomalies, (b) sensible heat flux anomalies, and (c) turbulent heat flux anomalies for the first CSEOF mode. The dots represent the latitude of the maximum variance of PV.
Although the correlation increase is not significant, the presence of turbulent heat flux appears to induce a stronger divergence and vorticity response in the lower atmosphere. Such fortification of divergence and vorticity is not seen over the continent primarily due to the weak turbulent heat flux (figure not shown). Figure 9 shows that the variances of lower-level divergence and vorticity normalized by the PV variance are much stronger over the oceans than over the continent. Thus, it is not the magnitude of PV that is associated with the land/ocean difference of variances of lower-level divergence and vorticity. Figure 10 further indicates that the impact of the turbulent heat flux is confined to the proximity of the ocean surface. The much stronger near-surface variance of divergence over the oceans is due to the positive fortification of divergence by turbulent heat flux (Figure 10a). The much weaker near-surface variance of vorticity is due to the weakening of the vorticity by the turbulent heat flux; the vorticity is weakened because a negative vorticity anomaly is induced ahead of the positive PV, where positive vorticity is observed (Figures 5, 6 and 8). In the presence of surface friction, anticyclonic circulation near the ocean surface also results in the divergence of wind. As shown in Figure 10, however, the variance of divergence close to the surface is significantly different between the two oceans, whereas the variances at 925 hPa and 850 hPa are not significantly different. Friction alone cannot explain this asymmetry. Furthermore, a similar pattern is not seen over the continent, specifically over the eastern U.S., where the role of friction in inducing near-surface divergence should be more important than over the oceans. Thus, turbulent heat flux appears to be the primary reason for the near-surface variation of variance.

Figure 11 further shows the ageostrophic component of divergence in comparison with the latent heat flux for the first mode. Although the figure depicts only the first mode, similar pictures can be seen for the first ten modes. It is clearly shown that the increased latent heat flux enhances the
10 m wind divergence, particularly over the Atlantic Ocean (latent heat flux leading to 10 m wind divergence). Both Figures 10 and 11 indicate that it is the turbulent heat flux, rather than the friction, that causes the increased divergence over the oceans. Figure 12 together with Table 1 shows that to the east of a positive PV, geopotential height decreases throughout the atmospheric column (that is, the air column compresses), while the difference of geopotential height between two adjacent layers increases near the ocean surface. To the west of a positive PV, where the turbulent heat flux increases, the situation reverses. Thus, there is generally a downward motion in the atmosphere except for an upward motion near the surface due to the increased turbulent heat flux. This seems to explain why divergence increases and vorticity decreases near the ocean surface. To the east of a positive PV anomaly, the increased divergence is reflected in the higher correlation values between the PV aloft and other physical variables over the oceans than over the continent. The convoluted atmospheric circulation and diabatic heating associated with the complex terrain over the continent are partly responsible for this difference.

5. Summary and Discussion

[24] This study investigated the oceanic response to Rossby waves aloft and its feedback with the lower atmosphere. It also examined the detailed vertical structures of key physical variables in response to passing Rossby waves. Instead of using one-point regression, detailed space-time evolutions of key variables in association with Rossby waves were extracted from daily data sets via CSEOF analysis followed by regression analysis between the PC time series of the 250–200 hPa PV anomalies (target) and those of other variables (predictors).

[25] As is also the case over the continent, the propagation of Rossby waves induces intricate and concerted changes throughout the atmospheric column and over the surface of the ocean. One pronounced difference of the oceanic responses is that the changes in physical variables are much more organized and concerted than those that occur over land; this is reflected in the higher correlation values between the PV aloft and other physical variables over the oceans than over the continent. The convoluted atmospheric circulation and diabatic heating associated with the complex terrain over the continent are partly responsible for this difference.

[26] Another remarkable difference is the secondary atmospheric circulation near the sea surface, a result of the significant amount of latent and sensible heat fluxes released from the sea surface during the passing of Rossby waves. Increased heat fluxes for positive Rossby waves aloft results in the increased buoyancy of the air above the sea surface, pushing the isobaric surface upward near the ocean surface.
As a result, the vortex column near the surface shortens, leading to negative vorticity anomalies and positive divergence anomalies near the ocean surface. The magnitude of heat fluxes is also significant enough to change sea surface temperatures on the order of 0.02°C. This is reasonably consistent with a rough estimate of the SST change for a 20 m mixed-layer ocean based on one standard deviation of latent and sensible heat flux variations.

(Figure 13). The latent heat flux anomalies (contour) and the divergence of the ageostrophic component of 10 m wind anomalies (shade) for the first mode over (left) the eastern Pacific Ocean, (middle) the North American continent, and (right) the western Atlantic Ocean. The ageostrophic component of 10 m wind anomalies were computed by removing the geostrophic wind anomalies estimated from the sea level pressure anomalies. The divergence of the ageostrophic component of 10 m wind anomalies is 10–20 times larger than that of the geostrophic component.

(Figure 11). The latent heat flux anomalies (contour) and the divergence of the ageostrophic component of 10 m wind anomalies (shade) for the first mode over (left) the eastern Pacific Ocean, (middle) the North American continent, and (right) the western Atlantic Ocean. The ageostrophic component of 10 m wind anomalies were computed by removing the geostrophic wind anomalies estimated from the sea level pressure anomalies. The divergence of the ageostrophic component of 10 m wind anomalies is 10–20 times larger than that of the geostrophic component.

(Figure 12). Maximum vertical correlation of geopotential height anomalies with PV (red) and the corresponding correlation of geopotential height anomaly differences between two adjacent levels, $\Delta Z(z) = Z(z + 1) - Z(z)$, with PV (blue) over (a) the Pacific Ocean and (b) the Atlantic Ocean.
However, the change in SST due to the release of latent and sensible heat fluxes from the sea surface is much more complex than that of other variables because ocean dynamics with intricate land/ocean configuration redistribute SST in a complicated way (Table 2). Because the time and spatial scales of physical mechanisms in the ocean are, in general, quite disparate from those of Rossby waves aloft, a concerted oceanic response may be extremely difficult. The advection of heat by strong ocean currents, particularly the Gulf Stream, may effectively annihilate any hint of Rossby waves in the sea’s surface temperature changes [Zolina and Gulev, 2003]. The vorticity and divergence of surface wind triggered by latent and sensible heat fluxes may also modify upwelling along the coasts. In addition to these factors, the enormous heat capacity of the ocean also makes it difficult to detect sea surface temperature changes as a result of latent and sensible heat fluxes being released. Only a modeling study can confirm these conjectures.

While, in general, oceanic responses are similar between the Pacific Ocean and the Atlantic Ocean, the magnitude of heat fluxes and the corresponding secondary atmospheric circulations are substantially different between the two. The oceanic response to Rossby waves is much stronger along the coast of the western Atlantic Ocean with a larger change of heat fluxes. One plausible reason for this difference is the differential daily magnitude and location of latent and sensible heat fluxes in the two oceans. As a result, latent and sensible heat fluxes induced by Rossby waves would be different, as would the secondary atmospheric circulation.

Heat fluxes and secondary circulation induced by Rossby waves are not only substantial in magnitude, they also exhibit significant diurnal variations. Therefore, the detailed physical mechanisms of the oceanic response to Rossby waves should be taken into account in energy fluxes at the ocean–atmosphere interface, ocean observations from space, and the physics of the ocean and the atmospheric boundary layers.

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