Linkage between anomalies of pre-summer thawing of frozen soil over the Tibetan Plateau and summer precipitation in East Asia

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1. Introduction

As a core factor linking the land–atmosphere coupling system, the cryosphere is considered one of the most sensitive and significant indicators of global climate change. The cryosphere is highly vulnerable to climate warming; observations have shown considerable decline in snow cover (high confidence), glacial retreat (very high confidence) and permafrost degradation (high confidence) in recent decades. These cryospheric changes, in turn, threaten the climate system by regulating the surface energy budget, hydrological cycle, ocean current and atmosphere circulation (Portner et al. 2019).

Other than the high-latitude polar regions, the cryosphere also exists in some mid- and low-latitude
high mountain areas. The Tibetan Plateau (TP), also known as the ‘Third Pole’ due to its complicated topography and unique geological location, contains widespread and vital branches of the cryosphere, including glaciers, snow cover, permafrost and seasonally frozen ground (SFG) (Yang et al 2019). In response to climate change, the vanishing glaciers, snow cover and permafrost directly influence the water resources and hydrological cycle (Immerzeel et al 2010, Wang et al 2018). Many large rivers, including the Yangtze, Yellow, Lancang, Brahmaputra, Ganges and Indus rivers, originate on the TP and provide water resources to more than 20% of the world’s population (Tong et al 2016). Glacial melt has increased the runoff in western China in recent decades (Ren et al 2011), although this increase might not persist in the future (Huss and Hock 2018). Snowmelt can be a more important contributor than glacial melt for water supply in the TP region, and snow meltwater supply may decrease drastically in the future (Kraaijenbrink et al 2021).

In addition, cryospheric changes can also alter the water cycle through surface diabatic heating, which can affect the dominant atmospheric circulations in the TP, including the Indian monsoon, East Asian (EA) monsoon and westerlies (You et al 2020). Many studies have examined the impact of snow cover on the climate system. Henderson et al (2018) found that Eurasian autumn snow cover can change the atmospheric circulation in the Northern Hemisphere by generating planetary-scale atmospheric waves. As for the winter/spring snow cover in the TP, it is found to have a negative correlation with the summer precipitation in parts of northern and southern China, but a positive correlation with the summer precipitation in the middle and lower reaches of the Yangtze River (MLYR) (You et al 2020). Bo et al (2017) also found that heavier snow cover in the southern (northern) TP led to enhanced (weakened) precipitation in the Yangtze River basin. The profound influence of cryospheric changes in the TP on regional and global climate systems has attracted increasing attention and focus (Wang et al 2020a). Nevertheless, the impact of another vital component of the cryosphere, permafrost, on the climate system has received less attention and remains highly elusive.

Permafrost, defined as subsurface material below 0 °C for at least two consecutive years, occupies around 40% of the total area of the TP (Zou et al 2017, Wang et al 2020b). The TP has experienced great degradation of permafrost in past decades (Ran et al 2020), and this degradation is projected to further exacerbate in the future (Wang et al 2019). Persistent and pervasive degradation of permafrost has led to changes in hydrological processes, including enhanced surface water–groundwater interaction, more baseflow and changing streamflow seasonality (Walvoord and Kurylyk 2016, Wang et al 2018, Ma et al 2019). Although observations and simulations show that extensive degradation of near-surface permafrost would lead to abrupt increases in summer near-surface temperature and convective precipitation and decreases in relative humidity and surface runoff (Teufel and Sushama 2019), the mechanism of how soil freeze–thaw on the TP affects the land–atmosphere coupling process still needs to be further studied.

In EA, there is a northeastward-slanted rain belt called Meiyu in China and Baiu in Japan, extending from the valley of the Yangtze River in China to the east of Japan in June and July. Based on the all-season EA Atmospheric River (AR) catalog, the main route of the EA AR (southeast China–south Japan) in the summer monsoon period is related to the northward displacement of the Western North Pacific Subtropical High (WNPSH), coinciding with the stepwise propagation of the rain belt (Pan and Lu 2020). Rain belt anomalies are affecting the lives of around a billion people and are bringing colossal property losses. With the warming climate, many statistics and modeling analyses find that the TP surface diabatic heating is changing the EA summer monsoon (EASM) climate system, leading to movement of the EA precipitation belt. Land surface temperature (LST) cold (warm) anomalies in the TP are linked to drought (flood) events to the south of the Yangtze River and wet (dry) conditions to the north of the Yangtze River in late spring/summer (Xue et al 2018). Furthermore, TP heating anomalies were found to be related to the eastward extension of the South Asian High (SAH) and westward stretching of the Western Pacific Subtropical High, resulting in southwestern moisture transport and convergence and increasing summer precipitation over the MLYR (Ge et al 2019). Besides, soil moisture affects precipitation during the following summer in EA (Yang and Wang 2019a) as well as in South Asia (Ullah et al 2021) by influencing the sensible heat (SH) flux and latent heat and later regulating the diabatic heating profile over the TP. Hence, the elements regulating TP surface diabatic heating will significantly affect the near-surface energy exchange, atmospheric circulation, such as the SAH, WNPSH and EASM, and consequently water vapor transport as well as rain belts. Therefore, understanding the possible land–atmosphere mechanism between the TP and EASM climate system can promote the scientific formulation of countermeasures against future climate change, preventing extreme events and benefiting the regional economy.

In summary, cryospheric change on the TP is a critical indicator affecting the EASM climate system. However, there are still few studies using the direct indicator of soil freeze–thaw anomalies to assess the
impact of degradation of frozen ground on climate systems, mainly restricted by limited observed or simulated datasets (Ma et al 2017), leading to a knowledge gap between changes in frozen soil on the TP and the implications for downstream precipitation. A geomorphology-based ecohydrological model is capable of simulating the high-resolution TP frozen soil dataset (Zheng et al 2020). In this study, the above frozen soil dataset over the TP and precipitation reanalysis datasets were adopted to explore the relationship between the degradation of frozen soil on the TP and EA summer precipitation, especially in the Meiyu–Baiu rainy season. The findings are expected to provide insights into land–atmosphere teleconnections in the EASM climate system due to thermal processes induced by the thawing of frozen soil on the TP, which has not been included in current studies.

2. Data and methods

2.1. Data and conditions

The TP is a vast elevated plateau with an average altitude exceeding 4500 m, located across South, Central and East Asia. It contains the largest volumes of freshwater reserves in the form of ice, snow and frozen ground outside the polar regions. Spanning the middle and low latitudes of the Northern Hemisphere, the TP blocks both meridional airflow and the northward transport of water vapor. The dominant atmospheric circulations over the TP are controlled by the Asian monsoon and mid-latitude westerlies. Also influential are the Polar Vortex, North Atlantic Oscillation and Pacific Decadal Oscillation under the complex teleconnection interaction between atmospheric and oceanic components (Yao et al 2012, You et al 2020). Regarding permafrost distribution over the TP, the permafrost region and SFG region both occupy around half of the entire TP (figure 1(a)). Permafrost degradation starts with deepening of the active layer, followed by the formation of talik (a perennially unfrozen layer above the permafrost table that cannot be frozen again in winter) and finally the disappearance of permafrost (Connon et al 2018). The permafrost is mainly distributed in the northwestern TP, and SFG is primarily located in the southern and southwestern TP.

The distribution of the active layer thickness (ALT) of permafrost and the maximum thickness of SFG (MTSFG) over the TP from 1981 to 2019 with 1 km resolution on a monthly scale has been simulated by the Geomorphology-Based Ecohydrological Model (Zheng et al 2020). With the high-resolution ALT and MTSFG datasets, this study uses the annual thawing thickness of frozen soil in pre-summer (before June) as an indicator of TP freeze—thaw for analysis. In this study, we define the pre-summer thawing thickness as representing the melting of frozen soil before the monsoon season. For SFG, the thawing thickness before summer is calculated as the difference between the annual maximum frozen thickness in January/February and the maximum frozen thickness in May; for permafrost, the thawing thickness before summer is defined as the maximum thaw depth in May. Over the past 40 years or so, it can be seen that the annual mean of the pre-summer frozen soil thawing thickness is generally larger in the northern and western SFG regions, while there are apparent increasing trends in the regions where talik forms or permafrost has totally disappeared, lying in the transitional area between permafrost and SFG regions, indicating rapid thawing of permafrost (figures 1(b) and (c)). The other simulated variables, such as soil temperature and SH, are also derived in this modeling. The simulated 5 cm soil temperature is used to represent LST in the following analysis.

Subtropical EA is one of the most active monsoon regions in the world. The rain belt of most concern is the Meiyu–Baiu rain belt. It is accompanied by oceanic factors (Zhou et al 2021) such as warmer sea surface temperatures (El Niño phase) or colder sea surface temperatures (La Niña phase) (Ronghui et al 2004), large northeastward transport of moisture, westerlies from the eastern flank of the TP (Xie and Sampe 2010) and land–sea contrast that is in tandem with the seasonal progression of the EASM. The monthly observational 0.25° precipitation datasets are obtained from the Climate Prediction Center Merged Analysis of Precipitation (CMAP) and have been used in many previous studies (Zhou et al 2021; downloaded via www.esrl.noaa.gov/psd/data/gridded). Focusing on satellite rainfall over EA for a 39-year period from 1981 to 2019, this study selected the average monthly rainfall in the Meiyu–Baiu rainy season (June, July; JJ) as the EA indicator. It can be seen that the mean JJ precipitation strengthened the most in southern China and Kyushu, Japan, during 1981–2019 (figures 1(d) and (e)).

In order to further explore the physical driving mechanism between freeze–thaw of TP frozen soil and the EASM water cycle, hourly and monthly atmospheric data in four recent decades (1981–2019) were retrieved from the ERA5 reanalysis dataset with 0.25° × 0.25° spatial resolution (Hersbach et al 2020; downloaded via https://doi.org/10.24381/cds.6860a573; https://doi.org/10.24381/cds.bd0915c6). For water vapor transport, the daily vertical integral of northward water vapor flux (\(F_u\); kg (ms\(^{-1}\)) and the vertical integral of eastward water vapor flux (\(F_v\); kg (ms\(^{-1}\)) are used to compute integrated vapor transport (IVT; kg (ms\(^{-1}\))): IVT = \(\sqrt{F_u^2 + F_v^2}\). The geopotential height, the \(U/V\) components of wind, temperature and vorticity for 850 hPa, 500 hPa and 200 hPa were also downloaded for analyzing general circulation in the middle and upper troposphere.
2.2. Methods

2.2.1. Coupled manifold technique (CMT)

The CMT (Navarra and Tribbia 2005) can decompose a climate field into two portions of variability, where one portion denotes the variability forced by the other field and a second portion, free from the other field in a statistical sense. This technique has been used to study the interactions among land, ocean and atmosphere (Cherchi et al 2007, Alessandri and Navarra 2008). Given that many studies have confirmed that oceanic factors, such as the sea surface temperature anomaly, were identified as the dominant forcing factors of Meiyu–Baiu rainfall variability (Tanaka 1997), the effect of oceanic factors should be minimized as much as possible in this study. We used this method to decompose the EA JJ precipitation and calculate the variability forced by the pre-summer thawing of TP frozen soil, namely the EA-forced JJ precipitation manifold ($EA_{\text{Precip-for}}$), for further analysis.

2.2.2. Maximum covariance analysis (MCA)

As a robust statistical method, MCA can be used to investigate the relationship between groups of variables based on the patterns of maximum covariance between two datasets by performing singular value decomposition. The MCA results are assessed through the square fraction of covariance, meaning...
the fraction of the covariance explained by the first few modes. So far, the MCA method has been used in many research works to analyze the possible remote relationship between different variables (Barreto et al 2016, Xue et al 2018, Lian et al 2020). It can also be coupled with the CMT method (Wang et al 2011), which also gives insights into the use of observational evidence to explain the potential teleconnection of the relevant variables. The MCA was used to analyze the relationship between EA_{Precip-for} and pre-summer thawing thickness of frozen soil on the TP (TP_{Thawing}).

2.2.3. Composite analysis
Composite analysis with both surface and atmospheric fields can further explore the potential mechanism of land–atmosphere interaction. In many studies, this approach has been used to find out the large-scale mechanism and driving forces robustly but not the small-scale differences (Ge et al 2017, 2019, Ullah et al 2021). In this work we analyzed the land surface and atmospheric fields in JJ during the positive and negative anomaly years, as described in the next section. The surface and atmospheric fields of the positive anomaly years were selected as the positive composites, and those of the negative anomaly years were selected as the negative composites. It should be noted that the current study only explored the possible mechanism and dynamics of the thermal forcing induced by thawing of TP frozen soil in altering the EA precipitation, even though freeze–thaw of TP frozen soil may have an impact on global-scale climate activity.

2.2.4. Research route
In short, this study explored the possible linkage between pre-summer thawing of frozen soil on the TP and JJ precipitation in EA over the past 40 years (1981–2019). The research route is summarized as follows (figure 2). After removing the warming trend in the time series of TP_{Thawing} and EA JJ precipitation, EA_{Precip-for} was first calculated with the CMT method. Then, MCA analysis was performed on TP_{Thawing} and EA_{Precip-for} to derive the robust remote relationship. Finally, by performing empirical orthogonal function (EOF) analysis on EA_{Precip-for}, the possible physical mechanisms were analyzed with the land surface and atmospheric variables in positive and negative anomaly years.

3. Results
3.1. Linkage between TP frozen soil and EA precipitation
Based on the CMT method, the forced precipitation manifold is decomposed from the detrended EA mean JJ rainfall. Figure 3(a) shows the percentage of the precipitation variance forced by the variability in TP_{Thawing}. These values are derived from the ratio of the forced precipitation manifold to the original precipitation fields. It can be found that high values (10%–20%) are found over the high-frequency sites of JJ rain events, and the values gradually decrease southward and northward in the surrounding MLYR regions and southern Japan, suggesting considerable sensitivity of EA JJ precipitation to TP_{Thawing}. Little forcing (low values) is found over most parts of the north and south of the MLYR, such as the North China Plain and South China, indicating little influence from TP_{Thawing}.

The TP_{Thawing} and EA_{Precip-for} fields are further decomposed into the associated spatial patterns and time coefficient series with the MCA. Figures 3(b) and (d) show the patterns of TP_{Thawing} and its precipitation response from the first MCA mode, which contains 59% of the total squared covariance. It can be found that positive anomalies of pre-summer thawing of frozen soil in most southern TP regions correspond to wet conditions over the MLYR and southern Japan and dry conditions in the south and north of the MLYR in JJ. The time expansion coefficients of TP_{Thawing} and EA_{Precip-for} fields of this first MCA mode, shown in figure 3(c), have a correlation coefficient of 0.99, indicating that the derived TP_{Thawing}–EA_{Precip-for} mode is robust. It should be noted that the high time coefficient correlation derived here is due to data pre-processing with CMT, which only constructs the precipitation variability forced by TP_{Thawing} at a significance level of 1%. These statistical properties suggest that the responses of TP_{Thawing}–EA_{Precip-for} are very likely to be driven by the land–atmosphere interaction.
coupling mechanism, as suggested by earlier numerical studies, which is related to regulation of surface diabatic heating on the TP.

To further explain the above robust relationship driven by the MCA analysis, a composite analysis is performed on the relevant data. Based on the precipitation manifold forced by the thawing of frozen soil, EOF analysis was applied to find out the spatial and temporal patterns of change in \( \text{EA Precip-for} \) in the first three modes. Figures 4(a) and (b) show the spatial pattern of the first EOF mode along with the associated standardized principal component (PC) of the first leading mode, which explains 69.23% of the total variance of the \( \text{EA Precip-for} \) manifold. The first EOF mode shows a robust reversed triple pattern, meaning that changes in precipitation show less–more–less patterns in the south–north direction, while the second (11%) and third (6.4%) EOF modes did not show the obvious characteristics. In mode 1, the significant anomalies in the MLYR regions and southern Japan further indicate a linkage between EA JJ precipitation and pre-summer thawing of frozen soil in the TP.

Meanwhile, based on the \( \text{EA Precip-for} \) manifold the PC1 time series, the positive and negative anomaly years are further selected with a standardized PC1 (\( |y_{pc1}| > 1 \)) shown in figure 4(b) for the positive and negative composite analysis, respectively. In this study, the positive anomaly years are 1986, 1997, 1998, 2010, 2014, 2015 and 2016, while the negative anomaly years are 1981, 2001, 2008, 2013 and 2018. The results of the composite analysis are shown in figure 5. Based on the difference values between the positive and negative anomaly years, positive anomalous pre-summer thawing of frozen soil (0–0.3 m yr\(^{-1}\)) over the TP relates to more precipitation (1.5–6 mm d\(^{-1}\)) over the MLYR and southern Japan and less precipitation (−1.5 to −4.5 mm d\(^{-1}\)) in the north and south of above areas in JJ (figures 5(a) and (b)). Hence, from the above analysis, it can be concluded that different EA JJ precipitation patterns respond to pre-summer thawing of frozen soil on the TP, possibly forced by land–atmosphere mechanisms that will be discussed later.

3.2. Possible mechanisms

In this section, we further combine the land surface and atmospheric components to analyze the underlying mechanisms between thawing of frozen soil on the TP and the EA precipitation patterns, especially the patterns of the Meiyu–Baiu rainfall belt.

The necessary direct prerequisite for Meiyu–Baiu precipitation is a plentiful supply of water vapor. For EASM regions, it usually shows a southwest–southeast (SW–SE) source swing pattern (Cheng and Lu 2020). For the Meiyu–Baiu rainfall belt, which mainly moves from the MLYR to the southern part of Japan, moisture mainly comes from the western Indian Ocean, the Arabian Sea, the Bay of Bengal,
EOF analysis for the EA manifold. The first three modes are (a) mode 1 with an explained variance of 69% and its time series of PC1 (b), (c) mode 2 with an explained variance of 11%, and its time series of PC2 (d), and (e) mode 3 with an explained variance of 6% and its time series of PC3 (f). The principal components (PCs) are scaled to unit variance by dividing the standard deviation of corresponding coefficients.

Indochina and Southwest China. Figure 5(c) shows the difference in the IVT vector in JJ between the positive and negative anomaly years. It can be seen that along with the northeastward moisture transport traits, the positive difference of IVT spreads from southern China, the Yangtze River regions to the southern part of Japan with gradually higher values, indicating that more water vapor is transported northeastward under the large-scale circulation in the positive anomaly years. The negative difference with westward (southwestward) arrows in the moisture source regions, such as the Bay of Bengal and the Arabian Sea, reflects that water vapor was stuck under small-scale circulation in the negative anomaly years. This study further detected the EA AR traits with an image-processing based AR tracking (IPART) (Xu et al 2020) Python package. From the difference in AR occurrence frequency (figure 5(d)), it can be seen that the AR center with a high occurrence frequency shifts from the exit of the Yangtze River
Figure 5. Composite analysis. The difference between positive anomaly years and negative anomaly years on (a) annual pre-summer frozen soil thawing thickness (m), mean JJ (b) EA precipitation (mm d$^{-1}$), (c) integrated vapor transport (IVT) vector (kg (ms)$^{-1}$) and (d) atmosphere river (AR) occurrence frequency (coefficient). (e) Mean JJ geopotential height at 500 hPa (contours, m) with wind vectors (wind bar, knots) and temperature anomaly (color bar, K), as well as (g) mean JJ geopotential height at 200 hPa (contours, m) with wind vectors (wind bar, knots) and temperature anomaly (color bar, K) for the positive composite. Parts (f) and (h) as in (e) and (g), but for the negative composite. The baseline temperature is typically computed by averaging around 40 years of temperature data.

and the south of Japan to the southeast of Japan, suggesting a west–east extension change of the Meiyu–Baiu rain belt between positive and negative anomaly years.

Given that the SAH and the WNPSH are two crucial systems affecting EA summer rainfall, the underlying general circulation in the middle and upper troposphere in JJ was further investigated. In this study, by depicting the geopotential height at 500 hPa (figures 5(e) and (f)), there is an apparent difference in the location of the WNPSH with respect to the EA continental boundary. The center of the WNPSH is located in the east of EA. It can be seen that the WNPSH corresponding to the positive anomaly years extends further westward based on the location of the 5880 m contour (red curve) penetrating EA, along with the warm low-latitude areas (10° N–30° N). On the contrary, in the negative anomaly years, the WNPSH is more restrained both in strength and range. The center of the SAH is located southwest of the TP. At 200 hPa, a massive SAH and strong westerly winds are dominant over the Eurasian continent (figures 5(g) and (h)). In the positive anomaly years, the SAH extends eastward to the East China Sea based on the location of the 12 500 m contour (brown curve) with accelerating
mid-latitude westerly winds (30° N–40° N). Relatively speaking, there is a decelerating westerly jet and less active evolution of the SAH in the negative anomaly years. Based on the above analysis, the SAH and WNPSH are closely related in opposite zonal directions and affect the EASM climate system, consistent with previous studies (Wei et al. 2019). Besides, there is an underlying correlation between 500 hPa horizontal temperature advection and the Meiyu–Baiu rain belt (Okada and Yamazaki 2012). By further depicting the difference in the horizontal temperature advection and vorticity advection at 500 hPa (figures S1 available online at stacks.iop.org/ERL/16/114030/mmedia), the stronger warm 500 hPa horizontal temperature advection accompanied by positive vorticity advection shows the convergence of rising flows at lower levels and upper-level divergence in the MLYR extending to southern Japan. As a result, the above conditions promise a strong Meiyu–Baiu rain belt with westward extension, and vice versa.

The above analysis has proved a significant relationship between the pre-summer thawing of frozen soil in the TP and summer precipitation in EA based on statistical analysis and the atmospheric physical mechanism. In order to determine how the frozen soil freeze–thaw influences surface diabatic heating in the land–atmosphere interaction, we further calculate the difference between LST and SH for the positive and negative composites. Figure 6(a) shows the differences in the LST and SH annual cycle between positive and negative anomaly years. It can be seen that, from January to July, differences in LST are always positive (>0), indicating that a deeper pre-summer frozen soil thawing thickness is accompanied by a higher LST. As for the differences in SH, almost all of them are positive in the spring (March to May) and JJ, except for the differences in April related to the soil water phase change. During the freezing–thawing period, less water storage in the form of ice means more evaporation and less SH flux, but after the thawing period, SH increases (Yang and Wang 2019b). In short, the stronger SH over the TP is also accompanied by more thawing of frozen soil in spring and JJ, which is consistent with the former study on the signal of spring SH over the TP, that is a stronger (weaker) spring SH is usually followed by a stronger (weaker) atmospheric heat source in summer (Duan et al. 2013). By further depicting the spatial difference between LST and SH in JJ (figures 6(b) and (c)), it can be found that there are positive values of LST and SH in the southern TP, corresponding to the heterogeneous map of TPthawing with a significance level of 90% (figure 3(d)). This further suggests a linkage between thawing anomalies of frozen soil and LST and SH. Based on the analysis of the thermal and moisture dynamics of frozen soil, the deepening of the active layer and thawing of frozen soil allow water to drain more deeply with the increase in soil water infiltration capacity, which may reduce the surface soil moisture if it is not counteracted by increasing precipitation (Parazoo et al. 2018, Teufel and Sushama 2019, Wei et al. 2021). With the loss of soil water, drier soils with lower thermal conductivity can reduce the ground heat flux and increase SH and LST (Teufel and Sushama 2019). In the summer season, thawing is almost finished, the increase (decrease) of SH accounting for the warmer (colder) LST and drier (wetter) surface soil layer (Yang and Wang 2019b) leads to positive (negative) anomalies of surface diabatic heating over the TP (Wang et al. 2020a).

Hence, the possible mechanism can be summarized as follows. The positive anomaly of thawing of frozen soil in pre-summer, indicating enhanced surface diabatic heating from spring to JJ, will enhance surface diabatic heating over the TP. As summer heating progresses in JJ, the enhanced SAH and WNPSH extend eastward and westward, respectively. Then, under the control of the general circulation, northeastward transport of water vapor strengthens, corresponding to the westward extension of the Meiyu–Baiu rain belt, and more precipitation occurs in the MLYR and southern Japan.

4. Discussion

In previous studies on the teleconnection between TP surface elements and EA climate systems, the mechanisms of extreme climate events were summarized as follows. A warmer TP induced by the increased SH was related to the enhanced the EASM circulation system, including the low-level southwesterly jet, the WNPSH with westward extension, the mid-tropospheric subtropical westerly jet and the eastward stretching of the SAH (Duan et al. 2013, Ge et al. 2019). They also explained that the enhanced EASM circulation system intensified moisture convergence over the MLYR valley, which is consistent with our findings. However, some studies thought that the possible mechanisms were related to the anomalous subtropical westerlies and propagation of a stationary Rossby wave train in middle latitudes (Bo et al. 2017, Yang and Wang 2019a). Xue et al. (2018) suggested that the above two mechanisms may work together to give rise to extreme EA climate events. Since the EA precipitation patterns and movement of the Meiyu–Baiu rain belt are forced by intertwined land–atmosphere–ocean system factors, it is still challenging to unveil the possible mechanisms the remote relationship between TP surface elements, especially the cryospheric elements, and the surrounding climate systems.

Under climate warming in recent decades, degradation of frozen ground has changed the water cycle in the TP and downstream regions in multiple
ways. This influence will strengthen based on the projection that 25.9% and 43.9% of the current TP permafrost will disappear by the 2040s and 2090s, respectively, under the RCP 4.5 scenario (Wang et al. 2019). With a shortened freezing season and persistent degradation of permafrost in the Northern Hemisphere (Li et al. 2021), it is increasingly important to assess the impact of degradation of frozen ground on climate systems and downstream precipitation. The findings of this study prove a strong link between pre-summer thawing of frozen soil and EA JJ precipitation patterns. It can be conjectured that via the TP\textsubscript{Thawing}–EA\textsubscript{Precip-for} coupling mechanism, stronger MLYR JJ precipitation and increasing extreme events such as floods (droughts) in the south (north) of the MLYR may be expected in the future as permafrost continues to degrade in a warming climate.

There are also some limitations to this study. The soil freeze–thaw dataset adopted in this study was derived from a physically based soil freeze–thaw model. Although this model has proved to be effective in simulating the soil freeze–thaw dynamics in the TP (Zheng et al. 2020), uncertainties still exist due to deficiencies of the forcing data, especially

\begin{figure}[h]
\centering
\includegraphics[width=\textwidth]{figure6.png}
\caption{The difference between positive and negative anomaly years for (a) LST (°C) and SH (W m\textsuperscript{-2}) annual cycle and spatial distribution of (b) mean JJ LST and (c) mean JJ SH over the TP.}
\end{figure}
in the uninhabited western region of the TP. Despite the uncertainties, the simulated high-resolution (1 km) dataset is the best available dataset so far. It provides an opportunity to fill the gap on TP frozen soil-related teleconnection research considering the poor performance and coarse resolution of existing earth system models in estimating permafrost processes in the TP region (Guo and Wang 2016, Burke et al 2020). Besides, this study used an offline model and reanalysis datasets to analyze the linkage between TP pre-summer thawing of frozen soil and EA summer precipitation; in the future, a large-scale coupled land–atmosphere model with better ability to reproduce the highly heterogeneous soil freeze–thaw dynamics in the TP can be developed to further examine the impact of permafrost degradation in the TP on the precipitation patterns in the downstream regions, as well as the hydrological extremes such as floods and droughts. It should be noted that the relationship between frozen soil and atmospheric variables may be nonlinear, which is another limitation of this paper and should be considered in future research with the coupling mechanism model. Despite the above limitations, this study helps improve the understanding of land–atmosphere coupling in the context of degradation of frozen soil over the TP in a warming climate.

5. Conclusion

In conclusion, this study has explored the relationship between pre-summer thawing of frozen soil over the TP and EA precipitation patterns, especially the Meiyu–Baiu rain belt, using statistical analysis. We found that the EA JJ precipitation manifold ratio forced by thawing of frozen soil is 10%–20% in the MLYR and southern Japan, suggesting the critical role of frozen soil in influencing the EA climate system. Based on the MCA analysis, the increases (decreases) in pre-summer frozen soil thawing thickness in the southern TP relate to more (less) precipitation over the MLYR and southern Japan and less (more) precipitation in the south and north of the MLYR in the Meiyu–Baiu rainy season. The underlying mechanism is the regulation of freeze–thaw of frozen soil on surface diabatic heating. The positive (negative) thawing and accompanying higher (lower) LST and more (less) SH in the spring and JJ leads to increased (decreased) surface diabatic heating. As a result of positive surface diabatic heating, northeastward movement of water vapor is intensified with an enhanced SAH and WNPSH extending eastward and westward, respectively, leading to the westward extension of the Meiyu–Baiu rain belt with more precipitation.

This study provides a possible explanation for the non-local effects of TP freeze–thaw on EA precipitation. Given that the present study is still a statistical diagnosis drawing qualitative conclusions about one cryospheric element (frozen soil), the robust relations and physical mechanisms still need to be further confirmed with development of the land–atmosphere–ocean coupling model and studies of the cryosphere mechanism in the near future.

Data availability statement

The data that support the findings of this study are available upon reasonable request from the authors.

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