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Hydrological and geocryological response of winter streamflow to climate warming in Northeast China

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Abstract

An abrupt warming of regional climate with a 1.3 °C rise in annual air temperature, coupled with an increase of 20–40% in precipitation, has occurred in the 1990s in the permafrost region of Northeast China. The geocryological and hydrological responses of a river basin at high latitude and at altitude with some permafrost are detected based on monthly climatological and streamflow data for 40 years (1958–1998). The variation in depth of the active layer is estimated by an empirical model using annual air temperature, its annual amplitude and the maximum thickness of snow cover. Significant responses of winter streamflows to a 2.4 °C of air temperature warming during December to February were observed. This was especially true for the greatest warming (4.4 °C in February during the 1990s) when runoff increased by 80% in February and by 100% in March from the prior. These responses are caused by a change in depth and temperature of the active layer ranging from 1.5 to 3.0 m in areas where the drainage of the unfrozen water can occur when the ground temperature rises above 0 °C from − 0.8 °C in February and March. The depth of the seasonal frost has shrunk by about 30 cm and the active layer thickness increased by about 40 cm in permafrost in the 1990s because of the warmer climate. The hydrological response from winter streamflows in permafrost areas is more significant and quicker than that from the seasonal frost areas. The freezing and drainage of ground water at 2.0–3.0 m deep in March is very sensitive to the climatic warming.

Keywords: Permafrost; Seasonally frozen ground; Air temperature; Precipitation; Unfrozen water; Discharge; Ground temperature; Active layer

1. Introduction

Scientists have predicted that increasing greenhouse gases can cause change in regional air temperature and precipitation. Changes in climate may already have affected elements of the hydrologic cycle such as precipitation redistribution, snow accumulation and meltwater, evaporation, surface and subsurface water (Houghton et al., 1995). These potential effects may already have made some important impacts on the water resources and ecological environments of the cold region in Northeast China, Northwest China and the Tibetan Plateau (Shi and Li, 1990; Liu et al., 1999; Wang et al., 1996a,b), particularly in the Songhuajiang River Basin of Northeast China (Zhou et al., 1996, 2000; Luo, 1996).

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The seasonal pattern of river flow regime integrates the contributions of regional climatic and geographical conditions at a scale. This average pattern can be stable, demonstrating the same seasonality in river runoff from year to year, or unstable when the flow regime alternates between a number of different climates. Being dependent on climate and physical geography, the flow regime patterns respond to the interaction between atmosphere and land surface and the changes in these two main variables—air temperature and precipitation. The objective of this paper is to detect hydrological and geocryological response from subsurface water in winter in the geocryological region to the climate change, which has important significance for predicting change in water resources and the ecological environment. On the other hand, some representative and sensitive indicators among river discharges, such as the annual minimum water flow should show a correlation with the depth of the active layer.

2. Study area

Fig. 1 shows the Ganhe River (49°15′–51°30′N, 122°30′–125°10′E) in Northeast China and the extent of both the discontinuous permafrost from the south and the continuous permafrost in the north along the DaHingganling Mountains. The areal distribution and the depth of the active layer are closely associated with to the latitude and altitude of the mountains (Ji et al., 1994; Zhou et al., 1996). The soil types from the upstream to the downstream are gray forest, podzolic, chernozem and clay soil.

On the geocryology, the role of the active layer is used to evaluate water transportation because the bottom of the frozen ground is still in frost under the permafrost, and the unfrozen bottom under the seasonally frozen ground. The thickness of the active layer and depth of seasonal frost change with temperature, soil moisture, snowpack depth canopy cover and terrain from year to year. The mean thickness of the

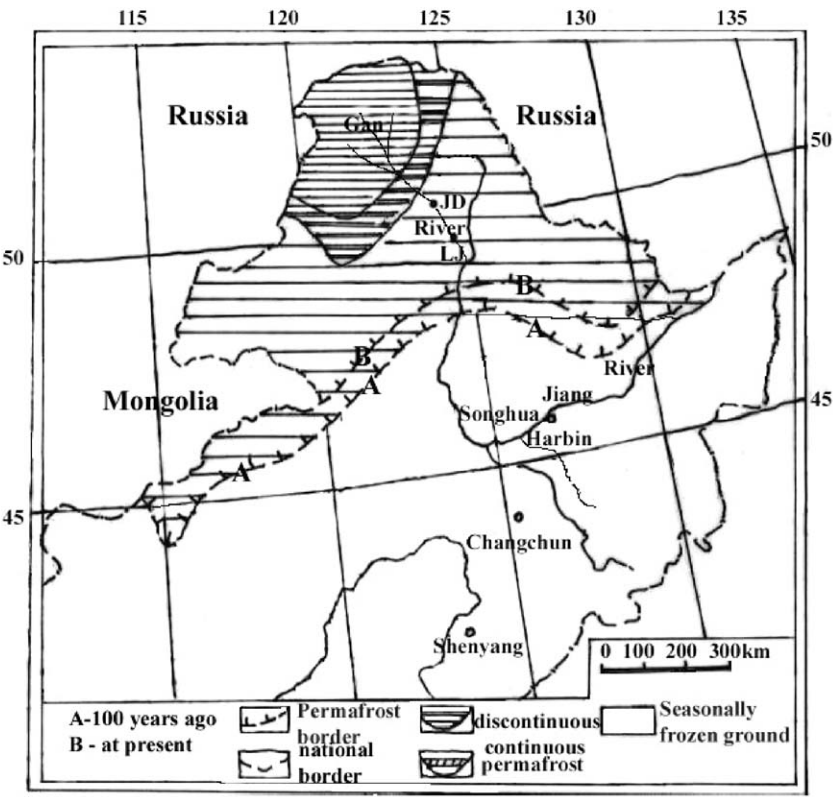

Fig. 1. Permafrost extent and the studied basin in Northeast China.

active layer in the Ganhe River Basin is in the range of 2.4–2.7 m in permafrost, while seasonally frozen ground ranges between 1.6 and 1.9 m (Luo, 1996; Zhou et al., 2000).

The Ganhe River has a drainage area of 9574 km^2 above the mid-stream gauge at station JD and a total area of 19,570 km^2 above the river outlet at station LJ. Local topography consists of broad southwest–northeast trend ridges and isolated flats and mountain peaks as high as 1490 m a.s.l. in the headwaters. Interspersed broad river valleys and relatively flat plains are found in the lower basin at 340 m a.s.l. Finally, the river flows southward into the Songhuajiang River Basin at 230 m a.s.l. Monthly discharge, air temperature and precipitation data in this study have been collected for 40 years (1958–1998) at two stations, JD and LD. Average annual air temperature at the catchment center JD is − 0.9 °C, and annual precipitation at stations JD and LJ is about 480–500 mm; the basin-wide annual precipitation is nearly homogenous ranging from 450 mm at downstream end to 600 mm at the mid-stream, with 60–70% falling during summer. Fig. 2 shows the hydrograph for 1988 illustrating a simple pattern of streamflow under a cold continental climate. The long and cold winter lasts from October to April annually, with the lowest daily air temperature below − 40 °C, usually in January, and the lowest monthly air temperature is about − 25 °C also in January, some time in December. In the northern high mountains, the January average air temperature can go below − 30.0 °C. There exist both the permafrost and the seasonal frost. The area above station JD in upstream of the catchment has island permafrost, and the areas between the two stations have sparse permafrost and seasonally frozen ground above station LJ. Snowpacks appear from October to June, especially at the high altitude. So the river in winter is recharged only by deep subsurface water and greatly affected by the freezing and thawing of the surficial soils. During the mild and rainy summer from June to September, the weather are controlled by

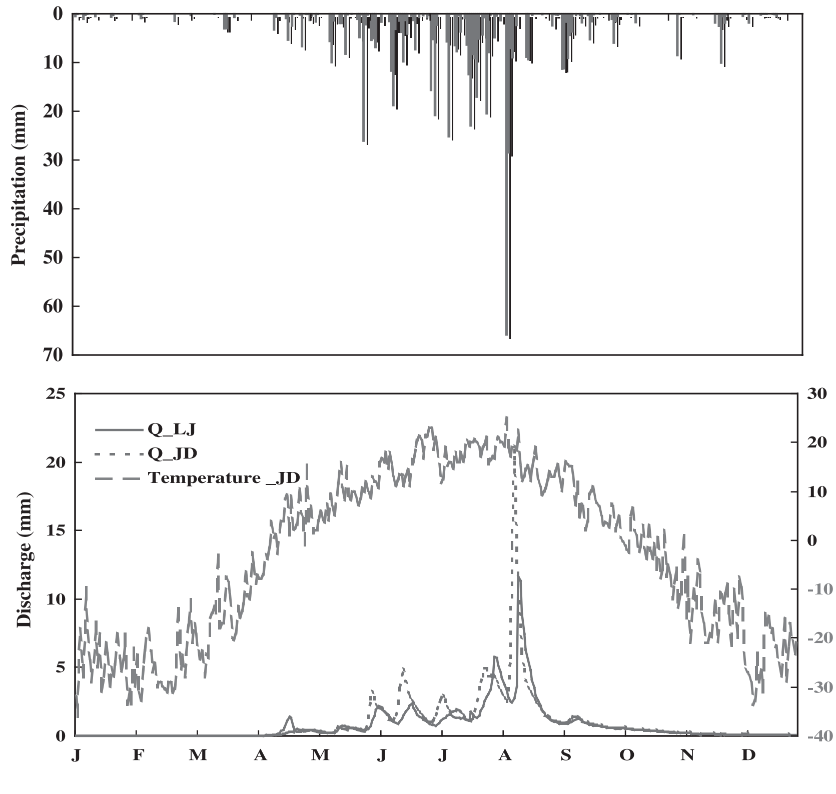

Fig. 2. Hydrographs of daily rainfall, air temperature and streamflow at the Ganhe River.

the Mongolian cyclones and Asian Monsoon, surface flow from the rainfall runoff is the main form of water cycling on the land surface.

3. Climate change in the recent 40 years

Chinese meteorologists have detected and estimated the impact of global warming to the regional temperature and precipitation in China. It has been found that the regional climate in Northeast China extreme warm since the 1980s, particularly in the northern Songhuajiang Basin where the climate variability was the largest in China (Lin et al., 1995; Li et al., 1996). The long-term climatic record at Harbin shows significant warming with temperature rising and precipitation decreasing during the past 80 years (1909–1989) (Luo, 1996). Although there is long-term evidence of the climate change in the studied area, we only have meteorological records for the most recent 40 years at stations JD and LJ.

3.1. Change in air temperature and precipitation

Fig. 3 shows the changes of annual and winter mean air temperature (data smoothed using a 3-year average) for a 40-year period at station JD. Trends are difficult to see in the first 10 years; for the next 20 years, there is a gradual warming, with a rapid step in warming since the late 1980s. The mean annual air temperature in the 1990s is approximately 1.3 °C higher than that during the 1960s up to the 1980s. The highest annual air temperature during the recent 40 years was recorded in 1990 and 1995, all annual air temperatures in the 1990s were higher than the interannual average, except for − 1.0 °C in 1996, which is not shown because of the smoothing.

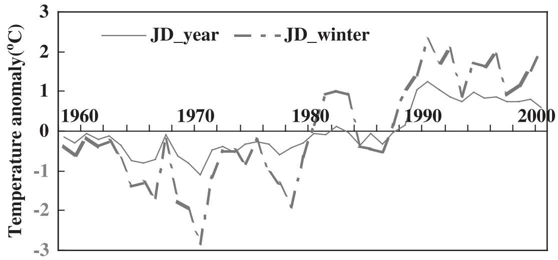

Fig. 3. Variation of annual and winter air temperature.

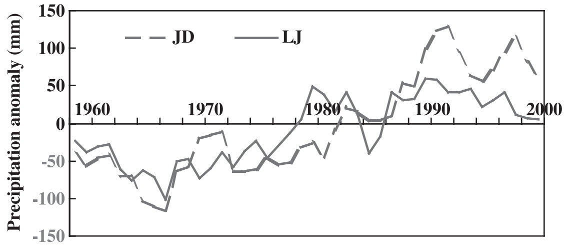

Fig. 4. Variation of annual precipitation.

Fig. 4 illustrates that precipitation from 1958 to the late 1960s was gradually decreasing, then it turned to increasing at the two stations on records of 40 years. The precipitation everywhere, in the plains and in the middle mountains, all significantly increased since the late 1960s and, furthermore, greatly exceeded the interannual average after the 1980s, especially at station JD. Annual precipitation in the 1990s averaged 80 mm or about 20% at station LJ, and averaged 160 mm or about 40% at station JD, more than the annual average during the 1950s up to the 1970s. Observed maximum annual precipitation at station JD was 752.6 mm in 1998 and 737.5 mm in 1991 and exceeded 700 mm in the 1950s up to the 1980s only once. It suggests even greater increases in precipitation in the high elevation in the catchment.

3.2. Change in season

The change of the regional climate in the 1990s, as compared to that before the 1980s, exhibits obvious seasonality during a year; the rise in air temperature and increase in precipitation are different in each month. The greatest rise in air temperature occurred firstly in the winter seasons as shown in Fig. 5. The

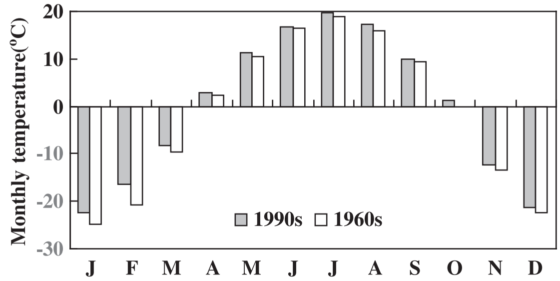

Fig. 5. Monthly variation of air temperature comparing in the 1990s and in the 1960s to the 1980s.

magnitude of the temperature rising is 4.4 °C in February, 2.2 °C in January, 1.4 °C in December and 1.5 °C in March and November, compared with the 1960s. Meanwhile, summer air temperature slowly rose by about 0.7–1.2 °C since the 1960s (Fig. 5). In other words, the climate warming is mainly caused by the rise of air temperatures in winter and spring.

Unlike the changes in air temperature, the seasonal increase in precipitation occurred mainly in the months from June to September; it increased by about 40–50 mm in summer months and by 5–20 mm in autumn months during the 1990s than that measured earlier. The increase in precipitation was much greater at station JD than that at station LD. Snowfall in winters in the 1990s, however, did not exhibit an evident increase. Overall, the climate in Northeast China seems to be becoming warmer in winter and wetter in summer since the 1990s.

4. Hydrological responses from winter streamflow

It is known that water in frozen soil generally exists in three phases: vapor, liquid water and ice. It is also accepted that the unfrozen water is mobile if a driving gradient of the ground temperature prevails. Hydrological observation and experimental research suggests that there exists a strong correlation between the flux of the unfrozen water and the gradient of total water content in frozen soil. The theoretical justification is made under the assumption that the soil water diffusivity is proportional to some power of water content. The unfrozen water content mainly depends more specifically on surface area of soil type and temperature (Nakano et al., 1982, 1983, 1984). When water movement in a frozen porous media is considered, it is reasonable to expect that this problem can be treated in manner similar to the problem of water transport in unfrozen soil. Under such condition, water transport in frozen soil corresponds to that in unfrozen soil in which the volumetric water content is less than 10%.

Anderson and Morgenstern (1973) first presented an empirical formula for estimating the unfrozen water transport with a decline in the soil temperature. Nakano et al. also have given similar correlation curve below -1.0 °C. It has been recognized that the phase-composition data of most frozen soils are conveniently represented as:

$$W_u = a(-T_g)^b \quad (0\ °C < T < -15.0\ °C) \tag{1}$$

where W_u is the unfrozen water content in % and T_g is the soil temperature in °C; $a>0$ and $b<0$ are constants depending on moisture and temperature of each soil whose values have been given by Lunardini et al. (1981).

It is clear that the unfrozen water content is an asymptotic curve that coincides with the decline in soil temperature, the greatest transport rate of soil water occur at the beginning of the freeze cycle when the temperature is below 0 °C down to -2 °C. Finally, the soils obtain a stable moisture content, which does not change with continuously declining temperatures. In our case, we are interested in the unfrozen water content around -1 °C where unfrozen water content increase with the rise of the ground temperature in freezing phase in the 1990s. A least-squared approximation for the marine clay was determined such that $W_u = 13.84(-T_g)^{-0.4861}$ when freezing and $W_u = 12.69(-T_g)^{-0.508}$ when thawing (Nakano et al., 1982).

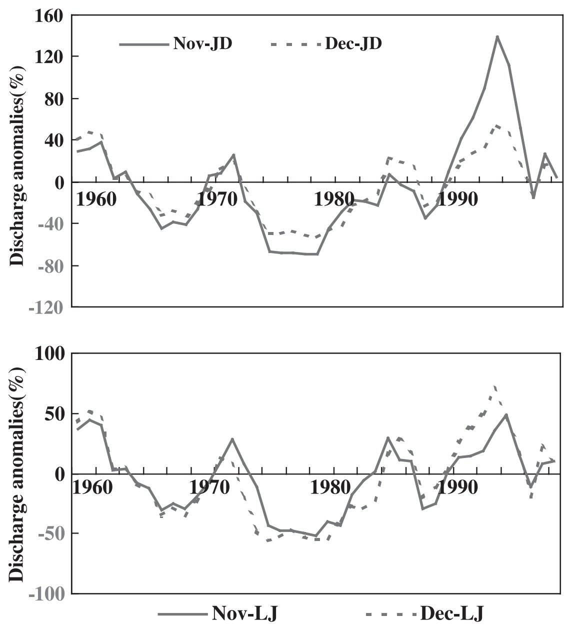

Fig. 6. Variation of streamflow in autumn.

The change in precipitation and air temperature in autumn and winter directly affects snow accumulation on land surface; the warming can enhance the drainage and weaken freezing of soil water at the active layer at the bottom. Figs. 6 and 7 show the varying graphs of monthly runoff in percentage in autumn and winter (November to March, there are rainfall and snowmelt in October) at the stations. It is interesting that the river discharge did not immediately produce an early hydrological response to the climate warming. In the 1980s, an increase in streamflow can be found, although it is greater in the 1990s. River runoff in November and December, however, indicated a significant increase since the 1990s delaying at least 10 years for the interannual series of the temperatures and lagging 1 month in a hydrological year. It coincides with possible changes in the depth of the active layer under water and heat conditions in the 1990s.

The annual minimum discharge occurs most in February, sometimes in March every year, in the form of groundwater flow without the infiltration supply of surface water and snowmelt water due to the freezing up. However, Fig. 7 shows the great variation of the winter streamflow (January through March) again for 40 years. It is very evident that the greatest variation

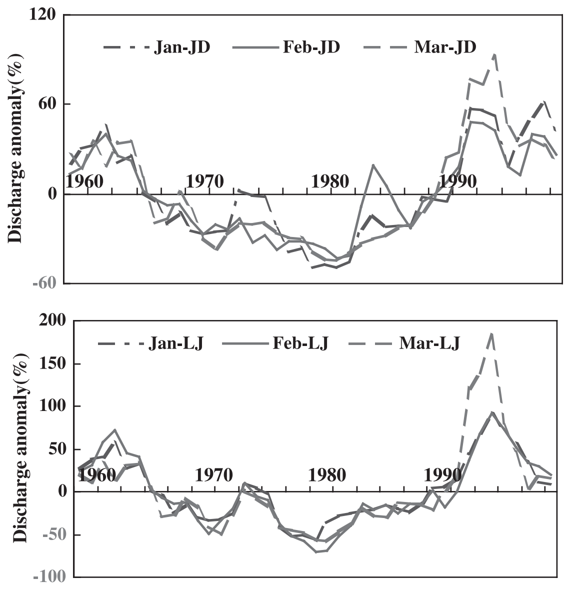

Fig. 7. Variation of subsurface flow in winter (January to March).

(increasing) in discharge occurred in March, instead of February, although the greatest warming occurred in February. The streamflows in the winter months exhibited great variation with an increasing tendency that coincides with the rise of winter air temperature in the 1990s. However, the responses of winter streamflow to the winter warming create a lagging in time, which reminds us that the changes in discharge discussed above are actually the results of change in the temperature and depth of the active layer where heat conduction lag to the climate warming.

5. Geocrylogical response

5.1. Response of the ground temperature

As is well known, soil temperature (T_g) depends on temperature at land surface (T_o), and that the ground temperature fluctuates periodically with air temperature as follows:

$$T_g(0,t) = T_o + A_o \sin \omega t \tag{2}$$

where T_o is the average surface temperature, A_o is the amplitude of fluctuating component, ω is the angular frequency and t is time. The ground temperature at a depth z and time t, $T_g(z,t)$, is obtained by solving the heat conduction equation as follow:

$$T_g(z,t) = T_o + A_o \exp\left(-z\sqrt{\frac{\omega}{2a}}\right)\sin\left(\omega t - z\sqrt{\frac{\omega}{2a}}\right) \tag{3}$$

where z is thickness of the soil layer; $a = \lambda/c$ is temperature conductivity of soil, which can be estimated from the thermal regime of the ground temperature as follows. If a stable gradient of the ground temperature T_{go} is maintained at the deep soil layer, $\frac{\partial T}{\partial z}\big|_{z\to\infty} = T_{go}$, then:

$$T_g(z,t) = T_o + T_{go}z + A_o \exp\left(-z\sqrt{\frac{\omega}{2a}}\right) \times \sin\left(\omega t - z\sqrt{\frac{\omega}{2a}}\right) \tag{4}$$

From this solution, the following important conclusions can be obtained:

(a) The amplitude of soil temperatures with the depth attenuates in the form of an exponential function as:

$$A(z) = A_o \exp\left(-z\sqrt{\frac{\omega}{2a}}\right) \tag{5}$$

(b) The fluctuation of the ground temperature with depth delays a time interval as:

$$t = \frac{z}{2}\sqrt{\frac{l}{a\pi}} \tag{6}$$

where $\lambda = 2\pi/\omega$ is the conductive period.

(c) Conductive depth of the ground temperature wave depends on the period of surface temperature as:

$$z_1 = z_2\sqrt{\frac{l_1}{l_2}} \tag{7}$$

(d) The soil depth with the amplitude of surface temperature A_o as:

$$A(z_o) = \ln\left(\frac{A_o}{A(z_o)}\right)\sqrt{\frac{ll}{c\pi}} \tag{8}$$

The deeper the soil layer, the slower the heat conduction is.

According to the equations above, T_o is estimated from the monthly surface temperature; A_o is the average annual temperature amplitude of about 43 °C; $\lambda = 12$ months is the conductive period; and $\alpha = \lambda/c = 0.024/\mathrm{m}^2$ month is the temperature conductivity of the soils induced from Eq. (5) as follows:

$$a = \frac{-z^2\pi}{t\ln\left(\frac{A(z)}{A_o}\right)} \tag{9}$$

where z is the depth at the investigated regime data of ground temperature at a depth of 1.5–3.0 m in 1974–1975 (Li and Cheng, 1995), t is the time in months and A_z and A_o are the amplitudes of ground temperature at 1.5 and 3.0 m plotted in Fig. 8. Fig. 8 illustrates that the lowest ground temperature at a depth of 3.0 m has risen to almost 1.0 °C with the lowest air temperature

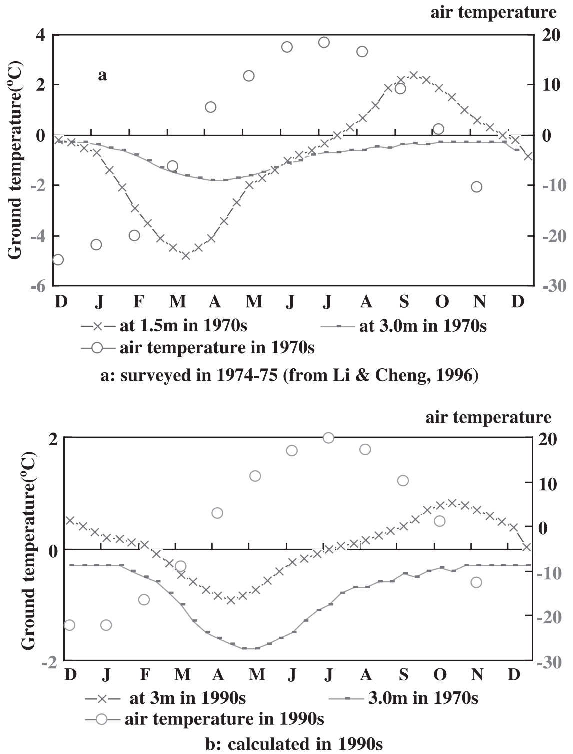

Fig. 8. Cycle curve of ground temperature at depths of 1.5 and 3.0 m below land surface in studied mountains.

occurring about 1 month earlier in the 1990s than that in the 1970s. The lowest air temperature in the winter frequently occurred in December during the 1990s, in 1992, 1994, 1995, 1997, 1999 and 2000, instead of January as it occurred before the 1990s. Similar conditions of climate and ground temperature profiles were reported in a neighboring river basin (51°15′–52°45′N, 122°–126°43′E), and at Harbin, the T_g value at a depth of 1.6 m was below 0° in December and reached the lowest ground temperature in late February (Ji et al., 1994, 1998).

Fig. 8a illustrates the ground temperatures at depths of 1.5 and 3.0 m investigated in 1974–1975 (Li and Cheng, 1995) and estimated from formulae above for permafrost area of the Ganhe River at a depth of 3.0 m (Fig. 8b). From Fig. 8a, the ground temperature at the depth of 1.5 m is below 0 °C from December to July and reaches the lowest value of about −5 °C in March; the highest and lowest ground temperatures lagged the air temperature extremes by about 2–3 months. From Fig. 8b, the ground temperature at 3.0

m below the surface ranged from − 0.3 to − 1.8 °C in 1975, and it is clear that the ground temperature is slightly affected by air temperature in the warm season. The lowest ground temperature occurred in April, thus lagging to the lowest air temperature in December by about 4 months. Since the 1990s, the average temperature at a depth of 3.0 m has yielded positive values, the lowest temperature has risen by about 1.0 °C from − 1.8 °C to − 0.8 °C, and the freezing period was shorten and removed ahead to April from May in 1975.

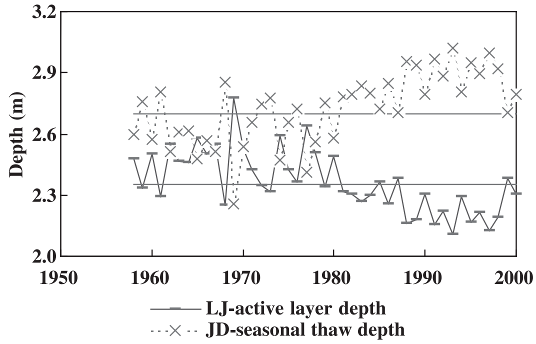

Fig. 9. Changes in depth of the active layer in the studied area.

5.2. Response of the active layer

Depth of the active layer in the frozen ground region is influenced by many factors such as soil temperature and moisture, snowpack and canopy (Nelson et al., 1997; Yamazaki et al., 1998). Generally speaking, the depth is determined in the first reference by regional climate. Most analytic treatments have used air and soil temperature measured at nearby point locations as its primary determinant. Broad spatial patterns of the temperature related directly to macroclimate can indeed be discerned at regional and local scales. Colin et al. (1999) had compared the single-year air and ground temperature records using reduced major axis analysis to demonstrate the limitations in using air temperature as indicators of ground temperatures in an Arctic alpine environment. Air temperatures predict some 10- and 50-cm-level ground temperatures with as much as 95% explanatory power at daily and monthly scales. In addition, seasonal snowpack has significant insulating effects on the thermal regime of frozen ground and the variation in the active layer depth (Zhang et al., 1997; Yamazaki et al., 1998; Yang et al., 2000), and as would be expected, the smoothing data from daily to monthly scales generally improves the correlation between air and ground temperature; this phenomenon is also due, in part, to lag effect associated with the heat conduction.

Based on the geocryological data and annual and monthly mean data observed at 71 meteorological stations in Northeast China in the past 40 years (1951–1990), we proposed in this study a model formulated as Eq. (10) and illustrated in Fig. 9, which can predict the maximum depth of the active layer D_a through the annual average temperature T_g that is from a linear multi-regression method as shown in Eq. (11) developed by Zhou et al. (1996) and Gao et al. (1996) correlating annual air temperature, its annual amplitude and the maximum thickness of snowpack in January. The monthly data at station JD can be used in this study for estimating the changes in seasonal frost and the active layer in the permafrost area by these models:

$$D_a = 300 - 20.5T_g \quad (10)$$

$$T_g = 8.82 + 0.619T_a - 0.129T_{mm} + 0.054S_t \quad (11)$$

where T_a, T_{mm} and T_g are the annual air temperature, its amplitude (amplitude between the lowest and highest monthly air temperature) and the temperature of the active layer bottom in °C; S_t (in cm) is the average thickness of snowpack in January; and D_a (in cm) is the maximum depth of the active layer at winter end.

The annual air temperature and the amplitude are available from the monthly data, the thickness of snow cover until every January can be estimate by total precipitation from October to the next January when average air temperature is below 1 °C, divided by 220 kg/m^3 of the average snow density (Ji et al., 1994; Yamazaki et al., 1998). By Eqs. (10) and (11), the annual depth of active layer at station JD can be estimated and is shown in Fig. 9.

From Fig. 9, it is clear that the depth of the seasonal frost has been decreased by about 30 cm from 200 cm in the 1960s to 170 cm in the 1990s. The active layer has increased by about 40 cm at station JD over permafrost from an average of 240 cm to an average of 280 cm.

According to the given warming variability of air temperature from climate data at station JD mentioned above and to Eqs. (3)–(8), the soil layer around 3.0 m below the ground has been should thawed into the active layer since the 1990s. The fluctuation of streamflow in winter, particularly in March and February, is affected by gravity drainage of the unfrozen water under a depth of 1.5 m. It is not occasional that according to Nakano et al.'s (1982, 1983, 1984) research, a rapid increase of the unfrozen water could occur when the ground temperature rises even below freezing temperature and from 0 to -2 °C in the marine clay. The unfrozen water in the active layer here should transport downward into deep unfrozen soils, then drain out in the form of groundwater flow in March and February. In other words, the drainage occurrence of frost squeezing depends on the timing and depth of the lowest temperature at the active layer bottom.

6. Conclusions

The climate warming since the 1980s, and particularly in the 1990s, has caused a significant rise in ground temperature, thickening depth of the active layer in permafrost areas, a thinning in the depth of the seasonal frost, enhanced transport of subsurface water through unfrozen soils that persist later into the winter and the contribution of this flow to winter base flow. The subsurface water did not immediately produce an active feedback to climate fluctuation in the 1980s but was delayed about 2 months, even no response at other times, during a period when the mean annual air temperature rose by about 1.3 °C and the precipitation increased by 20–40% on the average during the 1990s than in the previous 30 years. The rise of the temperature occurred mainly in winter, and the increase of precipitation in summer and autumn as snowfall. As a result of the interaction between atmosphere and land surface, enhanced groundwater flow resulted from the change in climate. Among all hydrological indices from field observation, the annual minimum discharge exhibited the maximum variability; this resulted from increasing subsurface flow in winter months because of climatic warming. From the point of the hydrological scale, the response to the climate fluctuations from a drainage basin with the continuous permafrost and in greater precipitation seems to be more sensitive to change in air temperature than that from the river basin with the discontinuous permafrost and in less precipitation.

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