Impact of Soil Freezing-Thawing Processes on August Rainfall Over Southern China

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Abstract The impact of soil freezing-thawing processes on August rainfall in Southern China (SC) during 1979–2008 and related physical mechanisms are investigated using the coupled land-atmosphere model Grid-Point Atmospheric Model version 2.0—Community Land Model version 2.0 (GAMIL2.0-CLM2.0). The moisture-budget analysis is employed to quantify the contributions of different factors to the change of precipitation in SC. The results indicate that the vertical moisture advection and its dynamic component play an essential role in the long-term trends of August rainfall. The possible physical mechanism is that the model which includes the supercooled water simulates much higher soil/air temperature in August, especially in the north of 40°N, which is not only weakened the meridional thermal contrast but also induced an anomaly “north high–south low” circulation pattern in the middle and high latitude of Eurasia. Both resulted in the 200 hPa zonal winds weakening, and 850 hpa northerly wind southward moved. Meanwhile, a convergence anomaly cyclone is initiated in the lower troposphere over SC regions, accompanied by enhanced ascending and strong moisture flux convergence, which is beneficial to the increased precipitation. This study provides a new interpretation of the “southern flooding” during 1979–2008 from the point of frozen soil changing.

1. Introduction

Southern China (SC), including regions south of the Yangtze River and east of the Tibetan Plateau (TP), is greatly affected by droughts and floods due to the long flood season and significant interannual and interdecadal variations in rainfall. During 1951–2010, most extreme precipitation events occurred south of 34°N and resulted in substantial economic losses (Chen & Zhai, 2013). In addition, the total number of records of flood disasters increased significantly at a rate of 77.4 times per decade from 1984 to 2010 (Shi et al., 2020). Improved understanding of summer rainfall in SC could enhance ability to predict rainfall in flood seasons, and increase capacity for preventing floods, controlling drought, and mitigating natural disasters.

More than half of the annual precipitation in SC is focused on April to September or early October. The April–June period is the “pre-rainy” season, and July–September is the “second flood” season. Monthly rainfall is variable during May–August (Su et al., 2014). Previous studies have suggested dividing the summer rainy season into early and late periods or treating each month separately when investigating variations in precipitation in SC (Wang et al., 2009; Yuan et al., 2019). August has the highest (and most variable) precipitation during the second flood season, but most previous studies have focused on early summer rainfall (Chen & Zhai, 2005; Li, Ren, et al., 2018). This study therefore focuses mainly on August rainfall in SC.

Numerous researches have shown that the summer rainfall over SC experienced a significant interdecadal change since the mid-1990s, which was very below-normal from the late 1970s to the mid-1990s and mutated to much higher than usual since then, according to the reanalysis and observational precipitation data of the China Meteorological Administration. This change contributed to the “southern flood and northern drought” rainfall pattern in Eastern China since the late 1970s (Ding et al., 2008; Kwon et al., 2007; Wu et al., 2010; Yao et al., 2008; Zhang et al., 2008; Zhu et al., 2011).

External forcing factors have contributed to the interdecadal variation and long-term trend. Previous studies have highlighted the role of sea surface temperature (SST) anomalies in different key regions (Liu et al., 2019; Wu et al., 2010), including the Atlantic Multidecadal Oscillation and the Pacific Decadal Oscillation (Si & Ding, 2016; Wu & Mao, 2017; Yuan et al., 2019; Zhu et al., 2015). In addition, tropical cyclones have contributed appreciably to precipitation in SC (Ren et al., 2006; Su et al., 2014). The increasing incidence of typhoons,
specially formed in the South China Sea, has been primarily responsible for the increased precipitation in southeastern China since the mid-1990s (Cao et al., 2019; Chen et al., 2012; Kim & Lee, 2012; Kwon et al., 2007; Ren et al., 2002). Recently, the impact on rainfall of external forcing by human activity is attracting attention, including land-use changes and anthropogenic aerosol emissions (Li et al., 2019). Modeling studies have shown that anthropogenic aerosols such as black carbon may contribute to the increased incidence of summer floods in SC (Jiang et al., 2017; Menon et al., 2002), while large-scale urbanization over eastern China may lead to weakened precipitation over southern East Asia in the late summer (Chen et al., 2016; Ma et al., 2015).

Other studies have shown that snow cover variations in spring and winter over the TP and Eurasia play an essential role in the East Asia precipitation following summer (Si & Ding, 2013; Wu and Qian, 2003; Wu et al., 2009; You et al., 2020; Zhang et al., 2013). Preceding winter and spring snow anomalies over the TP affect summer precipitation in eastern China. The correlation between winter snow cover and subsequent summer rainfall is negative in southern and far-Northern China areas and positive in the middle–lower reaches of the Yangtze River. Spring snow cover anomalies in the TP are considered important predictors of summer rainfall in China (Chen & Wu, 2000; Wu et al., 2010; Zhang et al., 2008, 2017).

Apart from snow cover, frozen soil is distributed widely over mid–high latitudes in the Northern Hemisphere. Transitions between freezing and thawing processes may cause changes in diabatic surface heating. Some studies have reported that the inclusion of soil freeze–thaw processes in land-surface and climate models simulate realistic soil temperatures, soil water contents, surface energy/moisture fluxes, and runoff during the spring melt season (Cox et al., 1999; Koren et al., 1999; Li et al., 2021; Niu & Yang, 2006; Schlosser et al., 2000; Viterbo et al., 1999). Meanwhile, many sensitivity tests show that the change of soil freezing-melting process significantly affects the atmospheric boundary processes, regional circulation, and regional climate (Poutou et al., 2004; Takata & Kimoto, 2000; Xin et al., 2012). However, there has been little study of the specific effects of frozen-soil changes on August rainfall in SC. Seasonally frozen soil at mid–high latitudes in the Northern Hemisphere has shown a clearly decreasing trend over recent decades, as indicated by using different data, such as observational soil freeze depth, freezing/thawing index, and so on (Frauenfeld & Zhang, 2011; Frauenfeld et al., 2007; Peng et al., 2017; Xia & Wang, 2020). This study aimed to investigate whether changes in frozen soil affect August rainfall in SC, investigate how the frozen soil variability affects the August rainfall over SC, and identify its mechanism. The remainder of this paper is organized as follows. Section 2 describes the model and related experiments, validated data, and methods used in this study. Section 3 presents the main results. Conclusions are given in Section 4.

2. Data and Methods

2.1. Model and Experimental Design

The Grid-Point Atmospheric Model version 2.0 (GAMIL2.0), developed by the State Key Laboratory for Numerical Modeling of Atmospheric Sciences and Geophysical Fluid Dynamics (LASG), Institute of Atmospheric Physics (IAP), Chinese Academy of Sciences (CAS), was applied with a horizontal resolution of about 2.8° × 2.8°, with high-latitude and polar regions having a weighted equal-area grid and other regions a Gaussian grid, and with 26-σ levels of vertical resolution with the topmost level at 2.194 hPa. The dynamic core of the model is based on a finite-difference scheme developed by Wang et al. (2004) and Wang and Ji (2006). Physical processes included in the older version (GAMIL1.0) were mainly from the CAM2.0 model (Collins et al., 2003), but GAMIL2.0 includes improvements related to deep convection, cumulus clouds, and cloud microphysical schemes, among others (Li et al., 2013). GAMIL2.0 takes part in the Atmospheric Model Intercomparison Project (AMIP II) and CMIP5 as the Flexible Global Ocean–Atmosphere–Land System Model's atmospheric component Grid-Point version 2 (FGOALS-g2).

In this work, the GAMIL2.0 is coupled with the land surface model of Community Land Model version 2.0 (CLM2.0) (Bonan et al., 2002) developed by NCAR, and it is referred to as "GAMIL2.0-CLM2.0". In CLM2.0, the critical temperature of soil freezing/thawing is 0°C; soil ice begins to melt at soil temperatures above 0°C, and liquid water freezes at soil temperatures below 0°C in each soil layer (Table 1, scheme A). According to the energy excess (or deficit) needed to change the soil layer temperature to 0°C, the rate of phase change is calculated and the soil ice/liquid content in each soil layer is readjusted. However, in fact, liquid water could coexist with ice
in the soil layer over a wide range of temperatures below 0°C. When the soil temperature is below 0°C and soil liquid water exists in the soil layers, the maximum liquid water content coexisting with ice in soil layer is calculated by using the freezing-point depression (Niu & Yang, 2006) as follow:

\[ \theta_{\text{liq,max}} = \theta_{\text{sat}} \left\{ \frac{10^4 L_f (T - T_{frz})}{g T_{frz} \rho_{\text{sat}}} \right\}^{-1/b} \]  

(1)

where \( L_f \) is the latent heat of fusion (J kg\(^{-1}\)); \( g \) is the gravitational acceleration (m s\(^{-2}\)); \( \rho_{\text{sat}} \) and \( \theta_{\text{sat}} \) are the soil-texture-dependent saturated matric potential (mm) and saturated volumetric water content, respectively; \( b \) is Clapp and Hornberger (1978) exponent; \( T \) is the soil temperature (K), and \( T_{frz} \) is the freezing temperature of water (273.16 K). Only when the soil liquid water content is above the maximum liquid water content of the soil layer at soil temperature below 0°C, excess liquid water freeze (Table 1, scheme B). The concept of supercooled soil water (i.e., liquid water coexisting with ice over a wide range of temperatures below the freezing point) is introduced in phase-change processes in subsequent versions of CLM3.5 (e.g., CLM 4.0).

The coupled land-atmosphere model GAMIL2.0-CLM2.0 was used to build two AMIP-type experiments to investigate the effect of freezing/melting processes on the precipitation, with prescribed the monthly Hadley Center Global Sea Ice and Sea Surface Temperature (HadISST) data (Rayner et al., 2003) for 1974–2008. The forcing data (including the solar constant, the concentration of greenhouse gases, aerosols, and ozone) recommended by CMIP5 are used in the two simulations (https://pcmdi.llnl.gov/mips/cmip5/forcing.html). The only difference between the two simulations is that they use the different freezing/melting schemes in the land surface model of CLM2.0. The control experiment (CTRL) uses scheme A and the NEW test uses scheme B, and the comparison between A and B is shown in Table 1.

The first 5 years were a spin-up, with the 30 years (1979–2008) being analyzed. The CPC Merged Analysis of Precipitation (CMAP, version 1811) (Xie & Arkin, 1997) and Global Precipitation Climatology Project (GPCP, version 2.3) data sets (Adler et al., 2003, 2018), which were provided by the NOAA/OAR/ESRL Physical Sciences Laboratory (PSL), were used to validate the simulations. Trends were calculated using the linear regression analysis, and the Student’s t-test was performed for the statistical significance of trends.

### 2.2. Methods

The moisture-budget analysis is widely used to separate dynamic and thermal contributions to precipitation variability (Akinsanola & Zhou, 2019; Chou & Lan, 2012; Li, Zhou, et al., 2018; Ma & Zhou, 2015; Peng & Zhou, 2017; Seager et al., 2010).

The vertically integrated moisture-budget equation is used as follows:

\[ P = -\left( \frac{\partial q}{\partial t} \right) - \left( \nabla \cdot q \overline{V_h} \right) + E + \delta \]  

(2)

where \( P \), \( E \) and \( \overline{V_h} \) represent precipitation, evaporation, specific humidity, and horizontal wind vector, respectively; and \( \delta \) is the residual term, which includes the influence of all nonlinear terms, contributions from surface processed due to topography and model bias (Seager et al., 2010). \( E \) is estimated according to the equation

\[ E = L/\lambda \]  

(3)

where \( L \) is the surface latent heat flux (W/m\(^2\)) and directly output from GAMIL2.0-CLM2.0, and \( \lambda \) is constant (\( \lambda = 2.501 \times 10^6 \) J/kg). The second term on the right of Equation 2 represents the horizontal convergences of moisture flux, and “( )” represents the vertical mass integration from surface to the tropopause (100 hPa), which can be expressed as: \( \langle \text{Variable}\rangle = \frac{1}{P_g} \int_{P_g}^{100} \text{Variable} dP \). The tendency term \( - \left( \frac{\partial q}{\partial t} \right) \) can be neglected for its monthly mean value is much smaller than that of other terms. At the surface and tropopause, \( \omega \) is approximate to 0 (\( \omega \approx 0 \)), leading to \( \langle \partial_h (\omega q) \rangle = 0 \), then the mass conservation equation can be transformed as \( - \langle q \nabla_h \cdot \overline{V_h} \rangle = - \left( \frac{\omega q}{\omega} \right) \),

### Table 1

| Scheme | Thawing conditions | Freezing conditions |
|--------|-------------------|-------------------|
| A      | \( T > T_{frz} \) and \( \theta_{liq} > 0 \) | \( T < T_{frz} \) and \( \theta_{liq} > 0 \) |
| B      | \( T > T_{frz} \) and \( \theta_{liq} > 0 \) | \( T < T_{frz} \) and \( \theta_{liq} > \theta_{liq,max} \) |

Note. \( \theta_{liq} \) and \( \theta_{liq} \) are the soil liquid-water and ice content, respectively; and \( \theta_{liq,max} \) is the maximum liquid-water content at \( T < T_{frz} \).
which means the vertical moisture advection is equal to the horizontal flow convergence of the moisture term. The vertical moisture advection is associated with the lower troposphere convergence, promoting the upward transport of moisture and leading to rain formation. The Equation 2 can be rewritten as:

\[ P = -\langle \mathbf{V}_h \cdot \nabla h q \rangle - \left\langle \frac{\omega_{\delta q}}{\partial p} \right\rangle + E + \delta \]  

(4)

Furthermore, with changes in the precipitation being expressed as:

\[ P' = -\langle \mathbf{V}_h \cdot \nabla h q \rangle' - \left\langle \frac{\omega_{\delta q}}{\partial p} \right\rangle' + E' + \delta' \]  

(5)

where \( \langle \rangle' \) indicates the linear trend from 1979 to 2008 over the area to be studied. The first and second terms on the right of Equation 5 represent horizontal and vertical moisture advection. The \( -\langle \frac{\omega_{\delta q}}{\partial p} \rangle \) term may be separated into three terms, as follows:

\[-\left\langle \frac{\omega_{\delta q}}{\partial p} \right\rangle' = -\left\langle \frac{\omega'\delta q}{\partial p} \right\rangle -\left\langle \frac{\bar{\omega}\delta q}{\partial p} \right\rangle -\left\langle \frac{\omega'\delta q}{\partial p} \right\rangle \]  

(6)

where \( \langle \rangle \) indicates the climatology during 1979–2008. The first and second terms on the right of Equation 6 are usually considered the dynamic and thermal components of vertical moisture advection. The dynamic component is associated with atmospheric circulation changes while specific humidity is unchanged; the thermal component reflects the influence of specific humidity changes on precipitation. The third term on the right of Equation 6 is nonlinear and is much smaller than the dynamic/thermal components, so it is neglected here. To simplify the equations, “\( vdq \)” and “\( wdq \)” represent \( \langle \mathbf{V}_h \cdot \nabla h q \rangle \) and \( \langle \frac{\omega_{\delta q}}{\partial p} \rangle \), respectively, and \( \langle vdq' \rangle \) and \( \langle wdq' \rangle \) stand for their trends; \( \langle vdq' \rangle \) and \( \langle wdq' \rangle \) denote the \( \frac{\delta q}{\partial p} \) and \( \frac{\omega_{\delta q}}{\partial p} \) terms, respectively.

### 3. Results

#### 3.1. Evaluation of GAMIL2 Simulations

The mean August precipitation over SC during 1979–2008 is illustrated in Figure 1. The gridded CMAP and GPCP observations indicate that precipitation increases gradually from north to south, with the main precipitation center being located in southeastern China (Figures 1a and 1b). The CTRL test simulates the spatial rainfall distribution well, for the pattern correlation coefficient \( r \) between CTRL and CMAP (or GPCP) of -0.80 (or -0.81) between CMAP and GPCP up to 0.93, and are statistically significant at the 95% confidence level. The NEW simulation captures rainfall variations around 1993 well, and significantly correlated with CMAP \( r = 0.80 \) and GPCP \( r = 0.85 \). However, the CTRL simulation fails to describe the abrupt changing process around 1993/94, showing only a weak correlation with observations \( r = 0.11 \) with CMAP and 0.21 with GPCP.
3.2. Moisture-Budget Analysis

A moisture-budget analysis was undertaken for the study regions (20°–32°N, 108°–122°E) to elucidate why the NEW test simulated the August rainfall much well than CTRL test in the mean and long-term trends, especially the latter.

Area-averaged moisture-budget components of precipitation over the study regions are shown in Figure 4. The climatology of August rainfall is mainly contributed by the local evaporation \( (E) \), indicating the importance of land–atmosphere interaction in the region (Figure 4a). The evaporation patterns simulated in the two tests (Figures 5a and 5b) are consistent with the rainfall pattern (Figures 1c and 1d). However, there is little difference in \( E \) between two simulations (Figures 4a, 5a and 5b), and the difference of precipitation simulated by two tests is mainly from the vertical moisture advection \(-<\text{wdq}>=\) (Figure 4a), which the more significant difference is in the north of GuangDong province and south of FuJian province (Figures 5c and 5d), respectively, consistent with their simulated mean August rainfalls. It can be seen that the changing of the freezing/melting scheme has a lesser impact on the \( E \) than the \(-<\text{wdq}>\), while the \( E \) has the most significant contribution in the mean rainfall. Therefore even the NEW test reduced the dry biases than CTRL, but both simulations still have the dry biases in mean August rainfall compared to CMAP (or GPCP).

The linear trend (Figure 4b) in moisture-balance components indicates that the increasing trend in August rainfall over the study area, as described by the NEW simulation, is caused predominantly by enhanced vertical moisture advection (statistically significant at the 95% confidence level). In contrast, horizontal moisture advection

![Figure 1](https://example.com/figure1.png)

**Figure 1.** Mean August precipitation (mm d\(^{-1}\)) in southern China during 1979–2008: observational data of (a) CPC Merged Analysis of Precipitation (CMAP) and (b) Global Precipitation Climatology Project (GPCP), and results of simulations (c) control experiment (CTRL) and (d) NEW. Rectangles represent the study area (20°–32°N, 108°–122°E).
and local evaporation contribute little to the rainfall trend (Figure 4b). The difference of vertical moisture advection (−<wdq>) causes the difference in rainfall trends between the CTRL and NEW simulations, and it was further decomposed into two components: the dynamic term −<w dq> which involves changes in pressure velocity ω and no changes in specific humidity q, and the thermodynamic component −<wdq> with changing in q but no changes in ω, with the −<w dq> predominating (Figure 4b). Spatial patterns of the linear trend in the dynamic term (Figures 6c and 6d) are consistent with those of vertical moisture advection (Figures 6a and 6b) and rainfall (Figures 2c and 2d). The differences in long-term trends for rainfall in the CTRL and NEW simulations over the study regions are thus induced by the dynamic component of vertical moisture advection relating to changing atmospheric circulation.

The time series of vertical moisture advection anomalies and its thermodynamic and dynamic components in the study rainfall regions are shown in Figure 7. The anomalies exhibit negative and positive trends of −0.18 mm d⁻¹ decade⁻¹ and 0.68 mm d⁻¹ decade⁻¹ for 1979–2008 for the CTRL (Figure 7a) and NEW (Figure 7b) simulations, respectively (statistically
Figure 4. Regional area-averaged moisture-budget components in the main rainfall regions: (a) Mean state (mm d$^{-1}$) and (b) linear trend (mm d$^{-1}$ decade$^{-1}$).

Figure 5. August mean evaporation (mm d$^{-1}$) for the (a) CTRL and (b) NEW simulations and vertical moisture advection (mm d$^{-1}$) for (c) CTRL and (d) NEW over the study area during 1979–2008.
significant at the 95% confidence level). Regarding long-term trends in the dynamic and thermodynamic components, the former has significantly decreasing and increasing trends of \(-0.18\) mm d\(^{-1}\) decade\(^{-1}\) and \(0.64\) mm d\(^{-1}\) decade\(^{-1}\) for the CTRL and NEW simulations (Figures 7c and 7d), respectively, with almost the same magnitude as that of vertical moisture advection, whereas the latter displays a weakly increasing trend of \(0.02\) mm d\(^{-1}\) decade\(^{-1}\) for both the CTRL and NEW simulations (Figures 7e and 7f). Thus, the discrepancies between the two simulations of long-term rainfall are also caused by changes in the dynamic component of vertical moisture advection (\(-\langle w dq \rangle\)).

Based on the continuity equation

\[
\nabla h \cdot \mathbf{V}_h + \partial_p \omega = 0
\]

(7)

the first term in the left of Equation 7 indicates the horizontal convergence, and \(\omega\) is the vertical velocity at the pressure coordination. If \(\omega\) is zero at the surface, its value at 500 hPa (\(\omega_{500}\)) approximates the convergence between the middle and lower troposphere. Spatial distributions of the linear trend in \(\omega_{500}\) and associated large-scale circulation at 850 hPa are shown in Figure 8. Compared with the CTRL simulation, the NEW simulation indicates a significantly increasing trend in \(\omega_{500}\) in the study region, accompanied by evident convergence in the lower atmosphere (Figure 8b).
3.3. Physical Mechanisms

To further understand the mechanisms causing the circulation changes between CTRL and NEW simulations, we compare the differences in two tests and analyze how the differences affect the atmosphere circulations.

3.3.1. Changes in Surface States

The CTRL and NEW simulations use different criteria for estimating soil freezing or melting (Table 1), which leads to differences in soil ice content between the two simulations, reflecting the response of frozen soil to global warming to a certain extent. In the CTRL simulations, all liquid water in the soil layers would freeze gradually according to the rate of phase change when the soil temperature is below 0°C, whereas the NEW simulations always hold some liquid water in soil layers over a wide range of temperatures below 0°C.

Typical frozen-soil regions of Eurasia at mid–high latitudes (45°–60°N, 0°–180°E) were considered here in examining the local impacts of the two simulations. Figure 9 shows the seasonal variations in the differences of soil ice content (kg m$^{-2}$) and soil temperature (°C) between NEW and CTRL simulation in each soil layer over Eurasia during 1979–2008. As air temperatures fall in autumn, soil temperature drops gradually to below 0°C,
liquid water freezes from the surface to the deeper soil layer. There is a negative difference of soil ice content between two simulations near the surface in October, and the significant negative differences appear in the deeper soil layer of 1 m on March–April, for the soil temperature in the deeper soil layer has time-lag compared with that in the shallower soil layer. Meanwhile, CLM2.0 discretizes the soil column into 10 layers based on an exponential function, with the thickness of layers increasing as depth increases; that is, layers are thinner near the surface. Therefore, differences in simulated soil ice contents increase with depth (Figure 9a).

Soil freeze–thaw processes accompany phase transitions of soil water and result in the respective release and absorption of latent heat (Yang et al., 2007). Differences in soil ice content thus lead to soil temperature disparities between the two simulations. A significantly negative soil temperature difference appears after November and spreads to deeper layers, reaching the 2.3-m level in April, before a positive difference develops and reaches a maximum in June–July. The soil temperature difference between the two simulations arises mainly because of the latent heat release or absorption during the respective freezing or melting processes. During autumn and winter, soil liquid water freezes and releases the latent heat of condensation, restraining the soil temperature decrease. The CTRL simulation thus provides a higher soil temperature than the NEW simulation in the cold season, when the soil ice content is higher. However, in the CTRL simulation, a higher soil ice content during the cold

Figure 8. Spatial distributions of the linear trend for the vertical velocity at 500 hpa ($\omega_{500}$ shaded area, $-\text{Pa s}^{-1} 30\text{-year}^{-1}$) and their associated large-scale circulation, as indicated by the 850 hPa wind (vectors, $\text{m s}^{-1} 30\text{-year}^{-1}$) (a) and (b) indicates the CTRL and NEW simulations, separately. The dotted areas are statistically significant at the $\geq 90\%$ confidence level.

Figure 9. Time (month)–depth plots of differences in (a) soil ice content (kg m$^{-2}$) and (b) soil temperature ($^\circ\text{C}$) between NEW and CTRL simulations at mid–high latitudes in Eurasia (45°–60°N, 0°–180°E). The dotted areas indicate differences significant at the 95% confidence level.
A season would consume more energy during melting, resulting in less energy being absorbed by the soil, with the CTRL-simulated soil temperature being slightly lower during May–August than that of the NEW simulation. Ice has a high thermal conductivity; therefore, during cold seasons, the difference in soil temperature could spread further into the soil than the ice content (Figure 9b).

Spatial distributions of the linear trend for 2 m air temperatures reflect the response of the lower atmosphere to the soil thermal state (Figure 10). Both simulations present the spatially non-uniform warming trends. Compared to the CTRL test, the NEW test simulates a significant warming trend in Western-Central Siberia (Figure 10).

3.3.2. Response of Atmospheric Circulations

We further analyze the changes of geopotential height and winds at various levels to explore the response of atmospheric circulations to the change of surface thermal state.

The spatial pattern of the linear trend of geopotential height in August during 1979–2008 is consistent from the lower to upper troposphere (not shown), indicating a quasi-barotropic structure of the atmospheric responses to the surface thermal forcing. Compared to the CTRL test, the change of 200 hpa geopotential height simulated by the NEW test presents a “north high–south low” pattern in the middle and high latitude of Eurasia. There is an increasing tendency in the east of Urals and west of Baikal, corresponding with the anomalous anticyclone (Figure 11), which coincides with the warming areas (Figure 10b). While in the south of anticyclone, it has an anomalous easterly flow in about 40°N and cyclone in central and eastern China, which weaken the westerly winds at 200 hpa and strengthen the interaction between low- and high-latitude atmospheres, and make the

Figure 10. Spatial distributions of the linear trend for 2 m air temperature (shaded areas, °C decade$^{-1}$) in August for the (a) CTRL and (b) NEW simulations during 1979–2008. The dotted areas are statistically significant at the 90% confidence level.

Figure 11. Linear trend of 200 hpa geopotential height (shadings, gpm decade$^{-1}$) and 200 hpa winds (vectors, m s$^{-1}$ decade$^{-1}$) in August 1979–2008.
northerly move southward (Figure 12). Previous studies have found correlations between 200 hPa winds and summer precipitation in SC. For example, Kwon et al. (2007) reported that the mean June–August zonal wind speed at 200 hPa had displayed a decreasing trend over East Asia since the mid-1990s, negatively correlating with summer precipitation in southeastern China. Chen and Lu (2014) reported that increased rainfall in SC around the mid-1990s coincided with decadal warming over northeastern Asia associated with a weakened subtropical westerly jet stream, which occurs only in summer (June, July, and August). Our findings are consistent with these previous works.

In addition, the anomaly anticyclone in NEW test is warm, for the significant warming trend of the whole atmosphere at 40°–60°N zonal averaging over 75°–125°E, corresponding with the anomalous descending flows at about 45°N (Figure 12), and a convergence anomaly cyclone is initiated in the lower troposphere over SC regions, accompanied by enhanced ascending (Figure 8). Meanwhile, the anomalous circulation results in strong moisture flux convergence at SC regions, and abundant water vapor from the South China Sea and the western North Pacific converge in this region (Figure 13), which provides adequate water vapor for precipitation. So the precipitation over SC shows an increasing trend in the NEW test.

To summarize, compared with the CTRL test, the NEW simulation has relatively little soil ice during autumn and winter due to the existence of supercooled water at temperatures below 0°C in seasonal frozen-soil regions. Because the phase change between water and ice accompanies the release and absorption of energy, the NEW simulation has a higher warming trend in 2 m air temperature in August, especially in Western–Central Siberia. It not only produces a north–south temperature gradient opposing the original gradient produced by solar forcing and weakening the meridional thermal contrast, and induces an anomaly “north high–south low” circulation pattern in the middle and high latitude Eurasia. Both of them result in the 200 hPa zonal winds weakened in the NEW simulation, with the blocking effect of the jet stream being reduced and the interaction between low- and high-latitude atmospheres strengthened (Zhu et al., 2011), and there is a northerly anomaly at 850 hPa in Northern China in August, with a weakening of the East Asian summer monsoon. Meanwhile, a convergence anomaly

Figure 12. Latitude pressure cross-section of air temperature trends (shaded, °C 30 years−1) and the trend in vertical velocity and meridional winds (vectors) in August, zonally averaged from 75°E to 125°E during 1979–2008 for the CTRL (a) and NEW (b) simulations. Dotted areas indicate the air temperature trend at the 95% confidence level.
cyclone is initiated in the lower troposphere over SC regions, accompanied by enhanced ascending, with precipitation increasing in that region in the NEW simulation.

4. Discussions and Conclusions

This study investigated the effects of soil freeze-thaw processes on August rainfall in SC by applying the coupled land-atmosphere model (GAMIL2.0-CLM2.0) with two different soil freeze–thaw schemes (the CTRL and NEW simulations) under the same climatological SST forcing. Moisture-budget analysis was applied to elucidate rainfall differences between the two simulations.

The intrinsic difference between the two soil freeze–thaw schemes was the diversity of the simulated soil ice content in soil layers, reflecting the response of frozen soil to global warming to some extent. The NEW simulation demonstrated that the inclusion of supercooled water in the soil freeze–thaw scheme reproduced the climatology and trends of August precipitation in SC, indicating that the variability of soil freeze–thaw processes influences August precipitation in SC. Furthermore, results indicate that the dynamic component of vertical moisture advection ($-\langle w dq \rangle$), related to atmospheric circulation changes, significantly influences the August precipitation trend.

The physical mechanisms were further explored to provide a simple explanation for how the changing freeze–thaw processes during autumn and winter affect the August rainfall in SC. The NEW simulation modeled the higher soil/air temperatures in August especially in the north of 40°N, for simulating lower soil ice contents during cold seasons, lead to the meridional thermal contrast weakened, and result in the 200 hPa zonal winds decreased, indicating the blocking effect of the jet stream reduced and the interaction between low- and high-latitude atmospheres strengthened, therefore the northerly wind at 850 hpa strengthened and was suppressed to the south of China, which is more beneficial to the convergence of the lower atmosphere and moisture flux, and thus causing vertical velocity increased and more precipitation. However, the changing of freeze–thaw processes likely induced a Rossby wave-like pattern, thus affecting August rainfall in SC. Previous studies have shown that the extension of stationary Rossby waves in the upper troposphere at mid–high latitudes from Europe to East China leads to a negative teleconnection between rainfall in SC and the East Europe Plain in July and August (Su & Lu, 2014). The Eurasian spring snow anomaly is associated with East Asian summer precipitation by triggering of an anomalous mid-latitude Eurasian wave train propagating eastward (Zhang et al., 2017). Further analysis of the mechanism related to the wave-train propagating is beyond the scope of this study.

To conclude, this study provides a new interpretation for the “southern flooding” during 1979–2008 from the view of the changing in frozen soil; meanwhile, attention should be paid to the frozen soil variability in the
mid-high latitude at the background of global warming, for their significant impact on the precipitation in SC regions, and take adequate measures to slow down the effect of the frozen soil changing.

Data Availability Statement

The data set of simulations by GAMIL2.0 was acquired from the website of (http://data.lasg.ac.cn/xiakun/). The CMAP precipitation and GPCP v2.3 precipitation datasets can be found at https://psl.noaa.gov/data/gridded/index.html. We thank the anonymous reviewers for their constructive comments and suggestions.

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