Characteristics of Summer Stationary Waves in the Northern Hemisphere

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ABSTRACT

Summer stationary waves in the Northern Hemisphere are separated by a midlatitude transition zone into the subtropical monsoon regime with a vertical phase reversal and the subarctic regime with a vertically uniform structure. The dynamics and maintenance mechanism of the subtropical stationary waves have been investigated in the context of monsoon circulation. Depicted in terms of streamfunction with 40-yr ECMWF Re-Analysis (ERA-40), the dynamic characteristics of stationary waves in the transition zone and the subarctic region are thus the focus of this study. The dynamics and maintenance mechanism of these waves were explored with the streamfunction budget and the velocity potential maintenance equations.

Stationary waves across the transition region consist of anticyclonic shear zones over the North Pacific and North Atlantic and a cyclonic shear zone in east Eurasia. These transition elements are linked to subtropical oceanic anticyclones and continental thermal lows. At high latitudes, a three-wave structure emerges with a weak central Eurasian trough aligned with two deep oceanic troughs. A longitudinal phase change occurs across the transition zone, but the direction of the east–west circulation associated with the transitional anticyclonic (cyclonic) zone is the same as that of the subtropical trough (high). This phase change is caused by the dynamics transition from the Sverdrup regime to the Rossby regime because of the increasing importance of relative vorticity advection. At high latitudes, relative vorticity advection becomes the dominant dynamic process in the upper atmosphere, but is negligible in the lower troposphere. This subarctic dynamic regime results in the vertically uniform structure of stationary waves. These waves are maintained by in situ diabatic heating (cooling) ahead of three subarctic troughs (ridges). Thus, the structure of the east–west circulation of subarctic stationary waves is opposite to that of subtropical stationary waves. These findings not only disclose more detailed structure and dynamics of summer stationary waves, but also provide a more complete basis to validate summer climate simulations and to search for the cause of interannual variation in summer climate.

1. Introduction

The Northern Hemispheric summer circulation is characterized by three monsoons: Asian, North American, and West African. In the upper troposphere, the first two monsoons are distinguished by the Tibetan and Mexican highs, which are separated by the North Pacific and North Atlantic oceanic troughs. The West African monsoon, which has a more complicated vertical structure than the other two monsoons, is reflected by the midtropospheric Saharan high, which is overlaid by the western part of the Asian monsoon high (Cook 1999; Chen 2003). Because the tropical summer circulation in the upper troposphere is well depicted by the streamfunction, Krishnamurti (1971) introduced the velocity potential to investigate the maintenance mechanism of this circulation. Using Krishnamurti’s 200-hPa wind fields, Holton and Colton (1972) examined the maintenance of the Tibetan high with a linearized streamfunction budget analysis. In their analysis, a parameterization of cumulus friction was included to balance streamfunction tendencies induced by vorticity advection and vortex stretching. More recently, depicting these monsoon circulations with summer stationary waves, the maintenance mechanisms of these monsoon highs were illustrated with differential heating through the east–west circulation and the interaction between the east–west circulation and these highs through the Sverdrup relationship (Chen 2003, 2005a). In the lower troposphere, continental thermal lows of these monsoons are juxtaposed with the North Pacific and North Atlantic anticyclones. It was suggested by Hoskins (1996) that the large size and
intensity of the oceanic anticyclones are caused by the monsoonal latent heat released over the neighboring continents to the east.

The dynamics and maintenance mechanism of stationary waves may be inferred from their vertical structure. Using the data analyzed on the National Meteorological Center (NMC) octagonal grid (Shuman and Hovemnale 1968), White (1982) examined the structure of Northern Hemispheric summertime stationary waves; these waves are characterized by a vertical phase reversal in the subtropics, and by a vertically uniform structure at high latitudes. Later, Ting (1994) investigated the maintenance mechanism of summertime stationary waves with a linearized general circulation model (GCM) and a 15-year simulation from the Geophysical Fluid Dynamics Laboratory (GFDL) GCM (Manabe et al. 1979). She showed that the vertical phase reversal of subtropical stationary waves is a response to global diabatic heating. On the other hand, Chen and Trenberth (1988) demonstrated that the equivalent barotropic structure of extratropical stationary waves is a result of the nonlinear interaction between thermally driven flow and orography.

The summer boreal-forest rainbelts are located along the latitude circle of 60°N. In search of the maintenance mechanism of these rainbelts, Yoon and Chen (2006) examined the structure of high-latitude stationary waves depicted by ERA-40 reanalyses (Uppala et al. 2005). Troughs emerge over the Aleutian and the Icelandic lows, which have been observed in previous studies (e.g., White 1982; Ting 1994), and another over Central Asia. Boreal-forest rainbelts are primarily formed by synoptic perturbations ahead of these subarctic troughs. Despite their vertically uniform structure, the favorable environment for perturbation genesis ahead of these troughs indicates that subarctic stationary waves differ dynamically from conventional barotropic waves. The longitudinal phase reversal of wintertime stationary waves at 30°N (Lau 1979) is caused by the wave dynamics transition from the subtropical Sverdrup regime to the middle to high latitude Rossby regime (Chen 2005b). This brings up the following question: How are summer stationary waves in subtropics transformed into the equivalent barotropic waves at high latitudes?

Numerous efforts have been made to seek the mechanisms for disastrous climate events, such as summer U.S. droughts–floods and the wet–dry conditions of the Asian monsoon. Nitta (1987) identified the Pacific–Japan pattern over the western Pacific. This anomalous climate pattern may be linked downstream to the North American summer drought—flood (e.g., Liu et al. 1998). The wet–dry condition of the Indian monsoon may be teleconnected with anomalous snowfall over the northwest Eurasian continent (e.g., Bamzai and Shukla 1999).

The interannual variation of the Saharan rainfall may be affected by North Atlantic sea surface temperature anomalies (e.g., Janowiak 1988). Despite these advances in climate research, a full understanding of the dynamics of summer stationary waves is still unclear, particularly in the extratropics. In view of salient features of summer stationary waves observed by previous studies, three aspects are of special interest to this study:

1) Winter stationary waves exhibit a clear-cut longitudinal phase reversal at 30°N (Lau 1979). What is the structure change of summer stationary waves from the subtropic to the subarctic region? Are there new features beyond White’s (1982) observations?

2) Accompanying this structure change, what is the transition of summer stationary wave dynamics from low to high latitudes? Is it only a transition from the Sverdrup to Rossby dynamics?

3) It was suggested by the linear Rossby wave theory that wave 5 is the most likely stationary wave configuration at 60°N (Hoskins and Karoly 1981). As observed in previous studies (e.g., Yoon and Chen 2006), boreal-forest rainbelts are maintained by synoptic disturbances ahead of subarctic troughs. The high-latitude stationary waves may not be barotropic in a conventional sense. What are the basic dynamics of these waves? Although White (1982) and Ting (1994) suggested that these subarctic stationary waves are generated–maintained by remote tropical forcing, what is their real generation/maintenance mechanism?

The ERA-40 reanalyses of northern summers [July–August (JJA)] for 1979–2002 are analyzed to explore these three aspects. This study is thus arranged in the following manner. The three-dimensional structure of summer stationary waves over the Northern Hemisphere is presented in section 2. To avoid possible mathematical redundancy, a brief summary of basic dynamics of stationary waves in different latitudinal zones, analytic formulation of the rotational–divergent circulation relationship of stationary waves, and the maintenance equation of divergent circulation are given in section 3. To keep this paper self-contained, a brief review of the dynamics of subtropical stationary waves related to monsoons is provided in section 4a. The dynamics of stationary waves in the transition region between the subtropical monsoon region and the high-latitude barotropic regime are illustrated diagnostically and analytically in section 4b. The search for the basic dynamics of the subarctic stationary waves and the exploration of their maintenance are presented in section 5. Finally, concluding remarks including a summary of new findings from this study are offered in section 6.
2. Summer stationary waves

a. Structure

The most conspicuous features of the upper-level circulation south of the midlatitude jets (Fig. 1a) are the Tibetan and Mexican highs juxtaposed with the North Pacific and North Atlantic oceanic troughs (Krishnamurti 1971; White 1982). The circulation north of these jets is characterized by upper-level troughs in the Bering Sea, eastern Canada, central Eurasia, and the Norwegian Sea, and ridges along the Rockies, northern Europe, and eastern Siberia. The central Eurasian trough, and the weak ridges between the Bering Sea trough and the North Pacific trough, and between the eastern Canada trough and the North Atlantic oceanic trough were not identified by previous studies (e.g., White 1982; Ting 1994). The lower-tropospheric circulation (Fig. 1b) is dominated by the North Pacific and North Atlantic anticyclones, which are juxtaposed with the continental thermal lows over Asia, North America, and North Africa. At high latitudes, major upper-level troughs and ridges have their counterparts in the lower troposphere, except in eastern Siberia where the continental thermal low of the Manchuria monsoon exists.

The spatial structure of summer (JJA) stationary waves is illustrated further by the eddy component of streamfunction ($\psi_E$) in Figs. 1c,d. In the upper troposphere, the high-latitude troughs in the Bering Sea and eastern Canada are meridionally juxtaposed with their counterparts in the subtropics over the North Pacific and the North Atlantic, respectively. The northern Eurasian and eastern Siberian highs, which are parallel with the Asian monsoon anticyclone, are separated by a col in central Eurasia. The ridge along the northern coast of Alaska is connected with the North American anticyclone. This almost in-phase meridional juxtaposition of summer stationary waves in the subtropics and at high latitudes is divided by a transition zone of opposite-phase anomalies. Winter stationary waves exhibit a longitudinal phase reversal at 30°N: a clear transition from the tropical Sverdrup regime to the middle–high latitude Rossby regime. The lack of such a phase change in summer stationary waves indicates that the dynamics of these waves north of jet streams may be different from winter stationary waves. Midlatitude stationary waves in the lower troposphere exhibit a puzzling eastward tilt with latitude (White 1982). This tilt is a result of connections between the Asian continental thermal low and the North Pacific oceanic trough, between the North Pacific high and the North American anticyclone, and between the North Atlantic high and the Northern European anticyclone (Fig. 1d).

The dynamics of summer stationary waves cannot be revealed immediately from their horizontal structure shown in Fig. 1. However, some light may be shed on these dynamics through the relationship between the vertical structure of these waves and the east–west circulation [$u_D$ and $\omega$ are zonal divergent wind speed and $p$ (pressure)–vertical motion, respectively] coupled with them (Fig. 2). In the subtropics (Fig. 2c), the upper-level monsoon anticyclone and the lower-level monsoon low are coupled by a counterclockwise east–west circulation, which is maintained by the east–west differential heating with cooling in the west and heating in the east of this anticyclone–low. In contrast, the upper-level oceanic trough and the lower-level oceanic anticyclone are coupled by a clockwise circulation, which is maintained by the east–west differential heating with cooling in the east and heating in the west (Chen 2003). Despite their vertically uniform structure, troughs (ridges) of high-latitude stationary waves (Fig. 2a) are spatially in quadrature with the associated counterclockwise (clockwise) east–west circulation (Fig. 2a). Even at high latitudes, it is still expected that these east–west circulations are maintained by east–west differential heating. As revealed from Figs. 2a,c, the relationships between the east–west circulation and the upper-level trough (or ridge) in the subtropics and at high latitudes are opposite.

Between the subtropic monsoon and high-latitude subarctic regimes exists the narrow transition zone characterized by the west–east-oriented Eurasian cyclonic shear zone ($L_T$, 60°–150°E), and the southwest–northeast-oriented North Pacific (150°E–150°W) and North Atlantic (80°W–0°), anticyclonic shear zone ($H_T$). A longitude–height cross section of [$\psi_E$, ($u_D$, $-\omega$)] 47.5°N of this transition zone is shown in Fig. 2b. Underneath these anticyclonic zones are the northeastern parts of the two oceanic anticyclones in the lower troposphere. It is evident that summer stationary waves in the subtropics yield their monsoon characteristics to the equivalent barotropic ones at middle–latitudes. This structure transition is reflected by an interesting feature; the direction of the east–west circulation associated with the transition anticyclonic (cyclonic) zone (Fig. 2b) is the same as that of the subtropical oceanic troughs (the Asian and Mexican monsoon highs; Fig. 2c). Nevertheless, this structure transition of stationary waves implies a dynamics transition from the subtropics to the subarctic region.

To facilitate our search for the basic dynamics of stationary waves, the relationship between $\psi_E$ and ($u_D$, $-\omega$) displayed in Fig. 2 are summarized by schematic diagrams shown in Fig. 3. Difference in the relationship between $\psi_E$ and ($u_D$, $-\omega$) in the subtropics, transition zone, and at high latitudes indicates that the dynamics of stationary waves in the latter region should belong to a different regime.
Based on the confluence theory of Namias and Clapp (1949), Blackmon et al. (1977) argued that the subtropical jet is maintained by the Coriolis acceleration of zonal flow caused by a direct cross-jet circulation upstream and the Coriolis deceleration by an indirect cross-jet circulation downstream. The speed of zonal flow is measured...
by the meridional gradient of streamfunction. Thus, any change in forcings maintaining the cross-jet circulation may result in modification of the streamfunction field embedded by this jet. The relationship between the subtropical jet and the cross-jet circulation offers an alternate perspective for the formation–maintenance mechanism of the transition zone. The high-latitude oceanic troughs are coupled with a counterclockwise east–west circulation (Fig. 2b), while the subtropical oceanic troughs are coupled with a clockwise one (Fig. 2c). The upward (U) and downward (D) branches of these east–west circulations are marked on the \( \psi_E \) (200 hPa) chart in Fig. 4a. Using these locations of upward and downward motion, and the midlatitude jets indicated by stippled areas, one can easily infer the structure of cross-jet circulations at different longitudes with the latitude–height cross sections of \( \psi_E \) in the upstream and downstream of the midlatitude jets (Figs. 4b–g) along lines b–g in Fig. 4a.

The Asian monsoon high and high-latitude anticyclones over Eurasia (marked as H) depicted with positive \( \psi_E \) anomalies are divided by an upward extension of the continental thermal low (marked as LT) denoted

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**FIG. 2.** Longitude–height cross section of eddy streamfunction \( \psi_E \) superimposed with the east–west circulation depicted by \( (u_{dp}, -\omega) \) at (a) 60°, (b) 47.5°, and (c) 30°N. The \( u_{dp} \) and \( \omega \) are the zonal divergent wind and \( p \) vertical motion \( (\frac{dp}{dt}) \), respectively. Positive values of \( \psi_E \) are stippled. Contour intervals of \( \psi_E \) are (a) \( 1.5 \times 10^7 \text{ m}^2 \text{ s}^{-1} \) and (b),(c) \( 10^7 \text{ m}^2 \text{ s}^{-1} \).
by negative $\psi_N$ anomalies and the midlatitude jet in Figs. 4b,c. The subtropical and high-latitude oceanic troughs are separated by the transition anticyclonic zones between 40° and 50°N, appearing to be upward and northward extensions of subtropic oceanic anticyclones (Figs. 4d–g). The distinction between the subtropical monsoon regime and the high-latitude equivalent barotropic regime is clearly revealed from these $\psi_N$ cross sections. As inferred from the east–west circulation coupled with the subtropical and high-latitude oceanic troughs, direct (indirect) secondary meridional circulations appear upstream (Figs. 4d,f) [downstream (Figs. 4e,g)] of the oceanic jets. The reverse relationship is found between the secondary meridional circulation and the continental jet. Jets thus couple with the northward (southward) branches of direct (indirect) secondary meridional circulations upstream (downstream). Because positive zonal wind is measured with the negative meridional gradient of streamfunction, midlatitude oceanic (continental) jets are located along the northern slope of anticyclonic (cycloic) transition zones. The relationship between jets and meridional circulation in coupling with the east–west circulation indicates the dynamic change across the transition zone.

### 3. Dynamics of stationary waves

The basic dynamics of summer stationary waves for different latitudinal zones were inferred in the schematic diagram in Fig. 3. The fundamental issue in this diagram is the transition of stationary wave dynamics from the subtropics to the subarctic region. The dynamics of winter stationary waves exhibit a clear transition at 30°N from the subtropical Sverdrup regime to the middle–high latitude Rossby regime. This dynamics transition was substantiated by the analytic formulation of the relationship between $\psi$ (streamfunction) and $\chi$ (velocity potential), in the Sverdrup and Rossby regimes. To avoid redundancy, this analytic $\psi$–$\chi$ relationship obtained by Chen (2005b) will be used to explore the transition of summer stationary dynamics suggested in Fig. 3.

The streamfunction budget equation for stationary waves may be written as

$$0 = \nabla^{-2} \left( -U \frac{\partial \xi}{\partial x} \right) + \nabla^{-2} (-u\beta) + \nabla^{-2} (-f \mathbf{V} \cdot \mathbf{V})$$

$$\psi_{A1} \quad \psi_{A2} \quad \psi_{\chi} \quad (1)$$

where $U$, $\mathbf{v}$, $\mathbf{V}$, $f$, and $\beta$ are mean zonal wind, meridional wind, wind vector, Coriolis parameter ($=2\Omega \sin\phi$; $\Omega$ and $\phi$ are the Earth’s rotational rate and latitude, respectively), and meridional gradient of planetary vorticity ($=2\Omega \cos\phi/a$; $a$ is the Earth’s radius), respectively. The $\psi_{A1}$, $\psi_{A2}$, and $\psi_{\chi}$ are streamfunction tendencies caused by relative vorticity advection by mean zonal flow, meridional advection of planetary vorticity, and vortex stretching, respectively. Note that $\psi_A = \psi_{A1} + \psi_{A2}$, and the interaction between rotational and divergent circulation is accomplished by vortex stretching. The velocity-potential maintenance equation is formed by the combination of the thermodynamic and continuity equations (Chen and Yen 1991a,b):

$$\chi = \nabla^{-2} \left[ \frac{\partial}{\partial p} \left( -\frac{1}{\sigma} \mathbf{V} \cdot \mathbf{T} \right) \right] + \nabla^{-2} \left[ \frac{\partial}{\partial p} \left( \frac{-1}{\sigma c_p} \dot{Q} \right) \right],$$

$$\chi_{HF} \quad \chi_{\dot{Q}} \quad (2)$$

where $\sigma$, $T$, $c_p$, and $\dot{Q}$ are static stability, temperature, specific heat with constant pressure, and diabatic heating, respectively. The divergent circulation portrayed by velocity potential is maintained by vertical differentiation of thermal advection and diabatic heating.

Based on the streamfunction budget analysis, one can express the dynamics of the Rossby and Sverdrup regimes by the following vorticity equations,

$$U \frac{\partial \xi}{\partial x} + u\beta = -f \mathbf{V} \cdot \mathbf{V}, \quad \text{Rossby regime} \quad (3a)$$

$$u\beta = -f \mathbf{V} \cdot \mathbf{V}, \quad \text{Sverdrup regime.} \quad (3b)$$
The dynamics of the subarctic regime will be presented in section 6.

The \( \psi - \chi \) relationship within these two dynamic regimes may be formulated in terms of normal mode expansions of the following variables:

\[
(\psi, X, X_H, X_Q) = \sum_{n=0}^{N} (\Psi^n, X^n, X^n_H, X^n_Q) e^{ikx},
\]

where the meridional structure of all variables in Eq. (4) are ignored, and \( k = n/a \), where \( n \) is the wavenumber.
along a latitudinal circle. The zonal-mean and wave components are denoted by $n = 0$ and $n = 1, \ldots, N$, respectively. Substituting Eq. (4) into Eqs. (3a) and (3b), one can find

$$\psi_{ana}^{Ra} = (1 - U_Z/C_\beta^n)^{-1}(-n \tan \phi \times \csc \phi) \chi^n e^{i\pi/2},$$

Rossby regime

$$\psi_{ana}^{S\ell} = (-n \tan \phi \times \csc \phi) \chi^n e^{i\pi/2},$$

Sverdrup regime

where $C^n_\beta = \beta/k^2$ and $k = 2\pi/L_x^u$ (where $L_x^u = 2\Omega a \cos \phi/n$). When $U_Z < C^n_\beta$ or $U_Z < 0$ (easterly), the $\psi^n - \chi^n$ relationship of stationary waves depicted by Eq. (5a) resembles that portrayed by Eq. (5b); that is, the Sverdrup regime. The spatially quadrature relationship between $\psi^n$ and $\chi^n$ is indicated by the factor of $e^{i\pi/2}$ in both Eqs. (5a) and (5b).

For the maintenance of divergent circulation, Eq. (2) may be expressed with the normal mode expansion as,

$$\chi^n = \chi^n_H + \chi^n_Q.$$  \hspace{1cm} (6a)

If the sensible heat advection is weak over some latitudinal zone, for example, subtropics, Eq. (6a) may be approximated by

$$\chi^n \approx \chi^n_Q.$$  \hspace{1cm} (6b)

Because the divergent circulation is primarily maintained by diabatic heating, the maintenance of rotational circulation by diabatic heating can be illustrated by combining Eqs. (5) and (6).

4. Dynamics of stationary waves south of the subarctic region

a. Subtropics

The monsoon circulation is characterized by the following features (Chen 2003):

1) A monsoon system consists of an upper-tropospheric anticyclone and a lower-tropospheric thermal low.
2) The monsoon anticyclone (low) is spatially in quadrature with centers of upper- (lower) level divergent circulation.
3) The monsoon divergent circulation is maintained by east–west differential heating with a heating (cooling) center located east (west) of the upper-level monsoon anticyclone.

The summer circulation in the subtropics consists of three monsoon systems (Asia, North America, and West Africa), which are well portrayed by stationary waves. Thus, the salient features of monsoon circulation are reflected by the basic dynamics of these waves. A brief overview of these dynamics will facilitate our search for the transition of stationary wave dynamics from the subtropics to the subarctic region. Because of the complicated vertical structure of the African monsoon, this overview will focus primarily on the other two monsoon systems.

Stationary waves depicted by the eddy streamfunction $\psi_E$ (Figs. 1c,d and 2c) in the subtropics are largely the result contributed by waves $(1 \sim 2)\psi^L$ (Krishnamurti 1971; Holton and Colton 1972). As revealed from our previous analysis (Chen 2003), 92% variance of the $\psi^L$ (30°N) longitude–height cross section is contributed by $\psi_{ana}^{SL}$ (30°N) derived from generated from Eq. (5b). Because the Asian monsoon circulation is well portrayed by $\psi_{ana}^{SL}$ (30°N), a vertical phase reversal of this monsoon is reflected by $\chi^L$, and the spatially quadrature relationship between this monsoon and its divergent circulation portrayed by $\chi^L$ is also indicated by the factor of $e^{i\pi/2}$. Following Eq. (2), it was found that $\chi^L_Q$ contributes over 90% variance to $\chi^L$ at 200 and 850 hPa. This is the expectation of Eq. (6b), $\chi^L \approx \chi^L_Q$. Based on this equation, it was shown that the divergent circulation of the Asian monsoon is maintained by the heating center over the western tropical Pacific–Asian monsoon region and the cooling center over North Africa.

The North American monsoon circulation is well depicted by waves $(2 \sim 8)\psi^M$. The mean-zonal easterlies ($U_Z < 0$) cover most of the tropics–subtropics, except a small strip of the troposphere south of 30°N and above 850 hPa (where $U_Z > 0$). Although the factor $(1 - U_Z/C^n_\beta^M)$ (30°N) for $n = 3$–8, [this wave regime is designated as $(\gamma^M)$], is negative in this strip of troposphere, positive values of this factor appear over the entire troposphere, even up to 60°N. Despite 92% of the $\psi^M$ (30°N) variance explained by $\psi_{ana}^{RM}$ (30°N), actually $\psi_{ana}^{SL}$ (30°N) contributes 81% of this variance. This fact leads us to conclude that the dynamics of subtropical stationary waves in the wave 2–8 regime belong to the Sverdrup regime. A vertical phase reversal of $\psi_{ana}^{SL} + \psi_{ana}^{RM}$ (30°N) appears over the longitudes (120°–60°W) of North America, and a spatial quadrature relationship exists between $\psi_{ana}^{SL} + \psi_{ana}^{RM}$ (30°N) and the divergent circulation $\chi^M$ across North America. Based on the analysis of Eq. (2), over 90% variance of $\chi^M$ is explained by $\chi^M_Q$ at 200 and 850 hPa. As expected by Eq. (6b), the divergent circulation of the North American monsoon is maintained by the warming center over the Caribbean Sea and the cooling center over the eastern tropical Pacific.

This brief overview for the dynamics of summer stationary waves in the subtropics may be summarized by the longitude–height cross section of $(\psi_{ana}^{SL} + \psi_{ana}^{RM})$.
superimposed with the east–west circulation \((u_D, -\omega)\) of waves 1–8 at 30°N shown in Fig. 5. Because close to 90% variance of \(\psi_E(30^\circ N)\) is explained by \((\psi_{S1}^{\text{ana}} + \psi_{Rm}^{\text{ana}})(30^\circ N)\), the salient features of monsoon revealed from \(\psi_E(30^\circ N)\) in Fig. 2c clearly emerge from \((\psi_{S1}^{\text{ana}} + \psi_{Rm}^{\text{ana}})(30^\circ N)\). Thus the basic dynamics of these summer stationary waves belong to the Sverdrup regime.

b. Transition region

The structure of stationary waves in the transition region (Fig. 2b) is characterized by:

1) The transition zone consisting of HTs and LT are embedded in ultra-long-scale stationary waves.
2) The anticyclonic transition zones (HTs) over the two oceans are coupled with the clockwise east–west circulation just like the subtropical oceanic troughs, but opposite to those coupled with the high latitude oceanic troughs. This relationship between the east–west circulation of anticyclonic transition zones and the subtropic–subarctic oceanic troughs are opposite the relationship between the east–west circulation of the cyclonic transition zone and the Tibetan high—the east Siberian ridge.

The dynamic implication of these features is examined. Because jets are located over the north (south) side of the cyclonic (anticyclonic) transition zone (Fig. 4a), the contribution of \(\psi_{A1}\) to the maintenance of stationary waves in the transition region may not be negligible. This concern is clarified by latitudinal variance distributions of \(\psi_{A1}\) and \(\psi_{A2}\) compared to \(\psi_{A1}\) (Fig. 6). Here, \(\psi_{A1}\) (850 hPa) is always negligible compared to \(\psi_{A2}\) (850 hPa), particularly in low latitudes. As inferred from this magnitude contrast, the dynamics of stationary waves in the lower troposphere belong to the Sverdrup regime.

For the upper troposphere, the magnitude of \(\psi_{A2}\) (200 hPa) diminishes north of 40°N, while that of \(\psi_{A1}\) (200 hPa) grows. This magnitude contrast between these two dynamic processes indicates the transition of the subtropical Sverdrup dynamics into the Rossby dynamics in the transition zone.

Inference of the dynamics change of stationary waves across the transition zone made above should be substantiated. The east–west circulations associated with transition highs (Fig. 2b) and the subtropical oceanic troughs (Fig. 2c) exhibit a similar structure. The \(\chi\) field does not exhibit a longitudinal phase change across the transition zone. Thus, dynamics of stationary waves in the subtropics—lower troposphere belong to the Sverdrup regime [Eq. (3b)], while those in the transition zone/higher latitudes fit the Rossby regime [Eq. (3a)]. The \(\psi''-\chi''\) relationship can be portrayed by Eq. (5b) for the former region and by Eq. (5a) for the latter region. As long as the factor \((1 - U_z/C_{m}^b) > 0\) and \(\chi''\) have no latitudinal sign change across the transition region, \(\psi^{\text{S}}_\text{ana}\) of the subtropics—lower troposphere exhibits a sign opposite to \(\psi^{\text{R}}\) over the transition—higher latitudes. A sign difference between these two solutions explains the latitudinal phase change of \(\psi''\) across the transition zone.

The \(\psi''-\chi''\) relationship across the transition region leads us to question what causes the latitudinal phase change of stationary waves between the subtropics and transition region. The difference between Eqs. (5a) and (5b) is:

\[
\psi^{\text{S}}_\text{ana} = [(U_z/C_{m}^u)(1 - U_z/C_{m}^b)](-n \tan \phi \times \csc \phi) \chi'' e^{i\pi/2},
\]

the contribution of relative vorticity advection \([-U_z(\partial \chi'/\partial x)]\) to \(\psi''\). Longitudinal–height cross sections of \(\psi^{S}_\text{ana}\) of
the subarctic regime, \( \psi^S \) of the Sverdrup regime, and \( \psi^R \) of the Rossby regime are shown in Fig. 7. The bias variance of \( \psi^R \) (47.5°N; Fig. 7c) against \( \psi_E \) (47.5°N; Fig. 2b) is about 8%. The contrast among these three cross sections reveals that the transition highs (HTs) and low (LT) are primarily contributed to by \( \psi^A \) (Fig. 7a). The positive (negative) relative vorticity advection ahead of midlatitude synoptic-scale troughs (ridges) is coupled with the counterclockwise (clockwise) east–west circulation (Holton 2004). It is not surprising that the structure of the east–west circulation associated with transition highs (lows) are the same as that with the subtropical oceanic trough (continental anticyclone). It is evident that the increasing importance of relative vorticity advection across the transition zone not only preserves the structure of the subtropical east–west circulation, but also leads to a longitudinal phase change of subtropical stationary waves.

5. Dynamics of stationary waves at high latitudes

a. Dynamics

The upward branches of the east–west circulations ahead of high-latitude troughs establish the preferred genesis environment of synoptic disturbances that maintain boreal-forest rainbelts along 60°N (Yoon and Chen 2006). As inferred from the quadrature relationship between stationary waves at high latitudes and associated east–west circulations, the vertically uniform structure of these waves does not imply that they are barotropic waves in a conventional sense. The basic dynamics of these stationary waves may be inferred from variance ratios of dynamic processes included in the \( \psi_E \) budget shown in Figs. 6a,b. Because the difference of \( f \) (North Pole) – \( f \) (60°N) is only approximately \( O(10^{-1}) \) of \( f \) itself; thus, \( -\nu \beta \) likely plays a minor role in total vorticity advection. This assessment is supported by a comparison

Fig. 6. Latitudinal distribution of variance ratios between \( \psi_{A1} \) (or \( \psi_{A2}, \psi_A \)) and \( \psi_x \) at (a) 200 and (b) 850 hPa, and vertical distributions of these variance ratios at (c) 47.5°N and (d) 60°N. Ratios of all wave components included are represented by solid thick–dashed thick lines, while those of low-wavenumber components are portrayed by lightly shaded solid–dashed lines.
of the following variance ratios (Fig. 6): $\text{Var}(\psi_{A1})/\text{Var}(\psi_A) \gg \text{Var}(\psi_{A2})/\text{Var}(\psi_A)$ in the high-latitude upper troposphere, although $\psi_{A2}$ is not completely negligible.

Dynamic properties of high-latitude stationary waves may be revealed further from the $\psi_E$ budget analysis in terms of a three-dimensional perspective. Longitude–height cross sections of $(\psi_{A1}, \psi_E)$ and $(\psi_{A2}, \psi_E)$ at 60°N shown in Figs. 8a,b, respectively, exhibit a vertically uniform structure: $\psi_{A1}$ (60°N) has its maxima–minima in the upper troposphere, while $\psi_{A2}$ (60°N) has its maxima–minima in the lower troposphere. This contrast is consistent with variance ratios for various terms in the $\psi$ budget [Eq. (1)] shown in Fig. 8. For barotropic waves, $\psi_{A1}$ and $\psi_{A2}$ are counterbalanced not only to maintain their stationarity, but also to satisfy the conservation of total vorticity. Actually, this is not the case for high-latitude stationary waves. The cancellation between $\psi_{A1}$ (60°N) and $\psi_{A2}$ (60°N) results in a vertical phase reversal of $\psi_A$ (60°N; Fig. 9a): $\psi_A$ (60°N) exhibits positive (negative) tendency east (west) of $\psi_E$ (60°N).
anomalies in the upper troposphere and west (east) in the lower troposphere.

The longitude–height cross section of $cx$ (60°N; Fig. 9b) exhibits a spatial polarity opposite to $cA$ (60°N). Because magnitudes of $cA$ (60°N) and $cx$ (60°N) are comparable, stationary waves are maintained by the counterbalance between these two dynamic processes. We should also explore whether this counterbalance is true north of 60°N. Because the circumference of a latitude decreases toward the poles, the National Meteorological Center octagonal grid (Shuman and Hovermale 1968) was introduced by Chen et al. (2008) to improve the depiction of meteorological variables in the polar regions. This approach is adopted by the present study.

At 200 and 850 hPa, $cA$ and $cx$ shown in Fig. 10 reveal the following features:

1) The $cE$ (200 hPa) and $cE$ (850 hPa) are out of phase (shown later), and so are $cA$ (200 hPa; Fig. 10b) and $cA$ (850 hPa; Fig. 10c).
2) The $cA$ (Figs. 10a,c) and $cE$ are comparable in their magnitudes and opposite in their spatial polarity at both levels. Because of feature 1, $cA$ (200 hPa) and $cA$ (850 hPa) are out of phase, as well.

3) Because $cE$ is spatially in quadrature with $cA$ and $cE$, it is expected that $cE$ is spatially in quadrature with $cE$, and so is $cE$ with the east–west circulation.

It is revealed from this $cE$ budget that subarctic stationary waves are maintained by the following dynamic processes: $cA$ (200 hPa) [$\approx cA$ (200 hPa)] = $-cE$ (200 hPa) and $cA$ (850 hPa) [$\approx cA$ (850 hPa)] = $-cE$ (850 hPa). The dynamics of high-latitude stationary waves in the upper troposphere reflected by the $cE$ budget analysis may be expressed by a simplified vorticity equation:

$$U \frac{\partial c^2}{\partial x} \approx -f \mathbf{V} \cdot \mathbf{V},$$

which is named as the subarctic dynamics. In the lower troposphere, stationary waves at high latitudes are still dictated by Sverdrup dynamics.

b. Analytic solution

High-latitude stationary waves are characterized by two basic properties: 1) vertically uniform structure and 2) the spatial quadrature relationship between stationary waves and associated east–west circulations. The dynamic
underpinning of these two properties inferred from the \( \psi_E \) budget analysis can be substantiated by analytic solutions of Eqs. (8) and (3b), given in Eqs. (7) and (5b), respectively: upper-troposphere \( \psi_{A1}^{\text{ana}} \) is primarily contributed by \( \psi_{A1} \), and lower-troposphere \( \psi_{A2}^{\text{ana}} \) by \( \psi_{A2} \). The spatial quadrature relationships between \( \psi^n \) and \( \chi^n \), and between \( \psi_E \) and the associated east–west circulations, is explained by the \( e^{i \pi/2} \) factor.

The high-latitude stationary waves depicted by \( \psi_E \) (60°N) in Fig. 2a are primarily formed by the low wave-number regime: more than 95% variance of \( \psi_E \) (60°N) is explained by waves 1–8. Values of \( (U_z/C_b^n)(1 - U_z/C_b^n)^{-1} \) for this wave regime at 60°N (Fig. 11) are nearly negative. Removing the common factor from Eqs. (7) and (5b), we may approximate the \( \psi^n \) relationship as follows:

\[
\psi^n(\text{upper troposphere}) \sim -\chi^n(\text{upper troposphere}),
\]

\[
\psi^n(\text{lower troposphere}) \sim +\chi^n(\text{lower troposphere}).
\]

As long as \( \chi^n \) in the upper and lower troposphere are out of phase, \( \psi^n \) should be in phase. The vertically uniform structure of \( \psi_E \) is a result of predominant \( \psi_{A1} \) in the upper troposphere and \( \psi_{A2} \) in the lower troposphere.

Let us validate the \( \psi_E \) vertical structure in terms of analytic solutions, Eqs. (5a), (5b), (6a), (6b), (7). With reanalysis \( \chi^n \) as forcing, solutions \( \psi_{A1}^{\text{ana}}, \psi_{A2}^{\text{ana}}, \text{ and } \psi_{R1}^{\text{ana}} \) at 200 hPa and \( \psi_{R2}^{\text{ana}} \) at 850 hPa are shown in Fig. 12. Salient features of these solutions include:

1) \( \psi_{A1} \) and \( \psi_{A2} \) are almost spatially out of phase at high latitudes (Fig. 8). As expected, \( \psi_{A1}^{\text{ana}} \) (200 hPa) and \( \psi_{A2}^{\text{ana}} \) (200 hPa) exhibit an opposite polarity north of 40°N. In this region, \( \psi_{A1} > \psi_{A2} \). Thus, spatial structure of \( \psi_{R2}^{\text{ana}} \) resembles that of \( \psi_{A1}^{\text{ana}} \).

2) In the region south of 40°N (Fig. 12b) and the lower troposphere over the Northern Hemisphere (Fig. 12e), one can obtain the approximation: \( \psi_{R2}^{\text{ana}} \approx \psi_{A2}^{\text{ana}} \).

3) To further validate the analytic solution of stationary waves, the longitude–height cross section of \( \psi_{R2}^{\text{ana}} \) at 60°N is shown in Fig. 12d. The error of \( \psi_{R2}^{\text{ana}} \) at 60°N is approximately 8% against \( \psi_E \) (60°N) in Fig. 2a.

In view of such a small error in depicting stationary waves by an analytic solution, this indicates that the dynamics of high-latitude stationary waves revealed from our diagnosis are accurate. The contrast among \( \psi_{A1}^{\text{ana}}, \psi_{A2}^{\text{ana}}, \text{ and } \psi_{R1}^{\text{ana}} \) at 200 hPa confirms once again that transition anticyclonic and cyclonic shear zones are mainly contributed to by \( \psi_{A1}^{\text{ana}} \).
c. Maintenance

According to the linear Rossby wave theory (Hoskins and Karoly 1981), the preferred stationary wave configuration is wavenumber 5 at 60°N. The high-latitude circulation structure, which is characterized by three troughs–ridges, does not agree with the preference of this wave theory. In addition, the subarctic stationary waves are not only separated from the subtropical ones by a transition zone, but also coupled with the east–west circulations whose structures are opposite with the latter waves. The origin of subarctic stationary waves seems to be independent of forcings at lower latitudes. As inferred from the streamfunction budget equation, these subarctic waves may be maintained by local forcings through vortex stretching (i.e., vorticity source), established by the divergent circulation.

If stationary waves can be maintained by the divergent circulation through the $\chi - \psi$ interaction, we should also explore whether the divergent circulation is maintained by diabatic heating. It was pointed out by Randall et al. (1998) that climate modeling in the Arctic region is complicated by radiative processes, cloud physics, and the air–ice–ocean interaction. Including these physical
processes in the diagnostic estimate of diabatic heating in this region is a formidable task. In addition to these complicated physical processes, it was shown by previous studies (e.g., Chen and Baker 1986; Nigam et al. 2000) that the diabatic heating estimation is sensitive not only to the diagnostic scheme used, but also reanalyzed data generated by different assimilation systems. To reduce these sensitive issues, and to make the effect of diabatic heating on the divergent circulation tractable, the indirect approach of Eq. (2) is applied to generate velocity potential contributed by diabatic heating. Our diagnostic analysis shows that \( \text{Var}(\mathbf{x}_Q)/\text{Var}(\mathbf{x}) \approx 95\% \) at 200 hPa and \( \approx 90\% \) at 850 hPa north of 50\(^{\circ}\)N; \( \mathbf{x}_Q \) is a good approximation at high latitudes. As inferred from upward branches of east–west circulations coupled with three subarctic troughs (Fig. 2b), three upper (lower)-level divergent (convergent) centers ahead of these troughs are expected. This expectation is confirmed by \( \chi_E \) (\( \mathbf{V}_D \): 200 hPa) and \( \chi_E \) (\( \mathbf{V}_D \): 850 hPa) shown in Figs. 13a,b, respectively, [convergent (C) and divergent (D) centers are marked]. Next, comparable magnitude and similar spatial structure between \( \chi_E \) (Figs. 13a,b) and \( \chi_{Q,E} \) (Figs. 13c,d) at 200 and 850 hPa indicate that the high-latitude divergent circulation is primarily maintained by vertical differential heating [Eq. (2)]. This argument is further confirmed by the coincidence between centers of \( \chi_{Q,E} \) and vertical differential heating (\( \partial \mathbf{Q}/\partial p \)) of opposite sign at 200 and 850 hPa.

Using the approximation \( \mathbf{x} \approx \chi_Q \), one may relate diabatic heating to the maintenance of stationary waves through the vortex stretching term in the vorticity equation,

- upper troposphere \( \mathbf{U}_E(\partial \mathbf{Q}/\partial x) \approx -f \nabla^2 \chi_Q \)
- lower troposphere \( \mathbf{U}_D(\partial \mathbf{Q}/\partial x) \approx -f \nabla^2 \chi_Q \).

The three upper-level troughs, divergent–convergent centers, and centers of \( \partial \mathbf{Q}/\partial p \) are collocated along 60\(^{\circ}\)N. The coincidence of these circulation centers supports our argument that subarctic stationary waves are generated by in situ forcings.

6. Concluding remarks

Two distinct regimes of summer stationary waves in the Northern Hemisphere were identified by White (1982): a subtropical monsoon regime with a vertical phase reversal and a subarctic regime with a vertically uniform (equivalent barotropic) structure. This finding leads to the following questions:

1) What is the structure of the transition between the two wave regimes?
2) What are the dynamics of stationary waves over the transition region?
3) What are the basic dynamics and mechanisms of subarctic stationary waves that make them vertically uniform?

The summer circulation is well portrayed by streamfunction (\( \psi \)) and velocity potential (\( \chi \)) (Krishnamurti 1971). The \( \chi-\psi \) relationship and interaction used by Chen (2005b) in exploring characteristics of winter stationary waves were applied to answer the questions about summer stationary waves posed above using the ERA-40 reanalyses (Uppala et al. 2005).

Major findings in this study may be summarized as follows:

1) Structure: The narrow transition region in the upper troposphere is characterized by two anticyclonic shear zones meridionally juxtaposed with the subarctic and subtropical oceanic troughs, and a cyclonic shear zone sandwiched in between the Asian monsoon anticyclone and the northwestern Eurasian and eastern Siberian ridges. The longitudinal–height cross section of eddy streamfunction shows that stationary waves in the transition zone consist of ultra-long-waves embedded by shortwaves. In fact, the transition anticyclonic and one cyclonic shear zones belong to some of these shortwaves. As revealed from latitude–height cross sections upstream and downstream of mid-latitude jets, the transition anticyclonic, and cyclonic
FIG. 12. Analytic solutions of stationary waves obtained from the vorticity equation including (a) only zonal advection of relative vorticity $\psi^A_{\text{ana}}$ (200 hPa), (b) only meridional advection of planetary vorticity $\psi^S_{\text{ana}}$ (200 hPa), (c) advection of total vorticity $\psi^R_{\text{ana}}$ (200 hPa) \([=\psi^A_{\text{ana}}(200 \text{ hPa}) + \psi^S_{\text{ana}}(200 \text{ hPa})]\), (d) longitudinal–height cross section of $\psi^A_{\text{ana}}$ (60°N), and (e) advection of total vorticity $\psi^R_{\text{ana}}$ (850 hPa). Areas with positive values of all variables are shaded; the contour intervals used in (a)–(d) and (e) are $1.5 \times 10^7$ m$^2$ s$^{-1}$ and $7.5 \times 10^6$ m$^2$ s$^{-1}$, respectively.
shear zones are linked to subtropical oceanic anticyclones and continental thermal lows in the lower troposphere, respectively. At high latitudes, the summer circulation exhibits a three-wave structure. One of the three subarctic troughs over central Eurasia is weaker than the other two over the oceans. This trough is related to the maintenance of the central Eurasian boreal-forest rainbelt.

2) East–west circulation and dynamics: The dynamics and maintenance mechanism of stationary waves were inferred from the spatial relationship between stationary waves and associated east–west circulations. Being spatially in quadrature with the transition shear element, the east–west circulation coupled with the transition anticyclonic (cyclonic) shear zone at mid-latitudes exhibits a direction as that of the east–west circulation coupled with the subtropical oceanic troughs (monsoon anticyclone). Because the midlatitude jets are located north (south) of the transition anticyclonic (cyclonic) shear zone, the cross-jet circulations both

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**FIG. 13.** Polar stereographic charts of the divergent circulation and its contribution from vertical differentiation of diabatic heating: (a) \((\chi_E, V_D)(200 \text{ hPa})\), (b) \((\chi_E, V_D)(850 \text{ hPa})\), (c) \((\chi_{q_E}, \partial \bar{Q}/\partial p)(200 \text{ hPa})\), and (d) \((\chi_{q_E}, \partial \bar{Q}/\partial p)(850 \text{ hPa})\). Magnitude of divergent wind vector \(V_D\) is measured by the arrow under the label of (c). Areas with positive values of \(\chi_E\) are shaded in (a),(b), values of \(\partial \bar{Q}/\partial p\) are shaded according to scale shown in the lower right corner of (c),(d). The contour interval of \(\chi_E\) and \(\chi_{q_E}\) is \(10^6 \text{ m}^2 \text{ s}^{-1}\).
upstream and downstream of a jet are consistent with the jet maintenance mechanism illustrated by Blackmon et al. (1977). As revealed from the streamfunction budget across the narrow transition zone, the increasing importance of horizontal relative vorticity advection transfers the dynamics of stationary waves from the subtropical Sverdrup regime to the midlatitude Rossby regime. In the subarctic region, stationary waves are spatially in quadrature with associated east–west circulations, but their relationship is opposite to that in the transition zone; namely, those coupled with troughs (ridges) have a counterclockwise (clockwise) circulation. Dynamically, the planetary vorticity (i.e., Coriolis parameter) becomes more uniform latitudinally in the subarctic region. Thus, the vortex stretching at high latitudes is counterbalanced by horizontal relative vorticity advection in the upper troposphere and by meridional planetary vorticity advection in the lower troposphere. These special high-latitude dynamics result in the vertically uniform structure of subarctic stationary waves. The east–west circulations coupled with the subarctic stationary waves are maintained by vertical–differential diabatic heating centers ahead of the three troughs—ridges.

Through better understanding of the structure and dynamics of summer stationary waves, we can develop more realistic global climate models and search for the cause of interannual variation in the summer climate. Because the intensity of atmospheric circulation is much stronger in winter, validation of climate simulations by global climate models are often focused on this season [e.g. the Climate Systems Model Special Issue (June 1998) in Journal of Climate was devoted to validate NCAR global climate model simulations]. Because summer is the major growing season and the peak season for human activity, it is equally or more important for global climate models to properly simulate the summer climate with various salient features of summer stationary waves; transition from the subtropical Sverdrup regime to the Rossby wave regime across a narrow transition zone to the subarctic regime.

Summer rainfall at high latitudes provides not only a water supply to boreal forests, but also freshwater to the thermohaline circulation in the Atlantic Ocean through the northbound river discharge (Chen et al. 1994). The boreal-forest rainbelts are essentially maintained by the northbound river discharge (Chen et al. 1994). The thermohaline circulation in the Atlantic Ocean through water supply to boreal forests, but also freshwater to the subarctic regime.

Any change in hydrological processes over the northern landmasses of Eurasia and North America may alter freshwater outflow into the Arctic Sea and impact the global climate through the thermohaline circulation in the oceans around the world (e.g., Delworth et al. 1993; Zhang et al. 1993). Therefore, effort should be devoted to explore the interannual variation of subarctic stationary waves.

It was suggested by Lau and Peng (1992) that summer teleconnection patterns in the North Pacific may be related to the U.S. summer drought/flood through two possible mechanisms: 1) a Rossby wave train generated by the tropical forcing located in the eastern Pacific (Trenberth et al. 1988; Palmer and Brankovic 1989) and 2) an arching wave train emanating from the subtropical Pacific across the Aleutian into North America (Nitta 1987, 1989). Two strong westerly zones appear ahead of the subarctic oceanic trough and the subtropical oceanic trough in the North Pacific (Fig. 1a). The two North Pacific summer teleconnection patterns are essentially shortwave trains of perturbations along these two strong westerly zones. Mechanisms generating these two summer teleconnection patterns were suggested well ahead of a full understanding of the structure and dynamics of summer stationary waves. New findings of these waves’ characteristics may facilitate the confirmation of summer teleconnection theories or the future search for new mechanisms.

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