Subseasonal Variations of Wintertime North Pacific Evaporation, Cold Air Surges, and Water Vapor Transport

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(Manuscript received 2 March 2017, in final form 20 August 2017)

ABSTRACT

This study addresses subseasonal variations of oceanic evaporation $E$ over the North Pacific during winter and the connection with the cold air surges (CASs) and atmospheric water vapor transport using the OAFlux and ERA-Interim daily data. By performing an empirical orthogonal function (EOF) analysis, two dominant modes of subseasonal evaporation anomaly $E_0$ are identified: a zonal wave train–like pattern (EOF1) and an east negative–west positive dipolar pattern (EOF2) in the midlatitude basin. Further analyses yield the following conclusions. 1) The Siberian high (SH)-related CAS has a crucial role in generation of the EOF1 mode of $E$. When the dry and cold air mass passes the region of the warm Kuroshio and its extension [Kuroshio–Oyashio Extension (KOE)], the increased air–sea temperature and moisture differences and intensified wind speed lead to the above-normal oceanic $E$, and vice versa. 2) The Aleutian low (AL)-related CAS contributes to the EOF2 mode of $E$. The intensified AL transports a dramatically colder and drier air mass toward the KOE region and a slightly warmer and wetter one toward the west coast of North America, leading to the east negative–west positive structure of $E_0$ in the midlatitude basin. 3) A quasi-linear relationship exists between $E_0$ and divergent water vapor transport anomalies over the KOE region. Positive (negative) $E_0$ is generally accompanied by anomalous vapor source (sink). 4) The divergent water vapor transport anomalies associated with the two EOFs are preliminarily decided by their individual lower-level wind field anomalies and second by the meridional inhomogeneity of subseasonal specific humidity anomalies. Hydroclimate effects on precipitation over the pan–North Pacific region are also discussed.

1. Introduction

Oceanic evaporation $E$ is a main hydrological cycle component, as shown in Fig. 1 of Trenberth et al. (2007). Water vapor is supplied to the atmosphere via oceanic $E$. Atmospheric circulation redistributes the vapor, and finally it condenses to form clouds and precipitation (Zhu and Newell 1998; Chen et al. 2004a; Sohn et al. 2004; Trenberth et al. 2007; Minobe et al. 2008; Newman et al. 2012; Knippertz et al. 2013; Rutz et al. 2014; Liu et al. 2016; Luo and Tung 2015; Mundhenk et al. 2016). Therefore, oceanic evaporation has a profound influence on weather and hydroclimate through thermal and hydrological ways (Frankignoul 1985; Cayan 1992; Wang et al. 2004; Trenberth et al. 2007; Yu and Weller 2007; Schneider et al. 2010; Kumar et al. 2017). The variability of oceanic $E$ varies in a broad time scale (Frankignoul 1985; Cayan 1992; Alexander et al. 2002; Zhu and Yang 2003; Wu and Liu 2005; Liang and Yu 2016). Previous studies have shown that robust subseasonal variations of $E$ occur around the Kuroshio and its extension [Kuroshio–Oyashio Extension (KOE)] region during winter (Qiu 2002; Kwon et al. 2010; Grodsky et al. 2009). This study focuses on these variations and their connections with the atmospheric circulation and water vapor transport over the North Pacific basin on a subseasonal time scale.

Cold air surge (CAS) activities over the Asian coastal water can cause subseasonal $E$ anomalies in the KOE region (Compo et al. 1999; Jeong and Ho 2005; Park et al. 2008; Fletcher et al. 2016). Parts of the CAS events are connected with the variations of the Siberian high (SH). The SH is a surface high pressure system on the winter mean sea level pressure (MSLP), inhabiting the region between 80° and 120°E north of 40°N. Previous studies have shown that the fluctuation of SH is a critical factor for the cold air mass activities (Gong and Ho 2002; Wu and Wang 2002; Takaya and Nakamura 2005). The

DOI: 10.1175/JCLI-D-17-0140.1
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SH-related CAS activities are anchored by propagation of wave train–like cold and warm airmass anomalies from the Eurasian continent toward the east coast of Asia (Zhang et al. 1997; Compo et al. 1999; Chen et al. 2004b). When the dry and cold air passes the warm KOE region, the noticeable air–sea temperature and moisture differences and intensified wind speed lead to an enhanced oceanic $E$ (Bond and Cronin 2008; Konda et al. 2010; Jiang and Deng 2011; Fletcher et al. 2016).

Other CAS events over the Asian coastal water are related with the subseasonal anomalies of the Aleutian low (AL). The AL is a surface low pressure system over $30^\circ$–$60^\circ$N, $160^\circ$E–$140^\circ$W on charts of MSLP. It is centered near the Aleutian Islands and most intense during winter (Trenberth and Hurrell 1994). When the AL is strengthened, the anomalously cold and dry air is imported from the northwestern North Pacific toward the KOE region (Compo et al. 2001; Newman et al. 2012; Song et al. 2016). The increased air–sea difference and stronger-than-normal wind speed are favorable to an enhanced $E$ around the KOE region.

Many studies have demonstrated the connections of latent and sensible heat fluxes in the KOE region with the atmospheric circulation and storm track over the North Pacific, on a synoptic scale or interannual time scale (Qiu 2002; Nonaka and Xie 2003; Joyce et al. 2009; Taguchi et al. 2009; Kelly et al. 2010; Kwon et al. 2010; Hotta and Nakamura 2011; Xu et al. 2011; Nakamura et al. 2015). Moreover, recent studies suggested an influence of the Asian coastal CAS on the downstream atmospheric hydrological cycle. For example, the North American west coastal atmospheric river (AR) activities and precipitation extremes were significantly modulated by the Asian CAS (Jiang and Deng 2011; Jiang et al. 2014). Compared to the above investigations, our understanding of the relationships between evaporation in the KOE region and the CAS and related atmospheric water vapor transport over the North Pacific basin is incomplete, especially on a subseasonal time scale. This makes up the core subject of the present study.

Atmospheric water vapor transport combines atmospheric circulation with vapor conditions in the air. The convergence of water vapor transport contributes to vapor gathering in the air and finally influences precipitation. Subseasonal anomalies in divergent water vapor transport contain multiscale features (Schneider et al. 2010; Newman et al. 2012; Knippertz et al. 2013; Liu et al. 2016). Thus we perform a scale decomposition on divergent water vapor transport. Namely, a variable is decomposed into its time mean and subseasonal and nonsubseasonal components. This method has been applied to understanding the dynamics of subseasonal modes, like the Madden–Julian oscillation (MJO) and the boreal summer intraseasonal oscillation over the Asian monsoon region (Zhang and Ling 2012; Ren et al. 2015).

Based on the scale decomposition, individual physical processes involving multiscale interactions of wind field and specific humidity are addressed in this study.

The paper is organized as follows: Section 2 describes data and methods, section 3 shows the features of subseasonal evaporation anomalies and the CAS, and section 4 investigates the associated anomalies in water vapor transport and its divergence or convergence. The accompanied precipitation anomalies are also addressed in section 4. Conclusions and discussion are given in section 5.

2. Data and analysis methods

a. Data

We use daily products from the objectively analyzed air–sea fluxes (OAFlux) project (Yu and Weller 2007) for the period of 1 January 1985–31 December 2015 with resolution of $1^\circ \times 1^\circ$. The goal of the OAFlux project is to develop an enhanced global analysis of air–sea latent heat, sensible heat, and net shortwave and net longwave radiation fluxes through blending various sources of meteorological variables. The OAFlux products have been used in understanding the issues related with air–sea interaction during the past decade (Yu et al. 2008; Joyce et al. 2009; Sugimoto and Hanawa 2011; Kumar et al. 2017; Liang and Yu 2016; Masunaga et al. 2016). The variables we used include latent heat flux (LHF), surface wind speed $W_s$ (at 10-m height from the sea surface), surface air specific humidity $q_{a2m}$ (at 2-m height from the sea surface), surface air temperature $T_{a2m}$ (at 2-m height from the sea surface), and sea surface temperature (SST). Evaporation is obtained by LHF divided by the latent heat of vaporization $L_e$.

Other atmospheric elements are from the European Centre for Medium-Range Weather Forecasts (ECMWF) interim reanalysis (ERA-Interim; Dee et al. 2011) daily analysis from 1 January 1979 to 1 December 2015 with resolution of $0.75^\circ \times 0.75^\circ$. Daily precipitation are the satellite–gauge blend of the Global Precipitation Climatology Project (GPCC) 1° daily precipitation (version 1.2), covering both ocean and land regions (Huffman et al. 2001). The time periods of GPCP precipitation are from October 1996 to the delayed present.

b. Method

The atmospheric water vapor transport $E$ and precipitation $P$ are connected via the moisture budget of an atmospheric column:
\[ \frac{\partial}{\partial t} \int_{p_s}^{p_t} q \, dp + \nabla \cdot \int_{p_s}^{p_t} q \mathbf{V} \, dp = E - P \quad \text{or} \quad (1) \]
\[ \frac{\partial \text{IWV}}{\partial t} + \nabla \cdot \text{IVT} = E - P, \quad \text{(2)} \]

where \( q \) is the specific humidity, \( g \) the standard gravity, \( p \) the pressure vertical coordinate, \( \mathbf{V} \) the horizontal wind vector \((u, v)\), and \( p_s \) and \( p_t \) the surface and 300-hPa pressure level, respectively. The term \( g^{-1} \int_{p_s}^{p_t} q \, dp \) or IWV on the left-hand side of (1) or (2) is called the total column-integrated water vapor (or precipitable water), and the term \( g^{-1} \int_{p_s}^{p_t} q \mathbf{V} \, dp \) or IVT is the integrated water vapor transport.

When a variable is decomposed into its time mean (denoted with an overbar), subseasonal (denoted with a prime), and nonsubseasonal components, the divergence of integrated water vapor transport on subseasonal time scale \( \nabla \cdot (\text{IVT}') \) can be expressed as

\[ \nabla \cdot \left( \frac{1}{g} \int_{p_s}^{p_t} q (\mathbf{V}' \cdot \mathbf{V}) \, dp \right) + \nabla \cdot \left( \frac{1}{g} \int_{p_s}^{p_t} (q' \mathbf{V}) \, dp \right) + \nabla \cdot \left( \frac{1}{g} \int_{p_s}^{p_t} (q' \mathbf{V}') \, dp \right) + \text{residue}, \quad \text{(3)} \]

where residue is the residual term that represents the contributions from nonlinear interaction between subseasonal and nonsubseasonal components. In (3), \( \nabla \cdot g^{-1} \int_{p_s}^{p_t} (q \mathbf{V})' \, dp \) is dominated by \( \nabla \cdot g^{-1} \int_{p_s}^{p_t} (q' \mathbf{V})' \, dp \)

\[ \nabla \cdot \left( \int_{p_s}^{p_t} (q' \mathbf{V})' \, dp \right) = \left( \text{A term}, \right. \nabla \cdot g^{-1} \int_{p_s}^{p_t} (q' \mathbf{V})' \, dp \quad (B \text{ term}), \quad \left. \nabla \cdot g^{-1} \int_{p_s}^{p_t} (q' \mathbf{V})' \, dp \quad (C \text{ term}) \right) \quad \text{according to our calculation.} \]

The residual term is much smaller and ignorable in this study. The A term is decomposed as follows:

\[ \nabla \cdot \left( \frac{1}{g} \int_{p_s}^{p_t} (q \mathbf{V})' \, dp \right) = \frac{1}{g} \int_{p_s}^{p_t} q \left( \frac{\partial u'}{\partial x} + \frac{\partial v'}{\partial y} \right) \, dp + \frac{1}{g} \int_{p_s}^{p_t} u \frac{\partial q'}{\partial x} \, dp + \frac{1}{g} \int_{p_s}^{p_t} v \frac{\partial q'}{\partial y} \, dp + \text{residue}_A, \quad \text{(4)} \]

where the A1, A2, and A3 terms contribute to \( \nabla \cdot (\text{IVT}') \), respectively, due to 1) divergence of subseasonal atmospheric circulation anomalies, 2) subseasonal zonal wind anomalies \( u' \) and zonal inhomogeneity of \( q \), and 3) subseasonal meridional wind anomalies \( v' \) and meridional inhomogeneity of \( q \). Similarly, the B term can be written as

\[ \nabla \cdot \left( \frac{1}{g} \int_{p_s}^{p_t} (q' \mathbf{V})' \, dp \right) = \frac{1}{g} \int_{p_s}^{p_t} q \left( \frac{\partial u'}{\partial x} + \frac{\partial v'}{\partial y} \right) \, dp + \frac{1}{g} \int_{p_s}^{p_t} u \frac{\partial q'}{\partial x} \, dp + \frac{1}{g} \int_{p_s}^{p_t} v \frac{\partial q'}{\partial y} \, dp + \text{residue}_B. \quad \text{(5)} \]

where the residue\(_A\) in (4) and residue\(_B\) in (5) represent the contributions from the surface component because \( p_s \) is also a function of \((x, y)\). Based on our calculation over the North Pacific region, the A1 and A3 terms dominate A, and B2 dominates B. The residue\(_A\) and residue\(_B\) are negligible terms in (4) and (5), respectively.

To extract the components of subseasonal variations, we use the method adopted by Krishnamurthy and Shukla (2000) and Ren et al. (2015). The procedure is as follows: 1) remove very high-frequency fluctuations in the daily data by applying a 7-day running mean, 2) deduce the daily climatology, and 3) remove the interannual signal by subtracting seasonal anomaly.

An empirical orthogonal function (EOF) analysis is performed on subseasonal evaporation anomaly \( E' \) in the North Pacific region \( 0^\circ \text{–} 55^\circ \text{N}, 120^\circ \text{E} \text{–} 120^\circ \text{W} \) during the winters (December, January, and February) of 1985–2014 (30 × 90 = 2700 days in total). Time-lagged regressions of subseasonal variables on the normalized first and second principal components of the EOF (PC1 and PC2) are calculated. In the following, “day 0” denotes simultaneous regression. Negative (positive) lag day indicates the variable leads PC1 (lags PC2) and is obtained by shifting backward (forward) the number of leading days.

3. Subseasonal variations of evaporation and the CAS

a. Subseasonal variations of evaporation

Before analyzing the feature of \( E' \), we first depict the climatology of evaporation and other related fields over
the North Pacific during winter in Fig. 1. The strongest subseasonal variance of evaporation is observed in the western North Pacific, especially in the KOE region (Grodsky et al. 2009; Kwon et al. 2010). The climatological IWV weakens toward the high latitudes. The meridional gradient of climatological IWV is large over the KOE region (Fig. 1b). There are three regions with large climatological precipitation and subseasonal variances in Fig. 1c: over the intertropical convergence zone (ITCZ), along storm track, and along the west coast of North America. The westerly wind at 850 hPa dominates over the midlatitude North Pacific (Fig. 1d). Big subseasonal variances of $q$ at 850 hPa are restricted over the band of 10°–35°N, where the equatorward side of the major SST front and westerly wind center is located.

The two leading modes of EOF analysis on $E^0$ account for 15% (EOF1) and 13% (EOF2) of the total subseasonal variance, respectively. The spatial pattern of EOF1 in Fig. 2a demonstrates a zonal wave train–like pattern in the band between 15° and 45°N. The pattern of EOF2 presents an east-west dipolar structure in the midlatitude basin (Fig. 2b). The positive lobe in Fig. 2b extends from the East Asian coastal water to the date line. The sock-shaped negative lobe is to the east of the positive one and near the west coast of North America. The power spectra of the whole PC1 and PC2 time series are calculated and plotted in Fig. 2c. The samples of the power spectra of PC1 and PC2 in four individual winters (1985, 1995, 2004, and 2014) are also shown in Fig. 2. Though there are large year-to-year changes in the dominant peaks, the PC1 and PC2 display a predominant periodicity of 10–20 and 10–30 days, respectively.

Figure 3 plots the regression fields of $E^0$ against PC1 and PC2, on days $-5$, $-3$, 0, and $+3$. The simultaneous regressions (day 0) assemble their individual EOF patterns. For EOF1, a small negative offshore center and a robust positive one to the southeast of Japan are seen on day $-5$ (Fig. 3a). After day $-5$, the zonal wave train pattern of $E^0$ is unequivocally located in the midlatitude North Pacific. The pattern propagates eastward and gets its peak on day 0. On day $+3$, it is vague but still can be seen. The negative $E^0$ that was along the East Asian coastal water on day
0 propagates to the open sea region on day +3. Meanwhile, a small positive $E'$ center occurs in the offshore region of East Asia between 30° and 50°N (Fig. 3d).

The east–west dipolar structure in EOF2 is robust during the days from −5 to +3 (Fig. 3, right). On day −5, the western lobe of the dipole with moderate positive amplitude of $E'$ is around the East Asian coastal water. It propagates eastward during the days from −3 to 0 with enhanced anomalies. Meanwhile, the eastern negative lobe gradually migrates toward the west coast of North America. On day +3, the positive lobe is near the date line with dramatically weakened values, while the negative one contracts eastward.

b. CAS and the corresponding atmospheric circulation anomalies

Figures 4 and 5 plot the lead–lag regression fields of $(T_{2m})$, subseasonal anomalies in geopotential height at 500 hPa, wind field at 850 hPa, and MSLP against PC1. Figures 4 and 5 exhibit the SH-related CAS activities and the corresponding atmospheric circulation anomalies associated with the EOF1 of $E'$. During the days from −7 to −5, the wave train–like cold and warm air-mass anomalies propagate toward the east coast of the Eurasian continent (Figs. 4a,b). After day −5, the air-mass wave train propagates into the oceanic region. On day 0, the air-mass wave train pattern occupies the mid-latitude North Pacific (Fig. 4d).

Accompanying the above cold air-mass activities are an extratropical cyclonic and anticyclonic wave train propagating southeastward toward the Asian coasts in the lower and middle troposphere (Fig. 5 and the contours in Fig. 4). The anomalous cyclonic circulation centered over Siberia area on days −5 and −3 suggests a decrease in the SH’s intensity (Figs. 5b,c). On day 0, the wave train is predominantly located over the midlatitude North Pacific. In particular, the regressed fields of MSLP and 850-hPa wind display an anomalous anticyclone occupying from Japan to the east of the date line between 30° and 55°N. Meanwhile, an anomalous cyclone is centered over southeastern China (Fig. 5d). The above pattern steers anomalously lower-level southerly flow to East Asia and its coastal waters (20°–45°N, 100°–140°E), contributing to the locally warmer air mass in Fig. 4d. After day +3, the extratropical cyclonic and anticyclonic wave train pattern over the North Pacific is not robust.

Figures 6 and 7 depict the regression fields of the atmospheric variables against PC2. Figures 6 and 7 display
the AL-related CAS activities and the corresponding atmospheric circulation anomalies associated with the EOF2 of $E'$. In general, the signals are mostly concentrated over the pan-North Pacific region. On the regressed ($T_{a2m}'$) fields, the robust signals are a cold center around Japan and a warm one over northeastern Asia during the days from $-7$ to $-3$ (Figs. 6a–c). After day $-3$, the abovementioned cold center moves eastward slightly and a new warm center appears over the eastern North Pacific and the west coast of North America. On the regressed fields of lower-level wind and MSLP anomalies, the gradual increase of cyclonic circulation anomaly over the North Pacific basin from days $-7$ to $-3$ indicates the persistent enhancement in the AL’s intensity (Figs. 7a–c). The AL’s circulation gets its peak on day 0. After day 0, the cyclonic circulation anomaly over the North Pacific basin fades away.

c. Subseasonal evaporation anomaly and the low-level atmospheric condition

Figure 8 plots longitude–time section of regressed $E'$, ($W_s'$), ($T_{a2m}'$), ($q_{a2m}'$), and SST anomalies averaged over 25°–40°N against PC1 and PC2. In general, an increased (a decreased) $E'$ is associated with an enhanced (a reduced) lower-level wind speed, and anomalously cold and dry
The role of subseasonal SST anomaly is limited, due to its small amplitude. For PC1, the decrease in SH’s intensity as shown in Figs. 5b,c suggests a weakened winter monsoon circulation and decreased CAS activities over East Asia and its coastal region, thus leading to the below-normal evaporation in the East Asian coastal region, and vice versa (Fig. 8, left). This is in agreement with previous studies (Cayan 1992; Bond and Cronin 2008). Because the cold and warm air masses and their corresponding extratropical circulation anomalies exhibit a wave train–like pattern (shown in Figs. 4 and 5), the induced $E'$ also displays the zonal wave train–like pattern over the midlatitude North Pacific during the days from −5 to +3.

For PC2 (Fig. 8, right), the intensified AL transports the anomalously cold and dry air mass toward the East Asian coastal waters and the anomalously warm and moist one toward the west coast of North America. Thus, the cold and dry air mass centered around Japan is persistent during the days from −6 to +3, and the warm and moist one appears later over the eastern North Pacific.
Pacific and the west coast of North America. The noticeably negative anomalies of \((T_{a2m})^0\), \((q_{a2m})^0\), and increased \((W_s)^0\) over the KOE region contribute to the enhanced evaporation locally. Meanwhile, the slightly positive anomalies of \((T_{a2m})^0\), \((q_{a2m})^0\), and decreased \((W_s)^0\) over the eastern North Pacific lead to the modestly decreased evaporation to the east of the date line. As a result, the east–west dipolar structure of \(E^0\) in the mid-latitude basin is produced in the EOF2 mode.

4. Associated anomalies in water vapor transport and precipitation

a. Terms \((IVT)^0\) and \(\nabla \cdot (IVT)^0\) associated with subseasonal variations of \(E^0\)

In this subsection, we reveal water vapor transport and its divergence anomalies associated with subseasonal variations of evaporation over the North Pacific, with emphasis on the simultaneous relationship. We will show that the regressed pattern of \(\nabla \cdot (IVT)^0\) is preliminarily decided by the pattern of \((u', v')\) and second by that of \(\partial q'/\partial x\). Meanwhile, we will demonstrate three major physical processes.

Figures 9a,b show regression fields of \((IVT)^0\) and \(\nabla \cdot (IVT)^0\) on day 0 against PC1 and PC2, respectively. We also plot the longitude section of regressed \(E^0\) and
\( \nabla \cdot (\text{IVT})' \) averaged over 15°–45°N on day 0 in Fig. 9c (PC1) and Fig. 9d (PC2). The patterns of \((\text{IVT})'\) in Figs. 9a,b bear a resemblance to the corresponding wind field anomalies at 850 hPa in Figs. 5d and 7d, respectively. This is because the water vapor transport is strongly dependent on the wind pattern in the lower level. For PC1, the zonal wave train pattern of \((\text{IVT})'\) builds up a similar wave train pattern of anomalous vapor source and sink. In particular, a prominent vapor sink is seen over the East Asian coastal waters between 20° and 40°N, where the negative \(E'\) occurs (Figs. 9a,c). For PC2, the intensified AL helps to build up an anomalous vapor source over the KOE region where the positive \(E'\) occurs and an anomalous sink over the northeastern North Pacific where the negative \(E'\) occurs (Figs. 9b,d).

Although Fig. 9 depicts the comprehensive image of \( \nabla \cdot (\text{IVT})' \) associated with the subseasonal evaporation changes, individual physical processes involving multiscale interactions of \(u, v,\) and \(q\) are still not clear. Therefore, the \(A, B,\) and \(C\) terms are computed according to (3)–(5). Figure 10 plots the regressed fields of the three terms on day 0. A comparison of Fig. 9 with Fig. 10 shows that the regressed pattern of \( \nabla \cdot (\text{IVT})' \) is preliminarily decided by that of the \(A\) term for both EEOFs. Meanwhile, the \(B\) term shows a certain modulation role. The details are as follows. The \(B\) term in Fig. 10 shows a zonal wave train pattern for both EEOFs. There is a spatial phase difference between the \(B\) and \(A\) terms, with individual centers of the \(B\) term leading those of \(A\). In addition, the positive and negative areas of the \(B\) term are smaller than those of the \(A\) term. Meanwhile, the \(B\) term’s amplitude is overall weaker than that of the \(A\) term. Hence, the \(B\) term has a role of modulating the spatial structure and moves each center eastward slightly on the \(A\) plus \(B\) image. The anomalies in the regressed fields of the \(C\) term are noticeably smaller than those of the \(A\) and \(B\) terms. They exhibit a small
damping effect on the A plus B image for both EOFs (Figs. 10e,f).

Figure 11 plots the regressed fields of $A_1$, $A_3$, and $B_2$ terms on day 0 against PC1 and PC2. The regressed fields of $\nabla \cdot (u', v')$, $u'$, and $q'$ at 850 hPa are also depicted in Fig. 11. The regressed patterns of $A_1$ and $A_3$ terms in Fig. 11 are similar to those of the $A$ term in Fig. 10, for both EOF1 and EOF2. This means the $A_1$ and $A_3$ terms contribute equally to the $A$ term. Furthermore, the regressed pattern of $A_1$ term is largely controlled by that of $\nabla \cdot (u', v')$ in the lower level. The physical processes are as follows: anomalous convergence flow can gather climatological water vapor and cause an anomalous vapor sink in the region with $\nabla \cdot (u', v') < 0$, while anomalous divergence flow can export climatological vapor to the surrounding region and lead to an anomalous vapor source in the region with $\nabla \cdot (u', v') > 0$. The regressed pattern of $A_3$ is mostly determined by that of $v'$. The reason is as follows: $\partial q/\partial y < 0$ is dominant over the midlatitude North Pacific, especially over the KOE region (Fig. 1b). Thus, an anomalous vapor source (sink) is produced by anomalously negative (positive) meridional wind transport.

During winter season, the westerly flow is prevalent over the North Pacific basin. That is why the regressed pattern of $B_2$ term is mainly dominated by $\partial q/\partial x$. The regressed $q'$ field against PC1 has a similar pattern to the EOF1 of $E'$, but they are opposite in phase over the midlatitude band. Therefore, the zonal advection of $q'$ by $\bar{u}$ for PC1 generates the wave train–like vapor source and sink, which is nearly a quadrature relation with the $A$ term. Consequently, the $B_2$ term promotes the vapor source/sink wave train to propagate eastward on the $A$ plus $B$ image. The physical process described above also exists for PC2.

b. Terms $E'$, $(T_{a2m})'$, and $\nabla \cdot (IVT)'$ over the KOE region

The above investigation demonstrated the anomalous pattern of CAS activities, $(IVT)'$, and $\nabla \cdot (IVT)'$ associated with subseasonal variations of $E$ over the North Pacific based on the EOF analysis. It is natural to ask whether the connections among them are also detectable in the daily observation. To verify again the robustness of the relationships among $E'$, CAS, and $\nabla \cdot (IVT)'$ over the KOE region, we identify the key regions of the subseasonal evaporation anomaly according to the significantly anomalous areas shown in Fig. 3c (EOF1) and Fig. 3g (EOF2). They are region E1 of 20°–40°N, 120°–140°E according to the EOF1 and region E2 of 25°–40°N, 145°–175°E according to the EOF2, respectively. Then, the daily indices, denoted as $E_1I$ (E2I), $T_{1I}$ (T2I), and $D_{1I}$ (D2I), are constructed by averaging the daily $E'$, $(T_{a2m})'$, and $\nabla \cdot (IVT)'$ averaged over 15°–45°N on day 0 for (c) PC1 and (d) PC2. The solid line for $E'$ and dotted line for $\nabla \cdot (IVT)'$. 

Fig. 9. Regression fields of $(IVT)'$ (kg m$^{-2}$ s$^{-1}$; vector, only values significant at 95% confidence level are shown) and $\nabla \cdot (IVT)'$ (kg m$^{-2}$ day$^{-1}$; shaded) on day 0 against (a) PC1 and (b) PC2. Longitude section of regressed $E'$ (kg m$^{-2}$ day$^{-1}$) and $\nabla \cdot (IVT)'$ averaged over 15°–45°N on day 0 for (c) PC1 and (d) PC2. The solid line for $E'$ and dotted line for $\nabla \cdot (IVT)'$. 


Figure 12 shows the scatter diagrams between E1I and T1I (E2I and T2I) and E1I and D1I (E2I and D2I). A simple quasi-linear relationship is shown between the E1I and T1I and E2I and T2I indices (Figs. 12a,c). That is, the enhanced $E_0$ in the two key regions is always accompanied by more active local CAS activities, and vice versa. These results demonstrate again the close conjunction of subseasonal evaporation anomalies with the CAS activities over the KOE region. The relationship between E1I and D1I and E2I and D2I also exhibits a quasi-linear feature. Namely, the positive $E'$ in the two key regions (especially in region E1) is associated with the anomalous local vapor source. These results are in agreement with those shown in Fig. 9.

c. Precipitation anomalies associated with subseasonal variations of $E'$

Figure 13 depicts regression patterns of GPCP precipitation anomaly $P'$ against PC1 and PC2 on days $-3$, $-1$, 0, and $+2$. We first analyze the precipitation anomaly associated with the EOF1. The regressed $P'$ against PC1 on day $-3$ shows a small area with positive $P'$ over southeastern China. On the same day, a large region with negative $P'$ extends zonally from Japan to 170°E (Fig. 13a). After day $-3$, the above positive $P'$ is intensified and moves eastward. After occupying East Asian coastal water between 30° and 45°N on day 0 (Fig. 13c), it continually propagates eastward on day $+2$ with almost invariable amplitude (Fig. 13d). The prior negative one on day $-3$ slightly moves

significantly small, indicating the independent variability of each other.

Figure 10. Regression fields of the (a),(b) $A$, (c),(d) $B$, and (e),(f) $C$ terms on day 0 against (left) PC1 and (right) PC2 (kg m$^{-2}$ day$^{-1}$). The contours for the $A$ and $B$ terms start from ±0.9 with an interval of 1.2 kg m$^{-2}$ day$^{-1}$ and for the $C$ term they start from ±0.3 with an interval of 0.3 kg m$^{-2}$ day$^{-1}$.
northeastward and decreases quickly. Another negative $P'$ exists around the 150°W meridian. It is robust between days 21 and 12 and moves slightly eastward.

The regressed $P'$ against PC2 displays an east positive–west negative dipolar structure over the midlatitude basin between 25° and 45°N (Fig. 13, right). On day −3, the negative lobe of the $P'$ dipole extends from eastern China through southern Japan to the west of the date line. Meanwhile, the positive lobe is located to the east of the date line centered at 150°W. The above dipolar pattern migrates eastward after day −3 and decays rapidly after day +2. In addition, a positive $P'$ is seen over the northwestern coast of North America. It is robust on days −1 and 0 (Figs. 13f,g). There are also anomalous rainfall signals to the south of 25°N, with a positive center to the east of the Philippines, a negative one between the date line and 150°W, and a small positive one over the tropical eastern Pacific on day 0 (Fig. 13g). These anomalous regions also migrate eastward slowly during the days from −3 to +2.

5. Conclusions and discussion

a. Conclusions

Strong subseasonal variations of oceanic evaporation are observed in the western North Pacific, especially around the KOE region during winter. This study addresses these variations and the connection with the CAS and related atmospheric water vapor transport on a subseasonal time scale by using the OAFlux and ERA-Interim daily data. Two dominant modes of oceanic evaporation anomaly are revealed by performing an EOF analysis on $E'$ in the region 0°–55°N, 120°E–120°W for 30 winters. The qualitative description of the connection is obtained by lead–lag regressing the
atmospheric variable and water vapor transport and its divergent fields in winter against the two principal components of the EOFs, respectively. Furthermore, three individual physical processes of $\nabla \cdot (IVT)'$ involving multiscale interactions of wind field and specific humidity are addressed based on the scale decomposition method.

The EOF analysis characterizes subseasonal variability of evaporation in the North Pacific during winter. The EOF1 exhibits a zonal wave train–like pattern of $E'$ in the band between 15° and 45°N. The EOF2 displays an east negative–west positive dipolar structure of $E'$ in the midlatitude basin. The PC1 and PC2 have a periodicity of about 10–20 and 10–30 days, respectively.

The regressed fields of $(T_{2m})'$, subseasonal anomalies in geopotential height at 500 hPa, wind field at 850 hPa, and MSLP reveal two types of CAS activities associated with the above two modes of $E'$. The EOF1 of $E'$ has a conjunction with the SH-related CAS activities, which feature a wave train–like cold and warm air mass and extratropical cyclonic and anticyclonic circulation anomalies in the mid-to-lower troposphere propagating from the Eurasian continent to the midlatitude North Pacific. The wave train pattern of the SH-related CAS has a crucial role in the generation of the EOF1 mode of $E'$. In particular, when the dry and cold (wet and warm) air mass passes the warm KOE region, the increased (decreased) air–sea temperature and moisture differences and intensified (reduced) wind speed lead to the above (below)-normal oceanic $E$.

The EOF2 of $E'$ displays a close relationship with the AL-related CAS activities, which feature a subseasonal change in the AL’s intensity. During the deepening episode of AL, the wind speed over the KOE region is robustly increased. Meanwhile, the intensified AL transports dramatically colder and drier air mass toward the KOE region and a slightly warmer and wetter one.
toward the west coast of North America. Consequently, the east–west dipolar structure of $E'$ in the midlatitude basin is produced, with larger amplitudes of $E'$ in the western lobe.

We investigate the simultaneous relationship between $(IVT)'$, $\nabla \cdot (IVT)'$, and the two EOF modes of $E'$. The pattern of $(IVT)'$ associated with the EOF1 bears a resemblance to the SH-related CAS’s wind field anomalies in the lower level. It builds up a similarly wave train pattern of anomalous vapor source and sink over the midlatitude North Pacific. Particularly, a prominent vapor sink is located over the East Asian coastal waters between 20° and 40°N, where the negative $E'$ occurs. The pattern of $(IVT)'$ associated with the EOF2 resembles the AL-related CAS’s wind field anomalies. The intensified AL helps to produce an anomalous vapor source over the KOE region where the positive $E'$ occurs and an anomalous sink over the northeastern North Pacific where the negative $E'$ occurs. The above connection of $E'$ with the CAS activities and $\nabla \cdot (IVT)'$ are verified again by scatter diagrams of the daily indices of $E'$, $(T_{2m})'$, and $\nabla \cdot (IVT)'$ over the KOE region.

Fig. 13. Regression fields of subseasonal anomaly of daily GPCP precipitation (mm day$^{-1}$) against (left) PC1 and (right) PC2 on days (a),(e) $-3$, (b),(f) $-1$, (c),(g) 0, and (d),(h) +2. The contours start from ±0.2 with interval of 0.4 mm day$^{-1}$. Areas exceeding the 90% significance levels are shaded.
The scale decomposition shows that the regressed pattern of $\nabla \cdot (\text{IVT})'$ is preliminarily decided by the pattern of $(u', v')$ and second by that of $\partial q' / \partial x$. Three dominant physical processes are involved. 1) An anomalous vapor sink is generated in the region with $\nabla \cdot (u', v') < 0$ through the anomalous convergence flow gathering climatological water vapor, and vice versa. 2) An anomalous vapor sink (source) is produced over the midlatitude North Pacific by anomalously southerly (northerly) wind transport because $\partial q' / \partial y < 0$ is dominant over the North Pacific. The above processes 1 and 2 contribute equally to $\nabla \cdot (\text{IVT})'$. 3) A zonal advection of $q'$ by $\pi$ generates the wave train–like vapor source and sink regions, which are nearly a quadrature relation with the vapor source and sink produced by processes 1 and 2. Thus, process 3 has a role of modulating the spatial structure and moves each center eastward slightly on the 1 plus 2 image.

The subseasonal evaporation anomalies and the associated atmospheric water vapor transport anomalies show hydroclimate effects over the pan–North Pacific region. On the precipitation aspect, the increased precipitation is observed over the North Pacific basin between 25° and 45°N associated with the EOF2. In particular, the enhanced precipitation is present along the northwestern coast of North America.

b. Discussion

Oceanic evaporation in the northwest Pacific region is the result of strong air–sea interaction. It contributes to the moisture budget of an atmospheric column. Thus, it can act as a vital link between air–sea interaction in the KOE region and basin-scale hydroclimate. The importance of air–sea interaction in the KOE region and the role in regional weather and climate variability cannot be overemphasized, just as many studies elaborated (Qiu 2002; Kwon et al. 2010; Kida et al. 2015; Nakamura et al. 2015). On interannual time scales, the dominant signals of surface sensible flux and evaporation over the Pacific basin are always associated with El Niño–Southern Oscillation (ENSO). The signals in the KOE region are only secondary and maybe contaminated by ENSO. However, the subseasonal variations of air–sea heat flux in the North Pacific basin are predominantly located in the KOE region, according to the observation. It is known that some high-impact events (e.g., persistent heavy rainfall or persistent disastrous freezing rain and/or snow) are often induced by anomalous atmospheric circulation and water vapor transport on a subseasonal time scale (Compo et al. 1999; Whitaker and Weickmann 2001; Wang et al. 2012; Frederiksen and Lin 2013; Ito et al. 2013; Jiang et al. 2014; Yu et al. 2015). Thus, examining surface sensible flux and evaporation on a subseasonal time scale is important to understand the key process in atmospheric hydrological cycle associated with those high-impact events.

For comparison purposes, surface latent heat flux from ERA-Interim twice-daily 12-h forecasts at 0000 and 1200 UTC is used in parallel. The dominant results are found to agree quite well with the OAFlux-based investigation. The differences are primarily in the amplitudes of the subseasonal variations of oceanic evaporation around the KOE region. ERA-Interim presents slightly stronger subseasonal standard deviations of evaporation near the East Asian coastal waters between 20° and 30°N and a slightly weaker one in the extension region of Kuroshio (to the east of Japan), compared with that of the OAFlux. The EOF1 and EOF2 of $E'$ from ERA-Interim bear resemblance to those of OAFlux’s, especially EOF1. The correlation coefficients between the PC1 (PC2) of OAFlux and those of ERA-Interim are above 0.90. The connected CAS activities present a small difference in the amplitudes around the KOE region between the two datasets. Future comparison of the two datasets should be the focus on this key region. It was reported that the OAFlux data have relatively unbiased value when compared with daily flux time series measurements at 107 locations (Yu et al. 2008). Meanwhile, OAFlux contains signals that are only in the oceanic region and are not contaminated by land area, which makes it a better fit for the purpose of this paper.

Acknowledgments. This work was jointly supported by National Natural Science Foundation of China (Grants 41330420, 41621005, 41675067, and 41675064), Jiangsu Province Science Foundation (Grant SBK2015020577), and Jiangsu Collaborative Innovation Center for Climate Change. ERA-Interim daily and monthly data are downloaded from http://apps.ecmwf.int/datasets/.

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