Climatology and Interannual Variation of the East Asian Winter Monsoon: Results from the 1979–95 NCEP/NCAR Reanalysis

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ABSTRACT

This paper presents the climatology and interannual variation of the East Asian winter monsoon based on the 1979–95 National Centers for Environmental Prediction/National Center for Atmospheric Research reanalysis. In addition to documenting the frequency, intensity, and preferred propagation tracks of cold surges and the evolution patterns of related fields, the authors discuss the temporal distribution of the Siberian high and cold surges. Further, the interannual variation of the cold surges and winter monsoon circulation and its relationship with ENSO were examined.

There are on average 13 cold surges in each winter season (October–April), of which two are strong cases. The average intensity of cold surges, measured by maximum sea level pressure, is 1053 hPa. The cold surges originate from two source regions: 1) northwest of Lake Baikal, and 2) north of Lake Balkhash. The typical evolution of a cold surge occurs over a period of 5–14 days. Trajectory and correlation analyses indicate that, during this time, high pressure centers propagate southeastward around the edge of the Tibetan Plateau from the mentioned source regions. Some of these high pressure centers then move eastward and diminish over the oceans, while others proceed southward. The signatures of the associated temperature, wind, and pressure fields propagate farther southward and eastward. The affected area encompasses the bulk of the maritime continent. Although the intensity of the Siberian high peaks during December and January, the frequency of cold surges reaches a maximum in November and in March. This result suggests that November through March should be considered as the East Asian winter monsoon season.

Two stratifications of cold surges are used to examine the relationship between ENSO and the interannual variation of the winter monsoon. The first one, described as conventional cold surges, indicates that the surge frequency reaches a minimum a year after El Niño events. The second one, defined as a maximum wind event near the South China Sea, shares the same origin as the first. The surge frequency is in good agreement with the Southern Oscillation index (SOI). A low (high) SOI event coincides with a low (high) frequency of cold surges.

The interannual variation of winter northerlies near the South China Sea is dominated by the South China Sea cold surges and is well correlated ($R = 0.82$) with the SOI. Strong wind seasons are associated with La Niña and high SOI years; on the other hand, weak wind seasons are associated with El Niño and low SOI years. This pattern is restricted to an area north of the equator within the region of (0°–20°N, 100°–130°E) and is confined to the near-surface layer. The SST variation in the same region is similar to the wind pattern but lags the wind for approximately 1–5 months, which suggests that the SST variation is forced by the wind. The surface Siberian high, 500-hPa trough, and 200-hPa jet stream, all representing the large-scale monsoon flow, are weaker than normal during El Niño years. In particular, the interannual variation of the Siberian high is in general agreement with the SOI.

1. Introduction

The East Asian winter monsoon, accompanied by a strong Siberian high and active cold surges, is among one of the most energetic monsoon circulation systems. Every winter, the dramatic shift of the northeasterlies and the outbreak of cold surges not only dominate the local weather and climate in the East Asian region but also exert a strong impact on the extratropical and tropical planetary-scale circulations (Chang and Lau 1982), and influence the convection and SST near the maritime continent (Chang et al. 1979).

General characteristics of the winter monsoon and cold surges, their link with tropical atmospheric and oceanic phenomena, and the interaction between the winter monsoon and other monsoon systems have been examined in many studies. For example, Chang and Lau (1980) have shown that the cold surges can force short-term changes in the Hadley and Walker circulations, the East Asian jet stream, and the large-scale deep convection over the equatorial western Pacific. Lau and Chang (1987) suggested that the interannual variation of the winter monsoon and cold surges may be related to
ENSO and tropical intraseasonal oscillation. Ding and Krishnamurti (1987) noted an eastward shift of the tropical planetary-scale divergent circulation in association with the cold surges that was very similar to the shift of the divergent circulation between El Niño and La Niña years. Davidson et al. (1983) presented a case that suggested a possible relationship between cold surges and the onset of the Australian monsoon. Shield (1985) also showed that the cold surges were linked to monsoon development in the Southern Hemisphere. More recently, Sun and Sun (1996) suggested that a strong winter monsoon usually precedes a drought summer season; likewise, a weak winter monsoon can induce a wet summer season.

These observational studies have firmly established a link between the winter monsoon and the tropical atmospheric and oceanic phenomena. They have also made a strong case that the winter monsoon is an important centerpiece in the complicated interactions among different monsoon systems. Therefore, the importance of studying the winter monsoon and cold surges cannot be overemphasized. Despite this, little attention has been given to the climatological aspects of the winter monsoon and cold surges except for limited discussions in a few studies. Boyle and Chang (1984) documented the mean winter circulation statistics based on the Navy Fleet Numerical Oceanography Center’s (FNOC) gridded analyses. Pan et al. (1985) compiled the statistics of the Siberian high and cold surges for the purpose of local weather forecasting. Ding and Krishnamurti (1987) summarized the climatological tracks of cold surges based on the data from the winters (DJF) of 1980 to 1984. Although the winter monsoon was much weaker than normal during the 1982–83 El Niño event (Lau and Chang 1987), there has never been a systematic investigation of the relationship between ENSO and the interannual variation of the winter monsoon.

The goal of this study is to present a complete climatology of the winter monsoon and cold surges based on the 1979–95 National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) reanalysis. We emphasize the temporal and spatial distribution of the cold surges and the Siberian high, their origin, propagation paths, and evolution patterns. Of particular interest is the interannual variation of the winter monsoon and cold surges. We want to determine if a consistent relationship exists between ENSO and the interannual variation of the winter monsoon and cold surges.

2. Description of the reanalysis

We use twice-daily surface and upper-air fields from the NCEP/NCAR reanalysis (Kalnay et al. 1996) for the period 1979–95. The surface quantities include temperature, wind, SST, and sea level pressure (SLP); upper-air fields are winds at 925, 850, 700, and 200 hPa and geopotential height at 500 hPa. The advantage of this dataset is that the reanalysis used a “frozen” state-of-the-art analysis and forecast system to perform the data assimilation throughout the whole period. Therefore, it circumvents problems of previous numerical weather prediction analyses due to changes in techniques, models, and data assimilation. Also, this dynamically consistent reanalysis offers good horizontal (2.5° × 2.5°) and vertical resolution (17 levels).

3. Winter monsoon circulation

a. Mean state

The mean monsoon circulation has been documented by many others (e.g., Boyle and Chang 1984; Lau and Li 1984; Lau and Chang 1987; Boyle and Chen 1987) and is summarized here to serve as background for this paper.

The key driving force for the winter monsoon, similar to its summer counterpart, is the available potential energy generated by the differential heating between land and sea. During the winter season, the main heating source is situated near the equatorial western Pacific, as manifested by the 200-hPa velocity potential minimum in Fig. 1. The latent heat release associated with intense convective precipitation fuels the gigantic planetary-scale overturning in the meridional direction. To the far north of this convective center lies a very strong cold dome near the Siberian region. This high pressure, as shown in Fig. 2, has a central pressure in excess of 1036 hPa. It covers the entire East Asian continent and yields a northeasterly flow over a large part of Asia. Superimposed on the SLP is the surface wind. This climatological wind is contributed by both the quasi-stationary Siberian high and the cold surges. North of 30°N, the anticyclonic flow is primarily a reflection of the Siberian high pressure. South of 30°N where the Siberian high exerts less influence, the wind maximum near the South China Sea can be attributed to the cumulative effects of the cold surges. One of the features of the wind field is that the westerly flow near Japan bifurcates. The eastward branch merges with the Aleutian low; the southern branch joins the trade wind. As expected, the wind pressure relationship breaks down near the Tibetan Plateau. This will not affect the following cold surge identification because the plateau region will be excluded.

The intensity of the Siberian cold dome is maintained by strong radiative cooling, large-scale descending motion, and persistent cold air advection throughout the troposphere (Ding and Krishnamurti 1987). The planetary-scale descending motion associated with the Siberian high and the ascending motion near the maritime continent form a strong local Hadley cell in the East Asian region.

Closely connected to the Hadley circulation is the 200-hPa East Asian westerly jet stream, as presented in
Fig. 1. Averaged 200-hPa wind (m s\(^{-1}\)) superimposed on the velocity potential (10\(^6\) m\(^2\) s\(^{-1}\)) for winter (DJF) of 1979/80–1994/95.

Fig. 1. This jet is associated with intense baroclinicity, large vertical wind shear, and strong cold advection. The Coriolis torque exerted on the ageostrophic southerly wind in the upper branch of the Hadley cell contributed to the maintenance of this jet. The Tibetan Plateau plays a crucial role in forming the circulation patterns of the winter monsoon. The plateau not only largely controls the flow configuration in the lower troposphere but also greatly influences the thermally direct Hadley cell and the 200-hPa jet stream (Murakami 1987). The huge east–west-oriented massif of extremely high elevation also acts as a natural barrier for the cold air near the Siberian region. The cold air damming effect is important for the buildup of the Siberian high and for the outbreak of the extremely cold air that spreads along the eastern edge of the Plateau (Yeh and Gao 1979; Luo 1992).

At 500 hPa, the prevailing flow pattern (not shown) near East Asia is dominated by the quasi-stationary coastal trough. The intensity of this trough is quasi-geostrophically linked to the surface Siberian high. The East Asian jet, the Siberian high, the 500-hPa trough, and the convection center near the western Pacific are inherently related to each other. These planetary-scale features characterize the three-dimensional winter monsoon circulation.

b. Variance of SLP and surface air temperature

The winter average SLP and surface air temperature variance are shown in Figs. 3a and 3b. The standard deviation of SLP indicates that the Siberian high is a stationary system. The largest variations occur to the northwest of the Siberian high and to the east where the variability near the Aleutian low dominates. Although the standard deviation of temperature is largest at high latitudes, substantial variability also occurs over eastern China. This is noticed as a pronounced southward extension over this region. A similar, but less well defined, southward extension is also seen in the SLP variation. The southward extension in both variables is an indication of the cold surge activity.

Figures 3c and 3d show the percent of total variance of SLP and surface air temperature for periods of 6–14 days, the typical lifetime of cold surges. Along the east coast of China, over 25% of the total variance of surface air temperature and SLP occurs on timescales of 6–14 days. The structure of the temperature variation exhibits more regionality than that of SLP and has a pronounced signal over eastern China. Power spectra of the surface air temperature and SLP in this vicinity exhibit many common spectral peaks on these timescales (not shown). These peaks suggest a close association between the temperature and SLP variation, which is consistent with the characteristics of cold surges. The substantial SLP
variance over the Pacific Ocean near 25°N and the one northward over Japan are not necessarily associated with cold surges. These variation maxima are intriguing features of the SLP, which will be investigated in future works. Cold air activity over the East Asian continent and the South China Sea will be of interest here.

4. Cold surges and Siberian high

The cold surges are the most important transient disturbances embedded within the mean monsoon circulation. The occurrence of a cold surge is characterized by a southward movement of a surface anticyclone and an associated abrupt surface air temperature decrease in the affected regions. The typical scenario of a cold air outbreak is as follows. The Siberian high and the coastal trough reach a certain intensity. Farther west over the continent, an upper-level short wave undergoes strong development as it moves eastward. Eventually, the short wave develops into a major trough and replaces the old quasi-stationary coastal trough. During this process, a surface anticyclone moves southward and a cold surge occurs (Staff Members of Academia Sinica 1958). The buildup and maintenance of the Siberian high is a necessary condition for the occurrence of cold surges. Radiative cooling, persistent cold air advection (usually blocking over the Urals) throughout the troposphere, and large-scale descending motion all contribute to the maintenance of the Siberian high (Ding and Krishnamurti 1987). However, cold surges should not be regarded as the simple southward expansion or temporal variability of the Siberian high. Rather, they are a separate dynamical entity. The Siberian high and the cold surge are the two most characteristic weather phenomena of the winter monsoon. The relationship between the two will be further discussed in section 4b.

a. Criteria for cold surges

Objective criteria for the identification of the cold surges are required in this study. Many cold surge definitions can be found in the literature. A summary of the commonly used definitions can be found in Boyle and Chen (1987). Some of the cold surges are purposely defined for the convenience of weather forecasting. Definitions vary depending on regions of interest. For example, cold surges defined over the South China Sea may have nothing to do with cold surges defined near Korea. The essence of the cold surges, as mentioned above, is the rapid south-southeastward movement of a
surface anticyclone and the associated significant surface air temperature decrease in the affected regions.

In defining cold surges, we average the necessary fields at particular regions, which encompass a nine-grid-point rectangle. Figure 4 shows all three regions. Region 1 is in southern Siberia, region 2 is in the middle of China, and region 3 covers part of southern China. Using information from these regions, we define a cold surge as follows.

1) A cold surge occurs when (a) a surface anticyclone can be identified (criteria in Zhang and Wang 1997) near region 1, and the SLP at the anticyclone center is greater than 1035 hPa; and (b) during the movement of this surface anticyclone, the 24–48 h surface air temperature decrease exceeds 9°C in region 2, or 6°C in region 3.

2) A cold surge ends when (a) a negative pressure tendency at the anticyclone center has persisted for 24 h and the center pressure is less than 1025 hPa, and (b) a positive surface temperature trend can be found over 50% of the grid points in the bulk of the East Asian continent 25°–50°N, 102.5°–117.5°E.

A cold surge is chosen when these criteria are met. The thresholds, mostly based on synoptic experience, can be altered within a certain range. Despite this, the overwhelming majority of surges will be identified because the key characteristics of cold surges are represented by the criteria. Also, because a cold surge affects the surface air temperature and SLP in a large area, it is unlikely that a nonsurge event will be chosen as a surge.

To verify the identified cold surges, time series of surface air temperature in regions 1, 2, and 3 ($T_1$, $T_2$, and $T_3$) of the 1987–88 winter are plotted in Fig. 5. According to the criteria, if a surface anticyclone has reached a certain intensity, a required decrease in either $T_2$ or $T_3$ will qualify the disturbance to be a cold surge. All the observed cold surges during this season (known from other sources) were found to be associated with a temperature decrease in either region 2 or region 3. Figure 5 illustrates how quickly the temperature can drop in a specific location and its southward propagation. One of the strongest cold surges ever observed occurred during late November 1987. The recorded maximum surface pressure is 1081 hPa; the temperature decreases are greater than 15°C in all three regions.

The cold surges, their timing, duration, and frequen-
cies identified using the above criteria are further compared with those defined operationally by the Beijing Meteorological Center (BMC) for the period 1979–84 (for which the data from BMC are available), and good agreement was found between the two. Thus, we have confidence that the cold surge definition is appropriate.

b. Statistics, trajectories, and spatial evolution of cold surges

Based on the above criteria, over 200 cold surges have been identified for the 1979–95 period. Table 1 gives the average frequency, intensity, and duration for surges and strong surges. Strong events are surges that are associated with at least a 10°C temperature decrease and a 4 m s⁻¹ northerly wind in region 3. With the addition of this wind criterion, the strong cases are assured to be cold surges that exert impact on the Tropics. The average maximum SLP at the anticyclone center is approximately 1053 hPa for cold surges, and it is approximately 1060 hPa for strong surges. Lifetimes of individual surges range from 5 to 14 days, with an average duration of approximately 7 days for cold surges and approximately 9 days for strong events.

The trajectories of the surface anticyclones associated

![Figure 4](image1.png)

**Fig. 4.** Selected subareas for cold surge identification in the East Asian region. The size of each area is 5° lat × 5° long.

![Figure 5](image2.png)

**Fig. 5.** Twice-daily time series of surface air temperature (°C, 2 m) over regions 1, 2, and 3 from November of 1987 to March of 1988. A five-point running average has been applied.

**Table 1.** Cold surge statistics based on 16 winter seasons (1979–95).

<table>
<thead>
<tr>
<th></th>
<th>Number of events per year</th>
<th>SLP intensity (hPa)</th>
<th>Duration (days)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Surges</td>
<td>13</td>
<td>1053</td>
<td>7</td>
</tr>
<tr>
<td>Strong surges</td>
<td>2</td>
<td>1060</td>
<td>9</td>
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with the cold surges are shown in Fig. 6. The circles indicate the origin of the pressure centers and the squares denote their termination points. The trajectories indicate that cold surges originate from two distinct regions. The first is located northwest of Lake Baikal near 45°–60°N, 95°–105°E, and the second is to the north of Lake Balkhash near 52°–60°N, 75°–90°E. The surges propagate southward or southeastward over the eastern China. Some of them end in the East China Sea and coastal regions while others propagate farther south.

Although the cold surge trajectories terminate around 30°N, the pressure, temperature, and meridional wind perturbations are observed farther south. This is indicated in Fig. 7, which shows the climatological evolution of SLP, temperature, and northerly wind associated with cold surges. For each winter (120 days, mid-November through mid-March), daily values of $T_o$ are lag correlated with each of these fields. At each lag, the average correlation from all 16 winters is calculated. Given that a cold surge is associated with a temperature decrease, the sign convention is such that 1) a negative correlation with SLP indicates high pressure, 2) a positive autocorrelation with surface air temperature indicates low temperature, and 3) a positive correlation with the meridional wind indicates enhanced northerlies. Correlations for $|r| > 0.2$ are contoured. This corresponds to the 95% confidence level for 100 independent time samples. The correlation cutoff used was very conservative given that 120 time points in each of 16 years of data were used.

At day −4, a massive high pressure covered most of Siberia, and temperatures were below normal. However, the signal in the meridional wind was weak at best. The centers of action in the surface air temperature and the SLP are located near Mongolia (50°N, 95°E) at this time. Note the below-normal SLP in the vicinity of the maritime continent. This indicated the existence of a meridional SLP gradient prior to the southward migration of the cold air and enhanced northerlies. Two days later, these centers of action shifted southeastward, and the correlations became stronger. The signal in the $v$ wind also strengthened, with the maximum northerlies located to the east of the SLP and temperature signals. By day 0, the cold air and pressure surge spilled southward along the eastern flank of the Tibetan Plateau, with the strongest correlations located over region 3. By day +2, the cold surges progressed farther south, with the centers of action located over southeastern China. The signal
Fig. 7. Average lag correlations of region 3 surface air temperature with sea level pressure, surface air temperature, and surface $u$ wind (10 m) for the period mid-November through mid-March for 1979/80–1994/95. Positive (negative) correlations greater than 0.2 ($< -0.2$) are shaded white (black). For sea level pressure, negative correlations are plotted for $R < -0.2$ at an interval of 0.1. For surface air temperature and $u$ wind, positive correlations are plotted for $R > 0.2$ at an interval of 0.1.
in the $v$ wind is located even farther south, with the strongest correlation (0.5) near $5^\circ$N, $110^\circ$E. Even on day +4, there still remained a close correspondence between the SLP and temperature centers, having now progressed to the southwest. The $v$-wind signature remained intact over the South China Sea, with the maximum correlation located on the equator at $105^\circ$E. The deeper intrusion of the northerlies into the Tropics than either the SLP or temperature signature is a consistent feature during individual years when the correlations may be even stronger.

The eastward extension is also a robust feature, although it may not appear as significant as the southward component. This is because the anticyclones that travel to the east, unlike those to the south, are subject to immediate and strong deformation by the warm and moist marine boundary once they reach the coast. Cold surges that end near the East China Sea belong to this category and comprise a substantial number of total cold surges (see Fig. 6). The patterns of lag correlations of $T_2$ with temperature, SLP, and surface wind (not shown) are very similar to those presented in Fig. 7.

The frequency, intensity, regionality, trajectory, and propagation patterns of the cold surges presented above have extended and substantiated the results of previous studies (Staff Members of Academia Sinica 1958; Zhu et al. 1981; Ding and Krishnamurti 1987). They are also consistent with the annual cold surge summaries of BMC (1990).

c. Temporal distribution of cold surges and the Siberian high

The temporal distribution of cold surges and the Siberian high is shown in Figs. 8a and 8b. The intensity of the Siberian high, as indicated in Fig. 8b, reaches a peak in December and January. The cold surge frequency, however, exhibits maxima in November and March. This result contradicts a well-accepted rule on the relationship between the intensity of the Siberian high and the occurrence of cold surges (Ding 1994). This phenomenon can be attributed to both dynamical and thermodynamical factors.

During December and January, the high-index flow is characterized by a strong Siberian high and an intense jet stream. Despite the abundance of cold air during this time, this regime is unfavorable for the occurrence of cold surges because the short waves tend to move rapidly through such a pattern without causing any disturbances (Boyle 1986). During the regime transition period of November and March, however, the flow is usually in a low-index mode (Zhu et al. 1981). The typical low-index pattern is characterized by a blocking over the Asian continent and deep troughing along the coast. This regime yields strong northwesterlies over Lake Baikal, which are highly unstable and more ready to release potential energy when triggered by a short wave (Boyle 1986).

Second, an important criterion for cold surge occurrence is the 24–48-h surface air temperature decrease. During December and January, the jet stream is farther south and the surface air temperature is relatively lower. Thus, it takes a very strong anticyclone to induce a substantial temperature decrease in an already cold background. While the surface air temperature is relatively warmer during November and March, cold air and anticyclones with modest intensity can significantly reduce the temperature. These surges, accompanied by the most abrupt temperature decrease, have a profoundly damaging effect on local agriculture and economy (BMC 1990).

Recently, a similar monthly distribution of cold surges was found by Ding and collaborators based on cold surges from six winters (Y. Ding 1996, personal communication). They suggest that, during December and January, the warm and moist troughs moving out of the
Bay of Bengal partly result in the relatively lower surge frequency.

As mentioned before, one of the necessary conditions of cold surge outbreak is that the Siberian high reaches a certain intensity. However, this does not imply that the cold surge frequency should monotonically increase as the Siberian high intensifies. The statistics shown here and Ding’s recent results do not support the suggestion that “the higher the Siberian high pressure, the greater the possibility of cold surge occurrence” (Ding 1994). This is an empirical rule that has been erroneously accepted in the forecasting community.

The temporal distribution of strong cold surges is examined. Of all the 26 strong cases identified, 20 occurred in December, January, and February, five in November, only one in March, and none in October or April. The majority of the strong cold surges occurs during very cold months. This implies that the strong meridional pressure gradient, which is usually associated with massive anticyclones in midwinter, is an important driving force to push cold air far into the Tropics. This result also suggests that the intensity of the Siberian high may be related to the occurrence of strong cold surges that affect the tropical regions.

Given that the cold surge frequency is high and the prevailing northeasterlies dominate the East Asian region in November and March, November through March (NDJFM) should be the most applicable months for studying the East Asian winter monsoon and cold surges, rather than December through February (DJF), which the majority of previous winter monsoon studies have emphasized.

5. Interannual variation

To investigate the possible link between the interannual variation of the East Asian winter monsoon and ENSO, the 1979–95 winter average Southern Oscillation index (SOI) is plotted in Fig. 9. Although the Southern Oscillation is not a standing phenomenon, the evolution of monsoon circulation during ENSO phases is ignored given that the winter monsoon is not examined on a monthly basis. The SOI is very low during the 1982–83 El Niño event. Negative departures of the SOI also occur during the 1986–87, 1989–90, and 1991–92 seasons. The 1988–89 La Niña event has an SOI in excess of 2 hPa.

a. Cold surges

The variation of cold surge frequency is examined first. An interesting feature in Fig. 10 is that the minimum frequencies occur 1 yr after the El Niño events. To test the robustness of this result, we investigated the sensitivity of the cold surge frequency to the SLP threshold in the cold surge definition. Aside from an expected decline of frequency with a stricter SLP criterion, the variation pattern did not change. Observation from BMC also indicated that the surge frequency reached a minimum a year after the 1957–58 and 1965–66 El Niño events. This phenomenon agrees with the 1- to 1.5-yr lag response of the northeast China surface air temperature to the SST near the eastern equatorial Pacific suggested by Bao et al. (1989). Li (1993) suggested that the cold surge frequency reached maximum a year before the El Niño events, and cold surges triggered the El Niño. As indicated in Fig. 10, the surge frequency reached maximum for the 1982–83 and 1986–87 El Niño seasons, although it is not the case for the weak El Niño event during the 1991–92 season.

The cold surges identified above, which are in good agreement with the in situ observation from BMC, typically influence the bulk of East Asia and the coastal regions. A subset of these cold surges can reach to the South China Sea and the maritime continent. This is illustrated in Figs. 11a and 11b, which show a time
series of meridional wind in three regions for the 1987–88 season, along with the region 1 SLP. The phases of SLP are the exact reverse of those of the meridional wind in region 3, which are followed chronologically by similar patterns near Taiwan and near the South China Sea. The lags between the patterns in the regions represent the southeastward propagation of the surface anticyclones.

The typical synoptic scenario of the propagation is as follows. Before the outbreak of a cold surge, the intensification of the Siberian high simultaneously strengthens the northeasterly flow near region 3. The outbreak of the cold surge pushes the anticyclonic flow southeastward and increases the northerlies near Taiwan. The wind then subsequently penetrates into the South China Sea or farther south.

According to the weather forecasters (Li 1996, personal communication) of Hainan Island (18°N, 110°E), whether a cold surge will affect the island and the surrounding oceans depends on its path and intensity. When this region is under the influence of a cold surge, the most conspicuous weather phenomenon is the strong northeasterly wind. The seven strong wind events (northerly wind greater than 7.0 m s$^{-1}$) near the South China Sea, as indicated in Fig. 11a, are a subset of the 11 cold surges identified during the season. Strong wind events for other years are also examined (not shown). We found that, with no exception, every event is associated with a surge outbreak from the extratropics.

Given that the surface meridional wind is the most important indicator of the cold surge activity near Hainan Island, we define the South China Sea cold surge as the days that maximum northerly wind is greater than 7.0 m s$^{-1}$ in the region of 10°–20°N, 110°–120°E. A similar definition has been used by Chang and Chen (1992). According to this definition, the frequency of the South China Sea cold surge is computed and shown in Fig. 12. The interannual variation of cold surge frequency is in good agreement with the SOI ($R = 0.85$). Low frequencies coincide with El Niño and low SOI events; high frequencies accompany La Niña and high SOI years. This agrees with observed strong wind record of Hainan Island (T. Li 1996, personal communication) for winters from 1950 to 80, which indicates that cold air activity is reduced during the years of 1957–58, 1965–66, 1968–69, and 1972–73, all of which are El Niño events.

### b. Meridional wind and SST

The frequency of the South China Sea cold surges not only bears good relationship with the SOI but also largely controls the meridional wind variation over the South China Sea and over even larger areas. Figure 13 shows time series of meridional wind for increasingly larger regions. The patterns of the three regions are
similar to that of the cold surges. Stronger northerlies occur during La Niña and high SOI seasons, and weaker northerlies occur during El Niño and low SOI years. The correlation coefficient between the SOI and the meridional wind near the South China Sea is 0.82. This SOI-like pattern is confined within the region of $0^\circ$–$20^\circ$N, $100^\circ$–$130^\circ$E and is restricted near the surface. Figure 14 is the wind variation near the South China Sea at the 925- and 850-hPa levels. Although the patterns at both levels still bear a certain resemblance to the SOI, the biennial oscillation is the dominant mode. At 700 hPa and above (not shown), the geostrophic flow is decidedly zonal over the entire East Asian continent, and the wind in the northern South China Sea is greatly influenced by the split jet stream south of the Tibetan Plateau. The signature of the SOI is totally lost.

The anomalies of SST and meridional wind (for all 12 months) in the region of $5^\circ$S–$25^\circ$N, $100^\circ$–$130^\circ$E are shown in Fig. 15. This figure is constructed by the monthly anomalies and is smoothed by a nine-point filter. Consistent with Fig. 12, the wind anomaly is in agreement with the SOI. Maximum meridional wind anomalies occur during the three major El Niño events. In the 1989 La Niña event, the wind anomaly is negative. A very interesting feature in this figure is that the SST and wind pattern are similar, and the wind lags SST anywhere from 1 to 5 months. Warm SST anomalies occur a few months after the positive wind anomalies during the three El Niño events; cold SST anomalies occur a few months after most of the negative wind anomalies. The correlation coefficient between the two series is 0.35, and it increases to 0.43 when the two fields are two-month lag-correlated. The lags between the SST and wind suggest that the SST anomalies near the South China Sea are forced by the meridional wind.

c. Large-scale features

Figure 16 presents the number of days that the regional ($45^\circ$–$55^\circ$N, $90^\circ$–$110^\circ$E) maximum pressure is greater than 1050 hPa. Although this is not the only measure of the Siberian high intensity, the variation pattern is in general agreement with the SOI ($R = 0.62$). The average meridional wind over the southeast periphery ($5^\circ$–$20^\circ$N, $160^\circ$–$180^\circ$E) of the Siberian high, another way to represent the intensity of the high pressure, shows the same result as Fig. 16 (not shown).

The variation of the 500-hPa geopotential height anomaly along $135^\circ$E is calculated. Although the overall variation pattern (not shown) is not comparable with the SOI, the trough is weaker than normal during the 1982–83, 1986–87, and 1991–92 El Niño events. The 200-hPa East Asian jet is also examined. This jet is subject to propagation or expansion in the east–west direction during different phases of ENSO. Nonetheless, the jet intensity (at $140^\circ$E) is abnormally weak during the three El Niño events (not shown), which is consistent with the weak Siberian high and 500-hPa trough.
6. Summary and discussion

The study of the East Asian winter monsoon and cold surges from the 1979 to 1995 NCEP/NCAR reanalyses has revealed and extended several noteworthy climatological features of the winter monsoon. According to the conventional surge criteria, about 13 cold surges occur every winter, of which two are strong cases. The average lifetime of the surge is about 7 days and the average intensity is 1053 hPa. The strong cases usually last for 9 days with an average intensity of 1060 hPa.

The cold surges originate from two source regions: one is near the northwest of Lake Baikal and the other is to the north of Lake Balkhash. The high pressure centers propagate southeastward and end in the east and southeast of China and surrounding oceans. Consistent with trajectories of pressure centers, correlation patterns indicate that the signature of the cold surge associated SLP, surface wind, and surface air temperature stretches farther south and east. The southward extent includes the bulk of the maritime continent; the eastward extent reaches 150°E.

Although the intensity of the Siberian high varies according to the season and reaches peak in January, the cold surges are most active during November and March. This result suggests that November through March should be considered as the East Asian winter monsoon season. The result also indicates that the occurrence of cold surges does not monotonically increase with the strengthening of the Siberian high, which disapproved an empirical rule that has been widely accepted in the forecasting community. Examination of monthly distribution of strong cold surges indicates, however, that the majority of strong cold surges occurs from December through February. This result suggests that the strong meridional pressure gradient is an important driving force to push the cold air far into the Tropics.

The two stratifications of cold surges are revealing when considering the relationship between ENSO and the interannual variation of the cold surges. The first stratification, as described in the cold surge criteria, yields a result that is consistent with the in situ observation from BMC. The interannual variation of cold surge frequency shows that minimum surge frequency occurs 1 year after an El Niño event. The second stratification, based on the maximum wind event near the South China Sea, indicates that the interannual variation of the South China Sea cold surges is in good agreement with the SOI. High surge frequency is found during La Niña and high SOI years, low frequency is associated with El Niño and low SOI years.

The winter meridional wind near the South China Sea hinges on the South China Sea cold surges. The interannual variation of the wind is well correlated with the SOI. Strong winds are associated with La Niña and high SOI events; weak winds are found during El Niño and low SOI events. This variation pattern is restricted north of the equator (0°–20°N, 100°–130°E) and confined to the near-surface layer. The SOI-like surface wind variation is a very intriguing phenomenon. The physical mechanisms that result in this feature and its implication deserve to be explored by further observational analysis and modeling studies. Such an effort is currently under way.

Throughout the year, the interannual variation of the meridional wind anomaly near the South China Sea is also related to the SOI. The SST variation is similar to the wind but lags the wind by approximately 1–5 months. The lags between the fields suggest that the SST variation is forced by the wind. If this is proved, the analysis of the winter monsoon intensity can be useful for the prediction of the Indian summer monsoon, given that the SST near the South China Sea is linked to all India monsoon rainfall (Tomita and Yasunari 1996).

The strength of the 200-hPa jet stream and the 500-hPa trough are weaker than normal during El Niño events, although their variation patterns do not agree with the SOI as well as that of the Siberian high. The interannual variation of the Siberian high, measured by means of both pressure and wind, is in general agreement with the SOI.

Given the close relationship between the SOI and the interannual variation of the monsoon wind, the cold surges, and the Siberian high, it is natural to wonder why some of the monsoon-related disturbances, either originated from or located in the extratropics, exhibit variation patterns similar to the SOI, and what the physical mechanisms are between the ENSO–winter monsoon interaction.

During a typical El Niño year, the convective activity in the maritime continent is usually reduced. The suppression of the convection should be responsible for the weaker than normal monsoon circulation during the El Niño events. With the available observed outgoing long-
wave radiation (OLR) data from the National Oceanic and Atmospheric Administration, we computed the OLR anomaly near the maritime continent (5°N–5°S, 110°–120°E) from 1979 to 1992. The overall variation pattern (not shown here) is not as close to the SOI as that of the South China Sea cold surges or the Siberian high. However, negative anomaly is found during the 1988–89 La Niña season; likewise, positive OLR anomalies are found during the 1982–83 and 1986–87 El Niño seasons. The correlation coefficient between the OLR anomaly and the SOI is |R| = 0.5.

On the other hand, evidence indicated that the winter monsoon and cold surges can greatly affect the tropical convection and SST (Chang et al. 1979). Whether a cold surge can significantly affect the tropical region, as has been discussed, depends on its path and intensity. It may also depend on the large-scale background circulation. Theoretical studies have suggested that the interaction between the tropical convection and midlatitude disturbances depends critically on the relative position of the convection and the extratropical quasi-stationary waves (Branstator 1983; Lau and Li 1984). It is conceivable that the tropical convection can be favorable or unfavorable for the southward penetration of cold surges on interannual and shorter timescales. We speculate that the ENSO-related variation of the tropical phenomena may be responsible for the interannual variation of the South China Sea cold surges, although the surges also directly affect the tropical convection and SST.

The short-term variation of the Siberian high intensity, as indicated in Fig. 11, largely regulates the strong wind event near the South China Sea. A remarkable feature in Fig. 11a is the intensification of wind from region 3 to the South China Sea. The significant increase in the wind magnitude also occurs during other seasons. This phenomenon indicates that the South China Sea cold surge is not purely an extratropical event, and the interaction between the cold surges and the tropical atmospheric and oceanic phenomena is responsible for the intensification of the wind. Apparently, on the interannual timescale, the SOI-like wind variation comes from the combined effect of the Siberian high and the air–sea interaction near the South China Sea. In other words, the air–sea interaction modified the interannual variation of the wind from a pattern similar to Fig. 16 into the pattern in Fig. 13. Two fundamentally important questions are: Why does the Siberian high behave the way it does on an interannual timescale (P. Webster 1997, personal communication)? In what manner did the surface wind interact with the SST near the South China Sea so that the interannual variation of the wind was modified? The answer to these questions should shed some light on the mechanisms of ENSO–winter monsoon interaction.

Finally, although the current study has indicated the close association between ENSO and the interannual variation of the winter monsoon and cold surges, physical mechanisms responsible for their interaction are still not well understood, especially questions of if and how the winter monsoon affect ENSO through its impact on tropical SST and convection near the maritime continent. The mechanisms that regulate the Siberian high on the interannual timescale and the air–sea interaction near the South China Sea are also very important problems that deserve more attention. Further studies are needed to understand the winter monsoon and its interaction with tropical atmospheric and oceanic phenomena with more reliable, high-quality observation as well as numerical experiments. Simulations with fully coupled atmospheric–ocean models and GCM simulations with specified SST, such as those from the Atmospheric Model Intercomparison Project, should provide an excellent opportunity to address these problems.

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