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Research Paper

Can forest water yields be increased with increased precipitation in a Qinghai spruce forest in arid northwestern China?

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ABSTRACT

Climate-induced changes in regional precipitation are projected to affect forest water yields, although the effects are expected to vary. Few studies, in fact, have examined the response of conifer forests to increases or decreases in precipitation, in arid regions. To answer the question posed above, we investigated the variability of forest canopy transpiration versus precipitation during the 2011–2013 growing seasons, and constructed a complete hydrological budget of an arid montane spruce forest by directly measuring its main component at the stand level, at long-term experimental catchments on Qilianshan Mountain, located in the upper Heihe River Basin, in the arid region of northwest China. It was found that total precipitation during the 2012 and 2013 growing seasons was 12.3% and 36.5% higher, respectively, than during the 2011 growing season, and total stand transpiration during the 2012 and 2013 growing season was 12.5% and 21.7% higher, respectively, than during the 2011 growing season. In the study period, transpiration, soil and moss evaportranspiration, canopy evaporation, and the drainage and change in soil water storage accounted for 71.1%, 19.9%, 5.3%, and 3.8% of the precipitation, respectively. Although the precipitation increased during this study period, the increase was not sufficient to increase the forest water yield. In the future, though, if the precipitation continues to increase in this forest, it may be sufficient to effect such an increase in forest water yield.

1. Introduction

Global warming has become an indisputable fact (IPCC, 2007), and changes in global and regional climate have raised concerns about the potential impact of precipitation and temperature on the water budgets of terrestrial ecosystems (Vitousek, 1994; Hanson and Wullschleger, 2003; Knapp et al., 2001; Weltzin et al., 2003). The spatio-temporal distribution of precipitation and its trends and variability, associated with climate change caused by global warming and human activities, have received much attention, particularly in arid and semi-arid areas where precipitation is an extremely important environmental factor. In these areas, the distribution and variation of precipitation can have a significant impact on the local ecological systems and environment (Lioubimtseva et al., 2005; Lioubimtseva and Henebry, 2009). Increases and decreases in precipitation are expected to alter surface evaporation, transpiration, and soil water content, which, in turn, will have implications for plant function, catchment water yield, and hydrologic budgets across broad spatial scales (Wullschleger and Hanson, 2006).

In arid regions, and especially in northwestern China, water is almost always a strong limiting factor for the continuous existence of biota, even when temperature, radiation, and nutrients are sufficient. In this area, the mountains—including ranges such as the Qilian, the Helan, and the Tianshan, experience higher precipitation than some of the lower-elevation lands in the region, and this precipitation plays an important role in supplying water to those lowlands. Some studies have found that annual precipitation in the arid zone of northwest China has been tending toward moister conditions than normal (Yang et al., 2014; Li et al., 2013; Wang et al., 2013; Lan et al., 2012; Wang et al., 2008; Xu et al., 2008; Zhai et al., 2005). The effects of this increased precipitation on water yield from riparian headwaters are of great concern, given their key role as sources of water supply (National Research Council, 2008). It is not known, however, whether the increased water yield from these headwaters has resulted from an increase in alpine precipitation. One must consider quantifying hydrologic components in the montane forested catchments of this area, as this information is essential for water resource management, especially considering that forest management in the 21st century may need to include provisions for multiple ecosystem services, including maintaining adequate water supplies (Likens and Franklin, 2009).

The Heihe River, which originates in the Qilian Mountains, is one of

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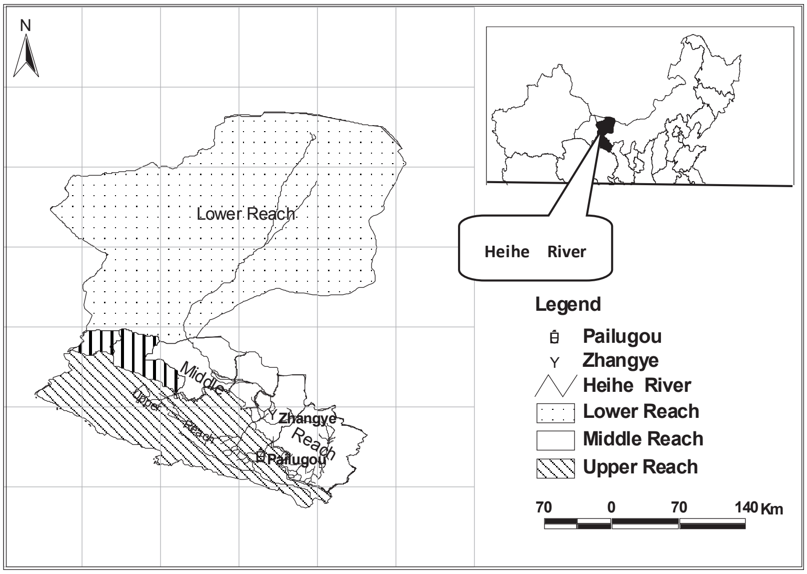

Fig. 1. Map of the Heihe River Basin and its location in the Northern of China.

the largest inland rivers in the arid zone of northwestern China. The forest vegetation of the Qilian Mountains not only serves as a valuable forest resource, but also performs the ecological function of water storage. In fact, this forest is designated as a water conservation forest. These water resources support the oasis ecosystem of the middle and lower reaches of the inland river system, and the adverse processes of deep drainage and runoff are thereby minimized.

Qinghai spruce (*Picea crassifolia*) is the dominant tree species in the Qilian Mountains. The upper Heihe River Basin comprises 43.16×10^4 hm^2 of forest area, with Qinghai spruce forests occupying about 25% of the total forest area and 78% of the arbor forests (Chang et al., 2001). Because of this species' critical role in maintaining favorable hydrological processes, it is important to understand its physiological processes and its contributions to maintaining the water balance. To gain a clearer understanding of the water budget of this forest, it is essential to develop a modeling framework for predicting the effects of possible land use and climate changes on water resources in the region, especially rainfall change, because alterations in the quantity, intensity, and frequency of precipitation, and the resulting changes in soil water availability, could have important implications for Qinghai spruce forest ecosystems. However, hydrological knowledge of this system remains poor. Therefore, a detailed estimate of all the components comprising this integrated flux is needed, in order to understand the controls over ecosystem responses to changes in precipitation and climate, namely, Qinghai spruce forests' transpiration (E_C), soil and moss evaportranspiration(E_s), and evaporation originating from precipitation intercepted by the canopy (E_I). This study aimed to fulfill this knowledge gap.

Transpiration by Qinghai spruce forests can be a substantial component of the total water budget of montane systems in the upper Heihe River Basin. Under the complex terrain and spatial heterogeneity of the forest environment, there are a considerable uncertainty in the estimates of transpiration in these systems. Because complex terrain and spatial heterogeneity do not limit its applicability, a sap-flow technique is most effective. This technique has proved to be a useful methodology for investigating forest water use at both temporal and spatial scales (Granier, 1985; Smith and Allen, 1996; Wilson et al., 2001; Green et al., 2003; Kumagai et al., 2007; Ford et al., 2007). Common methods for studying sap flow include heat pulse (Swanson and Whitfield, 1981; Edwards and Warwick, 1984; Edwards et al., 1996; Vertessy et al., 1997; Chang et al., 2006; Chang et al., 2014a, 2014b), stem segment heat balance (Čermák et al., 1998; Cienciala et al., 1992; Jiménez et al., 1996; Čermák et al., 2004), heat dissipation (Granier, 1985; Loustau et al., 1998; Meinzer et al., 2001; Bush et al., 2010), and heat field deformation (Čermák et al., 1998; Meiresonne et al., 1999; Čermák and Nadezhdina, 2000).

In the present study, the heat-pulse velocity method was used to characterize the temporal variability of transpiration from a Qinghai spruce forest in the upper Heihe River Basin of arid northwestern China, as the method can obtain reliable estimates of long-term forest stand transpiration. Estimates of whole-tree transpiration tend to differ by less than 15% between the heat-pulse velocity method sensors and cut-tree (potometer) measurements (Olbrich, 1991; Smith, 1991; Barret et al., 1995; Hatton et al., 1995).

The objectives of this study were to: (a) improve the understanding of the interannual variation of transpiration from Qinghai spruce canopies; (b) partition the water balance of the forest by determining all of its components at the stand level; and (c) provide greater insight into the links between precipitation variability and forest water yields in montane forested catchments of the arid zone of northwestern China.

2. Materials and methods

2.1. Experimental site

This study was carried out on Qilianshan Mountain, located in the upper Heihe River Basin, in pure stands of Qinghai spruce (*Picea crassifolia*) located on a bench within a north-facing slope at 2800 m elevation in the Pailougou watershed (100°17′E, 38°24′N) (Fig. 1), 50 km south of Zhangye, Gansu province, during the 2011–2013 growing seasons. From 1994 to 2010, the mean annual air temperature was 0.5 °C, and the mean maximum and minimum temperatures were 28.0 °C and −36.0 °C, respectively. Annual precipitation was between 290.2 and 467.8 mm, and the pan evaporation was 1051.7 mm. The frost-free period was about 165 days per year.

The forest in the north slope of watershed, 50% coverage of the forest canopy, consists mainly of Qinghai spruce ranging from 80 to 120 years old. The leaf area index (LAI) of the selected stand was 1.84 (measured with a Digital Plant Canopy Imager (CI – 110, CID, Inc., Washington, U.S.A.)). The stand density was 1100 trees ha^{-1} and the frequency distribution of tree diameters at breast height (D_{BH}, mm) is shown in Fig. 2. Tree height ranged from 5.0 m to 16.5 m; the average was 11.8 ± 2.8 m. Diameter at breast height ranged from 80 mm to 330 m; the average was 18.2 ± 6.5 cm. Moss (*Abietinella abietina*) covered the forest floor. The moss was 10 cm to 20 cm thick and at about 95% coverage. There were no Qinghai spruce seedlings or

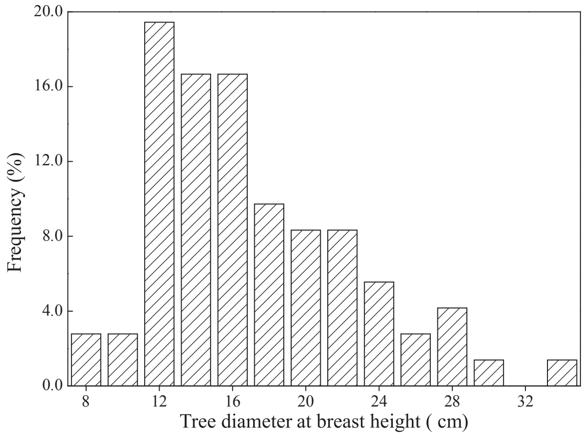

Fig. 2. Frequency distribution of Qinghai spruce tree diameters at breast height.

herbaceous vegetation on the forest floor. The soil was gray-drab with a field water capacity of 53.8%, a total porosity of 71.4%, a bulk density of 850 kg m^{-3}, and a soil depth of 0.7 m. A 20 m by 35 m (700 m^2) plot within the northern slopes(the shady slopes) was selected for the studies.

2.2. Meteorology measurements

Meteorological variables were measured from two weather stations, one positioned in the forest, and another in an open site at 100 m distance from the forest boundary. The global short-wave radiation (R, W m^{-2}) was measured in the open, 100 m distant from the forest boundary, with a pyranometer (CM7B, Kipp and Zonen, Delft, Netherlands). In the forest area, net downward radiation (Rn, W m^{-2}), air temperature (T, °C), humidity, and wind speed were measured. Net downward radiation was measured with a radiation balance sensor (Net radiometer 8110, Philipp Schenk, Wien, Austria) at the height of 1.6 m above ground. A capacitive relative humidity sensor (HMP35C, Campbell Scientific, Inc., Logan, Utah), which was installed in a shield, monitored relative humidity and air temperature at the height of 1.6 m above ground. Wind speed and wind direction were measured at the height of three meters with a Rotronic sensor (RS2 rotronic AG, Bassersdrof, Switzerland). These measurements were taken every 5 min, and the mean 30-min values were stored in a datalogger. The precipitation was manually recorded twice a day (at 08:00 and 20:00 local time) at a meteorological station in an open site 100 m distant from the forest boundary at 2750 m elevation above sea level.

2.3. Stand water balance computations

For the Qinghai spruce forest stand, the sum of inputs and outputs about the water balance of an area over a period of time is following:

$$P - R - E_C - E_I - E_s - \Delta S = 0 \tag{1}$$

where P is precipitation, I is irrigation, R is surface runoff, E_C is canopy transpiration, E_I was the canopy evaporation originating from precipitation intercepted at a particular time, E_s is the soil and moss evapotranspiration, and ΔS is drainage and water storage in the soil (all measurements in mm). Minor water-balance components that are routinely neglected include changes in plant storage, lateral flow, and capillary rise. Based on the long-term monitoring (since the 1970s) of surface runoff by Gansu Qilian Mountain forest ecosystems research stations, surface runoff was omitted because there is usually no runoff in the study area in this study. E_C, E_I and E_s was estimated as following.

2.4. Measurements of sap flow and transpiration

Eight Qinghai spruce sample trees were selected with differing diameters at breast height ranging from 127 mm to 320 mm (Table 1), a range that covered more than 91% of all the trees in the Qinghai spruce forest, according to our vegetation surveys.

We used two sets of heat pulse meters (SF-300, Greenspan technology Pty Ltd., Warwick, Queensland, Australia), with eight probes, to measure sap flows of individual Qinghai spruce trees. To install the probes, the bark of the sample tree was removed until the cambium was exposed. The heat-pulse velocity probes were implanted at a depth of 25 mm into the xylem of each sample tree (Chang et al., 2014a). The wound diameter was 2.2 mm. Before insertion, each probe was coated with silicone gel to ensure good thermal contact between the probe elements and the sapwood. After insertion, the exposed cambium was covered with silicon gel to reduce evaporation from the wood surface, and then covered again with aluminum foil to reduce the effects of ambient temperature fluctuations and solar radiation. Eight probes altogether were set firmly in eight sample trees during the 2011–2013 growing seasons.

Each probe utilized two sensor probes and a heat probe. The upstream sensor probe was located 5 mm below the heat probe; the downstream sensor probe, 10 mm above the heat probe. On each sensor probe were two thermistors. The first was positioned 5 mm from the end of the probe; the second, 5 mm behind the first. The thermistors were paired on a vertical plane to facilitate the measurement of sap flow velocity. Heat pulses lasting 1.6 s were produced by the heat probe. Sap flow velocity (V_s, mm h^{-1}) was calculated following the method of Edwards and Warwick (1984):

$$V_s = V'_h(0.505F_m + F_i) \tag{2}$$

where V'_h (mm h^{-1}) is the heat pulse velocity; F_m is the volume fraction of the woody material; and F_i is the volume fraction of water.

Heat pulse velocity is calculated from the heat pulse time. At this point the heat pulse has moved to a midpoint between the probe and the formula for heat pulse velocity(V'_h) is:

$$V' = \frac{x_1 + x_2}{2t_0} \tag{3}$$

Table 1
Biometric and physiological parameters of sap flow measurements.

No.	Diameter at breast height (mm)	Height (m)	Crown Width (m)	Sapwood radius (mm)	Bark depth (mm)	Heartwood radius (mm)	Sapwood area (mm^2)
1	142.0	13.5	3.3	32	7	32	9566
2	205.0	14.5	3.9	37	7	59	17,882
3	127.0	8	3.6	30	7	26	7892
4	164.0	12.5	4.4	33	7	42	12,224
5	211.0	12.5	4.5	38	7	61	18,792
6	231.0	14.4	4.6	39	7	69	21,992
7	181.0	12.5	3.2	35	7	49	14,454
8	276.0	15.8	3.3	44	7	87	30,186
9	320.0	16.2	4.8	40	7	113	33,410

Where x_1 and x_2 represent the distances of the upstream and downstream sensors respectively from the heater, respectively. A negative value is assigned to x_1 because of its position upstream from the heater probe. t_0 is the heat pulse time. Following the release of the heat pulse from the heater probe, the closer upstream ptobe(p_1) is warmed up with diffusion. Warming of p_1 by convetion can not occur because sapflow direction is from p1 to the heater probe. The upstream probe is warmed above the baseline temperature which both sensor probes are in equilibrium (as at T_1), and there is a positive temperature difference between the upstream probe and downstream probe(p_2). The temperature difference (p_1-p_2) reached a peak when the pulse moved downstream, with diffusion and any convect, and then declined because p_2 is warmed up. The temperature difference returned to zero at T_2, where the system has returned to a balance. t_0 is the time to T_2, or the heat pulse time (Closs, 1958).

The calculation of F_m and F_i for each tree required the following inputs: fresh weight (W_f, kg), oven-dried weight (W_d, kg), and the weight of water in the same volume as the sapwood sample (W_i, kg). On this basis, F_m is:

$$F_m = \frac{W_d}{1620W_i} \tag{4}$$

and F_i is:

$$F_i = \frac{W_f - W_d}{W_i} \tag{5}$$

Each sapwood sample was selected for five replications. First, the W_f was measured, then the sapwood sample was completely immersed in distilled water, into which the metering cylinder was placed. The volumes before and after immersion were compared, and the weight of water of the same volume as the sapwood sample (W_i), was calculated. Finally, the sapwood sample was oven-dried at 80 °C, and W_d, F_m and F_i were then calculated for the sample. The averages of F_m and F_i were used to calculate the V_s value. The average values of F_m and F_i were 0.36 and 0.27, respectively.

Sap flux (Q, $mm^3\,h^{-1}$) is a function of the velocity of sap flow and the area of conducting wood in which the flow occurs:

$$Q = V_s A_c \tag{6}$$

where A_c (mm^2) is the area of conducting wood (Closs, 1958).

The daily cumulative sap flow (Q_s, $kg\,d^{-1}$) in a tree is essentially equal to the daily sums of transpiration for time periods of one day or longer (Čermák et al., 1995). The dynamic response of the measured sap flow to atmospheric forcing occurred virtually immediately. Q was so tightly coupled to the climatic variables that no time shift was needed to fit a simple static microclimatic model to estimate Q (Cienciala et al., 2000).

Tree transpiration (E_t, $kg\,m^{-2}\,d^{-1}$) was expressed as sap flux on the sapwood area at breast height. In order to calculate sap flux ($kg\,m^{-2}\,h^{-1}$) for a given tree, V_s (Eq. (1)) was divided by the sapwood area of the tree.

Tree-level sap flow was scaled up to the stand level to calculate stand transpiration (E_s, kg) based on the trees' estimated sapwood areas. We assumed that the water velocity on a given date was the same per unit of sap flow area (i.e., that velocity did not vary within the sapwood in different trees (Wullschleger et al., 2001)). Therefore, E_s was estimated as:

$$E_s = \sum_{i=1}^{n} E_{si} = \sum_{i=1}^{n} E_{sti}\frac{A_{T,sw}}{A_{S,sw}} \tag{7}$$

where E_{si} (kg) represents the stand transpiration on day i; E_{sti} ($kg\,d^{-1}$) represents the sample tree transpiration on day i; $A_{T,sw}$ (mm^2) is the total stand's sapwood area; and $A_{S,sw}$ (mm^2) is a given sample tree's sapwood area.

The sapwood area was determined with the staining method. Because sapwood in Qinghai spruce cannot be visually distinguished from heartwood, a Fehling's Solution dye method was used to aid in the determination of sapwood depth (Kutscha and Sachs, 1962). An increment bore was used to drill each of Qinghai spruce tree selected with differing diameters according to our vegetation surveys in June 2011, and 5-mm increment cores were collected from each of the 17 trees. Fehling's Solution was carefully dripped onto every 5-mm increment core with a dropper. The difference in color, using this method, is clearly discernable; sapwood appears darker and wetter than heartwood. Every sapwood depth was then measured with a ruler, the sapwood area calculated and the statistical model established between the sapwood area and the tree diameter at breast height. The statistical model was established between sapwood area (A_{sw}, mm^2) and tree diameter at breast height (D_{BH}, mm) as:

$$A_{sw} = 12655.61 \cdot e^{\frac{D_{BH}}{226.72}} - 13306.78 (R^2 = 0.99,\ n = 17,\ p < 0.001) \tag{8}$$

We used the statistical model for sapwood area to extrapolate the values for all trees in the sample plot and to determine the sapwood area of the sample trees, too.

Canopy transpiration (E_C, $mm\,d^{-1}$) equals E_s weighted by the canopy's projected area. Crown projection areas were estimated from below the crown by sighting vertically at various positions around each tree. We measured the distances between the stem and the outermost projected point of the branches in all directions, and then drew the proportional lengths on standard cross-section paper to estimate the crowns' projection areas.

2.5. Measurements of canopy evaporation originating from intercepted precipitation

In theory, the canopy water storage capacity (S_c, mm) of a forest stand equals the maximum canopy evaporation originating from precipitation intercepted at a particular time (E_I, mm). That is, when the rainfall is equal to or greater than S_c, E_I is equal to S_c; and when the rainfall is less than S_c, E_I is equal to the total precipitation for the day. Therefore, E_I can be obtained as long as S_c can be estimated.

In this study, a simple method was used for estimating S_c using sap flow measurements; the crucial factor was the estimation of canopy drying time (t_{cd}) (or wet canopy duration) after rainfall. This time was estimated using sap flow measurements taken from Kume et al. (2006, 2008), which were conducted in a Bornean tropical rainforest. When the sap flow velocity after rainfall becomes equal to the value of the sap flow velocity assumed for bright (no-rainfall) days, it can be assumed that the effect of leaf wetness on transpiration has been removed and that the canopy is completely dry. After rainfall, the sap flow velocity is lower than on a bright day. It is assumed to be due to the canopy wetness at this site where cloudy weather after rainfall tends to abruptly change to sunny weather (Kume et al., 2006, 2008). This approach is a robust technique that can be conducted cheaply and interfaced readily through data loggers for long-term monitoring. In this study, three criteria were used to select target rain events to derive t_{cd}: (1) rainfall ceased during the daytime hours of 06:00 to 17:00; (2) the period of no rainfall between each rain event exceeded 2 h; and (3) total rainfall for each rain event exceeded a threshold of 1.5 mm.

In order for this method to work, it was essential to focus on rain events in which the canopy was fully wet at the time rainfall ceased. To address the concern that sap flow velocity may vary markedly throughout the day, the threshold of 1.5 mm of rainfall was selected because it would provide enough rainwater to guarantee a fully wet canopy at the time rainfall ceased (e.g., Lloyd et al., 1988; Schellekens et al., 2000). Furthermore, t_{cd} was defined as the period from the time rain ceased (A) to a specific time S in each rain event, where S is the time when sap flow velocity after rainfall exceeds the average minus the standard deviation of the sap flow velocity averaged over 20 bright days. This study assumed that a wet canopy had dried out by time S: hence, t_{cd} = S-A. Kume et al. (2006) also confirmed that almost the

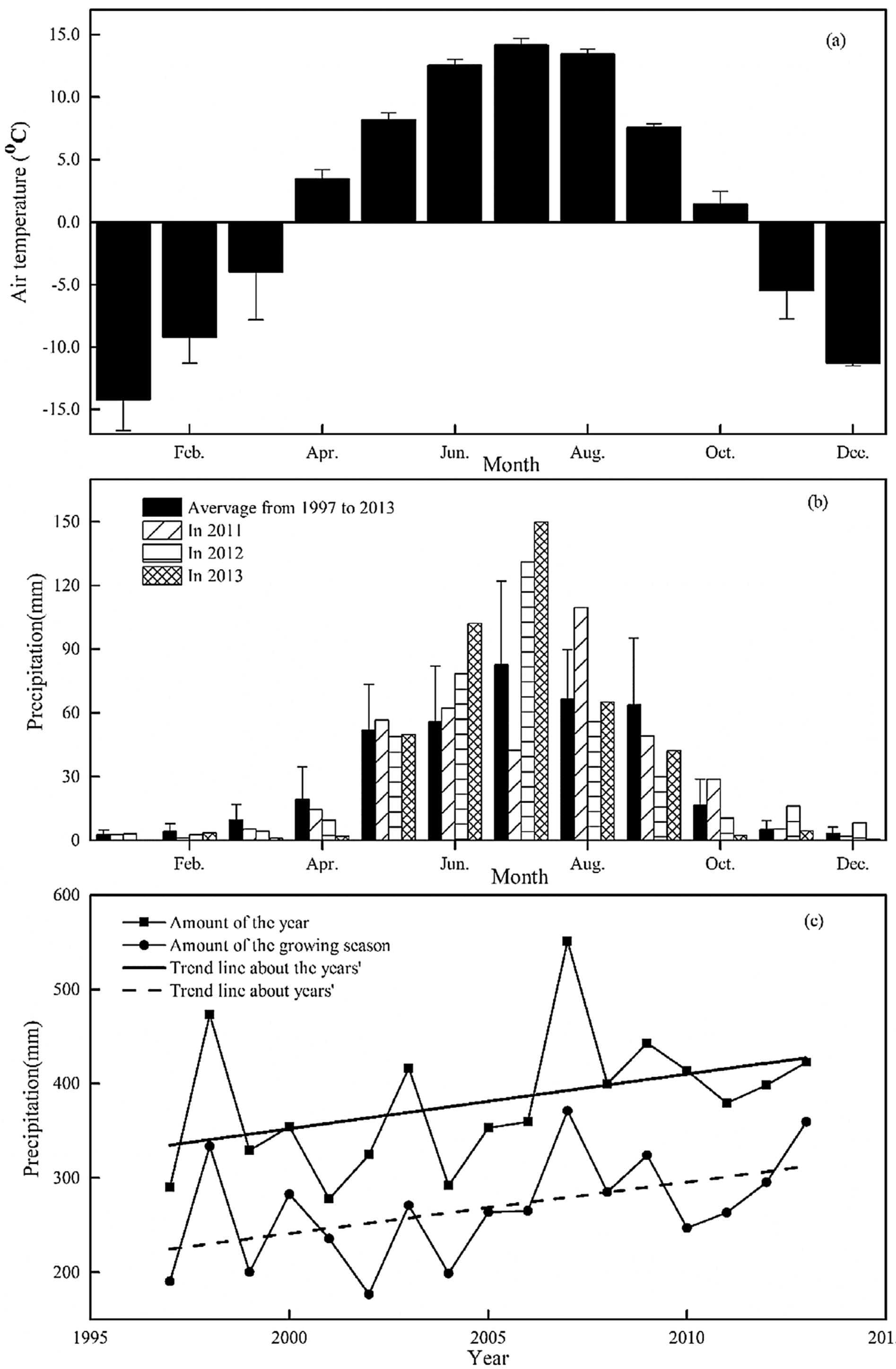

Fig. 3. Variations in average air temperature and precipitation at the Pailougou watershed on Qilian Mountain: (a) average air temperature during the study period; (b) precipitation during the study period; (c) annual and growing-season amounts of precipitation from 1997 to 2013.

same results were found using other definitions of S.

Canopy drying time after rainfall is the period needed to evaporate the water stored on leaves after rainfall. When we assume that evaporative energy is completely consumed to evaporate water stored on leaves during the period of canopy drying time after rainfall and that transpiration is negligible during that period, canopy drying time after rainfall can be expressed as follows:

$$S_c = \overline{E_i} \times t_{cd} \tag{9}$$

In which (Kume et al., 2008)

$$\overline{E_i} = \frac{\Delta(\overline{R_n} - G) + \rho \times C_p \times \overline{D_{VP}}/R_a}{(\Delta + \gamma)\lambda} \tag{10}$$

where t_{cd} is the calculated canopy drying time after rainfall (min); S_c is the canopy water storage at the time rainfall ceases (mm); $\overline{E_i}$ is the mean evaporation rate from a wet canopy after rainfall (mm/min); Δ is the slope of the saturation vapor pressure function (hPa K^{-1}); Rn is the mean net radiation during the canopy drying period (W m^{-2}); G is the soil heat flux (W m^{-2}); ρ is the air density (kg m^{-3}); C_p is the specific heat of the air (J kg^{-1} K^{-1}); D_{VP} is the mean vapor pressure deficit of the air during the canopy drying period (hPa); Ra is the aerodynamic resistance for heat and water from vegetation (s m^{-1}); γ is the

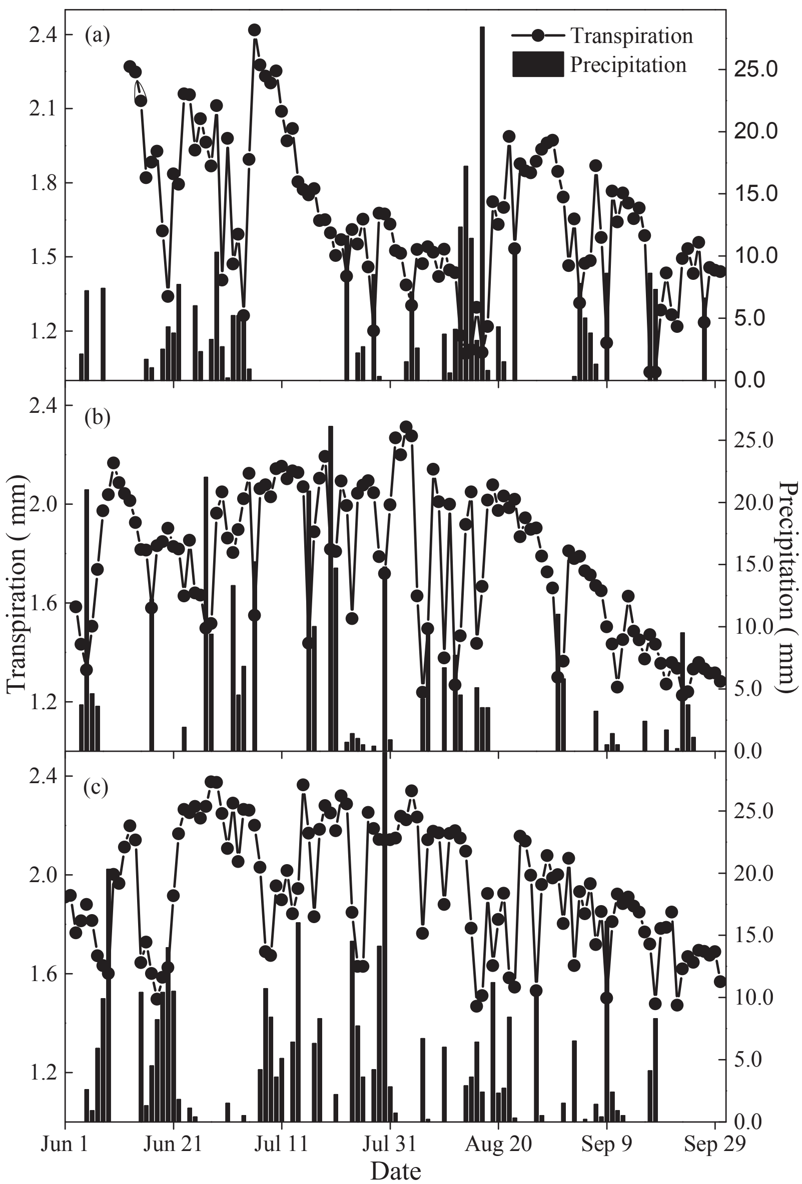

Fig. 4. Daily stand transpiration and precipitation during the 2011(a), 2012(b) and 2013(c) growing seasons.

psychrometric constant (hPa K^{-1}); and λ is the latent heat of water vaporization (J g^{-1}). G was negligible at this study site (Sato et al., 2004; Li et al., 2017). S_c was equivalent to the canopy water storage capacity (i.e., the maximum possible water storage after quick drainage has ceased) in an upper canopy in this study, because the rain events in this study were selected so that the canopy would be fully wet at the time rainfall ceased, under the three criteria. $\overline{R_n}$ and $\overline{D_{VP}}$ were the mean Rn and D_{VP} in the period from time A to time S. Aerodynamic resistance (R_a) was calculated from wind speed using the following equation (Schellekens et al., 2000):

$$R_a = \frac{\ln\left[{}^{(z-d)}/_{z_0}\right]^2}{k^2 W} \tag{11}$$

where W (m s^{-1}) is the wind speed and z (m) is the wind measurement height. The zero-plane displacement, d (m), and the roughness length, z_0 (m), were taken as 0.75 and 0.10 of the tree height (4.6 m) (Cienciala et al., 2000), respectively; k (=0.4) is the von Karman constant.

2.6. Measurements of soil and moss evaportranspiration

Daily soil and moss evaportranspiration was measured using five weighing mini-lysimeters (Shawcroft and Gardner, 1983; Walker, 1984; Boast and Robertson, 1982). The mini-lysimeter, which was made of rustless iron, consisted of two main parts: (i) the mini-lysimeter itself: a cylinder with a 29.8-cm diameter and 30.0-cm depth, with a pipe centered at the bottom, 5.0 cm long with a 1.0-cm diameter, for draining off the leakage; and (ii) a water-tight cylindrical tank, 30.0 cm deep and 30.0 cm in diameter, in which the mini-lysimeter was placed. Centered at the bottom of this tank was a small cylindrical container, 15.0 cm deep and 15.0 cm in diameter, which caught the slow leakage (via the pipe) from the mini-lysmeter. The mini-lysimeters were installed at three points, equal distances apart, along two diagonals of a 20-m by 35-m (700 m^2) area along the northern slope that had been selected for the sap flow studies. In order to accurately represent the field's soil, each mini-lysimeter was filled in with an original soil monolith cut from nearby the mini-lysimeter location. In order to minimize the influence of the equipment on micro-meteorological elements, the soil surface inside the lysimeter was kept flush with the surrounding field surface and to reduce structure size of underground compartment for weighing and measuring systems. The mini-lysimeters were manually weighed daily at 19:00 local time during the growing season, to determine water loss, using electronic balances with 1.0-g precision, sufficient for monitoring the day to day variations of the evapotranspiration. The leakage in the small cylindrical container from the bottom of the outside tank, when there was any, was also measured.

3. Results

3.1. Precipitation

The variations in annual and growing-season amounts of precipitation from 1997 to 2013 are shown in Fig. 3. The average monthly precipitation varied between 2.6 ± 2.3 mm and 82.6 ± 39.5 mm. The minimum monthly precipitation was observed in January and the maximum was recorded in July. During the study period, the monthly average air temperature vaired between −14.23 ± 2.49 °C (in January) and 14.17 ± 0.55 °C(in July) (Fig. 3a), and the monthly precipitation varied from 1.1 mm to 109.5 mm in 2011, from 2.7 mm to 131.2 mm in 2012, and from 0.3 mm to 149.8 mm in 2013 (Fig. 3b). From 1997 to 2013, the annual amounts of precipitations varied from a low of 277.7 mm in 2001 to a high of 550.9 mm in 2007, and the growing-season precipitation amounts ranged from a low of 176.9 mm in 2002 to a high of 371.2 mm in 2007. All these amounts showed a generally increasing trend (Fig. 3c). During the study period, the annual precipitation amounts in 2011–2013 were −0.5%, 4.6% and 11.0% higher, respectively, than the annual average (381.0 mm) from 1997 to 2013; and the growing-season precipitation amounts in 2011–2013 were −2.0%, 10.1%, and 33.8% higher, respectively, than the growing-season average from 1997 to 2013 (268.5 mm). From 1997 to 2010, about 69% of the total annual rainfall was observed during the growing season; however, this proportion rose to 69.4%, 74.2%, and 85.0% in 2011–2013, respectively. Total precipitation during the 2012 and 2013 growing seasons was 12.3% and 36.5% higher, respectively, than during the 2011 growing season.

3.2. Canopy transpiration

During the 2011–2013 growing seasons, the calculated stand transpiration ranged from 1.0 mm d^{-1} to 2.4 mm d^{-1}, from 1.2 mm d^{-1} to 2.4 mm d^{-1}, and from 1.5 mm d^{-1} to 2.4 mm d^{-1}, respectively, reaching averages of about 1.6 mm d^{-1}, 1.8 mm d^{-1}, and 1.9 mm d^{-1} (Fig. 4). The rain had a strong influence on canopy transpiration. After precipitation of 7.7 mm on 22 June 2011, the canopy transpiration rose from 1.8 mm d^{-1} to 2.2 mm d^{-1}, and after continuous rain from 28 June to 5 July and in the middle of August 2011, with cumulative precipitations of 33.4 mm and 87.5 mm, the canopy transpirations reached 2.4 mm d^{-1} and 2.0 mm d^{-1}, respectively (Fig. 4a). After continuous rain from 4 to 7 June 2012, the canopy transpiration increased from 1.5 mm d^{-1} to 2.0 mm d^{-1}; while during frequent rainfall in the middle of July, the canopy transpiration remained at 2.0 mm d^{-1} (Fig. 4b). The same result occurred during the 2013 growing season (Fig. 4c). The total stand transpiration was 195.2 mm, 219.6 mm and 2237.6 mm in 2011–2013 growing season, respectively. Total *E*c during the 2012 and 2013 growing seasons was 12.5% and 21.7% higher, respectively, than during the 2011 growing season.

3.3. Canopy evaporation originating from intercepted precipitation

Fig. 5 shows typical diurnal courses of sap flow velocity for rainfall events, and the average diurnal course of sap flow velocity for bright days. On 6 August 2013, total rainfall was 6.9 mm between 5:40 and 15:20. After the rainfall, sap flow increased slowly compared with the average diurnal course for bright days. At 16:00, diurnal courses of sap flow velocity for rainfall events met the average diurnal course of sap flow velocity for bright days, and then the moist Qinghai spruce forest canopy spent about 40 min drying out (Fig. 5). Hence the canopy water storage capacity (S_c) was estimated based on 40 min of canopy drying time after rainfall, and was calculated as 0.7 mm. The precipitation was therefore calculated based on this standard of 0.7 mm: if the precipitation was greater than or equal to 0.7 mm, the Qinghai spruce forest canopy water storage was estimated as 0.7 mm; if the precipitation was lower than 0.7 mm, the canopy water storage was estimated based on actual precipitation. So, during the 2011 growing season, there were 24 precipitation events, and the Qinghai spruce forest canopy evaporation originating from intercepted precipitation was calculated as 15.5 mm; during the 2012 growing season, there were 20 precipitation events, and the canopy evaporation originating from intercepted precipitation was calculated as 13.8 mm; during the 2013 growing season, there were 29 precipitation events, and the canopy evaporation originating from intercepted precipitation was calculated as 19.1 mm (Table 2).

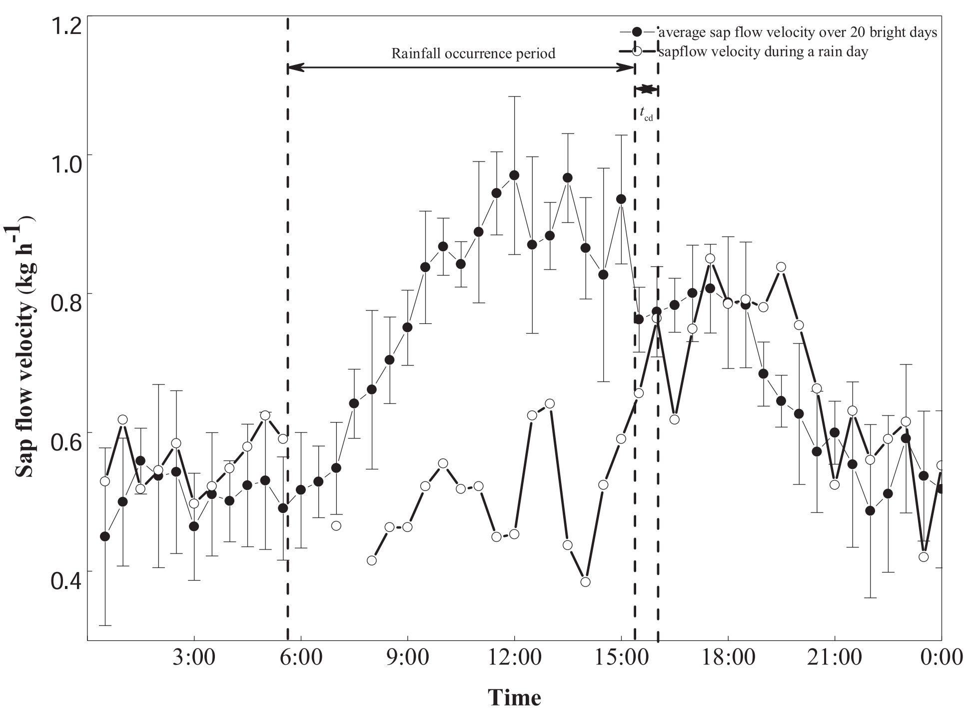

Fig. 5. Definition of estimated canopy drying time (t_{CD}). Line with white circles represents diurnal courses of sap flow velocity during a rainy day, and line with black circles shows averages over 20 bright days. Vertical bars represent the standard deviations about the mean. t_{CD} is defined in the text.

Table 2
Qinghai spruce forest canopy water storage from intercepted precipitation during the 2011–2013 growing seasons.

Year	Precipitation greater than or equal to 0.7 mm		Precipitation lower than 0.7 mm		Number of precipitation events	Canopy water storage (mm)
	Occurrences	Amount (mm)	Occurrences	amount (mm)		
2011	21	14.7	3	0.8	24	15.5
2012	19	13.3	1	0.5	20	13.8
2013	26	18.2	3	0.9	29	19.1

3.4. Soil and moss evaportranspiration

During the 2011–2013 growing seasons, observed soil and moss evaportranspiration averaged about 0.5 mm d^{-1}, 0.5 mm d^{-1}, and 0.6 mm d^{-1}, respectively, with a maximum of about 1.0 mm d^{-1}. There remained some differences in monthly and annual soil and moss evaportranspiration for the Qinghai spruce forest during the growing seasons of these three years (Fig. 6). During the 2011 growing season, the maximum monthly stand soil and moss evaportranspiration was 21.6 mm (in June), the minimum 10.7 mm (in September), and the total stand soil and moss evaportranspiration, 67.8 mm. During the 2012 growing season, the maximum monthly stand soil and moss evaportranspiration appeared in August (18.3 mm), and the minimum in September (11.8 mm), while the total stand soil and moss evaportranspiration was 65.8 mm. During the 2013 growing season, the maximum monthly stand soil and moss evaportranspiration was 20.4 mm (in August), the minimum 15.9 mm (in September), and the total stand soil and moss evaportranspiration, 70.1 mm.

3.5. Components of the hydrological budget

The values of each component of the hydrological budget of the Qinghai spruce forest during the growing seasons are specified in Table 3 and shown in Fig. 7. During the 3-year research period, the average precipitation was 306.1 ± 48.8 mm, the average transpiration 217.5 ± 21.3 mm, the average canopy evaporation 16.1 ± 2.7 mm, the average soil and moss evaportranspiration 61.0 ± 12.2 mm, and the drainage and change in soil storage 11.5 ± 15.6 mm. During the study period, transpiration, which presented the largest flux in the water balance, accounted for 71.1% of the precipitation. In this forest ecosystem, soil and moss evaportranspiration was a significant part of the water budget, accounting for approximately 19.9% of the precipitation, while canopy evaporation accounted for 5.3% of the precipitation. The drainage and water storage in soil made up the smallest proportion of the precipitation: only 3.8%.

In theory, when soil moisture is higher than the water capacity of a field, the water in the soil is likely to infiltrate through the soil and run down the slope, through rocky nooks and crannies, and eventually become river runoff. Thus, the soil moisture was higher than the field water capacity, 53.8% in this studied area, and i.e., the total soil water storage capacity in the forest stand was 32.0 mm, the water in the soil is likely to infiltrate. However, during the 2011 and 2012 growing seasons, the drainage and change in soil water storage was only 4.0 mm, too low to do anything other than simply improve the soil moisture. During 2013 growing season, the drainage and change in soil water storage was 29.4 mm, very close to the total soil water storage of 32.0 mm, but there was still no excess water for infiltration and runoff. Hence, although the precipitation was higher in 2013, the increase was not enough to increase the amount of forest water yield, during this study period.

4. Discussion

4.1. Precipitation changes in arid and semi-arid regions

Climate change is causing measurable changes in rainfall in many

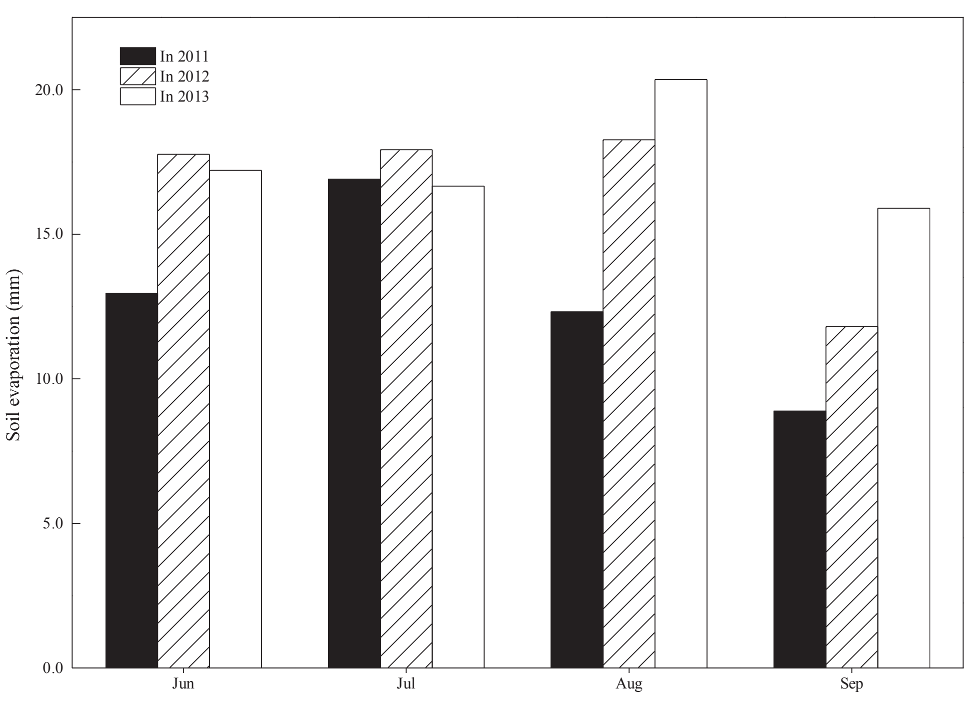

Fig. 6. Monthly soil evaporation for Qinghai spruce forest during the 2011–2013 growing seasons.

Table 3
Components of the hydrological budget of the Qinghai spruce forest during the 2011, 2012 and 2013 growing seasons.

Year	Precipitation (mm)	Transpiration (mm)	Soil and moss evaportranspiration (mm)	Canopy evaporation (mm)	Drainage and water storage in soil (mm)
2011	263.5	195.2	48.8	15.5	4.0
2012	295.5	219.6	61.0	13.8	1.1
2013	359.3	237.6	73.2	19.1	29.4
Average	306.1 ± 48.8	217.5 ± 21.3	61.0 ± 12.2	16.1 ± 2.7	11.5 ± 15.6
Proportion of precipitation (%)	100.0	71.0	19.9	5.3	3.8

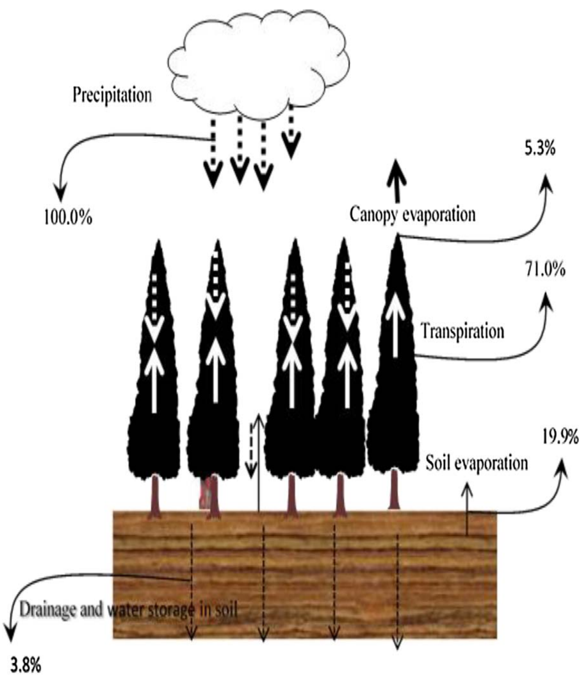

Fig. 7. Components of the hydrological cycle in the Qinghai spruce forest. Values are percentages of average precipitation during the 2011–2013 growing seasons.

regions, such as Mediterranean, Central Asia, and Northwest China. In recent years, there has been a general consensus among the scientific community concerning the variability and trends of precipitation and their effects on the environment (Karl and Knight, 1998; Folland and Karl, 2001; Zhang et al., 2001; Philandras et al., 2011). In general, total annual changes in precipitation have been variable, with some regions showing an increase and others a decrease or no change. Across the U.S., the mean precipitation has increased overall since 1900, and while some areas have increased at a higher rate than the national average, other areas have decreased (Groisman et al., 2004; Meehl et al., 2005; Anderson et al., 2015). Over the entire Mediterranean area, precipitation has decreased by as much as 30% (IPCC, 2007; Zhang et al., 2007). In Central Asia, one of the largest arid regions at the middle latitudes in the world, research by Xu et al. (2015) has indicated that annual precipitation has displayed an increasing trend, with an average increase of 3.9 mm per 10 year from 1950 to 2000. Arid Northwest China, which is characterized by hot, dry, and sunny summers, is basically a climatic intersection between the westerlies and the East Asian summer monsoons. Here there has been a significant inter-annual change in precipitation: the annual precipitation has increased significantly, by 17.4% (Xu et al., 2015). Specifically, Zhang et al. (2015) reported that annual precipitation showed a significant increasing trend in the upstream part of the Heihe River, while no such trend was detected for the middle and lower reaches of the river. This result is consistent with our results here. In the study area, both the annual and the growing-season amounts of precipitation were on the increase from 1997 to 2013.

4.2. Stand-level forest water use and its response to variations in precipitation

Stand-level forest water use is often estimated by indirect methods: catchment water balance, the eddy covariance technique, or, more simply, the Bowen ratio energy balance (Herbst, 1995; Ford et al., 2007; Sun et al., 2010). These estimation methods represent an integration of the main components of stand evapotranspiration, including tree transpiration, soil evaporation, and canopy interception. In addition, individual tree size and species play a dominant role in determining stand water balance (Granier, 1987; Meinzer et al., 2001). Therefore, a more accurate and reliable estimate of tree water use can be obtained using direct measurements of tree sap flow (Wullschleger et al., 1998). These measurements are very useful for examining the effect of changes in forest structure on stand water balance and for providing information on the physiological regulation of transpiration. Forest ecologists can utilize those estimates of tree water uptake derived from sap flow to evaluate the role of transpiration in forest hydrology (Barrett et al., 1996; Ewers et al., 2002), and to address issues of water resource management (Schiller and Cohen, 1995; Oishi et al., 2008). For the Qinghai spruce, the sap flow velocity was measured with the heat pulse method, and was overestimated by about 5.6% (Chang et al., 2014a).

Forest water use is significantly affected by tree physiology (Zotz et al., 1998; Becker et al., 1999; Field and Holbrook, 2000; Ocheltree et al., 2014), stand density (Alsheimer et al., 1998), stand age (Alsheimer et al., 1998; Köstner et al., 2002; Irvine et al., 2002; Ewers et al., 2005; Clausnitzer et al., 2011), and soil moisture. The natural variations in precipitation are vital in terms of their effects on the soil moisture (Trenberth, 2011), because soil water content determines forest water use. A number of studies have reported an observed decrease in canopy transpiration as a result of declines in soil moisture, for a variety of species (Pataki et al., 2000; Lagergren and Lindroth, 2002; Chang et al., 2006; Tromp-van Meerveld and McDonnell, 2006; Gartner et al., 2009; Llorens et al., 2010; Chang et al., 2014a). Pataki et al. (2000) reported an observed decrease in maximum sap flow for *Pinus contorta, Abies lasiocarpa, Populus tremuloides*, and *Pinus flexilis* later in the seasonal soil drought of 1996, when soil moisture declined from 0.35 to 0.24 $m^3 m^{-3}$ at 0–45 cm, in the Medicine Bow Mountains of southeastern Wyoming, U.S.A. Lagergren and Lindroth (2002) reported that Scots pine (*Pinus sylvestris* L.) and Norway spruce (*Picea abies* (L.) Karst) transpirations were reduced by 40 and 67%, respectively, during the dry summer of 1999 in the Norunda Forest in central Sweden. Chang et al. (2006) found that the sap flow of a Gansu Poplar (*Populus gansuensis* C. Wang and H. L. Yang) shelter belt was as much as 233 ± 82 $kg\, m^{-2} h^{-1}$ in the middle reaches of the Heihe River Basin in Northwest China after the shelter belt was irrigated. Tromp-van Meerveld and McDonnell (2006) reported that variations in soil depth and total soil water stored in the soil profile caused differences in soil moisture content and transpiration rates between upslope and midslope sections at the end of the wet season in the Panola Mountain Research Watershed, Georgia, U.S.A. Gartner et al. (2009) found that trees responded to drought stress conditions by reducing their transpiration, and that the mean sap flow of all trees was decreasing steadily, from

0.567 kg cm^{-1} to 0.258 kg cm^{-1}. Llorens et al. (2010) reported that transpiration was only 0.6 ± 0.04 and 0.7 ± 0.04 mm day^{-1}, respectively, during the drier summers of 1998 and 2000, a decrease from the wetter summer of 1997. For Qinghai spruce, Chang et al. (2014a) found a logistic functional relationship between sap flow and soil moisture content; the models of sap flow velocity relationship to soil moisture content explained 84% of the variations in sap flow velocity, and showed large sensitivities. In this study, because of the increase in precipitation during the 2011–2013 growing seasons, total *E*c values during the 2012 and 2013 growing seasons were 12.5% and 21.7% higher, respectively, than during the 2011 growing season.

4.3. Canopy water storage capacity

When the forest canopy is saturated by rainfall, most of the rain that falls on the forest canopy will drain to the ground, and only residual rain will remain on the canopy. Because the canopy is saturated at such times, it reaches its maximum water storage capacity, represented as S_c. To estimate S_c, four conventional methods are commonly used: net rainfall measurements on an event basis (Klaassen et al., 1998), scaling-up (Liu, 1998; Llorens and Gallart, 2000), remote sensing (Bouten et al., 1996), and sap flow (Kume et al., 2008). S_c depends on stand characteristics such as species composition and surface area index. Some researchers have reported that S_c ranged between 0.6 and 1.3 mm in tropical rainforests (e.g., Lloyd et al., 1988; Schellekens et al., 2000). Hutchings et al. (1988) quantified *Picea sitchensis* canopy storage capacity as 2.1 and 2.8 mm under windy and calm conditions, respectively, in a Sitka spruce canopy. Teklehaimanot and Jarvis (1991) found *Picea sitchensis* canopy storage capacities were between 0.1 and 1.1 mm. Llorens and Gallart (2000) reviewed 17 studies of canopy storage capacity in different stand densities of coniferous forests; capacity ranged from 0.3 to 3.0 mm. They also studied canopy storage capacity in a heterogeneous 40-year-old *Pinus sylvestris* stand in a Mediterranean mountain area of the Southeastern Pyrenees (Catalonia, Spain), using simplified direct methods; it ranged from 1.24 to 2.65 mm in still air and from 1.16 to 2.47 mm under windy conditions. Liu (1998) measured canopy storage capacities in cypress (*Taxodium ascendens*) wetlands and slash pine (*Pinus elliottii*) uplands in Florida flatwoods; the average was 0.94 mm in the wetlands and 0.43 mm in the uplands. Pypker et al. (2005) estimated the canopy storage capacity as 1.40 ± 0.27 mm in a young (25-year-old) Douglas-fir forest and 3.32 ± 0.35 mm in a nearby old-growth (> 450-year-old) Douglas-fir forest within the Gifford Pinchot National Forest in southern Washington state, U.S.A. Kume et al. (2008) estimated the canopy water storage capacity as 0.7 mm, from sap flow measurements, in a lowland tropical rainforest in the Lambir Hills National Park, Sarawak, Malaysia. Galdos et al. (2012) found that the canopy storage capacity showed great intra-annual variability and also large spatial differences, varying from 0.0 mm to 6.0 mm, using a simple conceptual model and a moderate-resolution imaging spectroradiometer (MODIS), in the North Mountains of Spain. Frischbier and Wagner (2015) found that the average canopy storage capacities for *Fagus sylvatica* L. (in its leaf-bearing period) and *Picea abies* (L.) Karst. forests were 3.5 mm and 5.8 mm, respectively, in the Tharandt Forest in Saxony, Germany, using a new regression approach. In this study, the Qinghai spruce forest canopy water storage was 0.7 mm.

4.4. Forests, climate change, and water yields

A significant proportion of the water supply for human consumption originates from forested catchments (e.g., 53% in the US; Brown et al., 2008). But the direction of climate-change impact on water yield from forests has been debated. Some argue that additional forest cover will reduce water yield, whereas others suggest it will increase water yield by intensifying the hydrological cycle (Ellison et al., 2012). In fact, forests must consume water in order to ensure their own survival, and these forces increase terrestrial interception, evaporation and transpiration. When annual precipitation is higher and annual forest evapotranspiration is lower, annual runoff tends to be higher (Zhang et al., 2001). Forest removal reduces evapotranspiration, elevates soil moisture, raises the groundwater table level, and increases overall watershed discharge (Sun et al., 2001; Brown et al., 2005). And a Brook hydrologic model was used to simulate streamflow response to possible variations in transpiration among species of hardwood trees from small, forested watersheds in the eastern United States, and it was found that a 10% increase in stand transpiration was sufficient to decrease simulated streamflow (Federer and Lash, 1978). Precipitation and forest evapotranspiration, however, are not the only determinants of runoff, and studies that take only these factors into account are not rigorously scientific, nor are studies that are based on too small a scale. After all, the water cycle is intensified by the very existence of a forest. Some researchers have found that broad expanses of forest (presumably significantly greater than 2 km^2) give rise to increased precipitation (Makarieva et al., 2009; Sheil and Murdiyarso, 2009), and a forest's evapotranspiration is one of the principal drivers of precipitation as well: without forests, precipitation will be significantly diminished (Ellison et al., 2012). In these ways, the climate regulatory function of forests has a beneficial impact on the water regime and the availability of water resources. In this study, the precipitation during 2011–2013 growing season increased from 263.5 mm to 359.3 mm, and the evapotranspiration also increased. The homochronous transpiration of Qinghai spruce increased from 195.2 mm to 237.6 mm, and the homochromous soil and moss evaportranspiration ranged from 48.8 mm to 73.2 mm (Table 3). And Qinghai spruce forest's evapotranspiration accounted for 96.2% of the precipitation, consuming a large amount of precipitation. Because stand evaportranspiration increased, the drainage and change in soil water storage was the smallest proportion of the precipitation, accounting for only 3.8%. This small amount left no excess water for infiltration or runoff. So, the increase of rainfall was not sufficient to increase the amount of forest water yield. But this work was carried out on a very small scale: a single forest stand. A similar study on a larger scale might better predict future forest hydrology and climate effects.

Mounting evidence has suggested that climate change has played an important role in controlling the water cycle, by affecting evaporation, transpiration and runoff (McCabe, 2002; Hamlet et al., 2007; Syed et al., 2010; Wang and Hejaz, 2011; Chien et al., 2013; Hegerl et al., 2014; Huntington and Billmire, 2014; McCabe and Wolock, 2014; Sun et al., 2014). However, different forest types have responded differently to climate change (Creed et al., 2014), rates of climate change vary geographically (Walther et al., 2002; Loarie et al., 2009), and forests of different types and ages may influence catchment responses differently (Brown et al., 2005; Ewers et al., 2005). Although over this study period the precipitation increased, the increase was not sufficient to increase the amount of forest water yield.

5. Conclusions

This study examined the response of a Qinghai spruce forest to increases in precipitation in an arid region over the period 2011–2013, and constructed a complete hydrological budget of an arid montane pine forest, by directly measuring its main component at the stand level. It contributed to the observed variability in water-yield responses to climate change, and may help predict changes in water balance partitioning in response to climate change. Although the precipitation increased in this forest, the increase was not sufficient to increase the amount of forest water yield over the study period. Further research into these factors is needed, at larger scales and over longer time periods, to detect and analyze the influences of forest composition, structure, and age on catchment water yields.

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