Atmospheric Circulation Anomalies and Key Physical Processes behind Two Categories of Anomalous Eurasian Spring Snowmelt

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(Manuscript received 18 January 2023, in final form 26 May 2023, accepted 26 May 2023)

ABSTRACT: Eurasian spring snowmelt plays an important role in the subsequent climate and hydrological cycle, however, the understanding of snowmelt itself and its causes remains insufficient. This study explored the basic characteristics of spring snowmelt in the eastern Europe–western Siberia (EEWS) region by classifying snowmelt anomalies into two categories based on the different factors that dominate spring snowmelt, and then investigated the associated atmospheric circulation anomalies and local physical processes. The first category of anomalous snowmelt (category 1) is controlled by both the initial snow mass and the later snowmelt process, while the second category of anomalous snowmelt (category 2) is mainly linked to the later snowmelt process. Specifically, category 1 is characterized by an anomalous trough in EEWS in winter, where water vapor transported and converged, accompanied by anomalous upward motion, which promotes snowfall and snow accumulation, providing initial conditions conducive to snowmelt. In April, this region is controlled by an anomalous ridge, with significant warm advection anomalies and subsidence promoting surface warming, thereby accelerating snow melting. In contrast, the winter circulation anomalies are insignificant in category 2, while the anomalous ridge in April is stronger than in category 1, accompanied by more intense snowmelt processes. In addition, from the surface energy balance perspective, atmospheric downward sensible heat transport is an important factor influencing the anomalous snowmelt in category 1, while shortwave radiation plays a secondary role. Conversely, the snowmelt in category 2 is dominated by shortwave radiation forcing, but the sensible heat effect is slightly weaker.

SIGNIFICANCE STATEMENT: Eurasian spring snowmelt significantly impacts the subsequent climate and hydrological cycle, but the understanding of snowmelt itself and its causes is still inadequate. The purpose of this study is to explore the monthly evolution of atmospheric circulation associated with anomalous snowmelt and its local physical processes associated by categorizing them based on snowmelt characteristics. Category 1 is jointly affected by winter snow accumulation and later warming, while category 2 is dominated by strong snowmelt process in late spring. These two categories are accompanied by different winter and spring circulation configurations. Our results provide a basis for further investigation of snowmelt precursor signals.

KEYWORDS: Atmosphere-land interaction; Snowmelt/icemelt; Energy budget/balance; Anomalies; Local effects

1. Introduction

As an important component of the cryosphere, snow cover exerts significant impacts on climate and hydrological cycle by altering the energy and water balance (Barnett et al. 1988, 1989; Groisman et al. 1994; Zhang 2005; Yang et al. 2007; Matsumura and Yamazaki 2012; Henderson et al. 2018). During winter and spring, snow covers most areas of the Eurasian continent, exhibiting significant seasonal and interannual variabilities. Previous studies suggested that the Eurasian snow can produce a crucial influence on the climate and has been regarded as an important predictor of short-term climate (Cohen and Rind 1991; Wu and Kirtman 2007; Yim et al. 2010; Xu and Dirmeyer 2011, 2013; Wu et al. 2014). It is found that spring snowmelt over Eurasia can induce soil moisture anomalies and produce delayed climatic effects through its hydrological effect, which can further affect summer monsoon, air temperature, and rainfall in the following summer (Zhao et al. 2007; Wu et al. 2009; Halder and Dirmeyer 2017; Shen et al. 2020; Cheng et al. 2022). Therefore, it is necessary to make the causes of the snow variations much clearer to advance our understanding of the climatic effects of snow anomalies and to improve the skill of climate prediction.

Snow itself is a product of the atmospheric motion, which exhibits an evident response to large-scale atmospheric circulation (T. Wang et al. 2015; Ye and Lau 2017). The geographic distribution of snow is also sensitive to climate elements such as surface air temperature, water vapor transport and wind (Eijima et al. 2007; Zhang et al. 2021), which are closely associated with the atmospheric general circulation (Trenberth 1995). Anomalous atmospheric circulation always provides important clues for understanding the reasons behind the snow variations. In addition, the atmospheric teleconnections can alter the temperature and precipitation together with the...
thermodynamic and dynamic processes of the atmosphere, and further affect regional snow changes (Ge and Gong 2008; Seager et al. 2010; Kim et al. 2013; Yeo et al. 2017). Studies suggested that the winter Eurasian snow cover bears a significant negative correlation with the North Atlantic Oscillation (NAO). The positive phase of NAO tends to enhance the zonal temperature advection, leading to increased air temperature and decreased snow cover over Europe (Hurrell 1995; Clark et al. 1999; Henderson and Leathers 2010; Kim et al. 2013). Besides the atmospheric internal variability, external forcings such as sea ice and sea surface temperature (SST) anomalies are also crucial factors that result in snow cover variations by modulating the atmospheric circulation (Corti et al. 2000; Ghatak et al. 2010; Cohen et al. 2012; Ghatak et al. 2012; Ye et al. 2015; Sun et al. 2019).

Previous studies paid more attention to the causes of snow cover change in winter, and there are relatively few studies on spring snow mass change. In fact, the spring snow hydrological effect is prominent. The delayed climatic effect of snowmelt is a significant contributor to the Eurasian summer climate. Hence, it is crucial to understand the variability of spring snow mass. Zhang et al. (2021) explored the main drivers of the interannual variations of Eurasian spring snow cover from the perspective of the seasonal mean. Meanwhile, the observations show that the spring snow melts rapidly and the snow mass varies widely in springtime (Zhang et al. 2017; Sun et al. 2019; Ye and Lau 2019). Therefore, it is far from sufficient to understand spring snow mass only from the perspective of the seasonal mean. In particular, snowmelt deserves further investigation, since the processes that regulate it are still not well understood. Ye (2019) investigated the influencing factors of March SWE, which represents the initial snow mass for snowmelt. These findings provide clues for exploring the causes contributing to snowmelt. As we know, spring snowmelt can be used to represent the hydrological effect of snow, and the snowmelt amount is closely related to both the initial snow mass and the intensity of the following snowmelt. However, a comprehensive consideration of its influencing factors is still inadequate, and a deeper exploration can help to promote a better understanding of the characteristics of snowmelt change, so as to enhance knowledge about the role of snowmelt in climate change. Studies suggested that spring snowmelt over eastern Europe–western Siberia (EEWS) exhibits evident interannual variability and has important influences on both summer temperature and rainfall over East Asia (Zhang et al. 2017; Sun et al. 2021). To further improve the understanding of the climatic effects of snowmelt, it is necessary to make the causes of spring snowmelt anomalies in the EEWS region clear.

This study aims to investigate the variation of spring snowmelt in Eurasia and its relationship with local atmospheric circulation, and categorically explores the characteristics of snowmelt anomalies under the influence of different dominant factors and their corresponding local physical processes. The article is organized as follows. Section 2 introduces the data and methods used. Section 3 presents the main results. Section 3a shows the classification of spring snowmelt anomalies and their corresponding characteristics. Section 3b presents the atmospheric circulation background associated with the snowmelt anomalies. The related local atmospheric dynamic and thermodynamic processes are explored in section 3c, and the local energy budget at the snow surface is discussed in section 3d. A summary and discussion are given in section 4.

2. Data and methods

The snow data used in this study are the monthly gridded GlobSnow v3.0 snow water equivalent (SWE) by the Finnish Meteorological Institute (FMI) (Luojus et al. 2020), whose detailed information is available at http://www.globsnow.info/. The original data are on the Northern Hemisphere 25-km Equal-Area Scalable Earth Grid (EASE-Grid) and were converted to regular 1° latitude–longitude grid data in our analysis. Constructed by combining satellite-based passive microwave radiometer data (Nimbus-7 SMMR, DMSP 5D2 SSMI, and DMSP 5D3 SSMIS) with ground-based synoptic snow depth observations, this snow product has relatively high spatial resolution and large time coverage, and well represents the real snow conditions (Xu et al. 2019a). Other surface and atmospheric fields are from the ERA5 monthly mean reanalysis data, including 2-m temperature, surface heat flux, geopotential height, air temperature, and so on (Hersbach et al. 2019a,b), which are available at https://cds.climate.copernicus.eu/#!/home. The ERA5-land (Muñoz Sabater 2019) monthly surface heat and radiant flux were also used to verify the results of energy balance analysis at the snow surface. To investigate the possible linkage between large-scale atmospheric teleconnection patterns and the anomalous atmospheric circulation patterns associated with snowmelt, the climate indices are acquired from the Climate Prediction Center (CPC) website (https://www.cpc.ncep.noaa.gov/data/teledoc/telecontents.shtml), which includes the monthly Scandinavia (SCAND) pattern and east Atlantic (EA) pattern indices. The study period is from 1981 to 2018, which is the common range of all these datasets. The linear trends were removed from all variables to avoid the effect of long-term tendency on correlation.

According to previous studies (Dey and Kumar 1982; Zhang et al. 2017), spring snowmelt amount was defined by the difference in monthly mean SWE between March and May (DSWE). Based on T. Wang et al. (2015), the energy balance at the snow surface was used to analyze the surface energy budget during snow melting. The net energy flux at the snow surface is described by the following equation:

\[ \Delta Q = \text{LW}_{\text{net}} + \text{SW}_{\text{net}} + \text{SH} + \text{LH} + G + Q_P. \]  

(1)

In which, \( \Delta Q \) is the energy used for melting snow at the surface. \( \text{SW}_{\text{net}} \) and \( \text{LW}_{\text{net}} \) indicate net shortwave radiation and net longwave radiation, which are defined as the difference between downward and upward radiation fluxes. \( \text{SH} \) and \( \text{LH} \) represent sensible and latent heat fluxes, respectively. Compared to the radiation and turbulent fluxes, the ground heat flux (\( G \)) and the energy supplied by rain (\( Q_P \)) are relatively small components in the energy balance of typical snow melting in spring (Male and Granger 1981). Thus, these two items are not considered in our study. In Eq. (1), all terms are defined positive downward.
3. Results

a. Basic features of spring snowmelt and its classification

The Eurasian snow begins to melt in spring, resulting in a large month-to-month variation in snow mass. Figure 1 shows the spatial distribution characteristics of the climatology and interannual variability of spring snowmelt in Eurasia. The spring snowmelt in Eurasia is widespread, being generally the largest around 60°N. There are two maxima snowmelt centers in eastern Europe and western Siberia, respectively. The spring snowmelt in northern western Siberia has the largest interannual variation. In addition, there is also a large snowmelt variability in the East European Plain. According to the basic characteristics of the spring snowmelt, the EEWS region (55°–70°N, 50°–95°E) is selected as the key area for the Eurasian spring snowmelt, which covers the main maxima of both spring snowmelt and its variability. Moreover, previous studies have shown that spring snowmelt anomalies in this region have important climatic effects (Zhang et al. 2017; Sun et al. 2021; Cheng et al. 2022). Therefore, the spring snowmelt anomalies in EEWS are mainly explored later, and the standardized area-averaged snowmelt time series in this region is defined as the spring snowmelt index (DSWEI) for subsequent analysis.

For spring snowmelt, the initial state of the snow mass and the subsequent melting process are important factors to determine the snowmelt amount. The initial snow state can be represented by the snow mass at the beginning of spring (SWE in March, SWE3). According to the definition of spring snowmelt, besides SWE3, the May SWE (SWE5) also determines the spring snowmelt anomalies, which indicates the final state of the spring SWE after melting. Figure 2a shows DSWEI and the standardized area-averaged SWE index in March and May (SWE3I and SWE5I) in EEWS during 1981–2018. The DSWEI can be divided into four quadrants according to positive and negative anomalies of SWE3I and SWE5I. We selected 0.4 standard deviations of DSWEI as the threshold of snowmelt anomalies and identified 29 years with high and low snowmelt anomalies for further analysis.

Both positive SWE3 anomalies and negative SWE5 anomalies contribute positive anomalies in spring snowmelt. As

![Fig. 1](image1.png)

**FIG. 1.** The (a) climatological mean and (b) interannual standard deviation of spring snowmelt (DSWE; mm) in Eurasia from 1981 to 2018.

![Fig. 2](image2.png)

**FIG. 2.** (a) Scatterplots of spring snowmelt anomalies vs March and May SWE anomalies in the EEWS region, (b) original time series of SWE3 (red solid line) and SWE5 (blue solid line) and spring snowmelt amount (shading between the two solid lines), and (c) standardized time series of spring snowmelt anomalies with anomalous snowmelt classification.
Table 1. Classification of anomalous snowmelt in spring.

<table>
<thead>
<tr>
<th>Types of DSWE anomalies</th>
<th>Descriptions</th>
<th>Details</th>
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<tbody>
<tr>
<td>Category 1</td>
<td>Affected by initial snow mass anomalies and snowmelt intensity</td>
<td>7 high years (quadrant IV) SWE3 &gt; 0, SWE5 &lt; 0 5 low years (quadrant II) SWE3 &lt; 0, SWE5 &gt; 0</td>
</tr>
<tr>
<td>Category 2</td>
<td>Affected by the intensity of the snowmelt process</td>
<td>6 high years (quadrant III) SWE3 &lt; 0, SWE5 &lt; 0 6 low years (quadrant I) SWE3 &gt; 0, SWE5 &gt; 0</td>
</tr>
<tr>
<td>Category 3</td>
<td>Affected by initial snow mass anomalies</td>
<td>2 high years (quadrant I) SWE3 &gt; 0, SWE5 &gt; 0 3 low years (quadrant III) SWE3 &lt; 0, SWE5 &lt; 0</td>
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Fig. 2a illustrates, the years with positive DSWEI anomalies are concentrated in the three quadrants of the lower-right region (I, III, IV), while the negative anomalies years are located in the three quadrants of the upper-left region (I, II, III). But more specifically, there are evident differences in the configurations of March and May snow anomalies corresponding to both positive and negative snowmelt anomalies. According to the different contributions of SWE3 and SWE5 to spring snowmelt, the snowmelt anomalies in EEWS can be classified into the following three cases (see Table 1). In the first case (category 1), the anomalies in SWE3 and SWE5 have the opposite sign, and therefore a synergistic effect on the snowmelt anomaly. In categories 2 and 3, the anomalies in SWE3 and SWE5 have the same sign, and their effect on the snowmelt therefore partly cancels out. In category 2, the anomaly in SWE5 dominates, whereas the anomaly in SWE3 dominates in category 3. The annual SWE3 and SWE5 characteristics along with the corresponding amount of spring snowmelt are displayed in Fig. 2b for further understanding of the classification. Figure 2c shows the DSWEI histogram and the categories of annual snowmelt anomalies. It is found that category 1 and category 2 are the main types of spring snowmelt anomalies in EEWS. In contrast, category 3 has fewer snowmelt anomalies and insufficient samples, which limits the analysis. Therefore, we focused on category 1 and category 2 snowmelt anomalies, and analyzed the characteristics of the two types of snowmelt anomalies.

To compare the differences between the two types of anomalous snowmelt, Fig. 3 illustrates the spatial patterns of spring SWE anomalies corresponding to category 1 and category 2. For category 1, the SWE anomalies in March are positive, while the SWE anomalies in April and May are negative. This indicates that there is a certain degree of anomalous snow accumulation in March and the preceding winter months, while April and May are periods of anomalous snow melting. The anomalous accumulation of snow in the earlier period and the anomalous snow melting in the later period jointly benefit more snowmelt in spring. For category 2, SWE in EEWS is anomalously low from March to May, and the negative SWE anomalies in April and May are more obvious. The intensity of the negative anomalies is greater than that in category 1, reflecting the stronger spring snowmelt.

b. Anomalous circulation patterns associated with two categories of anomalous snowmelt

The changes in SWE are closely tied to the atmospheric circulation, and the atmospheric circulation anomalies are an essential cause of snow mass anomalies. To reveal the circulation anomalies associated with the two categories of anomalous snowmelt, Fig. 4 presents the composite differences of the corresponding circulation fields. As the geopotential height anomalies in May are not significant over EEWS, the May circulation fields are not shown here. The results show that there are significant negative anomalies in 500-hPa geopotential height over EEWS in category 1 from January to March (JFM), and they turn into positive anomalies in April, accompanied by anomalous anticyclonic circulation at 850 hPa. For category 2, there are no significant geopotential height anomalies over EEWS in JFM, but the distribution of large-scale circulation anomalies is nearly reversed from category 1. In April, significant positive geopotential height anomalies appear in EEWS, whose position is slightly eastward from those in category 1. There is a good correspondence between the circulation anomalies and the SWE anomalies. In JFM, the anomalous trough accompanied by anomalous cyclonic circulation provides a favorable background circulation for snowfall, which benefits the snow accumulation in March. In April, the anomalous ridge over EEWS in category 1 and category 2 correlates to the anomalous low SWE in late spring. The local dynamic and thermodynamic processes accompanied by strong positive geopotential height anomalies can promote the snow melting process. In general, category 1 is influenced by the anomalous trough in JFM and the anomalous ridge in April, which is conducive to the snow accumulation in March and the accelerated snowmelt in the later period, leading to excessive snowmelt in spring. Category 2 is mainly dominated by the anomalous ridge in April, which further affects the snowmelt in late spring, resulting in larger snowmelt amount.
Furthermore, JFM and April atmospheric circulation anomalies were regressed onto SWE3I and SWE5I, respectively (Figs. 5a,b). When there is more SWE in March, similar to the category-1 circulation in JFM (Fig. 4a1), the geopotential height is anomalously low in EEWS. Meanwhile, anomalously high geopotential height appears in the northern Mediterranean and southern Baikal. When the SWE is lower in May, it corresponds to significant positive geopotential height anomalies over EEWS in April. In addition, over upstream of EEWS, a significant anomalous trough exists in northern Scandinavia and extends southward to the Black Sea, while anomalous ridge is found on the west coast of Europe, which is also generally consistent with the April circulation of the two categories (Fig. 4b).

We further investigated the relationship between SWE3I and SWE5I and teleconnection indexes, respectively. It is found that SWE3I is related to the EA pattern in JFM ($r = 0.34$, exceeds the 95% confidence level), which is characterized by anomalous low pressure centered over the eastern Atlantic Ocean and high pressure over northern Europe and western Russia. The circulation anomalies regressed on the JFM EA index (Fig. 5c) are generally similar to those in Fig. 5a, with the anomalies centers slightly shifted southward, implying the potential influence of EA teleconnection on snow accumulation in March. For May SWE, there is a close relationship between SWE5I and the April SCAND index ($r = 0.64$, exceeds the 99% confidence level). The SCAND pattern features a positive pressure anomaly over Scandinavia and negative anomalies over the North Atlantic and Siberia, and its spatial characteristics (Fig. 5d) resemble those of the circulation anomalies associated with the May SWE (Fig. 5b), indicating that the SCAND teleconnection in April may impact the snow changes in May in EEWS.

c. Local processes of snowmelt

Temperature and precipitation are the products of the atmospheric circulation, both of which can directly control snow mass changes (Shinoda et al. 2001; Ueda et al. 2003; Iijima et al. 2007; T. Wang et al. 2015; Ye 2019). Upon elucidating the circulation background that corresponds to the two anomalous snowmelt types, it is necessary to further investigate the role of precipitation and temperature in these two types and their respective local physical processes.
Figure 6 shows the basic characteristics of the snowfall corresponding to the two categories of anomalous snowmelt. For category 1, there is more snowfall in EEWS in JFM, with the largest positive anomalies concentrated in the southern part of EEWS, whereas snowfall anomalies are negative throughout EEWS in April. In contrast, category 2 has weak negative snowfall anomalies in JFM, but significant negative snowfall anomalies appear in April with stronger intensity than category 1. In fact, due to spring temperature rise, snow mass changes in late spring are more sensitive to temperature, and the snowfall in EEWS mainly occurs in winter, so the April snowfall is not discussed in focus. As for the JFM snowfall, the anomalously high snowfall for category 1 favors snow accumulation in March, providing sufficient initial snow mass for anomalous snowmelt, while category 2 is not closely related.

The Eurasian snow accumulates during winter, reaching a peak in March, and then begins to melt gradually (figure omitted). Temperature is the main factor dominating the intensity of snowmelt in spring. Figure 7 illustrates the surface temperature anomalies in spring under the two types of anomalous
snowmelt. No significant temperature anomalies are observed in EEWS in March for these two categories, while positive temperature anomalies in EEWS are more significant in April and relatively weak in May.

For category 1, there are significant positive surface temperature anomalies in the EEWS region in April, corresponding to the anomalies of April atmospheric circulation. In May, the positive temperature anomalies weaken and shift eastward. Due to the anomalously high surface air temperature, there is a significant increase in snowmelt amount during April and May, particularly with a larger contribution of April snowmelt to the total spring snowmelt anomalies (Fig. 8). This anomalously high snowmelt also further influences the anomalously low SWE in April and May (Fig. 3). Interestingly, it is found that the positive temperature and snowmelt anomalies are more significant in April than in May, while the above analysis indicates that negative SWE anomalies are more pronounced in May. This may be due to the excessive SWE in March for category 1. The large amount of snowmelt in April offsets positive SWE anomalies in the previous period, so SWE anomalies in April are small. However, with warm anomalies in both April and May, the SWE in May drops below the average.

The distribution of April surface temperature anomalies in category 2 is similar to that in category 1. The intensity of the anomalies in EEWS is stronger than that in category 1, and the largest positive anomalies occur further east, which corresponds to the circulation anomalies. The positive temperature anomalies in May are mainly located in the northern part of EEWS. Snowmelt amount anomalies in April and May are consistent with the characteristics of the temperatures, with larger snowmelt occurring mainly in April, and a northward contraction in the extent of positive snowmelt anomalies in May (Fig. 8). The SWE anomalies also change accordingly. For category 2, the slight negative SWE anomalies in March together with the acceleration of snowmelt by the anomalously high temperatures in late spring result in large negative SWE anomalies in April and May.

It is noted that there are high SWE anomalies in EEWS in March for category 1, but no significant anomalies in the surface temperature. This may be related to the fact that the temperatures in high latitudes are generally low at that time, making it easy to meet the temperature conditions for the snow accumulation, thus the temperature anomaly is not a primary factor affecting the SWE anomaly in March. On the other hand, there is feedback between the snow and the atmosphere. With less snowmelt in March, the snow mainly affects the atmosphere through the albedo effect. At this time, the EEWS area is almost completely covered by snow. Despite the high SWE anomalies, the snow cover extent changes little, and the corresponding snow albedo changes are weak, resulting in a weak impact of SWE change on the temperature in March.

Based on the above analysis, category 1 is related to the excessive JFM snowfall and later higher temperature under the background of anomalous circulation, while category 2 is mainly influenced by late spring warming. Further investigation is needed to understand the causes behind the corresponding anomalies in snowfall and temperature.

Figure 9a shows the vertically integrated moisture flux divergence and 850-hPa water vapor flux anomalies in JFM for category 1. There are significant negative divergence
anomalies of vapor flux over EEWS, and strong water vapor transport eastward and northward in the south and east sides of EEWS. The anomalous water vapor transport corresponds to the circulation anomalies. Under the background of the anomalous cyclonic circulation in JFM, the water vapor from the Atlantic Ocean is transported eastward to the south of EEWS along with the anomalous westerly winds and further transported northward with the anomalous southerly winds in front of the anomalous cyclonic circulation. Consequently, a large-scale water vapor convergence occurs in EEWS, providing favorable water vapor conditions for the snowfall.

The cross section of $50^\circ$–$95^\circ$E zonal mean horizontal divergence and vertical meridional circulation is shown in Fig. 9b. There is significant low-level convergence and upper-level divergence in the southern part of EEWS, accompanied by anomalous upward motion throughout the troposphere. The anomalous upward motion also has a southerly component and turns downward in high latitudes, inducing low-level divergence on the north side of EEWS. The lower-level convergence and upper-level divergence in the southern part of EEWS favor the maintenance of ascending motion, providing beneficial dynamical conditions for the snowfall. The stronger southerly airflow over the northern side of the EEWS also corresponds to the water vapor transport caused by the anomalous southerly winds (Fig. 9a).

Overall, category 1 is controlled by anomalous low 500-hPa geopotential height and cyclonic circulation maintained from winter to early spring. Water vapor from the Atlantic Ocean and lower latitudes is transported to the EEWS region, producing a strong convergence area in the lower levels accompanied by anomalous upward motion, which provides favorable water vapor and dynamic conditions for the snowfall, thus promoting anomalous snow accumulation, leading to positive SWE anomalies in March.

Further analysis on the causes of the temperature anomalies was performed. Given that the large-scale significant temperature anomalies mainly occur in April, we focused on April temperature changes under the two categories of anomalous snowmelt. Figure 10a displays the characteristics of 850-hPa temperature advection anomalies in April. For category 1, the temperature advection anomalies are significantly positive in most of the EEWS, indicating that this area is influenced by
warm advection, which is beneficial to the temperature increase here. For category 2, there are also significant positive anomalies of temperature advection in EEWS, which is stronger in intensity and more easterly in extent than category 1. This may be related to the fact that the anomalous ridge corresponding to category 2 is more eastward. The southerly wind anomalies on the west side of the ridge bring anomalous warm advection favoring the local temperature rise. In general, under the background of circulation anomalies, the anomalous warm advection in the EEWS region can promote warming and facilitate snow melting.

In addition to the temperature advection, vertical motion can also cause temperature changes. Figure 10b shows the meridional cross sections of the zonal mean temperature and vertical velocity at 50°–95°E under the two types of anomalous snowmelt. There are significant positive temperature anomalies in the lower and middle troposphere in EEWS and anomalous downward motion in its southern part for these two categories. The positive temperature anomalies are stronger in category 2 than in category 1, which is consistent with the surface temperature (Fig. 7b). And the anomalous subsidence of category 1 is slightly stronger than that of category 2. Generally, both types have anomalous downward motion in EEWS, which favors adiabatic warming and leads to an increase in near-surface temperature, thus further accelerating the spring snowmelt.

It is noteworthy that the anomalous subsidence occurs mainly in the southern region of EEWS, where the horizontal temperature advection anomalies are weak (Fig. 10a). This suggests that the downward motion and the associated adiabatic warming are the main factors contributing to the warming in this region. In contrast, in the rest of EEWS, the anomalous

FIG. 8. Composite difference of snowmelt amount (mm) for (left) category-1 and (right) category-2 anomalous snowmelt between high and low DSWEI years in (a) April and (b) May. The black (gray) dotted areas exceed the 95% (90%) confidence level.

FIG. 9. Composite difference of (a) vertically integrated moisture flux divergence (shading; $10^{-3}$ kg m$^{-2}$ s$^{-1}$) and 850-hPa water vapor flux (vectors; $10^{-3}$ kg kg$^{-1}$ m s$^{-1}$), and (b) 50°–95°E zonal mean horizontal divergence (shading; $10^{-6}$ s$^{-1}$) and vertical meridional circulation (vectors) in JFM for category-1 anomalous snowmelt between high and low DSWEI years. The dotted areas exceed the 90% confidence level.
warm advection appears to be the dominant factor in the surface warming.

d. Radiative forcing

Radiative forcing is a crucial factor affecting land surface temperature and plays a vital role in the snowmelt (Dong et al. 2001; Ohmura 2001; Shi et al. 2013; T. Wang et al. 2015). The budget analysis of the surface energy balance has been further carried out to disclose the specific roles of various energy fluxes in snowmelt for those two different types of anomalous snowmelt. Figure 11 presents the anomalies of the area-averaged energy flux at the snow surface [items of Eq. (1)] in EENS during the spring months for the two types of anomalous snowmelt and the characteristics of other elements affecting the energy flux changes. Both ERA5 and ERA5-land datasets were used to analyze the energy balance, and the characteristics of the flux anomalies obtained from these two datasets are consistent. For clarity purposes, only the results from the ERA5 dataset were chosen for presentation.

For category 1, negative $SW_{net}$ and positive $LW_{net}$ are found in March, while SH and LH anomalies are ignorable. There is a weak positive anomaly in $\Delta Q$, which favors anomalous snowmelt. Specifically, evidently increased total column water vapor (TCWV) and total cloud cover (TCC) in March can lead to an anomalous increase in downward longwave radiation ($LW_d$), thereby resulting in increased $LW_{net}$, providing an important energy source for the snowmelt in March. In April, SH and $SW_{net}$ anomalies are both positive, LH and $LW_{net}$ exhibit weak negative anomalies, and $\Delta Q$ shows a significant positive anomaly. Among them, SH plays a leading role, followed by $SW_{net}$. The strong positive SH anomaly indicates a continued heat transport from the atmosphere to the surface, which provides additional energy for the snowmelt. Negative LH anomaly may be associated with heat absorption during the snowmelt (Shi et al. 2013). $SW_{net}$ increases despite a slightly negative anomaly in downward shortwave radiation ($SW_d$), most likely due to the reduction of surface albedo during the snowmelt. In May, positive SH anomaly decreases, and the heat transfer from the atmosphere to the surface is weaker than that in April. At the same time, negative LH anomaly decreases, and the heat transfer from the atmosphere to the surface is weaker than that in April. At the same time, negative LH anomaly is enhanced, while $LW_{net}$, $SW_{net}$ and $\Delta Q$ have no significant anomalies, and the anomalous snowmelt weakens accordingly. Overall, the anomalous snowmelt process in category 1 occurs mainly in April, when it is mainly energized by the atmospheric SH transport to the surface, with $SW_{net}$ playing a secondary role, and anomalous high $LW_{net}$ in March may contribute to the initiation of the snowmelt (T. Wang et al. 2015).
For category 2, all surface flux anomalies in March are weak, and \( \Delta Q \) is close to zero, indicating no significant anomalous snowmelt. In April, the magnitude of positive SH anomaly is relatively consistent with that of category 1, while positive anomaly of SW\(_{\text{net}}\) is stronger than category 1, and \( \Delta Q \) anomaly is accordingly stronger, which is more conducive to the snowmelt. Meanwhile, the effect of SW\(_{\text{net}}\) on the snowmelt is primary, and SH has the second effect. The positive anomaly of SW\(_{\text{net}}\) is stronger in May than in April. It is found that SW\(_d\) anomaly is weak in May, and positive SW\(_{\text{net}}\) anomaly is mainly influenced by decreased upward shortwave radiation (SW\(_u\)) anomaly due to reduced surface albedo after the snowmelt. SW\(_d\) anomaly changes little between April and May while positive SW\(_{\text{net}}\) anomaly gradually increases from April to May, which reflects the positive feedback effect of snow cover–albedo during the snowmelt period. Except for SW\(_{\text{net}}\), SH, LH, and LW\(_{\text{net}}\) have all negative anomalies in May. Specifically, negative SH anomaly indicates a shift in the direction of anomalous heat transfer between the land surface and the atmosphere in May compared with that in April. Negative LH anomaly may be related to anomalous melting of snow or evaporation of soil moisture after the snowmelt. Net flux \( \Delta Q \) in May presents a negative anomaly, indicating weakening snowmelt process, which may be due to less SWE remaining to melt in May.

In summary, the anomalous snowmelt process of category 1 and category 2 mainly occurs in April, with SH and SW\(_{\text{net}}\) as the major influencing factors. For category 1, heat for snowmelt is provided mainly through enhanced atmospheric SH transport to the surface under April anomalous atmospheric circulation, with shortwave radiation playing a secondary role. For category 2, the atmospheric heat transport to the surface in April is comparable to that of category 1. Besides, the role of shortwave radiation occupies a more dominant role in the snowmelt anomaly in EEWS. This can be attributed to the consistently lower SWE during spring for category 2, which results in reduced albedo. As a result, there is a stronger positive feedback loop between snow and albedo, ultimately intensifying the melting process (Abe 2021).

4. Conclusions and discussion

a. Conclusions

Based on the GlobSnow v3.0 SWE data and the ERA5 reanalysis data, this article classified the spring anomalous snowmelt in the EEWS region according to two important factors, namely, the initial snow amount and the later snowmelt process. The characteristics of SWE changes in the two main categories of anomalous snowmelt together with the associated atmospheric circulation anomalies and their local dynamic/thermal processes were also explored. Further, we compared the similarities and differences between these two categories of anomalous snowmelt to deepen the understanding of spring snowmelt. The detailed processes of these two categories are shown in Fig. 12. The main conclusions are as follows.

The spring snowmelt anomalies are closely related to the initial snow mass and the following snowmelt process, which
can be characterized by the SWE in March and May, respectively. According to the specific contributions of March and May SWE to spring snowmelt anomalies, spring snowmelt anomalies in EEWS are divided into two dominant categories. The main factors for different types of anomalous snowmelt are diverse, accompanied by different intraseasonal variations of SWE and atmospheric circulation changes. Category 1 is affected by both the initial snow mass and the late snowmelt process, showing a positive SWE anomaly in March and a negative SWE anomaly in May, corresponding to anomalously low 500-hPa geopotential heights over EEWS in JFM and anomalously high geopotential heights in April. Category 2 is not related to the initial snow accumulation, and is mainly dominated by the later snowmelt process, which is manifested by a persistent anomalously low SWE in spring. The corresponding circulation has stronger geopotential height anomalies than category 1 over EEWS in April. Furthermore, anomalous atmospheric circulation influences the snowmelt and SWE changes in spring by controlling temperature and snowfall.

The March SWE anomaly in category 1 is linked to the anomalous circulation of the EA pattern in winter. In the context of large-scale circulation anomalies, the EEWS region is controlled by a local anomalous trough. Water vapor from the Atlantic Ocean and lower latitudes is transported to this region by anomalous westerly and southerly winds, forming a water vapor convergence zone here. Besides, the anomalous convergence of low-level wind fields in EEWS is accompanied by anomalous ascending motion. All the above factors enhance the snowfall and therefore the accumulation of snow during the winter, thus increasing SWE in March. Both category 1 and category 2 exhibit anomalous snowmelt process in spring. This is mainly associated with the local dynamic and thermodynamic processes accompanying the atmospheric circulation anomaly in April, when the snowmelt is stronger for category 2 than category 1. Under the background of a SCAND-like anomalous circulation pattern in April, the EEWS region is dominated by anomalous high 500-hPa geopotential height and anticyclonic circulation, with strong warm advection in the lower troposphere, in which anomalous southerly winds behind the anticyclone play the main role in transporting warm air. Meanwhile, the anomalous downward motion in the lower and middle troposphere in EEWS is conducive to adiabatic warming, boosting the snowmelt and the reduction of SWE in May.

Moreover, from the perspective of local energy balance, surface fluxes also play an important role in the snow melting process, and there are evident differences in the surface energy balance processes corresponding to different types of anomalous snowmelt. For category 1, the spring snowmelt is mainly influenced by LW_{net} in its early stage, which initiates the anomalous snowmelt. The snowmelt anomalies are significant in April, with downward atmospheric SH transport as the major factor promoting snowmelt, and SW_{net} playing a secondary role. The SH transfer is weakened in May, and the magnitude of the negative LH anomaly is comparable to that of the positive SH anomaly, and the two offset to make the anomalous snowmelt weaker. The anomalous snowmelt in category 2 occurs mainly in the late spring. There exists an atmospheric SH transport to the surface in April with similar intensity as category 1, while a stronger SW_{net} appears, which facilitates a large amount of snow melting. In turn, more snowmelt also further reduces the surface albedo and induces a stronger SW_{net} at the surface, favoring the positive snow–albedo feedback effect. However, the other energy balance components except for SW_{net} have negative anomalies in May, probably because there is little snow left to melt and this must be balanced by an enhanced upward energy flux from the surface.

b. Discussion

It is worth noting that the combined use of datasets, including GlobSnow SWE and ERA5, may lead to some uncertainties. In view of this, we further validated the present results using the ERA5 SWE data, and found that the results are generally consistent with those using the GlobSnow SWE, except for some differences in surface flux magnitude, which
may be attributed to different years chosen for category-1 and
category-2 composites. Overall, this suggests that the variable
characteristics and the associated physical processes of the
two categories of anomalous snowmelt are reliable and not
significantly influenced by the data. Besides, it should also be
noted that due to limitations in the time period and quality of
observational data, the sample size of anomalous snowmelt
events is limited. Therefore, further verification using larger
sample model data is needed in the future.

This study explored the possible influencing factors of spring
snowmelt and its related physical processes from a local per-
spective. Actually, snow mass changes are regulated by the
atmospheric circulation, and previous studies found that external
forcings such as SST and sea ice anomalies can modulate the
atmospheric circulation (Liu et al. 2012; Tylisz et al. 2019; Xu et al.
2019b). Then, it is also worth exploring the possible linkages be-
tween the spring snowmelt and these external forcings. Due to
the complex factors influencing spring snowmelt, there is still a
lack of clear understanding of the antecedent signal of snow-
melt anomalies, which is a direction for future research. To find
out the forcing factors in the previous period that affect the
snow accumulation and snow melting process and understand
their associated physical mechanisms can help to improve the
understanding of the snowmelt, thus clarify its linkage role in
influencing the later climate. In addition, some studies men-
tioned that the snow–atmosphere coupling is stronger during
spring snowmelt and the period afterward (Xu and Dirmeyer
2011, 2013). Our study only focused on the influence of the
atmosphere on snowmelt. The atmosphere can promote snow-
melt, and in turn, snowmelt can also change the thermal
properties of the land surface and further produce feedback to
the atmosphere, whose more detailed physical processes are
also worthy of in-depth exploration in the future. Moreover,
the snow melting can produce infiltration and runoff, affecting
local and nonlocal water availability (Novotny and Stefan 2007;
Tan et al. 2011; R. Wang et al. 2015; Horner et al. 2020), and the
corresponding hydrological processes under different types of
anomalous snowmelt also deserve to be investigated.

Acknowledgments. This work was jointly supported by the
National Key Research and Development Program of China
(2022YFF0801603) and the Postgraduate Research and Practice
Innovation Program of Jiangsu Province (KYCX21_0936). We
thank three anonymous reviewers for their constructive com-
ments and also the editor for helping with the review process.

Data availability statement. The GlobSnow v3.0 SWE data
are available at http://www.globsnow.info/, the ERAS and ERAS-
land reanalysis dataset are from https://cds.climate.copernicus.eu/
#!/home, and the climate indices are provided by CPC at https://

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