Atmospheric Circulation Associated with Anomalous Variations in North Pacific Wintertime Blocking

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(Manuscript received 1 April 2003, in final form 14 August 2003)

ABSTRACT

Atmospheric circulation associated with anomalous variation of North Pacific blocking during the northern winter (December to February) is described and examined using the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis data from 1948/49 to 1999/2000. The divergent wind and pressure vertical velocity are employed for the identification of atmospheric circulation cells. There are several atmospheric cells over the North Pacific associated with an anomalous blocking situation during winter. They are the zonal Walker cell along the equator (ZWC), the regional Hadley cell in the western Pacific (WHC), the regional Hadley cell in the eastern Pacific (EHC), the regional Ferrel cell over the eastern Pacific in the midlatitudes (EFC), and the midlatitude zonal cell (MZC) over the Pacific. During a strong blocking winter (SBW), the ZWC is strengthened and the anomalous EHC is opposite to the anomalous WHC and the thermally driven Hadley cell. The anomalous MZC is characterized by air rising in the west part of the North Pacific, flowing eastward in the upper troposphere, descending in the eastern North Pacific, then returning back to the east coast of Asia in the lower troposphere, modulating the mean MZC. It is also found that the anomalous regional Ferrel cell at midlatitudes exists in the eastern Pacific (EFC), whereas it is not apparent in the western Pacific. All anomalous atmospheric cells almost completely reverse during a weak blocking winter (WBW).

Evolutions of each cell are also investigated. The atmospheric cells over the tropical and subtropical regions (ZWC, EHC, and WHC) always emerge ahead of the anomalous blocking winter, and then lower troposphere signals propagate upward after the anomalous blocking winter. This may suggest a mid-to-low-latitude interaction of the response of the midlatitude atmospheric blocking to tropical SST variations and a feedback via the atmospheric cells to the Tropics. In contrast, the midlatitude cells (MZC, EFC) evolve very locally, with a simultaneous response to the blocking event and no propagation of signal.

1. Introduction

The term “blocking,” which denotes a breakdown in the prevailing tropospheric westerly flow at midlatitudes and is often associated with a split in the zonal jet and with persistent ridging at higher latitudes, is a large-scale, midlatitude atmospheric phenomenon that has a profound effect on local and regional climates in the immediate blocking domain (Rex 1950a,b; Illari 1984) as well as in regions upstream and/or downstream of the blocking event (Quiroz 1984; White and Clark 1975). Therefore, it has long been of interest to synoptic and dynamical meteorologists. Many studies have derived a comprehensive set of climatological statistical characteristics of blocking anticyclones using subjective or objective techniques, including location, frequency, duration, intensity, size, and distribution (Elliott and Smith 1949; Rex 1950a,b; White and Clark 1975; Lupo and Smith 1995; Lejenæs and Økland 1983). Preferred Northern Hemisphere locations for blocking are the northeastern boundaries of the Pacific and Atlantic Oceans (Elliott and Smith 1949; Rex 1950b; Lejenæs and Økland 1983; Dole and Gordon 1983; Shukla and Mo 1983).

The maintenance of blocking appears to be determined primarily by the internal dynamics of the atmosphere (Dole 1989; Mullen 1989) and may be relatively insensitive to boundary forcings. Some researchers using observational data analysis and numerical sim-
ulations have investigated the relationships between the frequency of occurrence of blocking over the North Pacific, the ENSO cycle, and the Pacific–North American (PNA) pattern (Wallace and Gutzler 1981). Van Loon and Madden (1981) found a relationship between mean sea level pressures (SLPs) and the ENSO cycle, such that during warm (cold) phases of the cycle, SLP over the North Pacific tends to be anomalously low (high), implying that blocking may be weakened (enhanced) over the Alaskan region, and the Aleutian low may be enhanced (weakened) in “warm” (“cold”) winters. Enhancement of the wintertime Aleutian low during the warm phase of the ENSO cycle is consistent with the observed association between ENSO and the PNA pattern. During the warm phase of ENSO, the positive polarity of the PNA pattern (negative height anomalies over the North Pacific) tends to occur relatively frequently (Horel and Wallace 1981; Simmons et al. 1983).

Data analysis and model simulations by Horel and Mechoso (1988) show an increase in the persistence of wintertime circulation anomalies over the North Pacific during the warm phase of ENSO, often associated with the positive polarity of the PNA pattern and with below-normal geopotential heights in the Gulf of Alaska. Mullen (1989) analyzed GCM simulations forced with a variety of tropical and midlatitude Pacific sea surface temperature (SST) anomalies and found that such anomalies do not significantly affect the frequency of occurrence of simulated North Pacific blocking, but that they do affect the preferred locations for block formation. When equatorial Pacific SSTs are enhanced in the model, as during an El Niño event, the region of most frequent North Pacific blocking moves eastward from the Aleutians toward the coast of British Columbia, Canada. Renwick and Wallace (1996) found that the occurrence of blocking is sensitive to the averaged polarity of the PNA pattern but is even more sensitive to the phase of the ENSO cycle. ENSO-related differences in blocking frequency are associated with changes to both the mean and variance of the circulation over the North Pacific.

Recently, Huang et al. (2002) also revealed the interannual and decadal variability of the North Pacific wintertime blocking and its connection with PNA and storm tracks. At interannual time scales, the North Pacific blocking is in some sense associated with ENSO, suggesting the response of the midlatitude atmospheric circulation to tropical SST anomaly, but the response to external forcing can lead to patterns distinct from those associated with internal variability (Straus and Shukla 2002). Thus, it is important to change the viewpoint from ENSO warm or cold phases to anomalous variability in midlatitude blocking when studying the interaction of the mid- and low latitudes. The large-scale circulation features associated with the North Pacific blocking have not been well studied, although atmospheric circulation cells evolving during ENSO have been analyzed recently (Wang 2002). The goal of this paper is to document the atmospheric circulation associated with North Pacific wintertime blocking.

The recently available data of the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis (Kalnay et al. 1996) include a considerably long time series (1948–2000) and so provide a good basis for analyzing the behavior of atmospheric circulation related to anomalous midlatitude blocking. The goal of this study is to further describe and investigate how the Walker, Hadley, Ferrel, and the midlatitude zonal cells vary during the northern winter when blocking anomalies appear over the North Pacific. Evolutions of each cell associated with the midlatitude anomalous North Pacific blocking are also studied.

The rest of the paper is organized as follows. In section 2 the datasets and the analysis techniques are described. Results are presented in section 3, and concluding remarks follow in section 4.

2. Data and methodology

The data source in this study is the NCEP–NCAR reanalysis from January 1948 to December 2000. Fields are extracted for 90-day periods covering the months of December–February (DJF) for the 52 winters 1948/49 through 1999/2000, inclusive. The NCEP–NCAR reanalysis uses a state-of-the-art global data assimilation system (see Kalnay et al. 1996 for details). Variables used in this study are monthly mean horizontal wind and vertical velocity at levels of 1000, 925, 850, 700, 600, 500, 400, 300, 250, 200, 150, and 100 hPa on a 2.5° latitude × 2.5° longitude grid. The vertical component of the wind in the NCEP–NCAR reanalysis is pressure vertical velocity. In our presentation, we change the sign of the pressure vertical velocity, so positive values of the vertical velocity indicate an upward movement of air parcels. As we are interested in the atmospheric circulation over the tropical and northern Pacific, the analyses in this paper are shown for 30°S–70°N, 60°E–60°W. Throughout this study, anomalies are defined as the monthly mean variation about the long-term climatological-mean seasonal cycle. “Wintertime” is defined as the mean climate state during December to February for each year pair 1948/49 through to 1999/2000.

The horizontal wind velocity can be divided into a nondivergent (or rotational) part and a divergent (or irrotational) part (e.g., Mancuso 1967; Krishnamurti 1971; Krishnamurti et al. 1973):

\[ \mathbf{v} = \mathbf{v}_d + \mathbf{v}_a = k \times \nabla \psi + \nabla \phi. \]

(1)

where \( \psi \) is a streamfunction, and \( \phi \) is a velocity potential (VP). The first part does not contribute to atmospheric divergent fields associated with atmospheric vertical motion (it is nondivergent). It is well known that the Walker and Hadley cells are thermally driven, associated with atmospheric convergence/divergence. Atmospheric
heating associated with convection induces atmospheric
convergence/divergence, which drives atmospheric vertical
motion and circulation. Therefore, what matters to
atmospheric cells associated with atmospheric conver-
gen/divergence is the divergent part of the wind, al-
though the rotational part is usually larger. In the search
for evidence of atmospheric circulation cells, it is es-
sential not to only isolate the divergent part of the wind
but also to ascertain the continuity following the flow
between the centers of upward and downward motion
(e.g., Krishnamurti et al. 1973; Hastenrath 2001). Anal-
yses of the divergent wind field and vertical motion are
relevant for the identification of atmospheric circulation
cells. Thus, we will focus mainly on the distributions of
atmospheric vertical motion and the divergent component
of the wind when we discuss atmospheric cir-
culation cells.

Although the blocking phenomenon is well known in
the meteorological community, there is no generally ac-
cepted definition of a blocking event (LejenaÈs and
Økland 1983). Commonly used definitions can be di-
vided into three categories of method to identify block-
ing. The first is a subjective technique first put forward
by Rex (1950a,b) in the 1950s. This was then inherited
and modified by Sumner (1954) and White and Clark
(1975). They identified the blocking events subjectively
by visual inspection and then used semiobjective criteria
to determine the exact dates of initiation and duration
of blocking based on examination of the daily weather
charts. The second is an objective criteria first used by
Elliott and Smith (1949) based on magnitude and per-
sistence of pressure anomalies. Later, Hartmann and
Ghan (1980), Dole (1986, 1989), Dole and Gordon
(1983), Shukla and Mo (1983), and Zhao and Chen
(1990) extended the analysis from sea level pressure
departure to the occurrence of persistent positive geo-
potential height anomalies at the upper level. The last
objective criteria to identify blocking events was de-
signed by Lejenäs and Økland (1983), using the north–
south geopotential height gradient based on the coher-
ence between the occurrence of persistent anomalous
midlatitude easterly flow and blocking. The advantage
of this objective method is its simplicity for automatic
calculation, and therefore it has been used widely (e.g.,
Tibaldi and Molteni 1990; D’Andrea et al. 1998) with
some variations in the choice of northern and southern
latitudes for calculating the north–south geopotential
height gradient and different criteria levels (negative
gradient or very small positive gradient) for blocking.

In this study blocking is defined in a manner similar
to that of Dole (1986): namely, as persistent positive
anomalies greater than 100 m at 500-hPa geopotential
height at grid points over the North Pacific–Bering Strait
region (180°–120°W, 35°–65°N). However, we also re-
quire persistence for at least 7 consecutive days. Ac-
cording to the study of Huang et al. (2002), four indices
(frequency index \(I_f\), area index \(I_a\), and two intensity
indices \(I_d\) and \(I_p\)) were defined in order to depict the
frequency, range, and intensity variation of blocking. Index \(I_f\) is defined as the number of days witnessing
blocking each winter, and \(I_a\) is defined as

\[
I_a = \sum_{i,j} \delta(\Delta H_{i,j} - 100.0),
\]

in which \(\delta(h)\) is a step function, \(\delta(h) = 1\) when \(h \geq 0\),
\(\delta(h) = 0\) when \(h < 0\). Here \(\Delta H_{i,j}\) is the geopotential
height anomaly at each grid over the North Pacific–
Bering Strait region. Index \(I_d\) is defined as

\[
I_d = \sum_{j=1}^{N} \sum_{i=1}^{M} (\Delta H_{i,j} - 100.0) \times \delta(\Delta H_{i,j} - 100.0),
\]

\(i = 1, \ldots, M, \quad j = 1, \ldots, N,\)

where \(M\) and \(N\) are the longitudinal and latitudinal grid
numbers over the region, respectively. Index \(I_p\) is de-
efined as

\[
I_p = I_p/I_a.
\]

The correlation coefficients between the four indices are
all larger than 0.92, confirming a significant (>99.9%)
coherence of blocking frequency, range, and intensity
during the 52 winters. This agrees with many previous
studies (e.g., Hartman and Ghan 1980; LejenaÈs and
Økland 1983; Lupo and Smith 1995).

Based on the variability of the four indices for the
North Pacific blocking (Fig. 1 in Huang et al. 2002),
10 winters with high blocking indices (all four stan-
dardized indices greater than 0.5) and 11 winters with
low blocking indices (all four standardized indices less
than −0.5) are chosen for composite analyses. Blocking
events occur frequently during winters with high indi-
ces, and so these winters are called strong blocking
winters (SBWs), and vice versa, weak blocking winters
(WBWs). Of the 52 winters used, 10 are classified as
SBWs (1949/50, 1951/52, 1956/57, 1961/62, 1962/63,
1967/68, 1970/71, 1984/85, 1988/89, and 1990/91), and
11 are classified as WBWs (1952/53, 1957/58, 1960/61,
1997/98, and 1998/99). The total number of blocking
events for the 10 SBWs is 23, while only 3 blocking
events occur in the 11 WBWs. It is worth noting that
9 of the 11 years of WBWs are El Niño years, while
half of the SBW years correspond to La Niña periods,
according to the classifications of Angell (1981) and
Rasmusson and Carpenter (1982). This may suggest that
the midlatitude blocking circulation is associated with
the atmospheric internal variability partly forced by the
tropical SST anomalies related to El Niño–Southern
Oscillation (ENSO). The extratropical atmospheric re-
sponses to ENSO warm and cold phase are different
and asymmetrical. The following results are based on the
10- and 11-yr composites of SBW and WBW years,
respectively.
3. Results

a. Climatological-mean circulation

To better understand anomalous variations of atmospheric circulation cells, we first consider the climatological-mean winter atmospheric cells (averaged from 1948/49 to 2000/01) over the tropical and midlatitude Pacific. Figures 1–3 show the boreal winter (December to February) climatologies of tropospheric circulation. Centers of high (low) velocity potential are associated with divergent outflow (convergent inflow) winds. Figures 1a and 1b show that the divergence (convergence) at the upper troposphere corresponds to convergence (divergence) at the lower troposphere (two levels of 1000 and 200 hPa are chosen as representative of the lower and upper troposphere, respectively). Large centers of upper-tropospheric divergence and lower-tropospheric convergence in the central and western equatorial Pacific regions are apparent. The equatorial eastern Pacific is associated with upper-tropospheric convergence and lower-tropospheric divergence. All of these findings are consistent with climate features of the
western Pacific warm pool and the equatorial eastern Pacific cold tongue (Wang 2002). Associated with these patterns is the east-west circulation cell along the equator—the zonal Walker cell (ZWC), as shown in Fig. 2a. The air ascends in the west and central Pacific, flows eastward in the upper troposphere, sinks in the east, and returns toward the west in the lower troposphere. Mid-latitudes upper- (lower) level divergence (convergence) in the central North Pacific, corresponding to upper- (lower) level convergences (divergences) at the west coast of America and at the east coast of Asia (Figs. 1a,b), are associated with a pair of midlatitude zonal cells (MZC), as shown in Fig. 2b. The MZC shows that the air rises in the central North Pacific, diverges eastward and westward in the upper troposphere, then descends over regions at the west coast of North America and the east coast of Asia, subsequently flowing back to the central North Pacific in the lower troposphere.

According to Figs. 1a and 1b, the meridional-vertical circulation in the western Pacific is different from that.
in the eastern Pacific. Therefore, we separately plot meridional-vertical circulations by averaging the meridional component of divergent wind and vertical velocity in the west (100°E–180°) and in the east (180°–100°W), as shown in Figs. 3a and 3b. In the west, a pair of almost equatorially symmetric strong local Hadley cells appears, with the air rising in the tropical region, flowing poleward in the upper troposphere in both hemispheres, and returning to the Tropics in the lower troposphere (Fig. 3a). The local Ferrel cell, with upward motion in the high latitudes and downward motion in the mid-latitudes, is not obvious here except for the strong descending flow at about 30°N. The circulation in the east region (Fig. 3b) is much more complicated than that in the west. The tropical circulation north of the equator has two cells with air rising in the ITCZ (Trenberth and Caron 2000), then diverging northward and southward in the mid- to upper troposphere. The branch flowing
northward descends at the midlevel (about 500 hPa) over the subtropical region, which looks like a local Hadley cell with upward motion at low latitudes and downward motion at midlatitudes. However, the direct cell is weak and does not extend that deep in the vertical. The other branch of air ascends to the upper troposphere at about the 300-hPa pressure level, and then diverges southward and downward, leading to an opposite atmospheric cell in the upper troposphere off the equator. This cell is very small and local, different from the typical Hadley cell. It produces a pair of antisymmetric cells adjacent to the equator, together with a small cell south of the equator. The pair of near-equatorial indirect cells at the upper level of the troposphere contributes to the convergence over the equator at the upper troposphere shown in Fig. 1b. A larger cell also appears south of the equator with its center at the midtroposphere, lower than that of the northern cell. In mid- and high latitudes a Ferrel-like cell seems more apparent than that in the west region of the Pacific, with strong downward flow at midlatitudes (about 30°N) and upward motion at high latitudes (about 60°N). The results above mostly agree with that of Wang (2002). Only a few differences exist, possibly because we analyze a slightly different range of longitudes averaged for the west and east regions of the Pacific, and a different time period of 3-month averages to represent boreal winter climatologies, as opposed to 1 month in Wang (2002).

b. Composites of anomalous circulation

1) VELOCITY POTENTIAL

Based on variations of North Pacific blocking (Huang et al. 2002), 10 SBWs and 11 WBWs are chosen for composite analysis as discussed above. Averaged anomalous zonal or meridional atmospheric cells are investigated in SBWs and WBWs, respectively. Figures 4a and 4b show the difference of VP between the SBWs and WBWs at the upper (200 hPa) and lower (1000 hPa) troposphere, respectively. The light and heavy gray shaded regions represent the confidence level exceeding 95% and 99% by a Student t test. It is found that at the lower pressure level (Fig. 4a), positive VP differences mainly appear in the eastern Pacific, and negative VP differences occupy the central and western Pacific as well as the eastern Indian Ocean. Two large positive centers are apparent, located at the Alaska–Aleutian region and the eastern equatorial Pacific. In addition, two relatively small negative centers are located over the Maritime Continent and over the central extratropical Pacific. Since a positive VP center means divergent air flow, and a negative VP center corresponds to convergent air flow, the pattern at 1000 hPa (Fig. 4a) indicates enhanced western tropical Pacific convergence, enhanced eastern equatorial Pacific divergence, and weakened Aleutian convergence during SBW. The opposite behavior persists during WBW according to the climatological VP in Fig. 1a. However, in the upper troposphere (Fig. 4b), the pattern of VP difference is almost the reverse of that in the lower level (Fig. 4a), except that the two positive and negative centers are larger than that at 1000 hPa and the two centers at mid-high latitudes show more of a dipole in the east–west direction. This configuration of the VP in the upper and lower levels suggests several anomalous vertical circulations between the eastern and western part of the Pacific in low and midlatitudes over the North Pacific, which will be discussed in the following sections.

2) ZONAL CELLS

An anomalous Walker cell of the tropical zone and a MZC along the midlatitude zone in SBW and WBW are shown in Figs. 5a,b and 5c,d, respectively. Here the averaged zonal component of divergent wind and vertical velocity between 10°N and 10°S are used to draw the vertical circulation representing the Walker cell. Anomalous upward motion over the western warm pool of the Pacific is obvious in SBW (Fig. 5a), which then flows eastward, descending in the central and eastern equatorial Pacific with the returning easterly at the lower troposphere, leading to a closed cell over the tropical Pacific. An almost completely opposite anomalous cell appears during WBW (Fig. 5b). Comparing with the winter mean Walker circulation in Fig. 2a, the winter mean Walker circulation is enhanced during SBW and weakened in WBW. These results are consistent with the relationship between the ENSO cycle and Walker circulation because in most of the WBW period an El Niño occurred, and vice versa. For midlatitudes (Figs. 5c,d), the most apparent characteristic of anomalous circulation is the strong descending (ascending) flow in SBW (WBW), with a pair of reversed anomalous zonal cells on either side of the vertical air flow. Over the North Pacific the anomalous MZC is characterized by air rising (descending) in the western part of the North Pacific, flowing eastward (westward) in the upper troposphere, and, with extremely strong descending (ascending) air in the northeast Pacific, returning back to the northwest Pacific in the lower troposphere during SBW (WBW). This modulates the MZC greatly, resulting in the eastern part weakening (strengthening) in SBW (WBW). The total effect of the anomalous cells is to enhance the winter mean MZC during WBW and weaken the climatological MZC during SBW (Fig. 5c).

3) MERIDIONAL CELLS

According to the distribution of winter mean VP (Figs. 1a,b) and anomalous VP (Figs. 4a,b) at the upper and lower levels of the troposphere, it is not difficult to see that the east and west parts of the Pacific have opposite divergence or convergence patterns occurring at different levels of the troposphere. The climatological meridional-vertical circulations in the western and east-
ern Pacific have been shown already in Figs. 3a and 3b. Therefore, here we mainly investigate the anomalous meridional-vertical circulation over the two regions in SBW and WBW, respectively. These results are shown in Figs. 6a–d.

As we know, the western Pacific is a typical monsoon region, with a northerly winter monsoon from the Asian continent to the west Pacific Ocean. The most significant difference in the anomalous meridional circulation in the western Pacific between SBW and WBW is the completely reversed cells, especially for the regional Hadley cells adjacent to the equator (Figs. 6a,b). The anomalous Hadley cells in SBW (WBW) consist of the upward rising (downward descending) flow in the Tropics, northward and southward divergent (convergent) flow in the upper troposphere, downward (upward) flow in extratropical areas, and equatorward (polarward) wind at the lower level of the troposphere. This means that the western Pacific regional Hadley cells on either side of the equator will be enhanced during SBW and weakened in WBW, while the lower troposphere winter monsoon is strengthened in SBW and weakened in WBW.
Fig. 5. Anomalous Walker circulation obtained by averaging anomalous divergent wind and vertical velocity between 10°S and 10°N in (a) SBW and (b) WBW. Anomalous MZC obtained by averaging anomalous divergent wind and vertical velocity between 35° and 55°N in (c) SBW and (d) WBW. The unit for zonal wind is m s$^{-1}$, and the unit for vertical velocity is hPa s$^{-1}$. The vertical velocity is scaled by 50 to clarify the circulation cell vectors.
As mentioned above, the Ferrel cell in the western Pacific is not very apparent. Here the anomalous Ferrel cells in SBW and WBW are also not closed cells but semiclosed cells with opposite directions of airflow. The Ferrel cell is enhanced in SBW and weakened in WBW. It is worth noting that the anomalous meridional cells in WBW are a little stronger than those in SBW.

Figures 6c and 6d show the anomalous meridional cells in the eastern Pacific. The meridional cells are almost antisymmetric in SBW and WBW except for the regional Hadley cell south of the equator. The dominant feature in the eastern Pacific is that there exist several strong closed cells from the Tropics to high latitudes. An inverse anomalous Hadley cell and Ferrel cell appear in SBW with upward flow at the subtropics, northward and southward divergent winds, downward motion at the equator and high latitudes, and return flows from the equator and high latitudes to the subtropical region (Fig. 6c). Inverted anomalous cells occur in WBW (Fig. 6d). These results reveal that the Hadley cell and Ferrel cell are all weakened in SBW while they are strengthened in WBW. Comparing with Figs. 6a and 6b, the regional Hadley cells and Ferrel cells in the western and eastern Pacific vary inversely, like a seesaw, and all the meridional cells have opposite circulation patterns in SBW and WBW, respectively.

c. Evolution of anomalous cells related to the blocking anomaly

Features of the atmospheric cells during the extremely strong or weak blocking winter over the North Pacific have been investigated above, focusing on the anomalous circulation associated with a simultaneous blocking anomaly. For a better understanding of the relationship between the atmospheric cells and North Pacific blocking, we now investigate the evolving nature of the atmospheric cells before and after the SBW or WBW. A composite analysis is also used here over a 24-month period, with the 3 winter months (December to February) of SBW (or WBW) in the middle of the 24-month time series. For convenience of notation, January to November before the winter of SBW or WBW are annotated as \(-1\), March to December subsequent to the winter of SBW or WBW are marked as \(+1\), and December to February of SBW or WBW are tagged with 0. Then, 10 samples of SBW and 11 of WBW for each month of the 24-month period are averaged for relevant variables. Since the anomalous winds of each atmospheric cell in the upper and lower troposphere are almost completely opposite in direction between SBW and WBW, it is easy to show the evolution of the atmospheric cells by drawing the vertical-time cross section. Figures 7a–e show the differenced cross sections of west–east and north–south components of divergent wind between SBW and WBW, indicating the evolution of the ZWC, the MZC, the regional Hadley cell in the western Pacific (WHC), the regional Hadley cell in the eastern Pacific (EHC), and the regional Ferrel cell over the eastern Pacific in the midlatitudes (EFC), respectively.

Figure 7a shows the difference of the west–east component of divergent wind averaged over the main ZWC region (140°E–140°W, 10°S–10°N) between SBW and WBW (SBW minus WBW). As mentioned above, the Walker cell is strengthened during SBW and weakened in WBW, which means an anomalous westerly appears in the upper troposphere and an easterly anomaly at the lower level over the central equatorial Pacific in SBW, and vice versa in WBW. The most obvious feature in Fig. 7a is the strong westerly anomaly at about 150 hPa and easterly anomaly at 850 hPa during Jan (0). It is notable that the anomalous westerly (easterly) in the upper troposphere and westerly (easterly) in the lower troposphere in SBW (WBW) appears from the former winter and lasts at least a whole year, with the strongest pattern in the winter when midlatitude blocking occurs frequently. The strongest westerly (easterly) anomaly in SBW (WBW) enhances considerably from the preceding autumn (Oct \(-1\)) and reaches two maxima at Jan (0) and Mar \(+1\). The negative area (meaning easterly anomaly in SBW and westerly anomaly in WBW) strengthens gradually from the preceding year and propagates upward from the middle and lower troposphere to the upper troposphere after the anomalous blocking winter (SBW or WBW), especially during the transition from the subsequent spring to summer (May \(+1\) to Jun \(+1\)). These results suggest a possible relationship between North Pacific blocking activity and the equatorial Walker circulation. That is, the midlatitude blocking circulation may respond to the anomalous tropical Walker circulation driven by the equatorial SST anomaly (Huang et al. 2002). Then, feedback to the ZWC in the winter of anomalous blocking results in the upward propagation of an easterly or westerly anomaly, leading to the Walker cell reversing. The reason for this may be associated with an interaction between the Tropics and midlatitudes, which has yet to be investigated.

The difference cross section of MZC [west–east component of divergent wind averaged over the midlatitudes (140°E–140°W, 35°N–55°N)] between SBW and WBW (Fig. 7b) shows a very strong westerly anomaly in the
upper troposphere and the opposite in the lower troposphere, but limited to the winter of SBW or WBW. This implies that the variation of MZC is mainly associated with the local blocking circulation anomaly. The opposite anomalous cells of the WHC and EHC (Figs. 7c,d) have similar evolution features to that of the ZWC, but the WHC reaches its peak about 1 month earlier and is weaker than that of the EHC, as expected. The EFC over midlatitudes evolves very locally and spontaneously as the blocking anomaly appears in each winter, with a southerly anomaly in the upper troposphere and a northerly anomaly in the lower troposphere in SBW (Fig. 7e). As mentioned above, no clear Ferrel cell appears over the western part of the midlatitude Pacific; thus, we do not discuss its evolution. The evolution of atmospheric cells over the tropical and subtropical regions (ZWC, WHC, and EHC) associated with the midlatitude blocking anomaly may suggest a mid–low latitude interaction. In contrast, the evolution of the midlatitude cells (MZC and EFC) seems to reflect a local effect of the blocking circulation.

4. Summary

Based on the NCEP–NCAR reanalysis field from 1948/49 to 1999/2000, the atmospheric circulation associated with anomalous variations of the North Pacific blocking during winter have been described and examined. The mean state and anomaly of the equatorial zonal Walker cell, the tropical meridional Hadley cell, the extratropical meridional Ferrel cell, and the midlatitude zonal cells associated with this blocking during winter are summarized in Fig. 8. Figure 8a shows a schematic diagram of the mean winter atmospheric circulation cells. Unlike the western Pacific, which shows a pair of regional Hadley cells symmetric about the equator, the wintertime mean meridional circulation in the eastern Pacific shows a much more complicated structure, namely, a pair of indirect meridional cells asymmetric to the equator in the mid- and upper troposphere, a smaller Hadley cell (compared to that in the western Pacific) centered at lower levels, and a semi-closed Ferrel cell (Fig. 8a). The zonal mean circulation along the equator shows a typical Walker circulation due to the strong tropical heat source over the western Pacific warm pool and the relatively weak heat sink over
Fig. 8. Schematic diagrams summarizing (a) the wintertime mean state of atmospheric circulation cells over the North Pacific, and anomalous atmospheric cells during (b) SBW and (c) WBW. The thickness of lines denotes the relative intensity of airflows. Dashed lines are used simply to illustrate the three-dimensional aspect of the diagrams.

the cold tongue of the tropical eastern Pacific. At mid-latitudes the divergent wind and vertical motion of the NCEP–NCAR reanalysis field shows the winter mean MZC in the North Pacific, characterized by air rising in the central North Pacific, diverging westward and eastward in the upper troposphere, descending over the regions of the east coast of Asia and the west coast of North America, then flowing back to the central North Pacific in the lower troposphere. This pattern is very similar to that of Wang (2002), although he chose a slightly different latitude band to our study and only the January mean instead of a 3-winter month mean.

During WBW, the ZWC (Fig. 5b) is weakened and the anomalous EHC (Fig. 6d) shows air rising in the Tropics, flowing poleward in the upper troposphere, sinking at mid-latitudes, and returning back to the Tropics in the lower troposphere, opposite to the anomalous WHC (Fig. 6b) in the western Pacific. These atmospheric cells and the velocity potential anomaly (Fig. 4) are all quite similar to that associated with the mature phase of El Niño in Fig. 10 of Wang (2002), which supports the linkage between ENSO and the midlatitude blocking activity over the Pacific. This is consistent with SST-related heating (Huang et al. 2002), which is a driving force for the Hadley circulation. According to the composite field of SST anomaly (Fig. 9b) from a retrospective analysis of the global ocean based on the Simple Ocean Data Assimilation (SODA, version 7) package of Carton et al. (2000a,b), negative SST anomalies appear in the tropical western Pacific and midlatitude central-east Pacific, while positive SST anomalies appear in the equatorial eastern Pacific and western subtropical Pacific in WBW. Corresponding to this SST anomaly distribution are lower-troposphere anomalous convergence (divergence), upper-troposphere anomalous divergence (convergence), and midtroposphere anomalous ascending (descending) motion in the tropical eastern (western) Pacific. Thus, the anomalous eastern and western Pacific Hadley cells exhibit an opposite circulation sense. The anomalous MZC is characterized by strong air rising in the eastern part of the North Pacific, flowing westward in the upper troposphere, descent in the central and northwest Pacific, then return flow to the northeast Pacific in the lower troposphere. This modulates the mean MZC greatly, resulting in a strengthening of the eastern part of the MZC in WBW. This also contributes to the midlatitude negative SST anomaly in the central Pacific in WBW (Fig. 9b), leading to a convergence in the upper troposphere and a divergence in the lower troposphere. However, we must keep in mind that, like the Walker and Hadley cells, the MZC in this paper is identified by the divergent wind and the pressure vertical velocity. If we also consider the rotational wind, the MZC is not a closed cell. It is also found that the anomalous regional Ferrel cell at midlatitudes only exists in the eastern Pacific and is not apparent in the western Pacific (Fig. 8c). In contrast to the anomalous Hadley circulation cell, the anomalous
Ferrel circulation cell is weak and incoherent. The reason for this is different driving mechanisms for the Hadley and Ferrel cells. The tropical Hadley cell is a thermal-driven cell, forced by diabatic heating, whereas the extratropical Ferrel cell is an indirect thermal-driven cell forced mostly by transient baroclinic eddy activity through associated poleward heat and momentum transports (e.g., Holton 1992). All anomalous atmospheric cells almost completely reverse in SBW (Fig. 8b).

Evolution of each cell associated with midlatitude North Pacific blocking exhibits anomalous signals of tropical and subtropical cells (ZWC, EHC, and WHC) preceding the blocking winter. The lower tropospheric signals then propagate upward after the anomalous blocking winter. In contrast, midlatitude cells (MZC, EFC) vary simultaneously with the blocking event, with no significant propagation. The SST differences between SBW and WBW (SBW minus WBW) occurring before, during, and after the strong and weak North Pacific blocking winters are shown in Fig. 9. This analysis suggests that the evolution of each cell is associated with the evolution of the SST field over the tropical and midlatitude Pacific. Almost 1 yr before the strong or weak North Pacific blocking winters (Fig. 9a), the tropical SST difference appears similar to a La Niña pattern, except that the SST anomalies are much weaker. This pattern is maintained but with markedly stronger SST anomalies during the strong or weak blocking winters (Fig. 9b). After the weak and strong blocking winters (Fig. 9c), a positive SST difference is maintained in the central Pacific at midlatitudes. In contrast, the negative SST anomaly in the equatorial eastern Pacific now appears as a near-equatorial positive anomaly. This may suggest a midlatitude–low latitude interaction, whereby the midlatitude atmospheric blocking responds to a tropical SST anomaly by changing the Walker and Hadley circulations, then interacts with midlatitude SST and feeds back to the low latitudes by variation of the local atmospheric cells, especially at the top of the troposphere. According to the vertical propagation conditions of planetary waves (e.g., Holton 1992; Charney and Drazin 1961), this may relate to the momentum exchange between the upper troposphere and the lower stratosphere when blocking occurs frequently in winter. Subsequent to such a winter, the feedback to the Tropics results in anomalous SST and anomalous westerly winds along the eastern equatorial Pacific. The mechanism of mid-to-low-latitude air–sea interaction over the North Pacific is, however, a complicated process and still requires further examination.

Acknowledgments. The authors would like to thank the Natural Science Foundation of China Geoscience Department, Nanjing Atmospheric Data Center (NADC) for supplying the NCEP–NCAR reanalysis data, which were obtained from the National Center for Atmospheric Research (NCAR). This research was supported by the National Natural Science Foundation of China (No. 40305009), the Australian Research Council and the China Scholarship Council.

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Fig. 9. Differences of composite SST fields between SBW and WBW (SBW minus WBW) that (a) precede (Apr), (b) occur simultaneously (Jan), and (c) occur after (Jul) the strong or weak North Pacific blocking winters, respectively. The contour interval is 0.3°C, and the dark (light) shaded areas denote SST values >0.3°C (<−0.3°C).


