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Does hillslope trenching enhance groundwater recharge and baseflow in the Peruvian Andes?

Lauren D. Somers[1] | **Jeffrey M. McKenzie**[1] | **Samuel C. Zipper**[1,2] | **Bryan G. Mark**[3] | **Pablo Lagos**[4] | **Michel Baraer**[5]

[1] Department of Earth and Planetary Sciences, McGill University, 3450 University Street, Montreal, QC H3A 0E8, Canada

[2] Department of Civil Engineering, University of Victoria, PO Box 1700 STN CSC, Victoria, BC V8W 2Y2, Canada

[3] Byrd Polar and Climate Research Centre, The Ohio State University, 108 Scott Hall, 1090 Carmack Rd, Columbus, OH 43210, USA

[4] Instituto Geofísico del Perú, Calle Badajoz #169, Mayorazgo IV Etapa, Ate Vitarte, Lima, Peru

[5] Départment de génie de la construction, École de technologie supérieure, 1100 rue Notre-Dame Ouest, Montreal, QC H3C 1K3, Canada

Correspondence
Lauren D. Somers, Department of Earth and Planetary Sciences, McGill University, 3450 University Street, Montreal, QC, Canada, H3A 0E8.
Email: Lauren.Somers@mail.mcgill.ca

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Abstract

As Andean glaciers rapidly retreat due to climate change, the balance of groundwater and glacial meltwater contributions to stream discharge in tropical, proglacial watersheds will change, potentially increasing vulnerability of water resources. The Shullcas River Watershed, near Huancayo, Peru, is fed only partly by the rapidly receding Huaytapallana glaciers (<20% of dry season flow). To potentially increase recharge and therefore increase groundwater derived baseflow, the government and not-for-profit organizations have installed trenches along large swaths of hillslope in the Shullcas Watershed. Our study focuses on a nonglacierized subcatchment of the Shullcas River Watershed and has 2 objectives: (a) create a model of the Shullcas groundwater system and assess the controls on stream discharge and (b) investigate the impact of the infiltration trenches on recharge and baseflow. We first collected hydrologic data from the field including a year-long hydrograph (2015–2016), meteorological data (2011–2016), and infiltration measurements. We use a recharge model to evaluate the impact of trenched hillslopes on infiltration and runoff processes. Finally, we use a 3-dimensional groundwater model, calibrated to the measured dry season baseflow, to determine the impact of trenching on the catchment. Simulations show that trenched hillslopes receive approximately 3.5% more recharge, relative to precipitation, compared with unaltered hillslopes. The groundwater model indicates that because the groundwater flow system is fast and shallow, incorporating trenched hillslopes (~2% of study subcatchment area) only slightly increases baseflow in the dry season. Furthermore, the location of trenching is an important consideration: Trenching higher in the catchment (further from the river) and in flatter terrain provides more baseflow during the dry season. The results of this study may have important implications for Andean landscape management and water resources.

KEYWORDS

Andes, baseflow, enhanced recharge, groundwater recharge, hillslope trenching, infiltration trenches

1 | INTRODUCTION

Mountain regions play an important role in global water supply and are highly sensitive to climate change (Barnett, Adam, & Lettenmaier, 2005; Bradley, Vuille, Diaz, & Vergara, 2006; Viviroli, Dürr, Messerli, Meybeck, & Weingartner, 2007; Viviroli et al., 2011; Rangwala & Miller, 2012). In the Peruvian Andes, communities and industries along the cordillera and on the arid coast depend on alpine watersheds for water resources. Glacial meltwater and stored groundwater supply consistent stream discharge during the dry season when precipitation is minimal (Baraer et al., 2012; Baraer, McKenzie, Mark, Bury, & Knox, 2009; Bury et al., 2013; Mark, McKenzie, & Gómez, 2005).

In Peru, glaciological and hydrological research in the Cordillera Blanca has explored the rapid recession of glaciers (Georges, 2004; Mark & Seltzer, 2005; Schauwecker et al., 2014) and their threat to water resources (Baraer et al., 2012; Mark, Bury, McKenzie, French, & Baraer, 2010), as well as the mediating influence and importance of groundwater discharge to alpine streams (Baraer et al., 2009; Gordon et al., 2015; Somers et al., 2016).

Similarly, the smaller and less studied Cordillera Huaytapallana in the central Peruvian Andes has undergone extensive glacial recession in recent decades (Instituto Geofísico del Perú, 2010; Autoridad Nacional del Agua [ANA] Ministerio de Agricultura y Riego, 2014; López-Moreno et al., 2014). Meltwater from the Huaytapallana glaciers feeds the Shullcas River, which in turn provides municipal water to the city of Huancayo and irrigation water to local agricultural operations (ANA Ministerio de Agricultura, 2010). Previous work has indicated that glacial melt accounted for less than 20% of dry season discharge in 2014 (Crumley, 2015), with the remainder coming from groundwater discharge to the Shullcas River. As the Huaytapallana glaciers continue to retreat, dry season stream discharge is expected to decrease, making groundwater discharge an increasingly significant source of water for this economically and socially important watershed.

Although dams or reservoirs are effective and are typically constructed in response to this type of seasonal water shortage, they can also be costly, induce evaporation, and be ecologically harmful (Nilsson, Reidy, Dynesius, & Revenga, 2005). Increasing groundwater recharge in times of excess surface water supply is one alternative method for increasing dry season water availability. Several different schemes have been proposed and are referred to in general as artificial aquifer recharge or managed aquifer recharge (MAR). Although some systems divert excess river water to injection wells (Bouwer, 2002, and references therein; Dillon, 2005), others divert excess water to infiltration fields, channels, or basins (Heilweil, Benoit, & Healy, 2015; Mastrocicco, Colombani, Salemi, Boz, & Gumiero, 2016; Heviánková, Marschalko, Chromíková, Kyncl, & Korabík, 2016).

The Peruvian government and not-for-profit organizations have installed infiltration trenches over large swaths of hillslope in the Shullcas River Watershed. Unlike most MAR schemes, the purpose of these trenches is to capture surface runoff (as opposed to diverted river water) during the rainy season and allow more time for infiltration. In theory, this passive aquifer recharge scheme could increase recharge to groundwater, thereby increasing groundwater baseflow to the river during the dry season with the additional benefit of reducing erosion (CARE Peru, 2013). It has been suggested that channels known as mamanteo were used in a similar manner by pre-Incan peoples in the Andes, but would drain into infiltration ponds (Bardales, Barriga, Saravia, & Angulo, n.d.; Fraser, 2015). However, direct measurement of recharge is challenging, and despite the large amount of resources invested in these types of projects, little research has been done to determine the effectiveness of this strategy.

Techniques for measuring groundwater recharge include use of lysimeters, tracers, and water table fluctuations (Scanlon, Healy, & Cook, 2002). However, these techniques are not easily applied to a hillslope scale application. Furthermore, recharge can be divided into distinct processes that occur heterogeneously in time and space and may not be adequately captured by point measurements. Meixner et al. (2016) suggest a four-fold classification of recharge processes: diffuse, focused, mountain system recharge, and irrigation. The addition of infiltration trenches essentially changes the proportions of diffuse and focused recharge. As direct measurement of groundwater recharge is difficult, researchers often rely on various estimation techniques, including numerical modelling of hydrological processes at the land surface and the unsaturated zone (Scanlon et al., 2002).

Furthermore, it is hypothesized that a change in recharge caused by the installation of trenches should affect groundwater baseflow to the stream. However, mountain hydrogeological systems are highly heterogeneous and still relatively poorly understood (e.g., Clow et al., 2003; Roy & Hayashi, 2009; Harpold, Lyon, Troch, & Steenhuis, 2010). Therefore, the impacts on the annual stream hydrograph are unknown.

The objectives of this paper are to better understand and quantify the groundwater flow system in the Shullcas River Watershed and to determine if and how infiltration trenches increase groundwater discharge during the dry season. We incorporate surface trenching into an infiltration and recharge model in order to estimate the difference in groundwater recharge between trenched and nontrenched terrain. The model is driven by high-frequency meteorological data and incorporates measurements of soil infiltration capacity and observations on vegetation and trench configuration. We then apply the resulting recharge rates to a groundwater model of a study catchment to estimate the change in quantity and timing of groundwater baseflow to the stream. Sensitivity analysis is performed on both models to quantify uncertainty and determine what conditions are favourable for infiltration trenches.

2 | STUDY AREA

The Shullcas River Watershed is a high-altitude proglacial watershed located partially within the Huaytapallana Conservation Area in the Cordillera Central and is a tributary to the Mantaro River in the Amazon Basin (Figure 1, inset). The river provides the city of Huancayo, Junín Region, central Peru (latitude ~ 12.1°S, longtitude ~ 75.2°W, population 466,000), with 60% of its municipal water as well as irrigation water for local agricultural projects (Crumley, 2015). Due to these diversions, the Shullcas River runs dry or almost dry before it reaches the city of Huancayo during the dry season. Average annual precipitation in the Shullcas Basin is approximately 800 mm and varies with elevation (ANA Ministerio de Agricultura, 2010). Precipitation is highly seasonal, with most of the annual precipitation during the rainy season from October to April. Conversely, air temperature stays almost constant throughout the year.

The watershed is mainly composed of steep alpine grasslands with some bedrock outcropping, flatter hummocky wetlands known as bofedales (see Maldonado Fonkén, 2015), and valley bottom alpine meadows known as pampas. Livestock grazing takes place throughout the study catchment. Aerial photography and satellite imagery have shown a 55% decrease in glacial area from 1984 to 2011 (López-Moreno et al., 2014). Additionally, Crumley (2015) performed a hydrochemical mass balance analysis for the watershed (using the hydro-chemical basin characterization method (HBCM) from Baraer et al., 2009) and found that glacier meltwater contributed approximately 9–16% of dry season stream discharge in 2014.

Between 2009 and 2012, the Peruvian Ministry of Agriculture, in collaboration with nongovernment organizations CARE and the World Bank, undertook a project entitled “Adaptation to the Impact of Rapid Glacier Retreat in the Tropical Andes,” known in Spanish as PRAA. Among other components, this project included the excavation of infiltration trenches in the Shullcas Watershed where 800 ha of land was

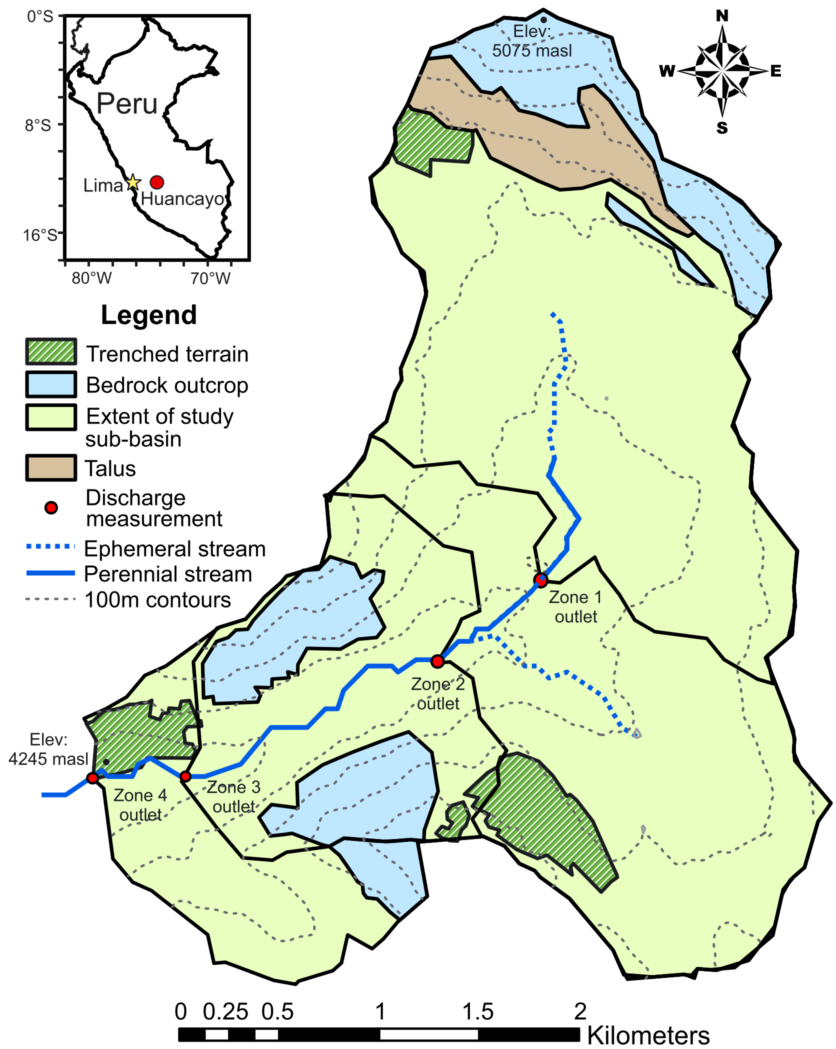

FIGURE 1 Map of study catchment within a nonglacierized basin of the Shullcas Watershed. The ephemeral streams shown are not included in the groundwater model as river nodes

covered in trenches (CARE Peru, 2013). Local communities were employed to manually dig trenches in several areas of the catchment. The trenches are trapezoidal in shape, roughly 30 cm deep, 40 cm wide at the top, and are spaced 9–10 m apart on steep grassy hillslopes (Figure 2).

In order to isolate the hydrologic influence of trenches from glacier melt, a nonglacial subbasin of the Shullcas River Watershed was chosen for this study (Figure 3). Table 1 shows total, trenched, and bedrock outcrop area of the study catchment, as determined from 2014 satellite imagery (Landsat imagery accessed through Google Earth™).

3 | METHODS

Our study employs both field and numerical modelling approaches to explore the recharge and hydrogeological regime in the study catchment.

3.1 | Data collection

Field data were collected between 2011 and 2017. In August 2016, stream discharge was measured at four locations along the stream (shown in red on Figure 1) to correlate stream discharge with the contributing trenched area. The hypothesis to be tested with the discharge measurements is that zones with more trenched area should contribute more baseflow to the stream during the dry season. (It should be noted that differential gauging can have large errors in estimating groundwater inflows; e.g., Briggs, Lautz, & McKenzie, 2012).

Infiltration capacity was measured with a double ring infiltrometer during the dry season (August 2016) and the rainy season (March 2017) at a variety of sites in the Shullcas Watershed. The infiltrometer had an outer ring diameter of 40 cm and an inner ring diameter of 25 cm. At each test site, the infiltrometer was inserted 5 cm into the ground, and infiltration measurements were carried out as described in Bodhinayake, Si, and Noborio (2004), mostly reaching steady state between 25 and 60 min.

High-frequency meteorological data were obtained from SENAMHI, the Peruvian national weather service, from January 2011 to June 2014. These data include precipitation, air temperature, humidity, wind speed, and shortwave and longwave radiation at 30-min intervals from the Lazo Huntay automatic weather station, located less than 2 km outside of the study catchment. The high frequency of these meteorological observations allow us to take into account the intensity of precipitation, not just the total precipitation amount. From July 2015 to August 2016, daily precipitation measurements for the Lazo Huntay station were obtained from

FIGURE 2 (a) Construction of infiltration trenches as part of the World Bank- and CARE-funded PRAA project. Photo taken from Comunidad Andina presentation. (b) Infiltration trench with ponded water. (c) Aerial view of trenched hillslope in the Shullcas Watershed, from Google Earth™

SENAMHI's online database (http://www.senamhi.gob.pe/?p=data-historica; visited October 10, 2016).

Because the Lazo Huntay weather station is located at an elevation of approximately 4,650 masl and most of the study catchment is between 4,200 and 4,700 masl with a midpoint of 4,450 masl, the temperature was corrected using the dry adiabatic lapse rate of the atmosphere (1 °C per 100 m × 200 m elevation difference), adding 2 °C over the entire record.

Stream discharge was recorded from July 2015 to August 2016 at 15-min intervals at our gauging station located approximately 700 m downstream of the outlet of the study catchment using a Solinst LTC Levelogger and Barrologger. A correction was applied to the discharge data to compensate for the difference in gauging location such that the study area outlet discharge is 82% of the discharge recorded at the gauging station, based on simultaneous discharge measurements at the two sites.

3.2 | Recharge model

We created and applied an uncalibrated, one-dimensional infiltration model, which is used as input to MODFLOW's Unsaturated Zone Flow (UZF) package to estimate recharge to the groundwater system (Figure 3). For each 30-min time step, incoming precipitation may first be intercepted by vegetation up to the available interception storage. Intercepted water stored on the vegetation is evaporated over time, creating space for interception in future time steps. Precipitation that exceeds the available interception capacity reaches the ground surface and infiltrates at a rate less than or equal to the infiltration capacity. Runoff is generated when the precipitation that reaches the ground surface exceeds infiltration capacity, as follows:

$$R = P - IS_f - I, \tag{1}$$

where R is runoff, P is precipitation, IS_f is available interception storage, and I is infiltration, all expressed as depth of water per unit area. In the

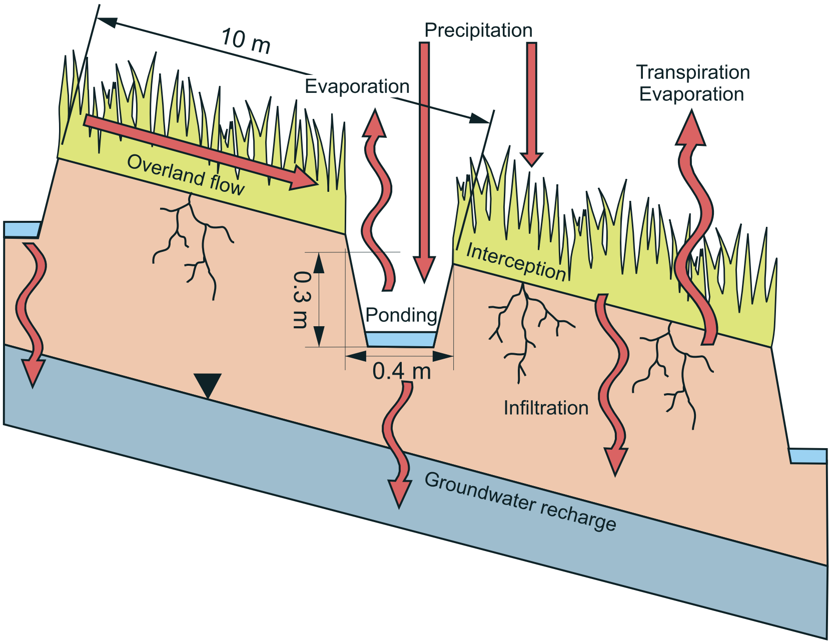

FIGURE 3 Recharge model schematic. Red arrows indicate movement of water through the relevant hydrological processes

TABLE 1 Basin and subbasin properties and discharge

Zone	Zone area (km^2)	Trenched area (km^2)	Outcrop area (km^2)	Stream discharge change (total) Aug 2016 (L/s)
1	4.39	0.09 (2%)	0.59 (13%)	14 (14)
2	2.85	0.23 (8%)	0 (0%)	0 (14)
3	2.02	0.03 (1.5%)	0.66 (33%)	4 (18)
4	1.14	0.14 (12%)	0.12 (10%)	5 (23)
Total	10.40	0.49 (5%)	1.37 (13%)	23

nontrenched, base case scenario, runoff is lost to the stream and therefore eliminated from the model domain. In the trenched scenario, runoff is ponded in ditches where it is evaporated and infiltrated according to

$$\frac{PV}{W_t} = P + R \times W_n - I - E, \tag{2}$$

where PV is volume of ponded water (per unit section of hillslope), W_t is the width of the trench, W_n is the width of hillslope between trenches, and E is evaporation from the surface of the ponded water. Potential evaporation from the ponded water in trenches is calculated using a Dalton-type equation for open water evaporation (Dingman, 2002, Eqs. 7–18a) and potential evapotranspiration from the grassland is calculated using the Penman–Monteith formulation as outlined by Dingman (2002, Eqs. 7–56).

The MODFLOW UZF package is then used for conditions representing an idealized hillslope to calculate recharge from infiltration. In the UZF package, a maximum evapotranspiration (ET) rate is assigned at the ground surface and decreases linearly with depth to an assigned extinction depth. Infiltrated water moves from the ground surface towards the water table according to a kinematic wave approximation of the Richards equation (Niswonger, Prudic, & Regan, 2006). The UZF package also requires a vertical hydraulic conductivity, and Brooks–Corey epsilon exponent value (Niswonger et al., 2006). Input parameters for the UZF package are calculated or estimated on the basis of field observations when possible or sourced from literature otherwise (Table 2). The UZF module is applied to a two-dimensional, 100-m wide, idealized hillslope that is representative of catchment terrain so that the results represent average recharge rate over the entire slope. More details on the UZF domain set-up are included in the Supporting Information.

This coupled recharge model is driven by 3.5 years of high-temporal-resolution meteorological data (including precipitation, temperature, solar radiation, longwave radiation, humidity, and wind speed) to estimate the proportion of precipitation that becomes groundwater recharge for two scenarios: the base case (no-trenching) and trenched case.

3.3 | Groundwater model

Once the net recharge for the base case and trenched scenarios is determined, a three-dimensional groundwater model of the study catchment is used to evaluate the impact of trenching on groundwater baseflow to the stream. The ASTER Global Digital Elevation Model with 30-m resolution (https://asterweb.jpl.nasa.gov/gdem.asp; Tachikawa et al., 2011) was used to define the model domain, and 2014 satellite imagery accessed through Google Earth™ was used to define the location of trenched hillslopes and bedrock outcrop.

The groundwater model was constructed using MODFLOW-NWT (Niswonger, 2011) and uses two MODFLOW Packages: the Recharge Package (RECH) and River Package (RIV). The surface of the model has 1,034 square grid cells with a side length of 100 m representing an area of 10.34 km^2. The model has two layers that are each discretized into two levels, for a total of 4,136 grid cells (4 × 1034). The top layer of the model is unconfined and represents unconsolidated silty sand and gravel glacial and fluvial deposits and has a depth between 1 and 30 m. Surficial layer depths are based on topography, satellite imagery of rock outcrops, and field observation of sediment depth, such as road cuts. The surficial layer is considered isotropic

TABLE 2 Model input parameters

	Value	Source
Infiltration input variable		
Infiltration capacity	0.2 m/day	Rainy season double ring infiltrometer measurements
Interception storage	1 mm	Burgy & Pomeroy, 1958; Dunkerley & Booth, 1999; Zou, Caterina, Will, Stebler, & Turton, 2015.
Width, spacing of trenches	0.4 m, 9 m	Field measurement, CARE Peru, 2013
Potential evapotranspiration time series	0–16 mm/day (average 1 mm/day)	Calculated using Penman–Monteith formulation. Eqs. 7–18a in Dingman, 2002
Open water evaporation	0–19 mm/day average (0.8 mm/day)	Calculated using Dalton-type equation. Eqs. 7–56 in Dingman, 2002
Recharge input variable		
Maximum ET	1 mm/day	Average of ET time series
ET extinction depth	2 m	Shah, Nachabe, & Ross, 2007 (approximate value for sandy loam with grass cover)
ET extinction water content	0.05	US Department of Agriculture, 1955 (approximate wilting point for sandy, silty loam)
Vertical hydraulic conductivity of the unsaturated zone	1 m/day	Table 3.7 in Fetter, 2001 (order of magnitude estimation for sandy, silty loam)
Brooks–Corey epsilon	3.5	Niswonger et al., 2006
Saturated water content	0.3	Table 3.4 in Fetter, 2001. Estimation for porosity of silty soil

because of the coarse and poorly sorted nature of the deposits. The bottom layer represents fractured bedrock and is simulated as confined to help with model convergence. Rock below 60 m in depth is considered impermeable. Bedrock fracture flow is considered isotropic in absence of detailed bedding information and to simplify calibration. An orthographic view of the model domain is included in Figure S2.

The model is run with daily time steps. The initial conditions are set by a steady-state spin-up that uses long-term average conditions followed by 1 year of transient spin-up to allow the model to reach dynamic equilibrium. Then the model is run for a period of 428 days between July 1, 2015, and August 31, 2016.

Because of limitations in the available data, the high-temporal-resolution meteorological data do not overlap with the stream discharge record, and only a daily precipitation record was available during this time period. Therefore, the overall relationship observed between precipitation and recharge for the high-temporal-resolution run of the recharge model is recreated empirically for the 2015–2016 period, preserving the overall recharge amount as a percentage of precipitation.

Ten-day intervals were used to average the relationship between amount of precipitation and recharge during the recharge model run. These coefficients were then applied to the 2015–2016 precipitation data to produce a recharge data set, also with 10-day intervals. The recharge data set was adjusted manually to smooth outliers resulting from periods of unusual or extreme weather in the 2011–2014 data, and lag time between precipitation and recharge was reduced to improve model fit. The average proportions of precipitation that becomes recharge for the base case and trenched case, as found in Section 3.2, were preserved over the entire model period and are used along with the daily precipitation record to create two daily recharge time series for the period from July 2015 to August 2016.

The base case and trenched case recharge time series are applied to the appropriate model area according to the presence or absence of trenching. The model output, groundwater baseflow to the stream, is calibrated to dry season stream discharge. Three calibration variables are used to achieve the best fit: hydraulic conductivity of the surficial layer, specific yield of the surficial layer, and hydraulic conductivity of the fractured bedrock layer.

A simple surface runoff component is added to the baseflow hydrograph in order to compare with the observed stream discharge. Precipitation depth is multiplied by the area of the catchment, combined into 3-day intervals and lagged by 2 days to smooth the runoff response. Several different intervals and lag times were tried to optimize the fit. The percentage of precipitation becoming runoff is then varied to find the best fit to the observed hydrograph. This simple runoff estimate is used because the goal is to roughly compare to the hydrograph and not to investigate in detail the precipitation-runoff response. Furthermore, the dry season discharge, which is of particular interest, is dominated by baseflow.

The ZONEBUDGET and MODPATH modules are used to postprocess the MODFLOW-NWT model results. ZONEBUDGET (Harbaugh, 1990) is used to analyse the water budget entering and leaving different zones of the model. MODPATH (Pollock, 2012) is used to analyse flow paths and travel times through the subsurface. This module tracks particles of water from some assigned launch point until they exit the model domain, in our case, through the river. Ten sites were randomly selected throughout the catchment, and two particles were released at different depths at each site, vertically distributed between the top-most active layer and the model bottom. The flow paths and travel times were analysed for each particle.

3.4 | Sensitivity analysis

To quantify uncertainty and determine the optimal setting for hillslope trenching, sensitivity analysis is applied to the recharge and groundwater models.

For the recharge model, a one-at-a-time sensitivity analysis is used (Hamby, 1994; Pianosi et al., 2016). A range of plausible values is selected for each input variable, and the model is run repeatedly, changing one variable while all others remained the same. For each sensitivity run, three results are recorded: the percentage of precipitation that becomes recharge for the base case scenario, the trenched scenario, and the difference between the two. Sensitivity is ranked on the basis of the difference between the two scenarios, which can be thought of as an indicator of the benefit of trenching.

For the three-dimensional groundwater model, sensitivity analysis is performed on the configuration of trenching within the catchment. Within the model domain, the trenched area is moved or expanded for consecutive model runs, and the impact on the timing and magnitude of groundwater discharge to the river is examined.

4 | RESULTS

4.1 | Data collection

Measured stream discharges at several points along the stream from August 2016 are shown in Table 1. However, there is no clear connection between the trenched area and the increase in stream discharge from a given zone of the catchment. We must, therefore, rely on the modelling results.

Thirteen infiltration measurements were taken with the double ring infiltrometer. The dry season measurements were performed between August 15 and 19, 2016, and ranged from 0.07 to 5.76 with an average of 2.4 m/day (n = 9). Rainy season measurements were taken between March 7 and 10, 2017, and ranged from 0.17 to 0.24 with an average of 0.20 m/day (n = 4). These measured infiltration capacities are within range of literature values for mountain grassland (Gaither & Buckhouse, 1983; Leitinger, Tasser, Newesely, Obojes, & Tappeiner, 2010; Roa-García, Brown, Schreier, & Lavkulich, 2011).

Meteorological data from the Lazo Huntay automatic weather station were obtained from SENAMHI. The data set from January 14, 2011, to June 17, 2014, was recorded at 30-min intervals. Precipitation followed the typical seasonal pattern with the majority of precipitation falling between November and April, whereas air temperature stayed relatively constant throughout the year (Figure 4a). Due to field malfunction, there is a data gap from January 29 to April 1, 2013. For the three complete years recorded, the total annual rainfall was 1,390, 1,240, and 1,250 mm for 2011, 2012 (starting January 14 of each year), and 2013–2014 (April 1, 2013, to April 1, 2014), respectively. These values were substantially higher than the literature value for average annual rainfall in the Shullcas Basin of 800 mm (ANA Ministerio de Agricultura, 2010), likely because of the location of the weather station higher in the catchment.

For the period between July 1, 2015, and August 31, 2016, daily precipitation measurements for the Lazo Huntay station were taken from SENAMHI's online database and showed the same seasonal pattern with a total annual precipitation of 736 mm recorded between July 1, 2015, and July 1, 2016. The measurement period coincided with El Niño, which is associated with anomalous weather in Peru. Although El Niño often results in excess precipitation in the northern coastal desert, the impact in Huancayo can be either an excess or

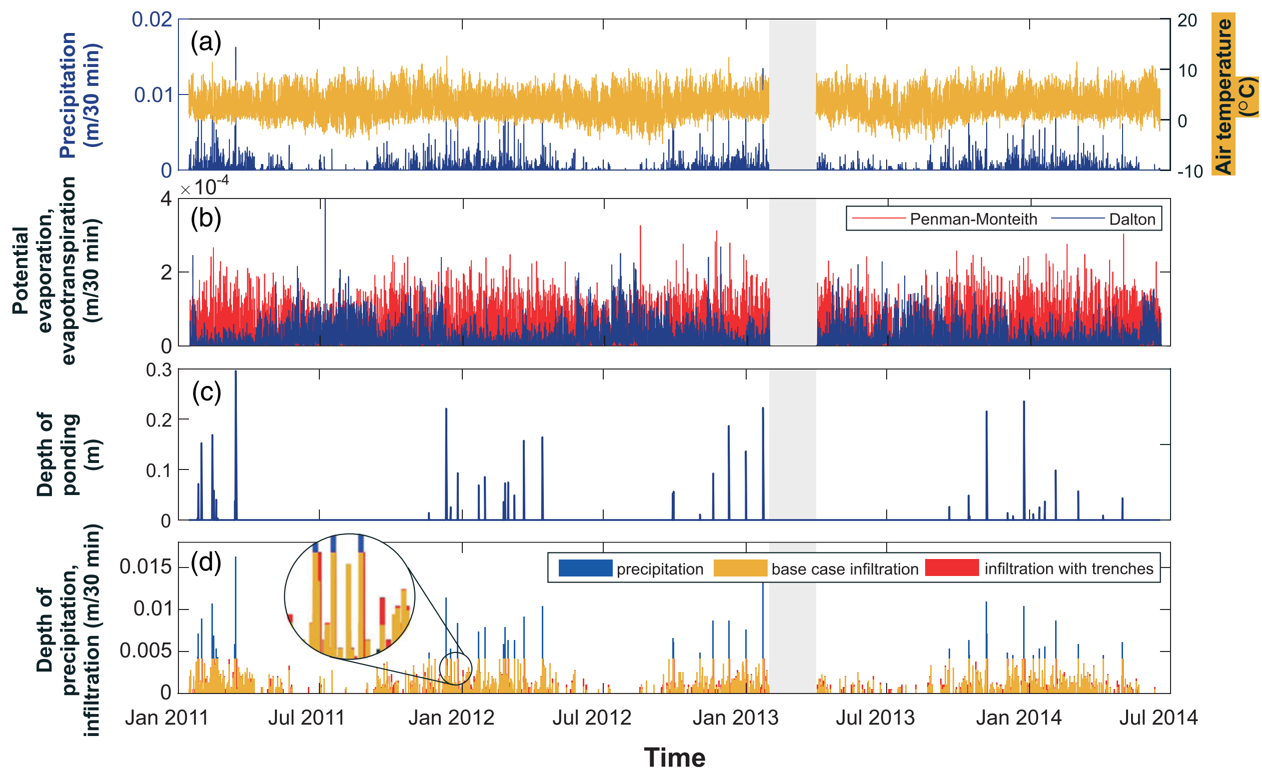

FIGURE 4 Input variables and results for the recharge model for an idealized grassy hillslope. (a) Precipitation and air temperature. (b) Potential evapotranspiration as calculated using the Penman–Monteith formulation (red) and potential evaporation as calculated using the Dalton-type equation (blue). (c) Depth of ponding. (d) Depth of precipitation and infiltration, where red indicates times where the trenched case infiltration exceeds the base case. Grey area indicates missing input data

deficit of precipitation (Kane, 2000) and may explain a drier rainy season than usual in the Shullcas Basin.

Stream discharge was recorded from July 21, 2015, to August 17, 2016, at 15-min intervals (Figure 5). The peak flow of 1.040 m^3/s was recorded on March 10, 2016, and the minimum flow of 0.016 m^3/s was recorded on August 20, 2015. Due to field malfunction, there is a gap in the data between January 3 and March 5, 2016. The stream discharge during this period was estimated using an empirical relationship between the study gauging station and a gauging station in an adjacent catchment (r^2 = .53).

4.2 | Recharge model

Input variables for the recharge model were selected on the basis of field observations and literature values (Table 2). The rainy season infiltration capacity was used because this is the only time that the precipitation intensity exceeds the infiltration capacity and runoff is generated.

The one-dimensional model was run for the 3.5-year period from January 14, 2011, to June 17, 2014, using 30-min interval meteorological data. The calculated potential evaporation from ponded water, calculated using the Dalton-type equation, and the potential evapotranspiration from the grassland, calculated using the Penman–Monteith formulation (Figure 4b), had average values of 0.8 and 1.0 mm/day respectively. Ponding occurred during only the highest intensity precipitation events (Figure 4c), and trench overflow occurred only once during the model run.

The pattern of infiltration closely follows that of precipitation (Figure 5). Infiltration for the trenched scenario exceeded the base case scenario at times of heavy precipitation when runoff is captured and at times when vegetation cover is dry. This is because ditches are not subject to interception, whereas grassland is (Figure 4d). The pattern of recharge is dampened and delayed relative to infiltration, with peak recharge occurring towards the end of the rainy season. Like infiltration, recharge in the trenched scenario exceeds the base case during times of high precipitation and particularly on the rising limb of the recharge curve.

The results of the recharge model indicate that in nontrenched hillslopes, 79.6% of precipitation infiltrates into the ground and 48.6% becomes groundwater recharge (reaches the water table), whereas for trenched hillslopes, 83.3% of precipitation becomes infiltration and 52.1% becomes recharge. This value is fairly high compared to that of some studies (Jódar et al., 2017) but is within range of the value of other studies that have estimated recharge relative to precipitation in alpine catchments (Crosbie, Jolly, Leaney, & Petheram, 2010; Voeckler, Allen, & Alila, 2014; Fan, Oestergaard, Guyot, & Lockington, 2014).

4.3 | Three-dimensional groundwater model

The model output, net groundwater discharge to the river, served as the target variable for calibration. The model was visually calibrated to the dry season stream discharge from July 21 to October 15, 2015, and May 1 to August 17, 2016 (Figure 5), by varying the hydraulic conductivity of the two layers and specific yield of the surficial layer. The calibrated hydraulic conductivities are 7 and 0.5 m/day for the surficial deposits and fractured bedrock, respectively. The calibrated specific yield of the surficial deposits is 0.09.

Once calibrated, modelled groundwater baseflow compares well with the measured dry season stream discharge, specifically during baseflow recession at the beginning of the dry season (May–June). The root mean square error of the model baseflow during the dry season was 0.011 m^3/s, and the normalized root mean square error was 7.5%.

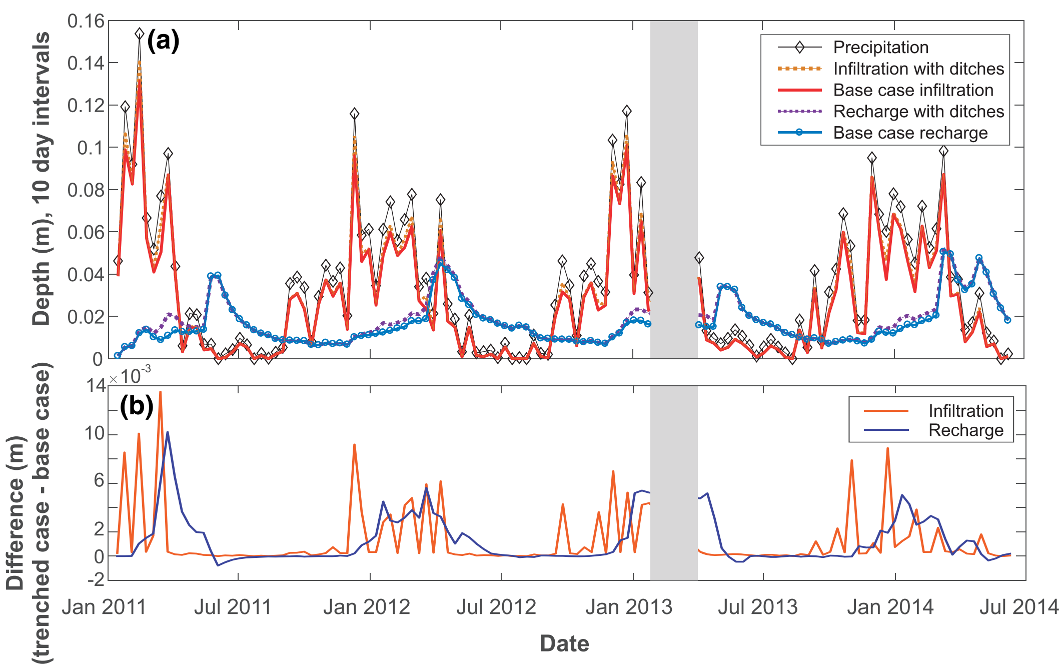

FIGURE 5 (a) Precipitation, infiltration, and recharge for the base case and trenched scenarios. (b) Difference in infiltration and recharge amounts for the base case and trenched case. Both plots are aggregated over 10-day periods. Grey area indicates missing input data

Three "peaks" in baseflow can be observed in the modelled baseflow hydrograph, the first in early January, the second and largest in early March, and the third in late April (indicated in Figure 6). These peaks correspond to periods of heavy precipitation and are followed by periods where baseflow becomes the dominant water source (baseflow recession), as is expected.

In order to compare to the measured hydrograph throughout the rainy season, a simple runoff component was added to the modelled baseflow (Figure 6). Using 20% of precipitation to represent aggregate runoff and shallow interflow over the study catchment provided the best fit to the measured hydrograph. This combined modelled hydrograph is generally slightly lower than the measured hydrograph during the rainy season and slightly higher during the dry season. This is likely because the simple runoff model does not include any treatment for antecedent moisture content of the soil and vegetation.

During the simulation, the groundwater table was generally shallow closer to the stream and deeper higher up on the hillslopes. In these higher and steeper areas of the catchment, the surficial layer was never saturated, and groundwater flow was channelled through the fractured bedrock. Once the groundwater flow meets the valley, part of it flows into the saturated surficial layer and then to the stream.

In this way, most of the groundwater flow reaching the river did so through the surficial layer due to its higher hydraulic conductivity. With the use of ZONEBUDGET, the river zone was delineated as the cell containing the river, plus one 100-m cell on either side, in all layers. Averaged over the modelled period, 72% of groundwater flow to the river zone was transmitted through the surficial layer, and 28% was transmitted through the fractured bedrock layer. Slightly more groundwater reached the river zone through the fractured bedrock layer during the early rainy season from December to January (maximum 33% bedrock flow and 67% surficial flow) and slightly less during the early dry season from May to June (minimum 25% bedrock flow and 75% surficial flow).

Flow paths were analysed for 20 particles using MODPATH. Two particles were launched at each of 10 locations randomly chosen throughout the catchment. Travel times ranged from 198 to 1,859 days (~5 years), from release until reaching the river. The average travel time for the top-most particle at each of the 10 locations was 545 days (1.5 years). A map of the resulting flow paths is included in Figure S3. This is compatible with a conceptual model based on hydrochemical

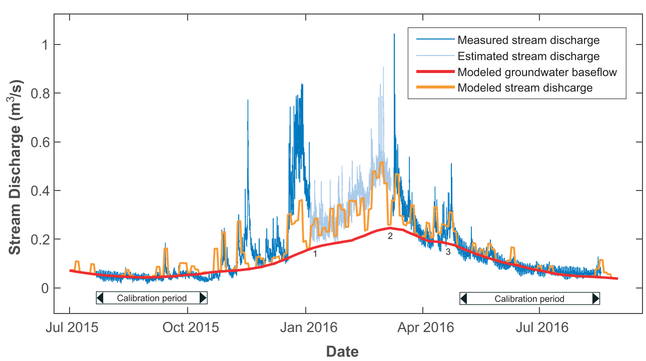

FIGURE 6 Measured stream discharge compared to modelled baseflow and modelled stream discharge over the 14-month groundwater simulation period. The estimated stream discharge fills the data gap by regressing to a nearby stream gauge

analysis by Baraer et al. (2015) in which the retention time is estimated to be long enough to maintain lateral springs through the dry season in glaciated valleys of the Cordillera Blanca.

4.4 | Sensitivity analysis

Sensitivity analysis was used for both models. For the recharge model, two types of sensitivity were evaluated: first, the sensitivity of the amount of recharge, in both scenarios, to the input variables, and second, the sensitivity of the difference in recharge between the base case and trenched scenarios to the input variables. For example, increasing the maximum ET decreases the amount of recharge for both the trenched and nontrenched scenarios (sensitive to max ET). However, increasing max ET does not greatly change the difference in recharge between the trenched and nontrenched scenarios (not sensitive to max ET).

Table 3 shows the range of likely values selected for each input variable and the resulting recharge rates and difference in recharge between the base case and trenched scenarios. The resulting recharge rates (Columns 3 and 4 of Table 3) are most sensitive to maximum ET, followed by infiltration capacity, ET extinction water content, interception storage, ET extinction depth, and the potential ET time series. All other input variables had little impact (<0.1% difference) on the overall recharge rate.

The difference in recharge between the two scenarios (Column 5) is most sensitive to infiltration capacity followed by the spacing of trenches and then the width of trenches. All other input parameters had minimal impact on the difference in recharge between the two scenarios, meaning that they affect the base case and trenched hillslopes similarly.

Sensitivity analysis was also performed on the groundwater model for different trenching configurations. Six different scenarios were tested and compared to a scenario with no trenching. Figure 7 shows the difference (increase) in baseflow between each different trenching configuration and the base case. Doubling the trenched area approximately doubled baseflow, as expected, and did not greatly change timing. Trenching low in the catchment or in steep areas made the baseflow increase flashier, with increases higher in the rainy season but then dropping off to almost zero in the dry season. Alternatively, trenching high in the catchment and trenching in flat areas both had the effect of delaying the peak baseflow increase and providing more baseflow longer into the dry season. A similar plot showing the difference in baseflow expressed as a percent difference is included in the Supporting Information (Figure S4).

5 | DISCUSSION

5.1 | Groundwater flow system

The calibration of the MODFLOW groundwater model yielded relatively high hydraulic conductivities (K) for both the surficial deposits and the fractured bedrock layer. These parameters are reasonable considering the coarse, gravel-rich nature of much of the surficial layer and the fractured nature of the shallow metamorphic bedrock. Our surficial deposit K is lower than the range reported by Langston, Hayashi, and

TABLE 3 Hillslope sensitivity analysis

1. Sensitivity variable (original value)	2. Range of values	3. Recharge base case (%)	4. Recharge with trenching (%)	5. Difference between base case and trenched case (%)
Unperturbed (from Section 4.2)	–	48.51	52.05	3.54
Infiltration capacity (0.2 m/day)	0.3 m/day 0.1 m/day	50.69 42.07	52.53 50.66	1.84 8.59
Interception storage (1 mm)	1.5 mm 0.5 mm	47.00 51.84	50.58 55.26	3.58 3.42
Width of trenches (0.4 m)	0.6 0.2	48.56[a] 48.56[a]	52.40 51.37	3.84 2.81
Spacing of trenches (9 m)	15 5	48.56[a] 48.56[a]	51.96 53.24	3.40 4.68
Potential evapotranspiration time series (−22–16 mm/day, average 1 mm/day)	+10% −10%	47.77 49.41	51.32 52.92	3.55 3.51
Open water evaporation (0–19 mm/day, average 0.8 mm/day)	+10% −10%	48.56[a] 48.56[a]	52.09 52.09	3.53 3.53
Maximum ET (1 mm/day)	1.5 mm/day 0.5 mm/day	40.70 57.97	44.15 61.49	3.45 3.52
ET extinction depth (2 m)	1 m 3 m	49.53 47.79	53.01 51.26	3.48 3.47
ET extinction water content (0.05)	0.075 0.025	53.55 48.56	57.02 52.03	3.47 3.47
Vertical hydraulic conductivity of the unsaturated zone (1 m/day)	1.2 m/day 0.8 m/day	49.09 47.85	52.57 51.33	3.48 3.48
Brooks–Corey epsilon (3.5)	4 3	46.32 51.13	49.79 54.60	3.47 3.47
Saturated water content (0.3)	0.35 0.25	48.79 72.38	52.27 75.81	3.48 3.43

[a]Graphical user interface for affect base case.

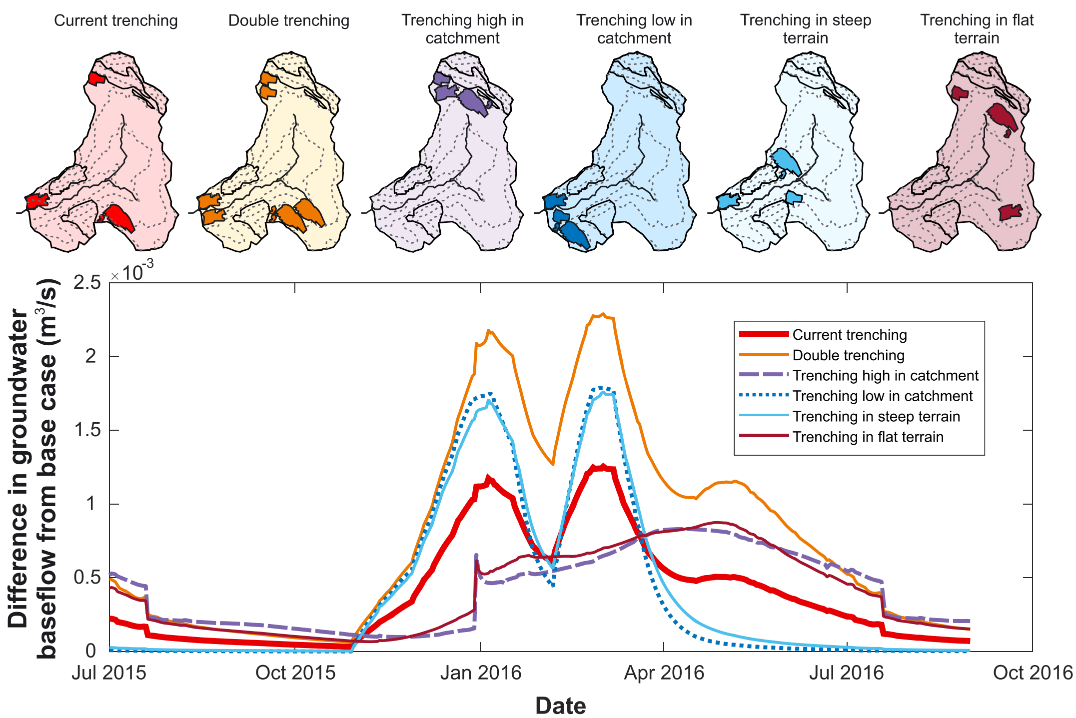

FIGURE 7 Sensitivity to trenching configuration demonstrating how location and area of trenching can affect groundwater baseflow generation. Six different trenching scenarios were compared to the base case (no-trenching) baseflow hydrograph, and the differences are plotted

Roy (2013) for a moraine in the Canadian Rockies ($0.3–3 \times 10^{-3}$ m/s) and is similar to that reported by Magnusson et al. (2014) for a glacier forefield deposit in the Swiss Alps ($0.8–5 \times 10^{-4}$ m/s). The specific yield value on the other hand was low compared to typical values for the mixed silt, sand, and gravel soil. Fetter (2001) suggests values ranging from 0.03 to 0.19 for silt and 0.20 to 0.35 for gravelly sand.

The calibrated model indicated that the catchment is dominated by relatively fast and shallow groundwater flow. This agrees with hydrogeological research done in mountain regions elsewhere. McClymont, Hayashi, Bentley, Muir, and Ernst (2010) used geophysics to examine groundwater flow paths in a talus and meadow complex in the Canadian Rockies. They found that unconsolidated deposits were less than 10 m thick and that precipitation inputs well exceeded the groundwater storage capacity of the small headwater catchment, meaning that the stored groundwater was replenished on a subannual basis. Voeckler et al. (2014) used a coupled groundwater–surface water model (MIKE SHE by DHI) to investigate the role of deep groundwater flow in a headwater catchment in the Okanagan highlands of British Columbia, Canada. They found that outward groundwater flux through the deep bedrock layer amounted to only 2% of the annual water budget. However, other research has emphasized the importance of deep groundwater flow in mountain environments (Gleeson & Manning, 2008; Graham, Van Verseveld, Barnard, & McDonnell, 2010). Welch and Allen (2012) simulated groundwater flow in different mountain topographic scenarios and looked at how the groundwater recharge is partitioned into baseflow and mountain block recharge (MBR). They estimated 12% to 15% (reported as BF/MBR ratios from 5.8 to 7.3) of groundwater recharge goes to MBR.

During model development, incorporating deeper groundwater flow through intact bedrock with a realistic hydraulic conductivity dampened the annual pattern in baseflow such that the model's prediction of dry season baseflow well exceeded the measured value (not shown). Therefore, it was determined that shallow subsurface flow through the surficial and fractured bedrock layers dominates baseflow to the stream.

Our characterization of the montane groundwater flow regime suggests that the Shullcas system does not have significant multiyear groundwater storage and may be sensitive to changes in precipitation and temperature. This agrees with Carey et al. (2010) who looked at 10 cold regions catchments in North America and Northern Europe and found that steep catchments had stronger correlations between monthly precipitation and stream discharge, meaning that they were less resistant to hydrologic perturbations. Furthermore, as the Huaytapallana glaciers continue to retreat, glacial meltwater contribution to streamflow will decrease, strengthening the coupling of precipitation and stream discharge and decreasing the ability of the catchment to resist hydrologic change. Conversely, the strong seasonality of precipitation in the Shullcas Basin could mean that the shallow aquifers are fully replenished each rainy season and that changes in rainy season precipitation do not greatly affect dry season stream discharge. In this case, the addition of trenches would provide no benefit. Multiyear streamflow analysis is needed to tackle this question.

5.2 | Effectiveness of infiltration trenches

The recharge model showed that the difference in recharge between the base case and trenched scenarios was small: only 3.5% relative to precipitation. This small increase in recharge over the trenched area of the study catchment results in increased groundwater discharge to the stream (Figure 7). For current trenching conditions, the maximum increase in groundwater contribution is 1.3 L/s (112,300 L/day) and occurs in early March during the late rainy season in the Peruvian Andes. The dry season increase is smaller at approximately 0.1 L/s (8,640 L/day). Given that the average Peruvian uses 175 L of water per day (United Nations Development Report, 2006), the dry season baseflow increase discharge could theoretically supply 49 more people with water during the dry season. Furthermore, the study catchment is only a small part of the Shullcas Basin, which has a total of 8 km^2 of trenches. Multiplying our result over the total trenched area of the Shullcas Basin, 806 more people could be served during the dry season, ignoring losses in the distribution system, which are often significant in municipal water systems.

Although direct field measurement would be ideal to demonstrate the effectiveness of these infiltration trenches and compare to our modelling results, this is very difficult in practice. We hypothesized that one way of doing this could be to install a series of soil moisture meters below adjacent trenched and nontrenched hillslopes for a comparison. However, heterogeneity in soil textures, installation technique, and even incoming precipitation mean the noise would almost certainly be greater than the difference between the two slopes. Furthermore, a large number of sensors would be required to achieve a statistically significant result if a difference was present. A similar problem was encountered by Mastrocicco et al. (2016) when attempting to measure the resulting groundwater mound from a MAR scheme in Italy. The calculated expected mound was well within natural variations in the water table, and it was not possible to distinguish the impact of the enhanced recharge.

The effectiveness of the trenches is also likely to change over time. For example, vegetation will recolonize the trenches as was observed in some areas of the catchment, such that interception will return to pretrenching levels. Mastrocicco et al. (2016) also suggested that the effectiveness of MAR schemes may be affected by pore size reduction from clogging and biologic activity.

5.3 | Best settings for infiltration trenches

Sensitivity analysis indicated that infiltration trenches provided differing benefits depending on the setting. Most prominently, trenching is more effective in areas with low infiltration capacity and therefore more overland flow for the trenches to intercept, provided that the density of trenches and infiltration capacity are sufficiently high to accommodate all the overland flow generated without additional spillover. This is the opposite of MAR systems, which require high infiltration capacities to work (Bouwer, 2002). For the same reason, areas that receive higher intensity precipitation are better candidates for hillslope trenching. Otherwise stated, hillslopes that do not generate overland flow would receive no added benefit from this type of hillslope trenching.

However, if the infiltration capacity is too low, trenches may be filled with water for extended periods of time, overflowing during precipitation events and limiting their effectiveness. For example, in March 2016, the higher elevation trenched area in the northwest corner of the study catchment was observed to have ponding in trenches at a time when the lower elevation trenched area in the south west corner of study area was not. This likely has to do with differing infiltration capacities where the high elevation trenched area has a siltier soil and low vegetation cover whereas the lower trenched area has a rockier soil and tall grass. It should be noted that the difference in ponding is probably also a function of elevation difference between the two sites, resulting difference in orographic precipitation and general heterogeneity of mountain weather.

Trenching configuration is also an important consideration. Trench spacing and width should correspond to the anticipated volume of runoff based on land cover and soil properties. Our recharge model (or a similar code) could serve as a tool for selecting the width and spacing of trenches. Furthermore, the sensitivity analysis of the groundwater model indicated that trenching higher in the catchment, further from the stream, or in flatter terrain is better for delaying the peak and increasing dry season baseflow. Differentiating between the impact of the trenching location and the flat terrain is difficult because of the overlap in these terrain characteristics. This is consistent with Smith, Moore, Weiler, and Jost (2014), who found that the spatial distribution of inputs, in their case snow melt, is an important control on stream response.

Here, we have outlined some practical considerations for hillslope trenching, but many additional concerns exist including the risk of water logging, changes to slope stability, ecosystem damage from modifying large areas of the land surface (Dillon, 2005), and impact on local herding communities.

5.4 | Assumptions and limitations

Infiltration capacity is a highly heterogeneous physical property (Haws et al., 2004). One limitation of the recharge model is that the results represent a small margin of increase compared to the uncertainty on the input variables. Although many infiltration observations were performed in order to obtain a representative value, it should be emphasized that the uncertainty in this input parameter has the potential to change the model results significantly. However, the one-dimensional model was only sensitive to three input variables, increasing confidence in our results.

Additionally, our precipitation input data do not vary spatially. Mountain weather is highly heterogeneous, but we depended on a single weather station to drive the model and therefore could not apply an orographic correction on the basis of multiple stations. However, this source of error is partially mitigated by the fact that the study area is fairly small (10.4 km^2) and the weather station is close by (<2 km).

Roy and Hayashi (2009) showed that multiple complex flow paths exist in mountain groundwater systems, complicating the modelling process. One limitation of our groundwater model is that we lack detailed hydrogeological data including the depth of sediments and permeability field test information. Due to the remote location of the Shullcas Watershed inside a conservation area and the heterogeneity of the hydrogeological materials in the catchment, it was not feasible to execute a field program to gather this information in detail. Therefore, the groundwater model is nonunique, meaning that there is more than one possible combination of input parameters that could provide the same or similar solution. However, fast and shallow groundwater flow is a consistent result across different parameter sets.

Furthermore, our groundwater model set-up assumes that all water leaving the model domain flows out through the river. However, in reality, there is likely some minor groundwater flow out of the catchment through the valley bottom sediments below the river and through the intact bedrock that we did not attempt to account for.

Although these limitations may affect our ability to duplicate exactly what was observed in the field, the recharge and groundwater modelling exercises do allow us to better understand the groundwater system and the role of trenching on the hydrological regime as per the goal of this study.

6 | CONCLUSIONS

We used a recharge model to evaluate the potential impact of hillslope trenching on groundwater discharge to a stream in the Shullcas River Watershed in the Peruvian Andes. Simulations showed that trenched hillslopes received approximately 3.5% more recharge, relative to precipitation, compared with unaltered hillslopes. We then applied the calculated recharge rates to a groundwater model of the study basin. The MODFLOW groundwater model indicated that incorporating trenched hillslopes (~2% of study catchment area) slightly increases baseflow in the mid–late rainy season but has only a small impact on dry season baseflow. If multiplied over the entire trenched area of the Shullcas Watershed, it could result in water service for an additional 800+ people during the dry season, neglecting losses to the distribution system, which are substantial for most municipal water systems. Therefore, the effectiveness of the trenches in augmenting dry season baseflow is limited, and the hydrogeological characteristics of the area should be considered in installing similar technology.

To our knowledge, this constitutes the first scientific study on hillslope trenching as a passive aquifer recharge technology. Additionally, the results indicate that the groundwater flow system in this mountain catchment is relatively fast and shallow and is an important contributor to stream flow, which should not be neglected in modelling efforts. Improving our understanding of mountain groundwater systems will allow better projection of the impact of climate change and allow better design of climate change adaptation strategies. The results of this study may have important implications for Andean landscape management and water resources.

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ORCID

Lauren D. Somers http://orcid.org/0000-0002-5382-1625
Jeffrey M. McKenzie http://orcid.org/0000-0002-0469-6469
Samuel C. Zipper http://orcid.org/0000-0002-8735-5757
Bryan G. Mark http://orcid.org/0000-0002-4500-7957
Michel Baraer http://orcid.org/0000-0003-4138-3354

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SUPPORTING INFORMATION

Additional Supporting Information may be found online in the supporting information tab for this article.

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