HYDROLOGICAL PROCESSES
Hydrol. Process. (2011)
Published online in Wiley Online Library
(wileyonlinelibrary.com) DOI: 10.1002/hyp.8400

Effects of rainfall on soil moisture and water movement in a subalpine dark coniferous forest in southwestern China

Qing Xu,[1] Shirong Liu,[1]* Xianchong Wan,[2] Chunqian Jiang,[3] Xianfang Song[4] and Jingxin Wang[5]

[1] *Laboratory of Forest Ecology and Environment of State Forestry Administration, Institute of Forest Ecology, Environment and Protection, Chinese Academy of Forestry, Beijing, China*
[2] *Institute of New Forest Technology, Chinese Academy of Forestry, Beijing, China*
[3] *Institute of Forestry Research, Chinese Academy of Forestry, Beijing, China*
[4] *Institute of Geographic Science and Natural Resources Research, Chinese Academy of Science, Beijing, China*
[5] *Division of Forestry and Natural Resources, West Virginia University, Morgantown, WV, USA*

Abstract:

Water content and movement in soil profile and hydrogen isotope composition (δD) of soil water, rainwater, and groundwater were examined in a subalpine dark coniferous forest in the Wolong National Nature Reserve in Sichuan, China, following rainfall events in 2003–2004. Light rainfall increased water content in the litter and at soil depth of 0–80 cm, but the increased soil water was lost in several days. Heavy rainfall increased soil water content up to 85% at depths of 0–40 cm. Following the light rainfall in early spring, the δD of water from the litter, humus, illuvial, and material layers decreased first and then gradually reached the pre-rainfall level. In summer, light rainfall reached the litter humus, and illuvial layer, but did not hit the material layer. Heavy rainfall affected δD of water in all layers. The δD of soil interflow slightly fluctuated with rainfall events. The δD of shallow groundwater did not differ significantly among all rainfall events. Light rainfall altered the shape of δD profile curve of water in the upper layer of soil, whereas heavy rainfall greatly affected the shape of δD profile curve of water in all soil layers. Following the heavy rainfall, preferential flow initially occurred through macropores, decayed plant roots, and rocks at different depths of soil profile. With continuing rainfall, the litter and surface soil were nearly saturated or fully saturated, and infiltration became homogeneous and plug-like. Forest soil water, particularly in deeper soil profile, was slightly affected by rainfall and, thus, can be a source of water supply for regional needs, particularly during dry seasons. Copyright © 2011 John Wiley & Sons, Ltd.

Supporting information may be found in the online version of this article.

KEY WORDS hydrogen isotopic composition; soil water movement; precipitation; subalpine coniferous forest; Wolong Nature Reserve

Received 8 September 2010; Accepted 27 October 2011

INTRODUCTION

Water movement in unsaturated soils is an inhomogeneous, nonlinear process (Gehrels *et al.*, 1998; Song *et al.*, 2009) and is influenced by climatic and environmental factors, such as precipitation, evaporation, vegetation, and soil properties (Hsieh *et al.*, 1998; Reynolds *et al.*, 2000; Dawson *et al.*, 2002; Lee *et al.*, 2007). In a forest ecosystem, the distribution of soil water is controlled by precipitation and evaporation and is related to air temperature and humidity (Löffler, 2007). Vegetation can serve as a shield intercepting rainfall and may cause desiccation because of evapotranspiration in forest ecosystems. The effect of vegetation on soil water movement strongly depends on vegetation type and density (Gehrels *et al.*, 1998). Soil physical properties have a profound influence on soil water movement. The heterogeneity of soil texture and porosity makes the processes of unsaturated flow in soils complicated, causes changes in the state and content of soil water during flow, and is thus difficult to describe quantitatively (Hillel, 1998).

*Correspondence to: Shirong Liu, Laboratory of Forest Ecology and Environment of State Forestry Administration, Institute of Forest Ecology, Environment and Protection, Chinese Academy of Forestry, Beijing 100091, China.
E-mail: liusr@caf.ac.cn

Precipitation intensity and frequency play an important role in determining soil water movement in terms of infiltration and percolation processes (Lee *et al.*, 2007). Thus, the magnitude, timing, and translocation of precipitation are critical factors influencing the movement and availability of soil water and ecosystem dynamics (Stephenson, 1990; Ferrio *et al.*, 2005). The infiltration and percolation of rainfall can be changed with soil heterogeneity, texture, porosity, and water content (Beven and Germann, 1982; Kung, 1990; Fravolini *et al.*, 2005). For example, at high water content, the downward water movement often is dominated by the flow in soil macropores, whereas at low water content, the flow is maintained in soil micropores (Bengtsson *et al.*, 1987). In rooted zones and below the rooted zones, infiltrating water from rainfall can flow preferentially along highly permeable pathways because of heterogeneous distribution of roots and rocks in forest soils (Gehrels *et al.*, 1998). Considering the interactions among precipitation, vegetation, and soil properties could help understand hydrological processes in forest ecosystems.

Stable hydrogen isotope compositions (δD) of soil water are influenced by atmospheric fractionation and water sources such as rainfall, surface runoff, and upward movement of groundwater (Dansgaard, 1964; Zimmerman *et al.*, 1967; Allison and Hughes, 1983; Hsieh *et al.*, 1998; Reynolds *et al.*, 2000). Besides, water uptake by roots that distribute in different layers of the soil profile also modifies the isotope composition (Ellsworth and Williams, 2007; Xu *et al.*, 2011). Therefore, the hydrogen isotope composition of soil water reveals information on hydrological processes in soil, including infiltration, percolation, and evapotranspiration (Ehleringer and Rundel, 1988; Ferrio *et al.*, 2005). Because δD of precipitation varies seasonally and is event specific, the altered δD of soil water can provide information about translocation, mixing, and residence times of rainwater along soil profile (Gazis and Feng, 2004). By comparing δD values of water in soil profile with other water sources, hydrological processes and roles can be quantitatively analysed and assessed in forest ecosystems (Lin *et al.*, 1996; Williams and Ehleringer, 2000).

The subalpine dark coniferous forest at the Wolong National Nature Reserve is one of the major forest ecosystem types in southwestern China. The forest plays a significant role in mitigating regional floods, protecting headwaters, and conserving water and soil in the upstream of the Yangtze River (Liu, 2002). However, limited information is available on water distribution and movement in the soil profile and the effect of precipitation on soil water dynamics in the forest (Xu *et al.*, 2005, 2006). In this study, we examined the following: (i) how the rainfall intensity affected the water spatial distribution and movement in the soil profile traced by hydrogen isotope composition in comparisons to soil water, rain, and shallow groundwater; and (ii) what relations existed among soil water, soil interflow, rain, and shallow groundwater in the subalpine dark coniferous forest.

METHODS

Study area

This study was conducted in a subalpine dark coniferous forest that is located in Dengsheng Forest Ecosystem Research Station (102°58''21'' E, 30°51'41'' N; 2805 m above sea level) at the Wolong National Nature Reserve, which is situated at the southeastern extension of the Qinghai-Tibet Plateau in southwestern China (Figure 1). The climate is characterized with frigid temperate with cool, humid summers and is influenced by the southern branch of westerlies and the southeastern monsoon. The mean annual precipitation was 884 mm, with the mean annual evaporation of 772 mm, according to its meteorological record of 2001–2003 from the weather station adjacent to the study site. More than 80% of annual precipitation fell between July and September. The lowest mean monthly rainfall occurred during January (6 mm) and the highest during July (193 mm), with an average annual relative humidity of 80%. Mean annual temperature was 4.3 °C with an average of −5.2 °C in December and 20.4 °C in July. Prevailing wind direction was north and northeast.

Soil in the subalpine dark coniferous forest is mountainous brown coniferous forest soil originating from limestone, phyllite, and basalt. The overstory of the forest was dominated by *Abies faxoniana* with *Betula albo-sinensis*, *Betula utilis*, and *Tsuga chinensis* with tree height ranging from 10 to 35 m and canopy coverage of 50–70%. The sub-overstory was dominated by *Bashania fangian* and *Fargesia nitida* with *Lonicera tangutica,*

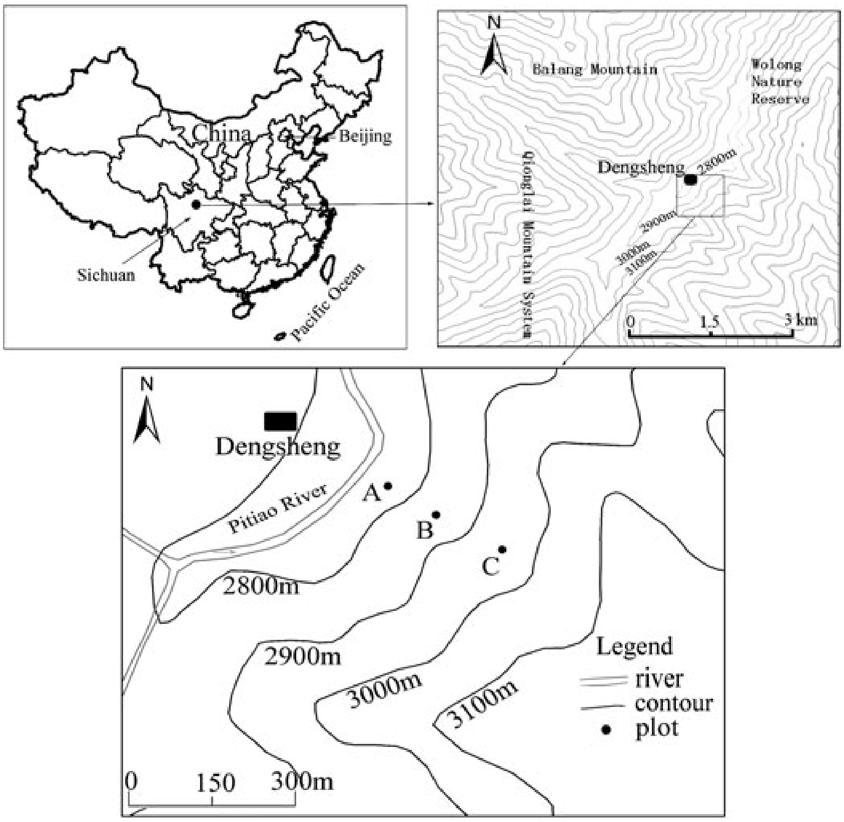

Figure 1. Location of the subalpine dark coniferous forest in the Wolong Nature Reserve of Sichuan, China, showing the study site

DOI: 10.1002/hyp

Hemiphragma wall, Viburnum betulifolium, and *Smilix sians* ranging from 0.5 to 10 m in height and with the coverage of 40–60%. Understory was predominantly grass with species such as *Carex lehmani* and *Cacalia latipes* with the coverage of 30%. Moss layers were dominated by *Hylocomium splendens* with *Abietinella abietina* with the depth ranging from 2 to 6 cm and the coverage of 70%. The thickness of forest floor litter layer was 3–15 cm.

Three 20 × 20-m permanent plots (A, B, and C at 30°51′21″N, 102°58′19″E; 30°51′16″N, 102°58′20″E; and 30°51′20″N, 102°58′22″E, respectively) were established along an elevation gradient with an interval of 100 m in the subalpine dark coniferous forest (Figure 1). At each of the three plots, a representative soil profile was excavated to 80 cm underneath the soil surface (Table I). Bulk density, porosity, field capacity, and retention capacity were measured in each plot (Table II) according to the methods described by Huo *et al.* (2009) and Shi *et al.* (2005).

Field sampling and data collection

Following each of rainfall events, the litter layer and soil samples in the soil profile were collected daily during 8:00–10:00 AM for seven to nine consecutive days from July to September 2003 and in March 2004. The soil sample in the soil profile was sectioned to 0–5, 5–10, 10–20, 20–30, 30–40, 40–50, 50–60, 60–70, and 70–80 cm. All soil samples were then placed in glass bottles, sealed with Parafilm®. The bottles were temporally stored in a portable cooler (0–5 °C) before being brought to the laboratory and stored in the refrigerator at −20 °C until laboratory analysis. Water in the litter and soil samples was extracted using cryogenic vacuum distillation, and then, the extracted water was stored at 0–5 °C until for hydrogen isotope analysis.

After each rain event, at the downhill profile, below the lowest plot (A), two PVC tubes were horizontally installed into a depth of 120 and 60 cm, respectively, and were connected to collecting buckets. The collected water from the two soil depths was mixed together as the interflow sample. The shallow groundwater samples were collected from three wells of approximately 1.5 m deep before and after the rain events. The wells were naturally formed by a ground spring outlet near each of the three plots. The collected samples were mixed before analyzing δD. During the rainfall events, 11 rainwater samples from the rainfall events were collected on open site, which was 100 m away from the study forest. All soil interflow, rainwater, and shallow groundwater samples were collected around 8:00 AM of each sampling day. All water samples were placed in plastic bottles, sealed with Parafilm®, and stored in a portable cooler (0–5 °C) before being brought to the laboratory and stored at −20 °C until hydrogen isotope analysis. A clean, upper portion of all water samples was carefully taken for the isotope analysis.

Data on precipitation, evaporation, and temperature were collected from two weather stations. One was located within the study forest, and the other was on the open site that was 300 m away from the study forest.

Isotope analysis

For all water samples including rain, soil interflow, shallow groundwater, and soil water samples, δD was measured using a Thermo Finnigan MAT Delta ^{Plus} XP isotope ratio mass spectrometer (Thermo Finnigen,

Table I. Characteristics of soil profile in the subalpine dark coniferous forest

Soil layer (cm)	Plot A	Plot B	Plot C
Litter	4–7 cm	5–10 cm	7–15 cm
0–10	Humus horizon, many roots	Humus horizon, many roots	Humus horizon, many roots
10–20	Illuvial horizon, many roots	Illuvial horizon, many roots	Humus and Illuvial, many roots
20–30	Illuvial horizon, many roots	Illuvial horizon, many roots	Illuvial horizon, many roots
30–40	Illuvial horizon, few roots	Illuvial horizon, few roots	Illuvial horizon, few roots
40–50	Parent material horizon, few roots	Parent material horizon, insert some detritus, few roots	Parent material horizon, few roots
50–80	Parent material horizon, insert some detritus, few roots	Parent material horizon, insert some detritus, few roots	Parent material horizon, insert some detritus, few roots

The number in litter row is referred to thickness of the litter layers.

Table II. Soil physical properties (mean ± SD, n = 3) in the subalpine dark coniferous forest

Soil depth (cm)	Bulk density ($g\,cm^{-3}$)	Total porosity (%)	Capillary porosity (%)	Non-capillary porosity (%)	Field capacity (%)	Max retentive capacity (%)
0–10	1.07 ± 0.30	58.50 ± 9.90	42.89 ± 4.55	15.59 ± 6.85	0.54 ± 0.21	0.71 ± 0.24
10–20	1.35 ± 0.42	49.50 ± 13.70	39.31 ± 8.08	12.40 ± 2.21	0.56 ± 0.24	0.73 ± 0.34
20–30	1.21 ± 0.18	54.10 ± 5.80	41.24 ± 3.13	12.86 ± 2.69	0.47 ± 0.11	0.50 ± 0.10
30–40	1.20 ± 0.14	54.30 ± 4.80	42.41 ± 6.40	11.91 ± 10.70	0.44 ± 0.18	0.62 ± 0.08
40–50	1.19 ± 0.10	54.40 ± 3.50	41.96 ± 2.79	12.45 ± 5.39	0.52 ± 0.06	0.62 ± 0.03
50–60	1.18 ± 0.07	54.50 ± 2.20	41.51 ± 2.38	12.99 ± 0.37	0.60 ± 0.06	0.61 ± 0.05

DOI: 10.1002/hyp

San Diego, CA, USA) coupled with a high temperature conversion elemental analyser (TC/EA) at the Stable Isotope Laboratory for Ecological and Environmental Research, the Institute of Botany of the Chinese Academy of Sciences. The precision of δD was ± 2‰ based on the two internal standards calibrated against a Vienna Standard Mean Ocean Water (NIST, Washington, DC). Hydrogen isotopic composition was expressed as the hydrogen isotope ratio:

$$\delta D = \left[R_{\text{sample}}/R_{\text{standard}} - 1\right] \times 1000 \quad ‰$$

where R_{sample} and R_{standard} are the hydrogen stable isotopic composition (i.e. the D/H molar ratio) of the sample and the Vienna standard mean ocean water (V-SMOW), respectively (Craig, 1961; Ehleringer and Rundel, 1988).

Statistical analysis

Data were statistically analysed using *SPSS* (*SPSS*, Inc., Chicago, IL, USA). An ANOVA analysis of field data showed that elevation did not significantly affect soil water stable isotope ($p = 0.66$) and content ($i = 0.18$) in the three soil profiles, and thus, the three plots are treated as replicates for further statistical analysis. A three-way ANOVA was used to test changes of water δD and content among soil layers and over time following each of rainfall events. Furthermore, one-way ANOVA with mean comparison was used to test for differences in soil water δD and content over time in each soil layer following rainfall events. Pearson's correlations were used to investigate relationships between soil physical properties. All statistically significant differences were tested at $\alpha = 0.05$ level.

RESULTS

Site characteristics

During the study period, a variety of rainfall events occurred (Figure 2). The δD of the rainfall showed seasonal variations, with lower values at the end of August and in the beginning of September 2003. The amount and intensity of the five sampled rainfall events are shown in Table III. After rainfall, evaporation on the forest floor was lower than that on adjacent open site (Tables IV and V). In the early spring, evaporation loss from soil was very low. Air and soil temperatures varied

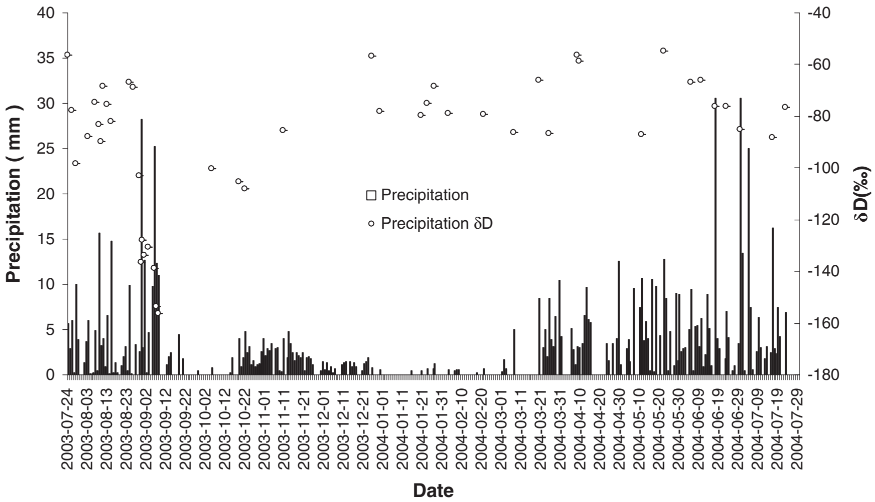

Figure 2. Rainfall and δD composition from July 2003 to July 2004 in the Wolong Nature Reserve of Sichuan, China

Table III. Rainfall intensity of the five sampled rain events

	Rain 1		Rain 2		Rain 3		Rain 4		Rain 5	
day	date	Rain (mm)	date	Rain (mm)	date	Rain (mm)	date	Rain (mm)	date	Rain (mm)
−1	Mar.5	0.0	Jul.27	0.0	Aug.8	0.0	Aug.14	0.1	Aug.29	2.0
1	Mar.7	5.0	Jul.28	9.9	Aug.9	15.7	Aug.15	14.8	Aug.30	28.3
2	Mar.8	0.0	Jul.29	3.9	Aug.10	3.2	Aug.16	0.2	Aug.31	3.0
3	Mar.9	0.0	Jul.30	0.0	Aug.11	4.0	Aug.17	1.3	Sep.1	12.7
4	Mar.10	0.0	Jul.31	0.0	Aug.12	0.9	Aug.18	0.2	Sep.2	0.2
5	Mar.11	0.0	Aug.1	1.4	Aug.13	6.6	Aug.19	0.1	Sep.3	4.6
6	Mar.12	0.0	Aug.2	3.7	Aug.14	0.1	Aug.20	1.1	Sep.4	0.0
7	Mar.13	0.0	Aug.3	6.0	Aug.15	14.8	Aug.21	0.0	Sep.5	9.8
8	Mar.14	0.0	Aug.4	0.1	Aug.16	0.2	Aug.22	0.0	Sep.6	25.3
9	Mar.15	0.0	Aug.5	0.1	Aug.17	1.3	Aug.23	0.0	Sep.7	12.3

The day 6, 7, 8, and 9 of rain event 3 were overlapped with the day 0, 1, 2, and 3 of rain event 4.

DOI: 10.1002/hyp

Table IV. Relationships between soil physical properties

	Bulk density	Total porosity	Capillary porosity	Non-capillary porosity	Field capacity
Total porosity	−0.93 (<0.01)				
Capillary porosity	−0.631 (<0.01)	0.726 (<0.01)			
Non-capillary porosity	−0.671 (<0.01)	0.671 (<0.01)	−0.0225 (0.93)		
Field capacity	−0.513 (0.03)	0.53 (0.02)	0.683 (<0.01)	0.0343 (0.89)	
Max retentive capacity	−0.639 (<0.01)	0.701 (<0.01)	0.738 (<0.01)	0.222 (0.38)	0.773 (<0.01)

The numbers are Pearson's correlation coefficients with *p*-values in parentheses.

Table V. Mean daily evaporation of the subalpine dark coniferous forest before and after each of rainfall events

	Evaporation (mm)									
	Rain 1		Rain 2		Rain 3		Rain 4		Rain 5	
Day before and after rain	Forest floor	Bare land	Forest floor	Bare land	Forest floor	Bare land	Forest floor	Bare land	Forest floor	Bare land
−1	0	1.5	1.1	4.4	0.4	4.5	0.3	2.4	0.2	1.1
1	0	0.4	0.4	4.0	0	0	0	0.9	0	0
2	0	2.9	0.7	5.2	0	0	0.4	2.0	0	0.3
3	0	5.1	1.0	6.5	0	0	0.3	2.6	0	0
4	0	6.0	1.3	6.7	0	1.3	1.0	5.0	0	0.4
5	0	0.7	1.8	5.6	0.2	1.1	1.0	5.2	0.2	1.2
6	0	3.2	1.1	3.7	0.3	2.4	0.6	1.4	0.4	1.6
7	0.2	1.6	0.4	2.1	0	0.9	0.8	1.7	0.4	1.6
8	0.3	3.4	0.9	5.0	0.4	2.0	0.7	3.0	0.3	1.0
9	0.5	4.3	1.3	6.5	0.3	2.6	0.8	3.4	0.5	1.0

seasonally (Table VI). Air temperature changed rapidly following rainfall compared with soil temperature. Changes in soil temperature decreased with soil depth.

The soil profile typically contained humus, illuvial, and parent material horizons (Table I). The thickness of the humus, illuvial, and parent material horizons in the soil profile was 4–11, 10–40, and 40–80 cm, respectively. Plant roots were distributed mainly at a depth of 0–60 cm with soil bulk density of 1.1–1.3 g cm^{-3} (Table II). Total soil porosity averaged 54%, and micropores or capillary pores accounted for 77% of total pore space. Soil field capacity and maximum retentive capacity averaged 52% and 63%, respectively. There were significant correlations between soil physical properties, except for the three relationships between capillary and non-capillary porosity, non-capillary porosity and field capacity, and non-capillary porosity and maximum retentive capacity (Table IV). Rocks were found at a depth of 60–80 cm or deeper below the soil surface.

Changes of soil water content following rainfall events

Prior to a rainfall event, on average, water content was 70% and 52%, respectively, in the litter layer and at 0–80 cm of soil depth in summer (from July to September 2003) and was 55% and 36% in the litter layer and at 0–80 cm of soil depth in early spring (March 2004, dry season). Soil moisture content decreased with soil depth (Figure 3, Table VII). Soil water content in all soil layers significantly increased with rainfall intensity (Tables VII and VIII, Supplemental Table I). For example, following a 5 mm rainfall on 7 March 2004, an increase in soil moisture content mainly occurred in the litter and at 0–40 cm depth and returned to its previous level within 4 days (Figure 3a). In the wet season, following a 10-mm or greater rainfall, large increases in soil water content occurred in all measured soil layers, including 40–80 cm of soil depth (Figures 3b, 3c, 3d, and 3e). Following a 28.3 mm rainfall on August 30, the litter and soil of 0- to 10 cm depth were nearly saturated, and soil water content in 10 to 40 cm layers was 85% for the first 3 days, and after then, moisture content gradually declined (Figure 3e). A subsequent rainfall event on September 7 caused an increase again in water content of the litter and soils. In the wet season, following each rainfall event, there were several smaller rainfalls (Table III). Rainfall event 3 (Figure 3c) was partially overlapped with the rainfall event 4, and the moisture content dramatically increased again in 5 days.

Temporal changes in soil water δD following rainfall events

Temporal changes in δD of soil water following rainfall events depend on the precipitation intensity and δD signature in the rainfalls (Figure 4, Tables VII and

DOI: 10.1002/hyp

Table VI. Air and soil mean daily temperature (°C) in the subalpine dark coniferous forest before and after the rainfall event

		Day before and after rain event									
Rain event	Air, soil layer (cm)	−1	1	2	3	4	5	6	7	8	9
1	Air	−1.0	−0.5	2.7	10.5	9.5	7.0	6.5	5.0	8.0	9.3
	0	3.3	2.0	1.8	3.2	2.2	6.0	5.6	5.5	5.2	6.0
	10	2.8	1.8	1.8	1.8	1.8	2.3	2.8	3.0	3.0	3.3
	20	3.0	2.5	2.8	2.2	2.3	2.5	2.7	2.5	3.0	3.0
	40	3.3	2.8	3.0	2.7	3.0	3.0	3.0	2.7	3.0	3.0
2	Air	12.0	10.0	11.3	11.7	11.0	13.3	11.0	10.3	10.7	12.7
	0	19.7	17.3	15.0	15.3	15.0	15.5	15.2	15.3	15.3	15.5
	10	18.0	16.5	15.5	14.7	14.2	14.2	14.0	14.0	13.8	14.0
	20	16.2	16.0	16.0	14.5	14.0	13.7	13.7	13.7	13.5	13.7
	40	9.8	9.9	10.0	10.0	10.2	10.4	10.4	10.4	10.4	10.4
3	Air	15.0	13.7	10.7	9.3	9.3	9.7	10.0	10.0	10.0	9.0
	0	14.8	13.7	12.3	11.2	11.3	10.7	10.8	10.7	10.3	10.2
	10	12.3	12.5	12.3	11.0	10.8	10.3	10.2	10.0	10.2	10.3
	20	12.3	12.5	12.5	11.3	10.0	10.7	10.3	10.5	10.2	10.3
	40	10.6	10.8	10.8	10.6	10.0	10.5	10.1	10.2	10.2	10.2
4	Air	14.0	10.3	10.0	9.0	11.3	12.3	11.7	12.7	13.3	13.3
	0	14.8	10.7	10.3	10.2	10.8	11.7	11.8	12.2	13.0	13.3
	10	12.3	10.0	10.2	10.3	10.2	10.2	10.3	10.5	11.0	11.3
	20	12.3	10.5	10.2	10.3	10.0	10.0	10.2	10.3	10.5	11.3
	40	10.6	10.2	10.2	10.2	10.0	10.0	10.0	10.2	10.0	10.1
5	Air	13.0	11.0	10.0	9.0	8.5	9.0	11.7	11.3	10.0	8.0
	0	13.3	11.0	10.7	10.7	9.5	8.8	10.7	11.5	10.8	9.0
	10	11.0	12.0	10.3	11.0	10.0	9.0	9.8	10.5	10.5	9.7
	20	11.0	12.0	10.7	11.0	10.5	10.0	10.0	10.5	10.5	10.0
	40	10.3	10.6	10.3	10.0	10.0	10.0	10.0	10.0	10.0	10.0

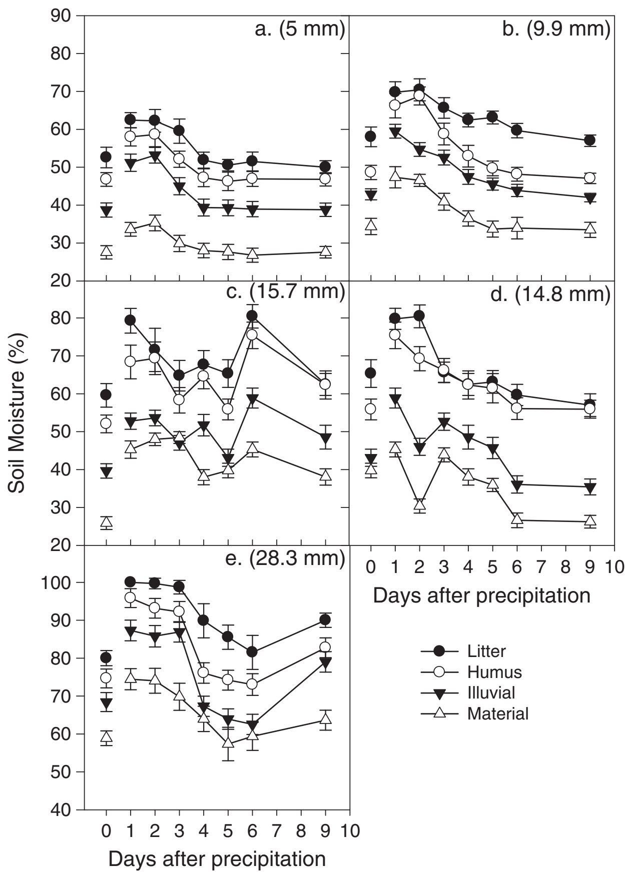

Figure 3. Soil moisture content (mean ± SD, n = 3) in the subalpine dark coniferous forest before and after the five rainfall events

VIII, Supplemental Table II). Following a 5.0-mm rainfall on 7 March 2004, the δD of water from the litter and all soil layers decreased first and then gradually went back to its previous level (Figure 4a). In the wet season, following 9.9-, 15.7-, and 14.8-mm rainfall events, the δD of water from the litter, humus, and illuvial layers first decreased and then slightly increased or fluctuated, except for the litter of rainfall event 2, whereas the δD of water from the parent material layer remained unaltered (Figure , 4b, 4c and 4d, Supplemental Table II). In contrast, 28.3-mm rainfall with the low δD signature greatly affected the δD of water in the litter and soil profile (Figure 4e, Supplemental Table II). The δD of soil water in humus and illuvial layers first dramatically decreased, following the rainfall of 28.3 mm. The δD of water from the litter

Table VII. Summary of three-way ANOVA in testing effects of rainfall, soil depth, and time after rainfall events on content and δD of soil water in the subalpine dark coniferous forest

		Water content (%)		δD (‰)	
Factor	d.f.	*F*-value	*p*-value	*F*-value	*p*-value
Rainfall	4	5998.11	<0.0001	4573.29	<0.0001
Depth	3	6027.39	<0.0001	5160.78	<0.0001
Day	6	763.51	<0.0001	63.01	<0.0001
Rainfall × depth	12	44.26	<0.0001	1316.71	<0.0001
Rainfall × day	24	89.58	<0.0001	19.14	<0.0001
Depth × day	18	14.98	<0.0001	10.55	<0.0001
Rainfall × depth × day	72	9.69	<0.0001	7	<0.0001

DOI: 10.1002/hyp

Table VIII. Summary of one-way ANOVA in testing dynamic changes in content and δD of soil water over time after rainfall events in each soil layer

Rainfall event	Soil layer	Water content (%)		δD of soil water (‰)	
		F-value	*p*-value	*F*-value	*p*-value
1	Litter	101.20	<0.001	13.43	<0.001
	Humus	50.07	<0.001	7.61	<0.001
	Illuvial	78.60	<0.001	12.72	<0.001
	Material	27.27	<0.001	3.28	0.008
2	Litter	162.36	<0.001	1.66	0.148
	Humus	118.06	<0.001	6.71	<0.001
	Illuvial	124.09	<0.001	8.75	<0.001
	Material	64.28	<0.001	1.09	0.379
3	Litter	24.71	<0.001	6.25	<0.001
	Humus	31.31	<0.001	1.10	<0.001
	Illuvial	41.80	<0.001	5.66	<0.001
	Material	38.36	<0.001	0.33	0.921
4	Litter	45.92	<0.001	3.67	0.004
	Humus	41.84	<0.001	13.08	<0.001
	Illuvial	69.90	<0.001	28.07	<0.001
	Material	113.45	<0.001	0.64	0.696
5	Litter	47.54	<0.001	69.52	<0.001
	Humus	121.27	<0.001	12.38	<0.001
	Illuvial	154.44	<0.001	11.23	<0.001
	Material	35.36	<0.001	3.47	0.006

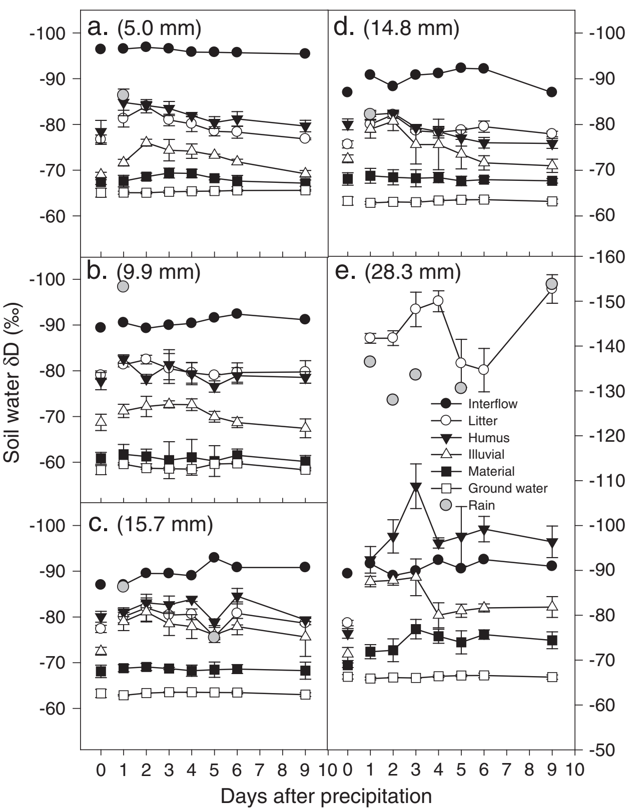

Figure 4. Daily δD (mean ± SD, n = 3) of water in the litter layer and soil profile, soil interflow, rainwater, and shallow groundwater in the subalpine dark coniferous forest before and after the five rainfall events. Days 5, 6, and 9 of rain event 3 were overlapped with days 0, 1, and 4 of rain event 4

greatly decreased to the level of the rainwater. A henceforth rainfall event on September 6 caused the δD of water from the litter to become lower on day 9, although it did not affect the δD of soil water (Figure 4e). In addition, the δD of soil interflow was relatively stable and around −90‰ except for rainfall event 1 in the dry season (Figure 4). Similarly, the δD of shallow groundwater did not change with the rainfall events during the study period, and the values were all below −60‰ except for rainfall event 2 in which the values were just above −60‰.

δD profile and vertical movement of soil water following rainfall events

Following the 5-mm rainfall in dry season of 7 March 2004, there were relatively synchronous profiles along the soil depth over the 9-day period (Figure 5a). The δD of water from 0- to 10-cm layers of the soil had closer δD signature of rainfall. Following a 9.9-mm rainfall, a relatively smooth curve of the δD profile of soil water was observed (Figure 5b). Two rainfall events with similar precipitation intensity of 15.7 and 14.8 mm that occurred in August had different effects on the δD profile of soil water (Figures 5c and 5d). Following the 15.7-mm rainfall, the δD of water in the litter and the soil at a depth of 0–30 cm decreased on August 11–14, as indicated by a

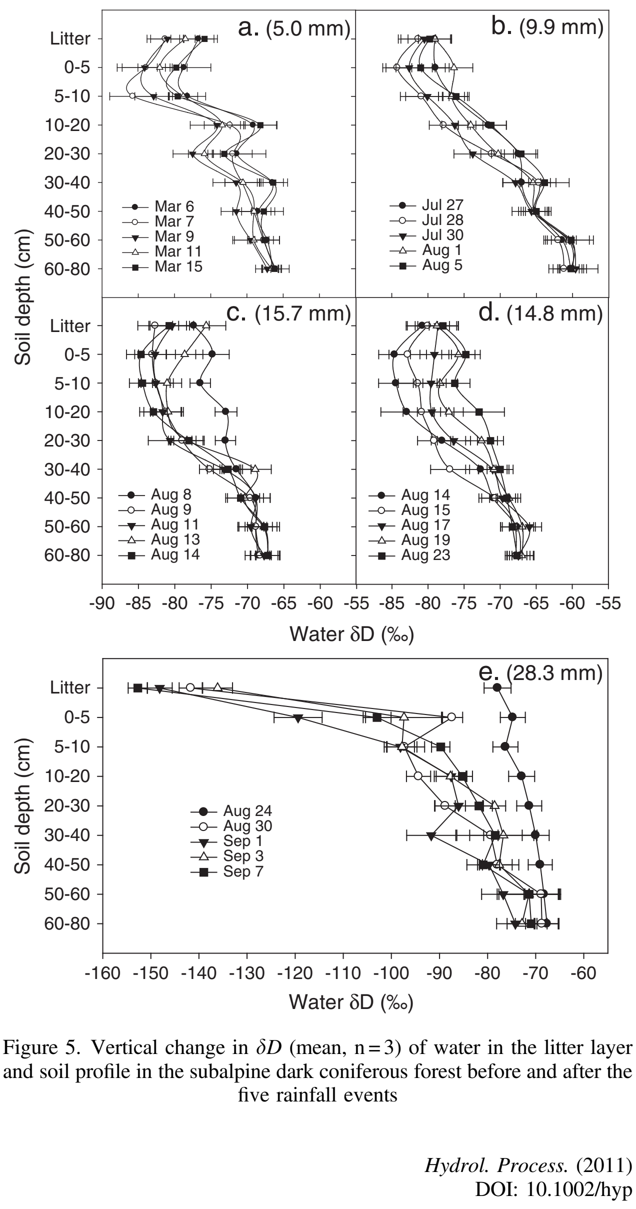

Figure 5. Vertical change in *δD* (mean, n = 3) of water in the litter layer and soil profile in the subalpine dark coniferous forest before and after the five rainfall events

DOI: 10.1002/hyp

shift of the upper portion of the δD profile curve of soil water toward the left, away from the δD profile curve of August 8 (Figure 5c). In contrast, the δD profile curve of soil water at a depth of 0–30 cm shifted toward the right following the 14.8-mm rainfall on August 19 and 23 (Figure 5d). The 28.3-mm rainfall greatly affected the shape of the δD profile curve of soil water, especially in litter and at 0- to 5-cm top soil layers, and the high intensity rainfall had significant impact on soils at 60- to 80-cm depth (Figure 5e, Supplemental Table II).

DISCUSSION

Temporal changes in δD of soil water following rainfall events

Temporal changes in δD of soil water following rainfall events depend on the precipitation intensity, δD signature in rainfalls, and plant phenology. We observed that the effect of rainfalls on δD of soil water differed between early spring and summer (Figure 4). Although rainfall intensity was low in early spring, the spring rainfall could infiltrate into deeper soil layers and, thus, had significant contribution to soil water. At that time, the temperature in air and soil was low (Table VI), and there was low plant evapotranspiration with plants inactive. Hydrological processes are partially determined by the period of activity of the plants (Reynolds *et al*., 2004). Therefore, the rainfall in early spring could even recharge groundwater (Bengtsson *et al*., 1987). On the other hand, summer rainfalls with intermediate intensity temporally increased soil water content but contributed little to water δD in deeper soil layers likely because water consumption by plants with deep root system partially neutralized the rainfall effect. However, heavy rainfall events (e.g. 28.3 mm) greatly increased soil water content and significantly affected hydrogen composition of the soil water, and thus, high-intensity summer rainfall could have recharged the groundwater.

The δD of soil interflow fluctuated with rainfall events, particularly following the 28.3-mm rainfall event (Figure 4). Thus, the source of the soil interflow was unlikely from rainfall. The δD of soil interflow in dry season was lower than that in wet season, suggesting that the source of the soil interflow might be susceptible to evaporation. Similarly, the δD in shallow underground water at 1.5 m did not change regardless of amount of rain.

δD profile and vertical movement of soil water following rainfall events

Changes in δD profile of soil water following rainfall events (Figure 5) indicated that the vertical movement of soil water was affected by rainfall events, evapotranspiration, and environmental conditions in the forest ecosystem.

The depleted δD of soil water from a depth of 0–10 cm (Figures 5a and 5b) suggested that pre-event 'old' water at a depth of 0–10 cm was replaced by rain water following low-intensity rainfall events (i.e. 5–10 mm). Gradual increases in δD of soil water with depth at 10–20 cm (Figure 5a) also indicated that rain may disperse through the soil profile. The rain likely moved through soil layers by piston displacement (Kendy *et al*., 2003; Gazis and Feng, 2004), with some water mixing between horizons. In contrast, rapid, great increases in δD profile curve of soil water following high-intensity rainfall (particularly Figure 5e) indicated that preferential flow occurred through macropores, decayed plant roots, and rocks at different depths of soil profile, and the infiltration process was likely heterogeneous during the initial period of rain (Beven and Germann, 1982; Kung, 1990). With continuing rainfall, the litter and surface soil were saturated or nearly saturated, and the wet front expanded gradually to all layers, and the infiltration process became homogeneous and plug-like. In addition, the decreased δD of soil water at a depth of 60–80 cm indicated that rainwater infiltrated at least to 80 cm of soil depth. Such high-intensity summer rainstorms have the potential to recharge regional groundwater stores. However, contribution of δD signature in rainfalls cannot be excluded, that is, a very large difference between the deuterium ratio of the precipitation and the soil moisture also would contribute to the abrupt changes in δD profile curve of soil water in rainfall event 5.

Initial decrease in δD of soil water and a subsequent return to the pre-rainfall level at a soil depth of 0–50 cm (Figures , 5a–5d) suggest that the infiltrated rain in soil was lost in several days, probably because of surface evaporation (Table V) and transpiration. Similarity of δD of water at a depth of 50–80 cm between pre-rainfall and post-rainfall suggests that the rainfall might not affect δD of soil water at soil depth of 50 cm or deeper.

In addition, comparisons between 5 and 9.9 mm rainfall events indicated that the effect of rainfall on δD of soil water differed between early spring and summer (Figures 5a and 5b). Although the rainfall intensity was low in spring, the spring rainfall could infiltrate into deeper soil layers and, thus, could greatly be attributed to the soil water, partially because of lower evaporation (Table V). Summer rainfall temporally increases soil water content, and its contribution to soil water is accordingly short lived because of high evapotranspiration.

Difference in δD profile of soil water following two rainfall events with similar intensity (Figures 5c and 5d) indicated the effect of forest evaporation on vertical water movement. The upper part of the δD profile curve of soil water shifted toward the left (Figure 5c), which indicated that more rain infiltrated into the soil profile over times. It was likely because rainfall was held in the forest, moss, and floor layers, because of low forest evaporation (Table V), and gradually moved downward in the soil profile. This finding also indicated that the forest, moss, and litter layers play an important role in holding rainwater. In contrast, the δD profile curve of soil water at a depth of 0–40 cm shifted toward the right following the 14.8-mm rainfall (Figure 5d), suggesting that the rain water was lost because of stronger evapotranspiration in the forest.

Forest soil water particularly in the deeper soil profile and shallow groundwater is a long-term, mixed, accumulative

DOI: 10.1002/hyp

product from previous precipitation and also is an important water source in the sub-alpine dark coniferous forest (Liu, 2002). Its release through springs and seeps maintains the flow of streams and rivers. Thus, forest soil water, as a pool in the hydrological cycle, can be a source of water supply for regional water needs, particularly during dry seasons. Afforestation increases actual evapotranspiration and reduces annual streamflow, but it may, through increased infiltration, increase groundwater release in the study region (Ma *et al*., 2010).

CONCLUSIONS

The effects of precipitation on soil water content and δD of soil water depend on rainfall and evapotranspiration intensity, as well as infiltration. Light and intermediate rainfall events (i.e. 5–15 mm) increased water content in the litter and soils at a depth of 0–80 cm, but the increased soil water content was lost in several days because of evapotransration and/or infiltration. Heavy rainfall events (e.g. 28.3 mm) greatly increased soil water content. The surface soil layer was nearly saturated, following the heavy rainfall, and soil water content in the upper layers (0–40 cm) was up to 85%. Therefore, heavy summer rainfall could recharge the groundwater and increase streamflow.

Light rainfall in early spring affected δD of water from the litter, humus, illuvial, and parent material layers, whereas intermediate rainfalls in summer did not change δD of water from parent material horizon, probably because the water uptake by plants with deep root system neutralized the rainfall effect in summer. However, the heavy rainfall greatly affected δD of water in the litter and soil profile. The δD of both interflow and shallow groundwater did not differ significantly following both light and heavy rainfall.

Light rainfall temporally altered the shape of δD profile curve of soil water. Following a rainfall event, infiltration mainly occurred in the upper soil depth, and the rain water was depleted through evaporation and likely plant uptake. In contrast, heavy rainfall greatly affected the shape of δD profile curve of soil water. Following a heavy rainfall, preferential flow initially occurred through macropores in soils and rocks at different depths of soil profile.

ACKNOWLEDGEMENTS

This study was jointly supported by China Natural Science Foundation (No 30771712, No. 31170661 and No. 30590383), China State Forestry Administration (No 200804001 and No. 201104006), CFERN and GENE Award Fund and National Basic Research Projects of China (No 2002CB111504).

REFERENCES

Allison GB, Hughes MW. 1983. The use of natural tracers as indicators of soil –water movement in a temperate semi-arid region. *Journal of Hydrology* **60**: 157–173.

Beven K, Germann P. 1982. Macropores and water flow in soils. *Water Resource Research* **18**(5): 1131–1325.

Bengtsson L, Saxena RK, Dressie Z. 1987. Soilwater movement estimated from isotope tracers. *Hydrological Sciences Journal* **32**(4): 497–520.

Craig H. 1961. Isotopic variation in Meteoric waters. *Science* **133**: 1702–1703.

Dansgaard W. 1964. Stable isotopes in precipitation. *Tellus* **XVI**: 436–468.

Dawson TE, Mambelli S, Agneta HP, Pamela HT, Kevin PT. 2002. Stable isotopes in plant ecology. *Annual Review of Ecology and Systematics* **33**: 507–559.

Ehleringer JR, Rundel PW. 1988. *Stable Isotopes in Ecological Research*. Springer-Verlag: New York; 1–15.

Ellsworth PZ, Williams DG. 2007. Hydrogen isotope fractionation during water uptake by woody xerophytes. *Plant and Soil* **291**: 93–107.

Ferrio JP, Resco V, Williams DG, Serrano L, Voltas J. 2005. Stable isotopes in arid and semi-arid forest systems Isotopes estables en sistemas forestales de ambientes aridos y semiaridos. *Investigacion Agraria: Sistemasy Recursos Forestales* **14**(3): 371–382.

Fravolini A, Hultine KR, Brugnoli E, Gazal R, English NB, Williams DG. 2005. Precipitation pulse use by an invasive woody legume: the role of soil texture and pulse size. *Oecologia* **144**: 618–627.

Gazis C, Feng X. 2004. A stable isotope study of soil water: evidence for mixing and preferential flow paths. *Geoderma* **119**(1–2): 97–111.

Gehrels JC , Peeters JEM, De Vries JJ, Dekkers M. 1998. The mechanism of soil water movement as inferred from ^{18}O stable isotope studies. *Hydrology Science Journal* **43**(4): 579–594.

Hillel D. 1998. *Environmental Soil Physics: Fundamentals, Applications, and Environmental Considerations*. Academic Press: San Diego, CA, USA. 771.

Huo XP, Li XW, Zhang J, Fan C, Dai J, Zhang XH, Zhang LH. 2009. Soil permeability capability of subalpine coniferous forests in western Sichuan, China. *Research of Soil and Water Conservation* **16**: 192–195.

Hsieh JCC, Chadwick OA, Kelly EF, Savin SM. 1998. Oxygen isotopic composition of soil water: quantifying evaporation and transpiration. *Geoderma* **82**(1–3): 269–293.

Kendy E, Gérard-Marchant P, Walter TM, Zhang Y, Liu C, Steenhuis TS. 2003, A soil-water-balance approach to quantify groundwater recharge from irrigated cropland in the North China Plain. *Hydrological Processes* **17**: 2011–2031.

Kung KJS. 1990. Preferential flow in a sandy vadose zone, 1: field observation. *Geoderma* **46**: 51–58.

Lee KS, Kim JM, Lee DR, Kim Y, Lee D. 2007. Analysis of water movement through an unsaturated soil zone in Jeju Island, Korea using stable oxygen and hydrogen isotopes. *Journal of Hydrology* **345**: 199–211.

Lin GH, Phillips SL, Ehleringer JR. 1996. Monsoonal precipitation responses of shrubs in a cold desert community on the Colorado Plateau. *Oecologia* **106**: 8–17.

Liu Q. 2002. *Ecological Research on Subalpine Coniferous Forest in China*. Sichuan University Press: Chengdu; **33–98**: 217–233.

Löffler J. 2007. The influence of micro-climate, snow cover, and soil moisture on ecosystem functioning in high mountains. *Journal of Geographical Sciences* **17**(1): 3–19.

Ma X, Xu J, van Noordwijk M. 2010. Sensitivity of streamflow from a Himalayan catchment to plausible changes in land cover and climate. *Hydrological Processes* **24**: 1379–1390.

Reynolds JF, Kemp PR, Ogle K, Fernández RJ. 2004. Modifying the 'pulse–reserve' paradigm for deserts of North America: precipitation pulses, soil water, and plant responses. *Oecologia* **141**: 194–210.

Reynolds JF, Kemp PR, Tenhunen JD. 2000. Effects of long-term rainfall variability on evapotranspiration and soil water distribution in the Chihuahuan Desert: a modeling analysis. *Plant Ecology* **150**: 145–159.

Shi H, Chen FQ, Liu SR. 2005. Macropores properties of forest soil and its influence on water effluent in the upper reaches of Minjiang River. *Acta Ecologica Sinica* **25**: 507–512.

Song XF, Wang SQ, Xiao GQ, Wang ZM, Liu X, Wang P. 2009. A study of soil water movement combining soil water potential with stable isotopes in two sites of shallow groundwater areas in North China Plain. *Hydrological Processes* **23**: 1376–1388.

Stephenson NL. 1990. Climatic control of vegetation distribution: the role of the water balance. *The American Naturalist* **135**(5): 649–670.

Williams DG, Ehleringer JR. 2000. Intra- and interspecific variation for summer precipitation use in pinyon-juniper woodlands. *Ecological Monographs* **70**: 517–537.

Xu Q, An SQ, Liu SR, Jiang YX, Cui J. 2005. Hydrogen stable isotope characteristics of the precipitation allocation processes in Sub-alpine Dark Coniferous Forest in Wolong, Sichuan Province. *Scientia Silvae Sinicae* **41**(4): 7–12.

Xu Q, Liu SR, An SQ, Jiang YX, Wang ZS, Liu JT. 2006. Allocation of precipitation in a sub-alpine dark coniferous forest of western Sichuan using stable oxygen isotopes. *Acta Phytoecologica Sinica* **30**: 83–89.

Xu Q, Li HB, Chen JQ, Cheng XL, Liu SR, An SQ. 2011. Water use patterns of three species in subalpine forest. Southwest China: the deuterium isotope approach. *Ecohydrology* **4**: 236–244.

Zimmerman U, Ehhalt D, Munnich KO. 1967. Soil water movement and evapotranspiration: changes in the isotopic composition of the water. In *Proceedings of the IAEA sysposium on the use of isotopes in hydrology*. IAEA: Vienna, 1966; 567–584.