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# Impacts of soil conservation on groundwater recharge in the semi-arid Loess Plateau, China

John B. Gates · Bridget R. Scanlon · Xingmin Mu · Lu Zhang

**Abstract** Soil conservation measures undertaken to address land degradation can alter the hydrologic cycle by changing partitioning of water fluxes at the land surface. While effects on runoff are well documented, impacts of soil conservation activities on fluxes to groundwater are poorly understood. The goal of this study was to examine fluxes to groundwater in a semi-arid area of China's Loess Plateau that has been subject to extensive soil conservation activities. Unsaturated zone pore-water pressures and concentrations of chloride show that impacts on deep drainage differ between ecological and structural soil conservation approaches. High matric potentials and low chloride beneath cultivated terrace and gully sites are consistent with deep drainage occurring at these sites. Estimated recharge rates for dryland cultivated upland sites were approximately 55–90mm/year (11–18% of mean annual rainfall) based upon chloride mass balance. In contrast, results suggest that mature tree and shrub plantations prevent deep drainage. Stable isotope signatures of unsaturated-zone moisture and

groundwater indicate that focused infiltration through gullies and other topographic lows is likely to be the primary recharge mechanism. The results of this study highlight the potential for inadvertent effects of some soil conservation approaches on regional water resources.

**Keywords** Groundwater recharge · Unsaturated zone · Loess plateau · China · Agriculture

## Introduction

Efforts to reduce erosion of agricultural soils are critical to maintaining food security and environmental quality. Human-induced soil erosion is estimated to affect approximately 1.6 million km<sup>2</sup> globally and significantly reduce world agricultural productivity (Biggelaar et al. 2003; Lal 2001). Eroded sediments and nutrients impair surface-water quality in agricultural catchments worldwide. Erosion is also a concern in the context of climate change because removal of soil organic carbon currently results in emission on the order of 1 petagram of carbon per year to the atmosphere (Lal 2003).

Common soil conservation practices have the potential to affect terrestrial water cycles in addition to soil transport. These practices are intended to mitigate soil loss by altering land-surface characteristics such as topography, soil structure, and vegetation cover (Unger 1995), which are also key factors affecting the partitioning of precipitation into evapotranspiration, runoff, and sub-surface drainage (Zhang et al. 2007a). In this way, large-scale soil-conservation efforts are capable of impacting water resources. Impacts of many other types of land-use changes on rivers and aquifers have been documented globally (Foley et al. 2005; Jackson et al. 2005; Scanlon et al. 2007). For example, deforestation is known to have increased river discharges in many basins worldwide because of reduced transpiration (Brown et al. 2005; Zhang et al. 2001). Similarly, conversion of native vegetation to cropland in semi-arid regions can increase recharge into underlying aquifers and lead to secondary impacts such as water-table rise and salinization (Allison et al. 1990; Leduc et al. 2001; Scanlon et al. 2005). The current expansion of bio-energy production is also projected to impact local and regional water cycles

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(Schilling et al. 2008; Uhlenbrook 2007). Soil-conservation activities result in similarly widespread land-use changes, yet little information is available on their hydrological impacts.

China's Loess Plateau has among the highest rates of agricultural soil erosion in the world (Fu 1989; Lal 2003) and is susceptible to water scarcity because of high agricultural water demand and semi-arid climate. Loess is highly erodible and, in combination with climate and agricultural land-use factors, results in extremely high erosion rates over much of the region. The Loess Plateau has been subject to a wide range of soil conservation measures over the past 40 years, and about 20% of its 640,000-km<sup>2</sup> area is now under some type of soil conservation (Upper and Middle Yellow River Bureau of the Yellow River Conservancy Commission 2001). A diverse range of soil conservation approaches have been applied in the Loess Plateau, including structural approaches like terracing and construction of check-dams across drainages and gullies (~30% of total conservation area), as well as ecological approaches (~50% of total conservation area) involving the introduction of trees, shrubs and grasses to promote soil stability and reduce runoff from steeply sloping terrain.

Because of the possibility for soil conservation activities to adversely affect water resources in the Loess Plateau, it is desirable to examine the impacts of soils conservation techniques on agricultural water cycling. The soil conservation methods used in the Loess Plateau are intended to reduce runoff, and have done so effectively in many cases (Chen et al. 2007; Fu et al. 2004; Huang et al. 2003a; Pan and Shangguan 2006; Zheng 2006). Water balance dictates that the volume of runoff reduction must be balanced by either increased subsurface drainage or evapotranspiration, or a combination of these. Discharges from river basins across the region have been in decline and this has been primarily attributed to reduced inputs as a consequence of soil conservation measures (He et al. 2003; Huang and Zhang 2004; Huang et al. 2003b; Mu et al. 2007; Zhang et al. 2008a, 2008b). For example, the Jialuhe Catchment (1,117 km<sup>2</sup>) experienced a 32% reduction in stream flow from 1967 to 1989 that was primarily attributed to soil conservation practices (Huang and Zhang 2004).

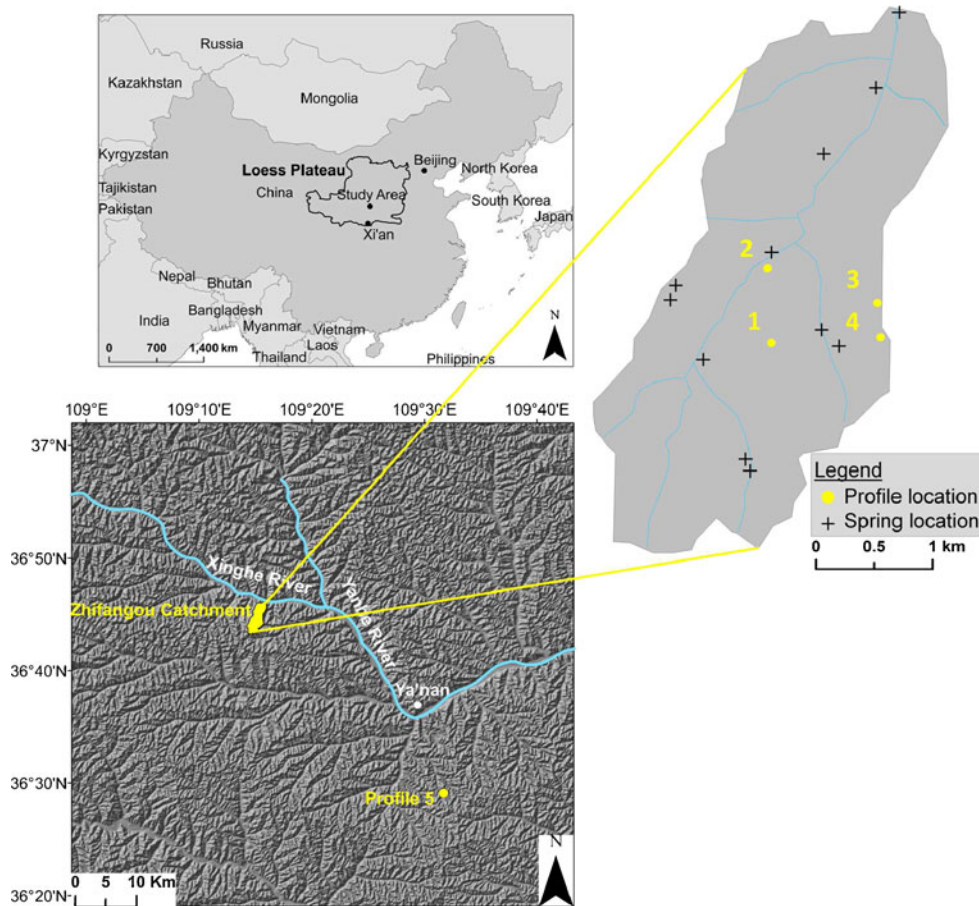
In contrast to surface-water impacts which are widely recognized, impacts of soil conservation on fluxes to groundwater in the Loess Plateau are poorly understood. Previous information on groundwater recharge in the Loess Plateau is limited and estimates of recharge rates are only available for a few locations. Tritium peak-displacement rates yielded recharge estimates of 47 and 68 mm/year, or approximately 12–13% of mean annual precipitation) for non-vegetated sites in Inner Mongolia and Shanxi Province (Lin and Wei 2006). Chloride mass balance suggested recharge rates of 33 and 55 mm/year for two sites beneath winter wheat cultivation in Gansu Province and Ningxia Province, respectively (Huang and Pang 2010). Modeling studies of agricultural water balance suggested that

recharge rates of 18 mm/year beneath wheat fields in Shaanxi Province are reduced to 9 mm/year when the fields are converted to apple orchards (Huang and Gallichand 2006). Several authors have also noted the likelihood that tree and shrub plantations may result in recharge reductions based on persistent soil moisture reductions, which have been extensively documented beneath afforested areas in the Loess Plateau (Chen et al. 2008b; He et al. 2003; Hou et al. 1999; Huang and Gallichand 2006; Huang et al. 2003b; Li 1983; McVicar et al. 2007; Mu et al. 2007; Shangguan and Zheng 2006) and attributed to the relatively high water demands of introduced vegetation (Chen et al. 2008b, 2007; Fu et al. 2003; He et al. 2003; Huang and Gallichand 2006; Li et al. 2008; Wang et al. 2010; Cao 2010, Hou 2000, and Sun 2006).

This study examined groundwater recharge in a semi-arid area of the Loess Plateau that has been subject to extensive soil conservation activities. This topic has relevance to resource management in the region because baseflow is an important contribution to rivers, particularly in dry seasons when runoff is low (Huang and Zhang 2004). The objectives of the study were to assess (1) how unsaturated-zone conditions compare among sites that are affected by differing soil conservation practices, and (2) the mechanisms of groundwater recharge in a 9 km<sup>2</sup> catchment. The study involved a parallel use of geochemical and physical techniques that has been used in previous applications (Scanlon and Goldsmith 1997; Scanlon et al. 2009), and was applied in the Loess Plateau for the first time. This approach provides evidence on net hydrological patterns integrated over time, which can be an advantage over monitoring where there is large temporal variability in hydrological flows and low flux rates. This is a data gap that has been previously recognized for dryland cropping systems of the Loess Plateau (Zhang et al. 2007a).

## Study area

Field investigations took place in an agricultural catchment of Shaanxi Province in the central Loess Plateau (Zhifangou Catchment; 36°44' N, 105°15' E; elevation ~1,000–1,200 m above sea level; Fig. 1) in April 2008. The catchment was a demonstration area of intensive soil conservation activities beginning in the 1970s and continues to use a wide range of conservation practices (Liu 1999). Cropland covers 20% of the catchment and 85% of cultivated area is located on terraces. Soil erosion rates declined from ~14,000 t/km<sup>2</sup> in 1975 to ~1,600 t/km<sup>2</sup> in 1997 (Liu 1999). Mean annual precipitation is approximately 500 mm in this region, about 70% of which is associated with the East Asian Summer Monsoon. The landscape is characterized by thick loess deposits (commonly >100 m) with steeply incised valleys. Unsaturated sediments are dominantly comprised of silty loess materials (mean



**Fig. 1** Location of Loess Plateau, the study area catchment, and sample locations

sediment texture was 35% sand, 50% silt and 15% clay). Water tables are mostly deep (up to hundreds of meters below surface) in upland areas because of thick loess deposits, and reach the surface near stream channels within incised valleys. In addition, the presence of localized perched groundwater is suggested by some springs emanating from hillsides at high elevation. Typical of upland catchments in the Loess Plateau, agricultural production in the study area is rain-fed (no groundwater irrigation).

Five sites were selected for unsaturated-zone profile studies to compare hydrological impacts of soil conservation practices (Table 1). Profile settings included two cultivated fields, one of which is located on a terrace (profile 1; Fig. 1) and one is located in a gully immediately upstream of an earthen check-dam (profile 2). Both sites were cultivated under maize and winter wheat rotation. Profile sites also included a tree plantation (*Robinia pseudoacacia*; profile 3), and a hillslope with introduced shrubs (*Caragana korshinskii*; profile 4). Each of the selected sites had been under the same land cover for approximately 30 years according to local records. In addition, one site with natural forest vegetation site was examined (profile 5). This site was selected from outside of the Zhifangou Catchment (from a location 35 km southeast in the vicinity of

Ya'nan; Fig. 1) because no sites with natural forest vegetation were present in the Zhifangou Catchment.

## Materials and methods

Unsaturated-zone sediment cores were collected for analysis of chloride ( $\text{Cl}^-$ ) concentrations, matric potentials and pore-water moisture contents. Dissolved  $\text{Cl}^-$  in pore water was utilized as a conservative tracer that is excluded by evapotranspiration and not subject to geochemical interactions with sediments (Cook et al. 1992; Eriksson and Khunakasem 1969; Hem 1985; Scanlon et al. 2002). Steady state  $\text{Cl}^-$  mass balance was used to estimate rates of deep drainage (e.g. potential recharge). It is typically formulated as:

$$P \cdot C_P = R \cdot C_R \quad (1)$$

where  $P$  is precipitation rate,  $C_P$  is  $\text{Cl}^-$  concentration in precipitation adjusted for additions from dry deposition,  $R$  is potential recharge rate, and  $C_R$  is  $\text{Cl}^-$  concentration of potentially pore waters (Allison and Hughes 1978; Edmunds et al. 1988; Gates et al. 2008a).

The equation was modified for this study to account for  $\text{Cl}^-$  fluxes in overland flow which may be an important

**Table 1** Profile data summaries. Moisture content, chloride and matric potentials are expressed as profile depth-weighted averages

Profile	Setting	Sampling date (mm/dd/yy)	Location	Total depth (m)	Moisture content (% g/g)	Cl (mg/kg)	Cl (mg/L)	Matric potential (m)	No. of Samples
1	Cultivated terrace	04/09/08	109°15.129'N 36°44.373'E	11.8	21.1	2.6	13.6	-1.3	87
2	Cultivated behind check-dam	04/09/08	109°15.109'N 36°44.765'E	6.1	12.2	0.9	9.1	-2.9	19
3	Tree plantation	04/10/08	109°15.664'N 36°44.588'E	12.0	8.1	1.0	10.5	-93.4	88
4	Shrub plantation	04/13/08	109°15.693'N 36°44.415'E	20.0	7.9	0.6	6.3	-136.1	120
5	Natural forest	04/15/08	109°31.542'N 36°28.855'E	10.0	10.0	2.5	25.9	-142.4	66

component of water and solute balances for some sites. For the case of a site where  $\text{Cl}^-$  may be lost (runoff generated) or gained (runon received) by overland flow:

$$P \cdot C_P + (Q_{\text{on}} \cdot C_{Q_{\text{on}}} - Q_{\text{off}} \cdot C_{Q_{\text{off}}}) = R \cdot C_R \quad (2)$$

where  $Q_{\text{on}}$  and  $Q_{\text{off}}$  are runon and runoff fluxes of water, and  $C_{Q_{\text{on}}}$  and  $C_{Q_{\text{off}}}$  are their respective  $\text{Cl}^-$  concentrations. The expression is similar to that of Wood et al. (1997) but allowing for both losses and gains of  $\text{Cl}^-$  by surface flow.  $\text{Cl}^-$  accumulation is also used to estimate the residence time of water ( $t$ ) at a given depth by dividing the total mass of  $\text{Cl}^-$  in the profile above a given depth by the annual  $\text{Cl}^-$  input rate:

$$t = \frac{1}{Cl_{\text{in}}} \int_0^{z_i} \theta(z) C_R(z) dz, \quad (3)$$

where  $Cl_{\text{in}}$  is the  $\text{Cl}^-$  mass influx rate (left hand side of Eq. 2),  $\theta$  is moisture content and  $z_i$  is depth at interval  $i$  below surface (Cook et al. 1992). Mean  $\text{Cl}^-$  in precipitation was estimated to be 1.4 mg/L based on wet deposition monitoring data (EANET 2010). Data on dry deposition of  $\text{Cl}^-$  and  $\text{Cl}^-$  concentration in runoff were not available for the study area; therefore, we use the bounding assumptions that dry deposition represents a mass flux between 0 and a flux equal to wet deposition ( $\sim 700 \text{ mg/m}^2/\text{year}$ ), and that  $\text{Cl}^-$  concentration in runoff is equal to the concentration in rainfall (1.4 mg/L).

Unsaturated-zone matric-potential measurements are used to assess the direction of vertical unsaturated flow (upward or downward). Matric potential is the component of potential energy of pore water that is due to capillary and adsorptive forces between water and the sediment matrix. Downward unsaturated water flux requires gravitational force to overcome matrix forces. In cases where osmotic potential is negligible (e.g. pore water does not have high salinity), the vertical gradient of the sum of matric potential and gravitational potential, as an approximation of total water potential, serves as a good indicator of direction of water movement in the unsaturated zone (Jury et al. 1991; Scanlon and Goldsmith 1997). Matric

potentials in the unsaturated zone are a function of moisture content and the relationship between the two can be expressed by a water-retention function (Mualem 1976; van Genuchten 1980).

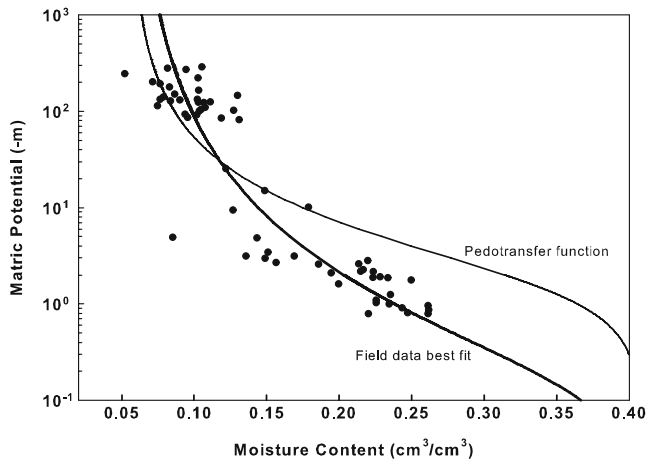
Numerical unsaturated-zone flow and solute transport modeling was undertaken to support profile data interpretation. Unsaturated water flow was simulated using the Richards equation and conservative solute movement by advection and dispersion with the Hydrus-1D model (Simunek et al. 2005). Soil hydraulic parameters (Table 2; Fig. 2) were estimated using laboratory moisture and matric potential measurements. The simulations included free drainage lower boundary conditions and differing upper boundary conditions for each simulation (described in section Discussion). Root water uptake was simulated using the Feddes (1974) model within Hydrus-1D.

Electromagnetic induction surveying was used as a reconnaissance tool to investigate spatial variability of soil moisture around profile sites. This step was taken to ensure that shallow moisture contents of point-based profile locations were broadly representative of the field scale. Transects were taken across fields containing profiles 1–4 plus one additional transect across a hillslope with scrub vegetation and one across a terrace containing a tree plantation. Electromagnetic induction survey measurements were collected on the same dates that unsaturated-zone cores were collected (04/09/2008–04/15/2008). (High vegetation density prevented surveying at the Profile 5 site). Bulk soil electrical conductivity ( $\text{EC}_a$ ) transects were taken with a Geonics EM-38 (Geonics Ltd, Canada) in the vertical dipole position. In this position the EM-38 has a nominal response depth of approximately 1.5 m (McNeill 1980). Variability in  $\text{EC}_a$  has been shown to be a function of moisture content, clay content and

**Table 2** Soil hydraulic parameters (Van Genuchten model) applied in unsaturated-zone flow simulations

Parameter	Field data best fit	Pedotransfer function
$\theta_r$ ( $\text{cm}^3/\text{cm}^3$ )	0.0562	0.0562
$\theta_s$ ( $\text{cm}^3/\text{cm}^3$ )	0.4076	0.4076
$\alpha$ ( $\text{cm}^{-1}$ )	0.0730	0.0058
$n$	1.3224	1.6055





**Fig. 2** Paired moisture content and matric potential data shown with soil-water retention functions. *Field data best fit* curve was used in profile modeling (see section Discussion). *Pedotransfer function curve* (based on grain-size distributions in the profiles; mean percentages: sand=34.4, silt=50.3, clay=15.3; Schaap et al. 2001) was not used in modeling and is shown for comparison

salinity (Reedy and Scanlon 2003). In this study,  $EC_a$  is interpreted as a qualitative indicator of moisture content. Response of  $EC_a$  to soil moisture was established by comparison with laboratory measurements of gravimetric water contents, which showed that  $EC_a$  and water contents were correlated ( $\theta=0.33 EC_a+9.36$ ;  $r^2=0.50$ ;  $n=12$ ). Based on  $Cl^-$  concentrations which do not exceed 100 mg/L we assume that salinity has minimal effect on  $EC_a$ . Low correlation strength between  $EC_a$  and  $Cl^-$  in laboratory tests supports this assumption ( $r^2=0.15$ ;  $n=12$ ). Correlation between  $EC_a$  and clay content was not determined because of low variability in percent clay content in unsaturated-zone profile samples (13.1–19.8%).

Stable isotopes of water ( $\delta^{18}O$  and  $\delta^2H$ ) were used as environmental tracers to infer recharge mechanisms in the study area. Stable isotopes of unsaturated-zone pore water in semi-arid regions typically follow a characteristically low  $\delta^{18}O/\delta^2H$  slope imparted by evaporation in the shallow subsurface where there is a large kinetic fractionation effect (Barnes and Allison 1988; Clark and Fritz 1997). Groundwater which is primarily derived from diffuse unsaturated flow is influenced by this isotopic pattern. Because no wells were available in the study area, samples were collected from spring discharge and the unsaturated zone.

Vertical unsaturated-zone profiles were collected using hollow-stem hand augers (models SA5010C and SOS5010C, Dormer Engineering, Australia). Profile depths ranged from 6 to 20 m (Table 1). Bulk sediment samples of approximately 300 g were collected at depth intervals ranging from 0.125 to 1.0 m, and sealed in airtight containers to prevent evaporation. Moisture contents of sediment samples were measured gravimetrically after drying for 24 h at 105°C. Unsaturated-zone moisture was extracted for chemical analysis by thoroughly mixing 50 g of moist sediment with 30 ml of deionized water. The slurry was shaken for 4 h for complete mixing and elution

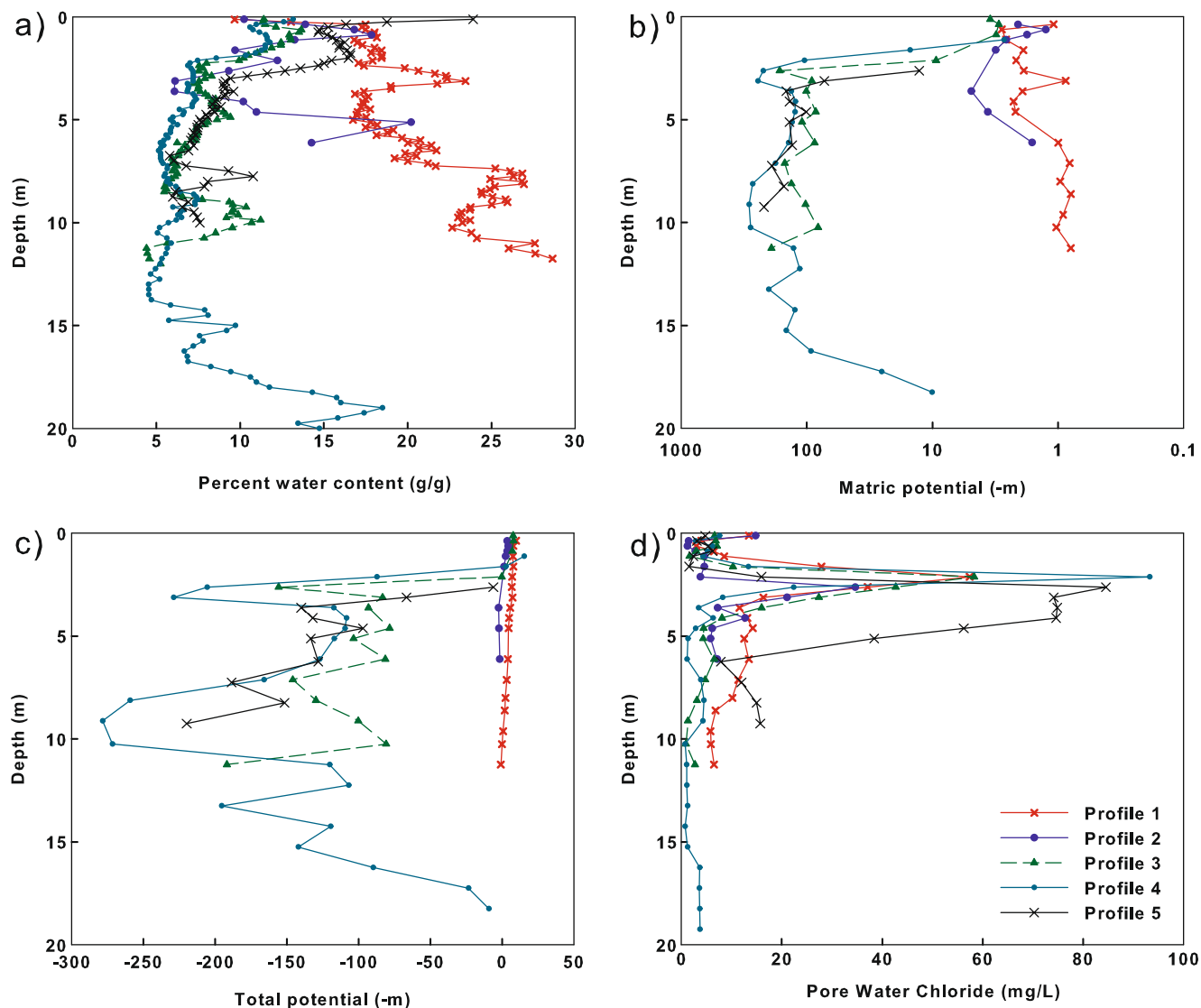
of soluble salts and then centrifuged. The supernatant was decanted and filtered (0.2- $\mu$ m pore diameter). Water for isotope analysis was extracted from the sediment samples with azeotropic distillation. Matric potentials were measured with laboratory tensiometers (Model T5, UMS, Munich, Germany) and dew point meters (Model WP4-T, Decagon Devices, Inc., Pullman, WA). Water samples from springs were collected in 30-ml HDPE bottles. Stable isotopes of water were measured with isotope ratio mass spectrometer (Thermo-Finnigan Delta XP, Reston Stable Isotope Laboratory, US Geological Survey) and results are expressed relative to Vienna Standard Mean Ocean Water (VSMOW) with  $2\sigma$  precision of  $\pm 2\%$   $\delta^{18}O$  and  $\pm 0.2\%$   $\delta^2H$ .  $Cl^-$  concentrations were determined using ion chromatography (Dionex IC-2100, Bureau of Economic Geology, University of Texas at Austin) with precision within 10% based on standards and replicate analyses.

## Results

Cultivated site profiles had higher depth-weighted mean moisture contents than plantations and forest (Fig. 3a). Moisture contents in profile 1 (terrace site) were approximately 10% near the surface and generally increased with depth. Profile 2 had high moisture contents near the surface and near the profile base ( $\theta_g=20\%$  at 5.1 m) and had a minimum of 6% at 3.6 m depth. Moisture contents in profile 3 and profile 4 had similar values and were more vertically uniform than profiles 1 and 2. Profile 5 had high near-surface moisture contents (up to 24% at 0.1 m) and lower moisture at depth. Moisture contents of individual samples ranged from 4% (3.0 m beneath the shrub plantation) to 19% (3.0, 7.9 and 8.5 m in the terrace site). Matric potentials were also significantly lower beneath natural forest, tree plantation and shrub plantation sites than cultivated sites (Fig. 3b). Near-surface moisture contents and matric potentials are likely to be highly variable in time in response to transient wetting and drying trends.

Below ~3 m depth, cultivated sites had matric potentials on the order of -1 m. Matric potentials in the cultivated terrace site ranged between -0.8 and -2.8 m and gradually decreased with depth. Matric potentials in the cultivated check-dam site were slightly more variable but within a similar range (-1.3 to -4.9 m). This contrasts with mean matric potentials of approximately -100 m beneath natural forest and tree and shrub plantations. In both plantation sites, matric potentials were  $> -5$  m in the near surface, but consistently  $< -100$  m below 2 m depth (profile minima -192 and -288 m for the tree and shrub plantations, respectively).

Mean  $Cl^-$  concentrations for the profiles (profiles 1–4) ranged between 0.6 and 2.6 mg/kg (6–14 mg/L; Fig. 3d).  $Cl^-$  in profile 5 (the forest site; 2.5 mg/kg; 26 mg/L) was also within this range on a mass basis but higher in terms of volumetric concentration. Each profile had a near-surface  $Cl^-$  peak which represents atmospheric-deposited solutes that were concentrated by evapotranspiration in the



**Fig. 3** Unsaturated-zone profile data including **a** water contents **b** matric potentials **c** total water potentials and **d** chloride concentrations. Profile settings and depths are described in Table 1

shallow subsurface and have not been flushed downward by incoming infiltration. Corresponding accumulation times may be estimated with  $\text{Cl}^-$  mass balance (Eq. 3). However, precision is strongly limited by uncertainties in  $\text{Cl}^-$  input fluxes (solute accumulation times and recharge rates are linearly related to  $\text{Cl}^-$  input rates). Therefore bounding calculations are presented based on differing  $\text{Cl}^-$  input flux assumptions (Table 3). Lower bounds for near surface accumulation time (upper bounds for profile mean

recharge rate) assumed  $\text{Cl}^-$  inputs from wet deposition, dry deposition, and net runoff of 100 mm/year with 1.4 mg/L  $\text{Cl}^-$ . Upper bounds for near surface accumulation time (lower bounds for profile mean recharge rate) assumed  $\text{Cl}^-$  inputs from wet deposition only and loss by net runoff of 100 mm/year with 1.4 mg/L  $\text{Cl}^-$ . Runoff/runon rates of 100 mm/year were selected as conservative upper boundary estimates based on published data from the region (Huang et al. 2003a, 2003b). Results corre-

**Table 3** Chloride-based recharge estimates using bounding assumptions

	Profile mean recharge rate (mm/year):			Near-surface solute accumulation time (years):		
	Lower bound	Wet deposition only	Upper bound	Lower bound	Wet deposition only	Upper bound
Profile 1	43	54	115	14	30	38
Profile 2	67	84	180	1	2	2
Profile 3	56	69	148	8	17	21
Profile 4	92	115	246	6	13	16
Profile 5	19	24	51	19	42	52

sponding to wet deposition only (no net runoff or runon) are also reported.

Assuming wet deposition of  $\text{Cl}^-$  only, near surface  $\text{Cl}^-$  accumulation times ranged from 2 years (profile 2) to 42 years (profile 5). Lower bound estimates range from approximately 1 to 19 years. Inputs from runon are likely at least in the case of profile 2 because it is in a topographic low, in which case the accumulation time may be near the lower bound of approximately 1 year. The  $\text{Cl}^-$  accumulation in profile 5 would have taken several years even with additional  $\text{Cl}^-$  inputs (for example, accumulation would take 19 years with wet deposition,  $700 \text{ mg/m}^2/\text{year}$  dry deposition, and  $100 \text{ mm/year}$  runon). Profiles 1, 3 and 4 (terrace and hillslope sites) each have several years of accumulation represented in near-surface  $\text{Cl}^-$  accumulations based on the same bounding assumptions.

Below the near-surface peaks,  $\text{Cl}^-$  concentrations become more vertically uniform. Using profile-averaged  $\text{Cl}^-$  values, bounding calculations for recharge rates ranged from a minimum of  $19 \text{ mm/year}$  (profile 5) to a maximum of  $246 \text{ mm/year}$  (profile 4). Estimates based on wet deposition range from  $24$  to  $115 \text{ mm/year}$ . Profile 1 (terrace site) and profile 5 (natural forest site) are not irrigated, and are not likely to receive significant runon or generate significant runoff. Therefore, atmospheric deposition is likely to represent the primary inputs for  $\text{Cl}^-$  mass balance, in which case their mean recharge rates would be  $54$  and  $24 \text{ mm/year}$ , respectively, assuming wet deposition only. Wet deposition may overestimate net  $\text{Cl}^-$  inputs to hillslope sites (profiles 3 and 4) if there was loss of atmospherically deposited  $\text{Cl}^-$  via runoff before infiltration (subsurface  $\text{Cl}^-$  concentrations in some of these samples are not much higher than rainfall). Lower bounds taking into account  $100 \text{ mm/year}$  of runoff for these sites are  $56$  and  $92 \text{ mm/year}$ . Because profile 2 is likely to receive runon, the recharge rate at this site may approach the upper bound estimate of  $180 \text{ mm/year}$ .

Transect  $\text{EC}_a$  values ranged from a minimum of  $34$  to a maximum of  $119 \text{ mS/m}$ . There was general consistency of  $\text{EC}_a$  values within each transect (coefficients of variation range from  $0.04$  to  $0.15$ ), suggesting that soil moisture contents are consistent across the field settings of the profile locations at the time of data collection. The transect data also suggest that moisture contents were uniformly higher within a non-vegetated terrace and a gully than in plantation sites (Fig. 4). This pattern is consistent with higher evapotranspiration losses from the plantations although the transect data provide no information on how persistent this pattern may be temporally. The gully transect had more variability in  $\text{EC}_a$  than the other four transects, which may be due to topographic effects or the effect of variable clay contents.

Stable isotope ( $\delta^2\text{H}$  and  $\delta^{18}\text{O}$ ) compositions of 10 springs were measured and compared with selected samples of unsaturated-zone moisture (Fig. 5). Individual springs were selected based on accessibility and to achieve a representative range of elevations within the catchment

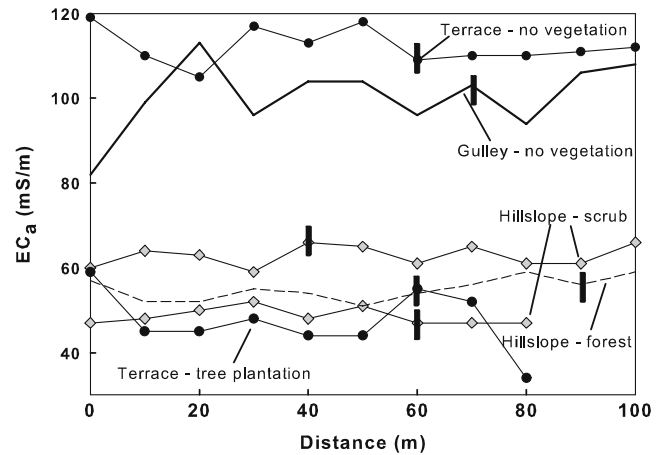
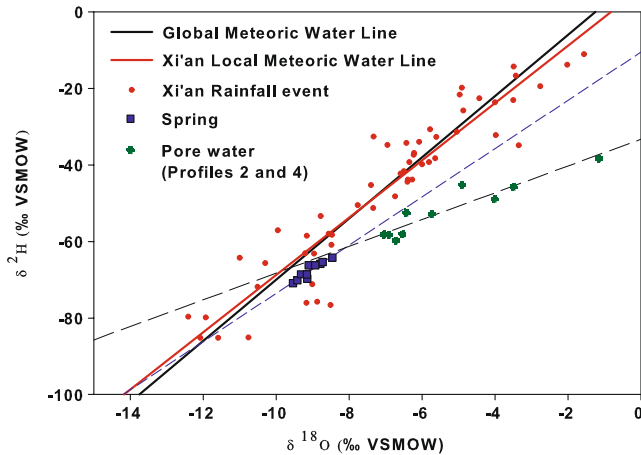


Fig. 4 Results of electromagnetic induction transects. Vertical bars denote sediment sampling locations

(spring elevations ranged from  $1,036$  to  $1,231 \text{ m}$  above sea level). The local meteoric water line is based on data from the nearest monitoring station in Xi'an, Shaanxi Province (1985–1992;  $\delta^2\text{H}/\delta^{18}\text{O}$  slope =  $+7.5$ ; IAEA/WMO 2010). Spring water isotope values follow an evaporation enrichment line with  $\delta^2\text{H}/\delta^{18}\text{O}$  slope of  $+6.3$ . The intercept between the regression line for the springs and the local meteoric line is approximately  $-12\%$   $\delta^{18}\text{O}$ . The regression line for the unsaturated-zone samples has a slope of  $+3.5$ , consistent with evaporative enrichment in the soil zone, and local meteoric line intercept of approximately  $-10\%$   $\delta^{18}\text{O}$ .

## Discussion

Unsaturated-zone  $\text{Cl}^-$  concentrations and matric potentials suggest that deep drainage is occurring beneath cultivated agricultural locations. Profiles 1 and 2 beneath cultivated land use had low matric potentials and downward gradients in total water potential that are consistent with gravity-driven flow (Fig. 3b and c).  $\text{Cl}^-$ -based recharge estimates using wet deposition data correspond to  $11$ – $18\%$  of mean annual precipitation. This range overlaps with ratios reported for other sites in the Loess Plateau ( $12$ – $13\%$  from Lin and Wei 2006;  $6$ – $12\%$  from Huang and Pang 2010), suggesting that this is a reasonable range of recharge to rainfall ratios in semi-arid areas of the Loess Plateau under dryland agriculture.  $\text{Cl}^-$  concentrations in the forest and plantation profiles (profiles 3, 4 and 5) were similar to those from cultivated sites.  $\text{Cl}^-$  mass balance suggests that this  $\text{Cl}^-$  concentration range corresponds to deep drainage rates of  $24$ – $115 \text{ mm/year}$  (using wet deposition data; Table 3). However, matric potentials beneath forest and plantation sites were significantly lower than the cultivated sites and had no obvious vertical pattern in total water potential gradients (Table 3 and Fig. 3b and c). The pore-water matric potentials beneath forest and plantation sites are not consistent with gravity-driven drainage, and suggest much lower water fluxes overall.



**Fig. 5**  $\delta^2\text{H}$  and  $\delta^{18}\text{O}$  plot showing data from **a** 10 springs, **b** unsaturated-zone pore water, **c** rainfall stable isotope data for the nearest monitoring station in Xi'an, Shaanxi Province (1985–1992) and **d** the global meteoric water line (GMWL). Local meteoric water line (LMWL) is defined for Xi'an data

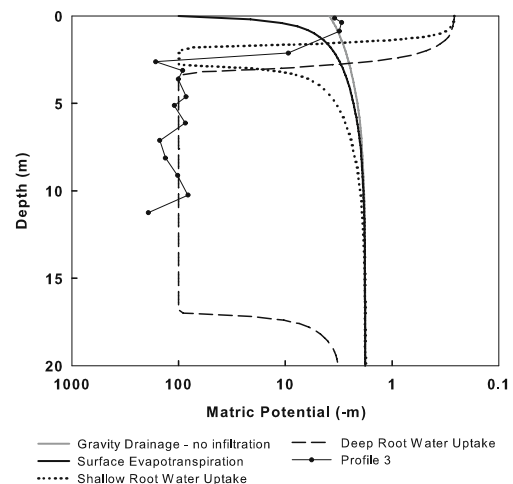
The co-occurrence of low  $\text{Cl}^-$  concentrations (indicating that the moisture was subject to relatively low solute evaporative concentration) and low matric potentials (indicating dry conditions and low advection rates) suggests that the infiltration of low- $\text{Cl}^-$  moisture preceded initiation of the low matric potential conditions beneath Zhifangou's plantations. Historical water-content data are not available to determine when profile drying initiated or how long it took to reach its current conditions. However, some constraint on the timing is provided by near-surface  $\text{Cl}^-$  accumulation peaks (Table 3), which indicate that solute accumulation times are on the order of 6–21 years beneath plantations sites (and up to 52 years beneath the forest site). The accumulation times suggest that drying most likely took place after the conversion of cropland to plantations (late 1970s for profiles 3 and 4). This is consistent with previous studies which show that soil-water desiccation has been a common consequence of plantations increasing net evapotranspiration when they replace agricultural land use or bare ground in the Loess Plateau (Chen et al. 2008a, 2008b; Wang et al. 2008).

Unsaturated flow and transport modeling was used to test the hypothesis that the plantations were capable of drying the profiles to field matric potential conditions ( $-100$  m) to a depth of  $\sim 17$  m within the time span of their 30-year existence. Unsaturated-zone drying beneath plantations was simulated under differing boundary condition assumptions, and the rates of drying were compared with field data from profile 3 (12 m). The simulations compared four constant-boundary scenarios, which are intended to provide a feasibility-level investigation of profile drying patterns following land cover change, rather than realistic prediction.

Simulation 1 represents profile evolution after an initial infiltration rate of 50 mm/year is reduced to 0 mm/year in the absence of any root-water uptake impacts, in order to provide a baseline case for comparison with root-water uptake scenarios. Initial moisture contents were specified using final moisture contents after 50 mm/year infiltration

through a 40-m profile until steady state was reached. (Infiltration rates of 10 and 100 mm/year were also tested to ensure that results were not affected by initial conditions.) Zero infiltration was simulated using a constant (zero) flux upper-boundary condition. No root-water uptake is included in this simulation. At steady state, moisture contents in simulation 1 approach the  $\theta_R$  value specified in the water-retention function ( $0.06 \text{ cm}^3/\text{cm}^3$ ; Table 2). Because no root-water uptake is simulated, simulation 1 represents an upper bound for water contents in a loess profile experiencing no infiltration or deep drainage. Simulations 2 and 3 were undertaken to evaluate the rate of drying front propagation over a 30-year period assuming that deep drainage is reduced to zero and a high rate of evapotranspiration extracts progressively deeper pore water over time. Simulation 2 had identical initial conditions to simulation 1 but included water removal by evapotranspiration at the soil surface using a constant head (suction) upper-boundary condition ( $-100$ -m constant upper-boundary value). Simulation 3 includes evapotranspiration within a root zone of 0–3 m depth that has a Gaussian vertical root density distribution. The constant potential evapotranspiration rate was specified to be ten times the study area's precipitation rate of 500 mm/year as an upper bound on the actual rate. Simulation 4 included deep-root water uptake (arbitrarily set equivalent to 17 m, which is the maximum depth of the low-moisture-content zone in profile 4) and was otherwise identical to simulation 3. In the absence of reliable root-water uptake information, the wilting point was considered a calibration parameter (adjusted to  $-100$  m to correspond with profile 4).

Simulation 1 demonstrates that restriction of infiltration exclusive of soil-zone water extraction cannot account for the moisture contents and matric potentials measured in profile 4 (Fig. 6). Gravity drainage yields final profile conditions that are significantly wetter than field data beneath the plantations (matric potential minimum =  $-4$  m). Simulations 2 and 3, which involve net discharge by evapotranspiration, demonstrate that



**Fig. 6** Modeled matric potential results for 30 year simulations using water retention parameters based on best fit to field data. Profile 3 data shown for comparison



the rate of drying front propagation is too slow to explain field profile water contents over the 30-year simulation. Drying to the measured field moisture conditions after 30 years is achieved only at the location of water loss, which is an artifact of the choice of wilting point values. Propagation of the drying front takes hundreds of years to reach 17 m, which is inconsistent with  $\text{Cl}^-$  data because a larger shallow bulge would be present if drainage was prevented for hundreds of years. Only simulation 4, in which root water uptake is permitted throughout the profile, sufficiently depletes water contents within the 30 year time period.

The modeling results suggest that drying to several meters depth would not be feasible within the time span of the Zhifangou plantations without root-water uptake >10 m. Assuming that uncertainties in water-retention function parameters are not responsible for major underestimation of drying front propagation rates, the results suggest that either the deep moisture desiccation predated the plantations, or that water uptake by tap roots of plantation vegetation is an important process in these settings.  $\text{Cl}^-$  mass balance peak accumulate times (maximum 21 years in plantation profiles) suggest that the desiccation did not predate the plantations. Rather, deep drying trends to ~10 m have been documented on similarly short timescales previously in the Loess Plateau (Li et al. 2008). As a result, it is suggested that effective root-water extraction of some introduced species such as *C. Korshinski* may exceed 10 m in some cases in the Loess Plateau, although this is deeper than has been previously suggested (Cheng et al. 2009; Li et al. 2004). Improved root physiological information under field conditions is desirable to evaluate this phenomenon in more detail.

Although inferences from vertical profiles are inherently limited because they are point-based and therefore difficult to generalize (Perkins et al. 2011), the profile data are considered likely to be reasonably indicative of field scale unsaturated-zone conditions beneath similar vegetation types. This is an assumption supported by (1) similar silt-loam soil textures measured across the study area and (2) consistency of  $\text{EC}_a$  values within EM induction transects, which suggests that soil-moisture contents are broadly consistent across the field settings of the profile locations (Fig. 4). Based on soil moisture declines extensively documented beneath plantations in the Loess Plateau (Fu et al. 2003; He et al. 2003; Li 1983; Wang et al. 2008, 2010), the decline in deep drainage shown in profiles 3 and 4 may be widespread. This phenomenon may have water-resource implications because reductions in deep drainage will ultimately reduce diffuse recharge. In upland areas with thick unsaturated zones and deep water tables, long transit times in the unsaturated zone mean that impacts of deep drainage reductions on the water table may be significantly lagged in time, so that full impacts of land-use change may be delayed by decades or more. However, discharge from high-elevation springs could also be affected and result in a more immediate impact on streamflow.

In addition to diffuse recharge through the unsaturated zone, recharge mechanisms may also include focused

(indirect) recharge through stream channels or topographic depressions. Focused recharge is considered a potentially important process especially considering the steep topography and extensive runoff generation in the Loess Plateau. The clear distinction between stable isotopic signatures of spring water and unsaturated-zone pore water suggests that drainage through the unsaturated zone is contributing little volume to spring discharge in the Zhifangou Catchment. Significant contributions from unsaturated-zone drainage on spring discharge would produce a lower  $\delta^2\text{H}/\delta^{18}\text{O}$  regression slope in spring waters. Rather, the spring water  $\delta^2\text{H}/\delta^{18}\text{O}$  slope is more indicative of evaporation from surface water, and therefore more consistent with indirect recharge sources.

The meteoric water line intercept (approximately  $-12\text{‰}$ ) is significantly lower than the mean annual rainfall data from the Xi'an record (approximately  $-8\text{‰}$ ). One reason for the depleted meteoric water line intercept may be depletion of rainfall in Zhifangou relative to Xi'an, considering that Zhifangou has an elevation 600–800 m higher than Xi'an (400 m above sea level). A  $4\text{‰}$   $\delta^{18}\text{O}$  depletion due to the 600–800 m elevation difference (or 0.50–0.67‰ per 100 m) and more northern location (300 km north) is possible but local rainfall isotopic data is needed for confirmation. Seasonal bias in recharge timing could be another potential factor affecting the depleted meteoric water line intercept. Xi'an monthly mean for July is approximately  $-12\text{‰}$ , and this corresponds with the monsoon rainfall season. The depleted isotopic signature is also consistent with isotopic compositions of paleowaters in northern China (Currell et al. 2010; Gates et al. 2008b; Zongyu et al. 2003) but groundwater dating would be required to invoke paleowater contribution in this case in view of the uncertainties in local meteoric water inputs.

## Conclusions

Unsaturated-zone results suggest that land-use impacts on diffuse recharge strongly differ between ecological and structural approaches to soil conservation in the study area. Results suggest that cultivated agricultural land use maintains deep drainage rates of approximately 11–18% of mean annual rainfall. In contrast, tree and shrub plantations have restricted deep drainage and depleted residual unsaturated-zone pore water within the past 20–30 years. The  $\text{Cl}^-$ -based recharge rate estimates beneath the plantation sites (profiles 3 and 4; Table 3) should be considered representative of pre-plantation conditions only because the matric potential data are inconsistent with that range of drainage rates occurring presently. Stable isotope signatures of unsaturated-zone moisture and springs indicate that diffuse recharge presently contributes little to spring flow in the study area. Focused infiltration through gullies and other topographic lows is likely to be the primary recharge mechanism. While the Zhifangou Catchment study area has been subject to extensive plantations as a means of soil conservation, and so not

necessarily representative of other areas of the Loess Plateau in terms of recharge rates and mechanisms, an earlier field study found similar agricultural recharge rates and indications that indirect recharge sources may be important (Lin and Wei 2006).

The results of this study emphasize the potential for impacts of soil conservation on groundwater resources, which is relevant globally considering the scale of agricultural soil loss. For the Loess Plateau, they also indicate that there is opportunity for achieving better balance between erosion mitigation and water-resource sustainability goals by diversifying conservation approaches. Ecological soil-conservation approaches have benefits of cost efficiency and carbon storage, but may not be sustainable where evapotranspiration losses are high. Semi-arid areas can expect diminished water tables and stream flows where non-native trees and shrubs are planted extensively, so structural soil conservation or less water-intensive vegetation could be emphasized where practicable. More humid areas may be able to better withstand the added transpiration losses so these areas may be able to integrate agro-ecological approaches to a greater extent. There is a need for monitoring recharge and groundwater storage in the Loess Plateau in order to better assess land-use impacts on water resources of the region.

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