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The hydrological regime of a forested tropical Andean valley

K. E. Clark[1], M. A. Torres[2], A. J. West[2], R. G. Hilton[3], M. New[4,1], A. B. Horwath[5], J. B. Fisher[6], J. M. Rapp[7,8], A. Robles Caceres[9], and Y. Malhi[1]

[1] School of Geography and the Environment, University of Oxford, Oxford, UK
[2] Department of Earth Sciences, University of Southern California, Los Angeles, CA, USA
[3] Department of Geography, Durham University, Durham, UK
[4] African Climate and Development Initiative, University of Cape Town, Rondebosch, Cape Town, South Africa
[5] Department of Plant Sciences, University of Cambridge, Cambridge, UK
[6] Jet Propulsion Laboratory, California Institute of Technology, Pasadena, CA, USA
[7] Department of Biology, Wake Forest University, Winston Salem, NC, USA
[8] now at: Department of Biology, Tufts Unviersity, Medford, MA, USA
[9] Facultad de Ciencias Biológicas, Universidad Nacional de San Antonio Abad del Cusco, Cusco, Peru

Correspondence to: K. E. Clark (kathryn.clark23@gmail.com)

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Abstract

The hydrology of tropical mountain catchments plays a central role in ecological function, geochemical and biogeochemical cycles, erosion and sediment production, and water supply in globally important environments. There have been few studies quantifying the seasonal and annual water budgets in the montane tropics, particularly in cloud forests. We investigated the water balance and hydrologic regime of the Kosñipata Valley (basin area 164.4 km^2) over the period 2010–2011. The valley spans over 2500 m in elevation in the eastern Peruvian Andes and is dominated by tropical montane cloud forest with some high elevation *puna* grasslands. Catchment wide rainfall was 3028 ± 414 mm yr^{-1}, calculated by calibrating Tropical Rainfall Measuring Mission (TRMM) 3B43 rainfall with rainfall data from 9 meteorological stations in the valley. Cloud water input to streamflow was 316 ± 116 mm yr^{-1} (~ 10 % of total inputs), calculated from an isotopic mixing model using deuterium excess (Dxs) and δD of waters. Field stream flow was measured in 2010 by recording height and calibrating to discharge. River runoff was estimated to be 2796 ± 126 mm yr^{-1}. Actual evapotranspiration (AET) was 909 ± 182 mm yr^{-1}, determined using the Priestley and Taylor – Jet Propulsion Laboratory (PT-JPL) model. The overall water budget was balanced within 10 %. Relationships between monthly rainfall and river runoff follow an anti-clockwise hysteresis through the year, with a persistence of high runoff after the end of the wet season. The size of the soil- and shallow ground-water reservoir is most likely insufficient to explain sustained dry season flow. Thus, the observed hysteresis in rainfall-runoff relationships is best explained by sustained groundwater flow in the dry season, which is consistent with the water isotope results that suggest persistent wet season sources to stream flow throughout the year. These results demonstrate the importance of transient groundwater storage in stabilizing the annual hydrograph in this region of the Andes.

1 Introduction

The routing of water from the eastern flank of the Andes determines the quantity and quality of this economically and ecologically valuable resource for the region (Célleri and Feyen, 2009; Barnett et al., 2005; Postel and Thompson, 2005) and impacts the biogeochemical cycles and ecology of the lowland Amazon (McClain and Naiman, 2008; Allegre et al., 1996; Stallard and Edmond, 1983). The Amazon River has the highest discharge of all of the world's rivers, at $6300\,\mathrm{km^3\,yr^{-1}}$, with a very high runoff of $1000\,\mathrm{mm\,yr^{-1}}$ over its watershed (Milliman and Farnsworth, 2011), and it contributes 20 % of the global water discharge to oceans (Beighley et al., 2009; Richey et al., 1990). The Andean portion of the Amazon Basin ($> 500\,\mathrm{m}$) represents an area of 623 217 $\mathrm{km^2}$ and covers $\sim 10\,\%$ of the Amazon River Basin (McClain and Naiman, 2008). Although the water input from the Andes to the Amazon is approximately proportional to areal coverage (10 %) (McClain and Naiman, 2008; Dunne et al., 1998), the Andes are the dominant source of the Amazon's dissolved load (McClain and Naiman, 2008; Gaillardet et al., 1999; Guyot et al., 1996), and contribute 80–90 % of its suspended sediment (Richey et al., 1986; Meade et al., 1985; Gibbs, 1967). Information about Andean river discharge, flow sources, and flow routing is thus critical for understanding the suspended sediment fluxes and chemical weathering processes of the Amazon River (Bouchez et al., 2012; Wittmann et al., 2011; Guyot et al., 1996), for quantifying how the Andes contribute to carbon and nutrient cycles (Clark et al., 2013; Townsend-Small et al., 2008), and for assessing the aquatic ecology of the region (Anderson and Maldonado-Ocampo, 2011; Ortega and Hidalgo, 2008). Hydrologic information is particularly important for understanding related responses to changes in climate and land-use.

Despite this importance, the dynamics of Andean hydrology are still incompletely characterized. This is especially true in Andean Tropical Montane Cloud Forest (TMCF), which comprises a small area but is likely to contribute disproportionately to the overall water balance of the region due to its topographic position that receives

high precipitation (Bruijnzeel et al., 2011; Killeen et al., 2007). The hydrology of TMCFs is of particular interest because these forests host valuable and diverse ecosystems (Bruijnzeel et al., 2010; Bubb et al., 2004; Myers et al., 2000) and have been shown to provide an important supply of water to downstream regions, due in large part to their relatively high water yield, i.e. high stream water output for a given precipitation input (Tognetti et al., 2010; Zadroga, 1981). TMCFs are unique hydrologic systems because of the additional water input from cloud water interception (CWI) and because frequent fog occurrence may lower incoming solar radiation, increasing humidity and potentially lowering evapotranspiration (ET) (Giambelluca and Gerold, 2011; McJannet et al., 2010; Zadroga, 1981). In many TMCFs high annual rainfall dominates over CWI (Holwerda et al., 2010a, b; McJannet et al., 2007, 2010; Schmid et al., 2010; Eugster et al., 2006) implying that rainfall is the primary source of TMCF streamflow. However, rainfall is often thought to be quickly exported via overland flow and shallow subsurface stormflow, leading to rapid rainfall-response times (Boy et al., 2008; Goller et al., 2005; Schellekens et al., 2004).

Transient groundwater storage may play a significant role in mountain hydrological systems (Andermann et al., 2012; Calmels et al., 2011; Tipper et al., 2006). The importance of groundwater in TMCF hydrology has recently been highlighted by studies in a Mexican TMCF, where groundwater was shown to stabilize the rainfall-runoff response (Muñoz-Villers and McDonnell, 2012), and in an Andean TMCF in Ecuador, where considering the effect of groundwater reservoirs was important for accurately predicting streamflow (Crespo et al., 2012). Improved understanding of the extent to which groundwater stabilizes Andean TMCF hydrology is likely to be important for accurately assessing how environmental change, such as land use change or shifting cloud base, might affect hydrological functioning in the Andes and downstream in the Amazon lowlands (Rapp and Silman, 2014; Crespo et al., 2012; Bruijnzeel et al., 2011; Mulligan, 2010; Ataroff and Rada, 2000).

In this paper we evaluate stream discharge of the Kosñipata River, in a well-studied region in the eastern Andes of Peru (Rapp and Silman, 2014; Halladay et al., 2012b;

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van de Weg et al., 2012; Salinas et al., 2011; Girardin et al., 2010; Malhi et al., 2010), over a one-year period. We compare discharge data to rainfall estimates in order to assess the water balance and hydrologic variability throughout the study year. We use the distinct water isotope composition of cloud and rain water to constrain the role of cloud water input. Stable water isotopes, i.e. δD (‰) and δ^{18}O (‰), can be used to distinguish water sources due to distinct fractionation that occurs during evaporation and condensation (Scholl et al., 2011; Froehlich et al., 2002; Gat, 1996; Rozanski et al., 1993; Craig, 1961). Stable water isotopes have been used in studies of cloud forest hydrology to estimate local water recycling (Scholl et al., 2007; Rhodes et al., 2006), trace water paths through soil layers in a catchment (Goller et al., 2005), evaluate water sources in stormflow (Muñoz-Villers and McDonnell, 2012), evaluate water mean transit time (Timbe et al., 2014) and examine ecohydrology (Goldsmith et al., 2012; Dawson, 1998). We extend this application to constrain the contributions of different precipitation sources to annual streamflow, and in the process we add valuable new water isotope data for a widely studied TMCF in the Andes.

We focus our attention on the following questions: (1) How well can the annual water budget of the Kosñipata Valley be closed and what are the uncertainties? (2) What is the importance of baseflow, i.e. the constant supply of water throughout the year, not associated with short term fluctuations due to storms? (3) Are there any significant seasonal lags between rainfall and stream runoff, and what are the causes of these lags? (4) What is the relative importance of rainfall and cloud water in sustaining streamflow throughout the year? and (5) What are the roles of soil and groundwater storage in determining seasonal patterns of river flow?

2 Study area

The Kosñipata Valley (13°3′37″ S, 71°32′40″ W) study area ranges from 1360 to 4000 m above sea level (m a.s.l.) (Fig. 1a). We focus on the Kosñipata River measured at the San Pedro gauging station, which drains an area of 164.4 km^2. In the

Supplement we present results from the nested Wayqecha sub-catchment that encompasses the headwaters of the Kosñipata River, draining an area of 48.5 km^2 (Table 1). Downstream of the study region, the river flows into the Alto Madre de Dios River which feeds the Madre de Dios River (Fig. 1b), a major tributary of the Amazon River (Fig. 1c).

The geology of the study area is dominated by meta-sedimentary mudstones covering $\sim 80\%$ of the valley with a plutonic intrusion comprising $\sim 20\%$ of the catchment (Table 1) (Carlotto Caillaux et al., 1996). The geological characteristics and vegetation of the valley are generally representative of the larger eastern Andean region of southern Peru and northern Bolivia (INGEMMET, 2013; Consbio, 2011; Carlotto Caillaux et al., 1996).

The climate of the Eastern Andes is influenced by the South American Low Level Jet (SALLJ), which carries humid winds west over Amazonia and then south along the Andean flank (Marengo et al., 2004). The Kosñipata Valley sits in a band of persistent cloudiness that runs along the Eastern Andes (Halladay et al., 2012b) and has high rainfall relative to the Andean regions to the north and to the south because of its location on an east-west kink of the Andean range that situates it perpendicular to the SALLJ (Killeen et al., 2007). Within the valley, rainfall decreases with increasing elevation, from 5300 mm yr^{-1} at 1500 m a.s.l. down to 1560 mm yr^{-1} at 3025 m a.s.l., near the treeline (Girardin et al., 2014; Huaraca Huasco et al., 2014), where down-valley winds collide with most air from Amazonia (Halladay et al., 2012b). Due to orographic effects, rainfall is highest from 1000 to 1500 m a.s.l. (Rapp and Silman, 2012). Note that lower total annual rainfall amounts were reported previously for this valley (Lambs et al., 2012), but the data used in this previous study were incomplete for the locations where we recorded highest rainfall. Orographic fog (cf. Scholl et al., 2011) plays an important role in the Kosñipata Valley. Cloud base varies in height throughout the year, with the cloud base at its lowest in the dry season (June to August) (Halladay et al., 2011). In July (mid-dry season) the cloud base is $> 60\%$ of the time > 1800 m a.s.l. and 30% of the time between 1500–1800 m a.s.l. (Rapp and Silman, 2014). Cloud water interception has not been directly quantified in the region, but ranges from ~ 20 to

$\sim 2000\,\mathrm{mm\,yr^{-1}}$ in other tropical montane cloud forest catchments (Bruijnzeel et al., 2011); similar ranges are expected in the Kosñipata Valley. Annual mean temperatures in the Kosñipata Valley range from $\sim 19\,^{\circ}\mathrm{C}$ at low elevations to $\sim 12\,^{\circ}\mathrm{C}$ at high elevations (Girardin et al., 2014; Huaraca Huasco et al., 2014) with an adiabatic air temperature
5   lapse rate of $4.94\,^{\circ}\mathrm{C\,km^{-1}}$ of altitude (Girardin et al., 2010). The wet season is generally defined to be December to March, the wet–dry transition season to be April, the dry season to be May to September, and the dry-wet transition season to be October and November (Table S1 in the Supplement). These terms are used in a relative sense in the Andes, since precipitation is still significant in the dry season.

10   The valley is dominated by forest ($\sim 80\,\%$) with the remainder of the catchments being high elevation grasslands called wet *puna* (Squeo et al., 2006) (Table 1). Small areas of bare bedrock are exposed at the highest elevations. In the forested area, vegetation consists of sub-alpine cloud forest at high elevations (3400 to 3600 m a.s.l.), upper montane cloud forest between 2000 to 3400 m a.s.l., and lower montane cloud
15   forest and sub-montane tropical rainforest between 1200 to 2000 m a.s.l.. The Kosñipata Valley is partially contained in Manu National Park, where logging is prohibited. The soils in the forested parts of the catchment are predominantly inceptisols (Asner et al., 2014) with a soil water content $> 25\,\%$ throughout the year (Girardin et al., 2014; Huaraca Huasco et al., 2014). Soils show little evidence of seasonal moisture stress
20   except at the highest treeline sites (Girardin et al., 2014). At lower altitudes there are only short periods at mid-day at the driest time of year which show some signs of moisture stress (Rapp and Silman, 2012). The timberline occurs at 3000 to 3600 m a.s.l. with *puna* grasslands and some shrubland above the timberline (Gibbon et al., 2010). The soils in the *puna* grasslands in the valley are usually saturated for ~ 8 months of the
25   year (November to June; I. Oliveras, personal communication, 2013) due to relatively high precipitation and low temperatures (Wilcox et al., 1988).

3 Materials and methods

3.1 Catchment wide rainfall estimates

Meteorological stations are located throughout the valley along an altitudinal gradient from 887 to 3460 m a.s.l. (Figs. 1a and 2a), distributed in various landcover types and on a range of slopes and aspects (Table S2 in the Supplement). Only data from the Wayqecha meteorological station (at 2900 m a.s.l.) was recorded over the full length of this river study, so rainfall was estimated using the long-term monthly record from $0.25° \times 0.25°$ merged Tropical Rainfall Measuring Mission (TRMM) data (TRMM, 2013) together with the long-term monthly rainfall data from nine meteorological stations (Girardin et al., 2014; Huaraca Huasco et al., 2014; ACCA, 2012; Rapp and Silman, 2012; SENAMHI, 2012).

The 3B43 v7a TRMM is a third level product with outputs in $mm\,d^{-1}$, which have been converted to $mm\,month^{-1}$ with an output each month from 1998 to 2012. The valley is situated entirely within one 3B43 TRMM tile, which covers an area of $\sim 730\,km^2$ centred at $12°7'48''$ S, $71°38'6''$ W (Fig. 1a). The raw TRMM 3B43 data underestimates rainfall in the Andes (Scheel et al., 2011; Bookhagen and Strecker, 2008). Indeed, in the case of the Kosñipata Valley, TRMM 3B43 rainfall is an underestimate compared to nearly all of the data from met station rainfall gauges in the valley and is most comparable to the met stations at high elevations with low rainfall (Fig. 2a). Because of the apparent systematic bias, we did not use the TRMM data directly but instead calibrated the TRMM data using meteorological data to make robust catchment wide rainfall estimates. This had the advantage of allowing us to use the long-term TRMM record that covers periods of time when data is not available from the met stations, since the latter only have sporadic coverage, ranging from 13 to 79 months (Table S2 in the Supplement). Details of the calibration procedure we used are provided in the Supplement.

3.2 Discharge and runoff measures

This study is based on measurements of Kosñipata River discharge made over a one year period (Figs. 1a and 3), focusing on the Kosñipata River gauging station located at San Pedro (13°3′37″ S, 71°32′40″ W), at 1360 m a.s.l.. Field measurements consisted
5 of river height, flow velocity, and cross-sectional area, which together allowed us to estimate discharge and runoff over the study period. For full details of the measurements and corrections see the Supplement; a brief summary is provided here. River stage height was measured from January 2010 to February 2011 using a river logger (Global Water WL16 Data Logger, range 0–9 m), recording river level every $\sim$15 min.
10 The instantaneous discharge associated with each height measurement was calculated based on calibrated stage-discharge relationships. Total monthly discharge was determined by summing over each month, and the monthly totals were converted into an instantaneous discharge ($m^3 s^{-1}$) for each month. Monthly, seasonal and annual discharge and runoff were determined from these values. There was a gap in the logger
15 data of 31 days in the dry season between mid-July and early-August (Fig. 3); these gaps were filled by linear interpolation. This interpolation misses storms, but these should have little influence on the annual discharge because of low flow throughout this period of time. Baseflow was determined from mean daily discharge ($m^3 s^{-1}$) using the method outlined in Gustard et al. (1992). Base flow index (BFI) was calculated
20 as the ratio of the total volume of baseflow divided by the total volume of streamflow.

3.3 Actual evapotranspiration estimates

Actual evapotranspiration (AET) was estimated using the ecophysiologically downscaled PT-JPL (Priestley and Taylor – Jet Propulsion Laboratory) AET method developed by Fisher et al. (2008). This method has been evaluated extensively throughout
25 the tropics (Fisher et al., 2009). The model is based on ecophysiological theory using traits that are measurable in the field or remotely. It takes a bio-meteorological approach incorporating the radiation based model from Priestley and Taylor (1972) to

determine rates of actual evapotranspiration. The model requires only four variables: normalised difference vegetation index (NDVI), net radiation (R_n), maximum air temperature (T_{max}), and minimum relative humidity (RH_{min}). Details of the parameter values selected for actual evapotranspiration estimates are provided in the Supplement.

3.4 Water isotope measurements

River water, rainfall, and cloud water were collected from 2009 to 2011 from a range of elevations throughout the valley. River water was collected from the river surface, passed through a 0.2 µm nylon filter, and stored unpreserved in containers that prevented evaporative loss (see Supplement). Rainfall samples were collected at the time of river water collection near the river gauge, with additional samples collected along an altitudinal transect in the valley between 1500 to 3600 m a.s.l. (Table S3a in the Supplement). Cloud vapour was collected along the altitudinal transect below the canopy using a cryogenic pump (Table S3b in the Supplement).

Isotopic analysis was carried out on the samples to determine δD (delta deuterium, $^2H/^1H$, ‰), $\delta^{18}O$ (delta 18-oxygen, $^{18}O/^{16}O$, ‰) and deuterium excess (defined as $Dxs = \delta D - 8 \times \delta^{18}O$, in ‰), all reported relative to Standard Mean Ocean Water (SMOW). Deuterium excess (Dxs), representing the offset from the meteoric water line (see Supplement), provides information about the source conditions of water vapour (Dansgaard, 1964). It is controlled by kinetic effects during evaporation, where a larger Dxs value is an indicator of enhanced moisture recycling and a lower value indicates an enhanced evaporative loss (Cappa et al., 2003; Salati et al., 1979).

River water, rainfall, and water vapour samples were analysed with a Picarro L1102-i cavity ring down spectrometer (CRDS). River water and rainfall from 2011 were injected 5 times and the final 3 samples averaged. Precision (1σ) was 0.2‰ for $\delta^{18}O$ and 1‰ for δD, though some samples showed larger uncertainties. VSMOW and VSLAP standards were analysed at the same time and were used to assess accuracy and precision of the instrument between runs. Rainfall from 2009 and water vapour were injected 9 times and the final 6 samples averaged. Precision (1σ) was < 0.1‰ for $\delta^{18}O$

and 1 ‰ for δD. Calibration of results to VSMOW was achieved by analysing internal standards before and after each set of 7 to 8 samples. Internal standards SPIT, BOTTY, and DELTA were used to calibrate against VSMOW. Additional analyses using Isotope Ratio Mass Spectrometry (IRMS) were used as a check on the CRDS results (see Supplement).

4 Results

4.1 Catchment wide rainfall

The estimated annual rainfall for the study period (February 2010 to January 2011) was $3028 \pm 414\,\mathrm{mm\,yr^{-1}}$ (Table 2). Based on the long-term calibrated TRMM record, the 15-year (1998 to 2012) mean annual rainfall was $2811 \pm 124\,\mathrm{mm\,yr^{-1}}$, indicating that our river discharge measurements were made in a year with slightly higher than average rainfall (Fig. 4; Table S4 in the Supplement). The total rainfall contribution over the study period was divided into 100 m altitudinal bins to evaluate how rainfall was distributed over the valley. Although most of the catchment area is located at mid to high elevation ranges (~ 2400–$3400\,\mathrm{m\,a.s.l.}$), the rainfall is predominantly sourced from $\sim 2400\,\mathrm{m\,a.s.l.}$ (Fig. 2c).

4.2 Discharge and runoff

The Kosñipata River basin at San Pedro, with a mean elevation of 2805 m a.s.l. and an area of $164.4\,\mathrm{km^2}$, was estimated to have a mean annual discharge of $14.6 \pm 0.7\,\mathrm{m^3\,s^{-1}}$ and an annual runoff of $2796 \pm 126\,\mathrm{mm\,yr^{-1}}$ (Table 2). This falls within the range of 2100 to $3070\,\mathrm{mm\,yr^{-1}}$ for 2 microcatchments in the Ecuadorian Andes, with very similar vegetation cover and elevation ranges (Crespo et al., 2011). In the Kosñipata Valley, 52 % of the annual flow was during the wet season, which covers only 33 % of the year (Table 2).

Baseflow was $2173 \pm 133\,mm\,yr^{-1}$ of the annual total runoff. The base flow index (BFI) is the ratio between the total baseflow volume and total streamflow volume. The BFI value for the Kosñipata (77 %) is consistent with a Mexican TMCF where 90 % of the annual flow was attributed to baseflow (Muñoz-Villers et al., 2012) and is also consistent with the 2 Ecuadorian catchments discussed above, where 80 % of annual flow was attributed to baseflow (Crespo et al., 2011).

4.3 Evapotranspiration

Actual evapotranspiration (AET) was estimated from the PT-JPL model (Fisher et al., 2008) at $909 \pm 182\,mm\,yr^{-1}$. In previous work in lowland tropical forests, AET was estimated to be $1000–1300\,mm\,yr^{-1}$ (Bruijnzeel et al., 2011; Fisher et al., 2009), while TMCF had lower ET values, between 800 and $1200\,mm\,yr^{-1}$ (Bruijnzeel et al., 2011). The AET in the Kosñipata Valley could differ from the PT-JPL model (Fisher et al., 2008) prediction because fog immersion in TMCFs reduces solar radiation and lowers the vapour pressure deficit, triggering a lower atmospheric evaporative demand (McJannet et al., 2010; Letts and Mulligan, 2005; Bruijnzeel and Veneklaas, 1998), while wet leaf surfaces lower transpiration and photosynthesis (Letts and Mulligan, 2005). On the other hand, despite reduced canopy transpiration, ET in TMCFs can be quite high due to an increase in interception evaporation (Giambelluca and Gerold, 2011; McJannet et al., 2010).

To validate the PT-JPL evapotranspiration estimates of the Kosñipata, we compared one month from mid-July to mid-August 2008 when sap flow was predicted to be $53\,mm\,month^{-1}$ using the Soil-Plant-Atmospheric (SPA) model based on Wayqecha forest plot (2900 m a.s.l.) forest canopy sap flow measurements (van de Weg et al., 2014). Sap flow is a proxy for canopy transpiration. For the same time period, the PT-JPL model predicted canopy transpiration at $49\,mm\,month^{-1}$ for the Wayqecha met station nearby the forest plot. This suggests that even though the PT-JPL model has

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not been deployed in TMCF previously, it provides a reasonable canopy transpiration result. Thus, we provide a maximum error of $\sim 20\%$ on AET.

4.4 Isotopic analyses and mixing calculations

Rainwater δD and δ^{18}O values display considerable seasonal variation whereas variation with elevation during a given season is less pronounced (Table S3a in the Supplement; Fig. 5). Rainwater δD and δ^{18}O values are highest during the dry season. Seasonal variation in Dxs is minimal (Fig. 5). The δ^{18}O and δD of Kosñipata cloud water vapours are not clearly distinct from rainwaters. This result departs from the isotopic enrichment found in cloud waters in non-orographic settings, but similarity between cloud and rainwater isotopes has also been found in the few cases of orographic cloud formation that have been studied (Scholl et al., 2011). Despite the overlap in δ^{18}O and δD, the cloud water vapour samples from the Kosñipata Valley have higher and more variable Dxs values than all of the rainwater samples (Fig. 5; Table S3b in the Supplement).

Streamwater δD, δ^{18}O and Dxs values ranged from -94.8 to -64.9‰, -14.5 to -10.9‰, and 19.1 to 22.6‰ respectively (Table S5 in the Supplement). A slight seasonality is apparent in stream water isotopic composition, with slightly higher δD values during the dry season and dry-to-wet season transition (Fig. 6a). A significant seasonal variation in Dxs in streamwater is not apparent (Fig. 6b). See the Supplement for full details on the δD, δ^{18}O, and Dxs isotope results.

Qualitative comparison between the Kosñipata River water isotope data and the rainwater and cloud water isotope data suggests that, throughout the year, wet season precipitation is the dominant contributor to river discharge (Fig. 5). As discussed below, this probably results from the storage of wet season precipitation in groundwater that is released to the stream over time. It is possible that isotopic enrichment may take place via evaporation as water makes its way from precipitation to streamflow, either associated with throughfall (e.g. Brodersen et al., 2000) or in soils (e.g. Dawson and Ehleringer, 1998; Thorburn et al., 1993). Such isotopic enrichment could bias

inferences about water sources using isotopic signatures. However, we note that any evaporative enrichment would act to decrease the relative contribution from wet season rainfall (the depleted source), supporting the qualitative inference that wet season rainfall is the dominant source of streamflow throughout the year. Moreover, the Kosñi-

5 pata streamwaters appear to have little geochemical imprint of evaporation. Chloride concentrations provide a conservative tracer that should be enriched during evaporation; in the Kosñipata samples, Cl concentrations are similar in rainwater (2–20 µM) and streamwater (2–12 µM) (Torres et al., 2014). Kosñipata stream waters also lie on the same local meteoric water line as rainwater (see Supplement), with no evidence for

10 relative D-depletion that may be expected during evaporation.

To quantitatively constrain the relative contributions of different water sources to river discharge, a three end-member mixing model was used (see Supplement for details). In this model, mixing between wet season precipitation, dry season precipitation, and dry season cloud water vapour is considered along with observed variability in the

15 isotopic compositions of each of these end-members (i.e. Phillips and Gregg, 2001). Since we assume minimal evaporative enrichment of water isotopes during runoff generation, the results of this model provide a minimum constraint on the contribution from wet season rainfall. Results of the three end-member mixing calculations are distributions of possible end-member contributions (Fig. 6c–f). For individual samples, mean

20 contributions of wet season rainfall, dry season rainfall, and cloud water vapour to river discharge range from 46–67, 19–33, and 7–31 % respectively (Fig. 6f; Table S6 in the Supplement). Similarly, the maximum likely contributions of each source to a single sample, which we define as the 95th percentile value of the distributions from our mixing calculations, range from 66–87, 38–60, and 19–52 % for wet season precipi-

25 tation, dry season precipitation, and cloud water vapour respectively (Fig. 6c–f). It is worth noting that only two samples ($n = 62$) show mean and maximum likely contributions of cloud water vapour greater than 18 and 40 % respectively (Fig. 6f; Table S6 in the Supplement). These contributions calculated from the water isotope mixing model reflect the ultimate source of the water to stream runoff, with storage and mixing in

groundwater likely to be an important intermediary but one which would not affect the source partitioning.

4.5 Cloud water in streamflow

Isotopic mixing calculations constrain the statistically most likely cloud water vapour contribution to between 7 and 31 % of streamflow, with only 2 samples > 18 % (Table S6 in the Supplement). All samples, except for the two with the highest analytical uncertainties, show this range of cloud water vapour contribution regardless of collection season (Figs. 6f and 7c). Based on our estimated monthly total river discharge and the average values for cloud water contribution in each month, we estimate that total cloud water contribution to streamflow was $316 \pm 116 \, \mathrm{mm\,yr^{-1}}$, using the 50th percentile values of the cloud water fraction and confidence intervals defined by the 5th and 95th percentiles (Tables 3 and S6 in the Supplement). Compared to our annual discharge of $2796 \pm 126 \, \mathrm{mm}$, this means cloud water contributed $11 \pm 4 \, \%$ to annual streamflow. Our estimated cloud water flux to the river falls within the (admittedly very broad) range of CW interception fluxes measured in TMCFs, which range from ~ 50 to $1950 \, \mathrm{mm\,yr^{-1}}$ (Bruijnzeel et al., 2011; Bendix et al., 2008).

Our results suggest that cloud water appears to play a non-negligible role in stream runoff in the Kosñipata River, but that it remains secondary to precipitation inputs even during the dry season when rainfall is at its lowest and cloud immersion is most frequent. Cloud frequency is high in the valley, with cloud cover > 70 % year round (Halladay et al., 2012b). In the dry season the cloud base was > 1800 m a.s.l. 40 % of the time (Rapp and Silman, 2014). Cloud immersion is a key characteristic of tropical montane cloud forest (Bruijnzeel et al., 2011), and provides an important water source to the forest canopy and the diverse epiphyte community (Rapp and Silman, 2014; Bruijnzeel et al., 2011; Giambelluca and Gerold, 2011). However, it is possible that much of the intercepted water is transpired or evaporated directly from the canopy. Overall, cloud water contribution to stream runoff supplies a relatively constant proportion of

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total flow throughout the year and never dominates water inputs to the Kosñipata River, even during times of the lowest flow (Table 3).

5 Discussion

5.1 Water balance

The annual water balance for the Kosñipata Valley can be described by the following equation (water inputs to the catchment on the left, losses from the catchment on the right):

$$\text{Rainfall} + \text{CW} = \text{AET} + \text{Runoff} + \text{Residual} \tag{1}$$

Rainfall was estimated catchment-wide from TRMM and met station rainfall at $3028 \pm 414\,\text{mm yr}^{-1}$. Cloud water (CW) was estimated from the isotope mixing model at $316 \pm 116\,\text{mm yr}^{-1}$. Actual evapotranspiration (AET) was estimated from the PT-JPL model (Fisher et al., 2008) at $909 \pm 182\,\text{mm yr}^{-1}$. Runoff was estimated from the gauging station at $2796 \pm 126\,\text{mm yr}^{-1}$. The residual of Eq. (1) sums to $-361 \pm 466\,\text{mm}$, which is $11 \pm 14\,\%$ of total annual water inputs through precipitation and CW, indicating that any imbalance within our budget is within the estimated uncertainties of the water balance calculation.

There are several additional structural uncertainties in the water balance. Rainfall was estimated for the catchment by calibrating TRMM rainfall with actual rainfall collected from 9 gauging stations. In the Kosñipata Valley there was a decrease of rainfall with an increase of elevation, corresponding to an average annual rainfall gradient of $\sim 148\,\text{mm}$ per hundred metres (Fig. 2a). It is possible that rainfall deviates from this trend along the altitudinal gradient because our results are limited to 9 meteorological stations dispersed over a large area (Fig. 1a). Intense localised storm activity also increases the chance of underestimating precipitation. Additionally, the types of rain gauges used in the valley (Table S2 in the Supplement) are not ideal for cloud forests

due to an underestimation of wind driven precipitation on steep slopes (Bruijnzeel et al., 2011; Frumau et al., 2011). Stream runoff can be overestimated in mountain rivers due to an overestimation of velocity by taking measurements predominantly near the surface of the channel (Chen and Chiu, 2004; Thome and Zevenbergen, 1985). We have taken this under consideration and corrected surface velocity to estimate mean channel velocity (following Eq. S1 in the Supplement), but it is possible that our runoff values remain overestimated. Taking these methodological uncertainties into consideration, our rainfall input value may be conservative, and stream runoff output value may be an upper bound.

5.2 Hysteresis

5.2.1 Characterizing hysteresis

A monthly breakdown of the water balance shows the distribution of annual residual when water is going into storage (+) and when water is coming out of storage (− ; Fig. 8). The mid-wet season (January and February) was a time of recharge with positive residual values. This store was subsequently drained as discharge to stream runoff in the wet–dry transitional season (April) and most of the dry season (May to August), both of which showed negative residual values (Fig. 8). This illustrates how rainfall stored during the wet season plays an important role in sustaining steady dry season runoff. The results of the isotope mixing analysis confirm this inference by showing that wet season rainfall is still prominent in contributing to streamflow in the dry season. Sources of streamflow from May to September 2010 were $61 \pm 9\%$ from wet season rainfall, $25 \pm 9\%$ from dry season rainfall, and $12 \pm 7\%$ from cloud water (Table 3).

At seasonal timescales, streamflow and baseflow in the Kosñipata Valley both follow an annual anti-clockwise hystereses pattern (Fig. 9). This is similar to that observed by Andermann et al. (2012) in the Nepalese Himalaya. In the wet season (December to March), the catchment wide rainfall in the Kosñipata Valley was greater than streamflow and baseflow (Fig. 9a and c). During the wet–dry transition season (April), and the

start of the dry season (May and June) however, there was a switch and streamflow and baseflow were greater than rainfall. By the middle of the dry season (July and August) rainfall was equal to streamflow and baseflow. By the time of the late dry season (September), rainfall started to increase and was greater than streamflow and baseflow. The dry-wet transition season (October and November) had higher rainfall than streamflow and baseflow. Finally, in the early wet season (December to January) the cycle was completed where the contribution of rainfall dominated streamflow and baseflow (Fig. 9a and c). The annual anti-clockwise hysteresis was even more pronounced when streamflow and baseflow were compared to the rainfall gathered over the study period at the Wayqecha meteorological station at 2900 m a.s.l. (Fig. 9b and d).

The water isotope data support the indication of seasonal hysteresis observed in the water balance estimates. The relationship between river discharge and δD showed a seasonal clockwise hysteresis, but this was not observed in Dxs (Fig. 7a and b). Considering the observed end-member δD and Dxs compositions, this implies that there was seasonal variation in the relative contributions of wet and dry season rainfall but not cloud water vapour (Fig. 7c–e). The Monte-Carlo derived confidence intervals on the mixing results provide large ranges. However, the mean results (Fig. 7), which best represent each end-member composition, show a seasonal anti-clockwise hysteresis between river discharge and the mean contribution of wet-season precipitation that is consistent with the hysteresis observed in the water balance. Dry season and dry-wet transitional season runoff appear to be sustained by relatively lower, but still dominant, contributions from wet season precipitation (Fig. 7d). A corresponding variation in the contribution of dry season precipitation with discharge is also evident, whereby dry season and dry-wet transitional season runoff is composed of a larger proportion of dry season rainfall (Fig. 7e). The hysteresis in the mixing model results is attributable to the seasonal hysteresis in streamwater δD. No seasonal hysteresis in the contribution of cloud water interception to river discharge is apparent (Fig. 7c), consistent with no seasonal pattern in the streamwater Dxs.

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The consistent, annual anti-clockwise hysteresis in both the water balance and the contribution from different sources inferred from the water isotopes indicate that there are important factors other than the storm runoff response that influenced hydrologic variability throughout the year in the Kosñipata Valley. The lag in runoff can be explained by a significant portion of wet season rainfall being stored, and then several months later discharged as runoff in the wet–dry transition and dry seasons (Fig. 9a and Table 2). The delay in rainfall to streamflow runoff helps provide water in the catchment at times of lower rainfall, stabilising dry season runoff.

5.2.2 Can soil water explain seasonal hysteresis?

There are several potential mechanisms causing a seasonal lag in streamflow. The water isotope data points to rainfall, rather than cloud water, as the primary source of water, but it is still unclear how rainfall is stored temporarily over the year. Shallow groundwater (i.e., lateral flow through soil layers) derived from accumulation of water in soils during the wet season may contribute to the delayed stream runoff. In the Kosñipata Valley, shallow groundwater may be sourced from drainage of saturated *puna* grassland soils. In *páramo* wetlands (a wetter mountain top biome) in the northern Andes of Ecuador, delayed groundwater has been shown to play an important role in dry season runoff (Buytaert and Beven, 2011). Tropical montane cloud forest soils, as found in a similar forest in Ecuador, can also be a potential source for delayed runoff over shorter periods of ~ 3.5 to ~ 9 weeks (Timbe et al., 2014).

In order to evaluate whether seasonal variations in soil water content are sufficient to account for the variability that would be needed to explain the seasonal lag in runoff in the Kosñipata, we use the following equation:

$$A_{\text{catchment}} \times \text{ED} = A_{\text{storage}} \times d \times \Delta V \tag{2}$$

where $A_{\text{catchment}}$ is the area of the drainage basin (m^2), ED is the excess discharge (mm) consisting of the sum of the monthly residual values from the wet–dry transitional

season and the dry season (April to September; Fig. 8), $A_{storage}$ is the area of the basin covered in soil (m^2), d is depth of soil layer (m), and ΔV is the variation in soil water content that needs to occur to account for the excess discharge. Since the area of the catchment and area covered in soil are approximately the same, the area variables in Eq. (2) cancel out. Equation (2) can be rearranged to solve for the variability in soil water:

$$\Delta V = \frac{ED}{d} \tag{3}$$

For our calculation, we assume mean soil depth (d) to be ~ 0.5 m, consistent with data from the Kosñipata catchment (Gibbon et al., 2010; Zimmermann et al., 2009). Using the mean CW monthly input into streamflow (Table 3), the excess discharge (ED) from April to September was 458 mm. Hence, from Eq. (3), volumetric water content would need to have a seasonal range of $\sim 90\,\%$ to fully account for the seasonal excess discharge. Typical soil water content in the TMCF and puna is around 40 % and shows a seasonal variation of $< 5\,\%$ (Girardin et al., 2014; Huaraca Huasco et al., 2014). This variability in soil moisture is an order of magnitude smaller than required to explain sustained streamflow; hence, the excess dry season discharge cannot be accounted for by seasonal storage and drainage of soils.

5.2.3 Importance of groundwater in hysteresis

If soil water content changes are insufficient to account for the excess dry season discharge, the source of this excess discharge is likely to be groundwater stored within the fractured bedrock below the shallow soil layer. In central eastern Mexico, groundwater in the TMCF was found to be an important component of dry season runoff (Muñoz-Villers et al., 2012). Groundwater occurs mostly in permeable bedrock and within fractures of impermeable bedrock (Jardine et al., 1999; Gascoyne and Kamineni, 1994; Todd and Mays, 1980). In the Nepal Himalayas, deep groundwater recharges through fractured bedrock containing aquifers several tens of meters deep and has a storage

residence time of ∼ 45 days (Andermann et al., 2012). Fracturing and the exposure of bedding planes through the process of uplift and erosion in the Kosñipata Valley (Carlotto Caillaux et al., 1996) could provide conduits that aid in deep groundwater flow. In the Kosñipata Valley, ∼ 80 % of the catchment area consists of sedimentary mudstones and shale, and ∼ 20 % consists of plutonic intrusions (Table 1). Shale has a very low porosity and permeability (Domenico and Schwartz, 1990; Morris and Johnson, 1967), but when fractured its porosity is greatly increased (Jardine et al., 1999). Plutonic intrusions, as found in lower parts of the catchment, also have increased porosity as a result of fracturing (Gascoyne and Kamineni, 1994). Thus we view deep fractured bedrock as the likely transient storage reservoir explaining the annual hydrograph in the Valley.

The observation of a significant role for seasonal groundwater storage and release in Kosñipata River has implications for understanding Andean water resources, predicting flooding, and quantifying biogeochemical fluxes, and particularly for assessing how these may respond to changing climate. The rate of warming over the next 100 years in the region of the Kosñipata Valley is expected to proceed an order of magnitude faster than the 1 °C increase in temperature per 1000 years during the Pleistocene-Holocene (Bush et al., 2004). The observation of upslope migration of plant species already indicates a dramatic pace of change in the Kosñipata (Tovar et al., 2013; Feeley et al., 2011; Hillyer and Silman, 2010). It remains unclear how patterns of rainfall and cloud frequency have been changing and will change in the future (Halladay et al., 2012a; Rapp and Silman, 2012), much less how the hydrologic system will respond, both to changes in magnitude and in seasonality of precipitation sources. The baseline of water isotope data, the partitioning of precipitation sources, and the conceptual framework presented in this study offer the potential to help understand what hydrologic responses might be expected if precipitation changes (e.g. as evaluated in Puerto Rico; Scholl and Murphy, 2014). Further exploration and verification of the observations in this study, for example by considering longer-term water budgets (Andermann et al., 2012) and/or detailed analysis of stream hydrochemistry (Calmels et al., 2011; Tipper et al., 2006),

would strengthen understanding of how Andean TMCFs function hydrologically today and how this function may evolve in the future.

6   Conclusions

An annual water budget for the Kosñipata Valley indicates that 3028 ± 414 mm was contributed to the catchment by rainfall, 316 ± 116 mm was supplied by cloud water, 2796 ± 126 mm was removed as streamflow, and 909 ± 182 mm was lost through actual evapotranspiration. The annual water budget balances within ~ 10 %, which likely reflects uncertainties in our measurements. Annual stream runoff was composed of 60 ± 5 % wet season rainfall, 26 ± 5 % dry season rainfall, and 11 ± 4 % cloud water. Baseflow contributed 77 % of the streamflow over the one year of study. Runoff followed an annual anti-clockwise hysteresis with respect to rainfall, exhibiting a lag in stream runoff that maintained stream water flow in the early dry season. Dry season runoff was composed of 61 ± 9 % wet season rainfall. The contribution from cloud water, although important to the TMCF ecology, plays a secondary role in river streamflow (~ 10 %) in this catchment, even during the low flow of the dry season. Excess discharge measured throughout the wet–dry transitional season and dry season (April to September) was ~ 460 mm. Deep groundwater in fractured rock is probably the cause of this seasonal lag in stream runoff, as seasonal variation in soil water content would be insufficient.

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Table 1. Descriptions of the Kosñipata Valley.

Catchment	Area (km^2)	Mean slope[a] (°)	Mean elevation[a] (m a.s.l.)	Elevation range[a] (m a.s.l.)	Landcover type[c] (∼ %)	Geology[b] (∼ %)	Gauge lat/long (S, W)	Gauge elevation (m a.s.l.)
Kosñipata at San Pedro	164.4	28	2805	1360 to 4000	TMCF (80), puna /transition (20)	mudstones (80), pluton intrusions (20)	13°3′37″, 71°32′40″	1360
Kosñipata at Wayqecha[d]	48.5	26	3195	2250 to 3905	TMCF (50), puna /transition (50)	mudstones (100)	13°9′46″, 71°35′21″	2250

[a] Based on Shuttle Radar Topography Mission (SRTM) data with a 90 m × 90 m resolution.
[b] Basin geology derived from Carlotto Caillaux et al. (1996).
[c] Landcover types were determined using 2009 Quickbird 2 imagery.
[d] Results presented in Supplement.

Table 2. Streamflow and rainfall for the Kosñipata Valley at the San Pedro (SP) gauging station, for the annual period from February 2010 to January 2011.

	number of months	Q $(m^3 s^{-1})$	Runoff $(mm\,d^{-1}, \%)$	Baseflow $mm\,d^{-1}$ (%)	BFI[a]	Rainfall[b] $mm\,d^{-1}$ (%)
Wet	4	23.1 ± 1.3	12.13 ± 0.68 (52)	9.41 ± 0.77 (52)	0.77 ± 0.04	14.59 ± 3.08 (58)
Wet-dry	1	19.6 ± 2.6	10.29 ± 1.37 (11)	8.75 ± 1.35 (12)	0.85 ± 0.02	6.76 ± 2.58 (7)
Dry	5	8.1 ± 0.9	4.31 ± 0.46 (24)	3.58 ± 0.48 (26)	0.83 ± 0.04	4.21 ± 0.73 (21)
Dry-wet	2	11.3 ± 1.5	5.94 ± 0.81 (13)	3.56 ± 0.73 (10)	0.60 ± 0.04	6.83 ± 1.95 (14)
Annual	12	14.6 ± 0.7	7.66 ± 0.35 (100)	5.95 ± 0.37 (100)	0.77 ± 0.04	8.30 ± 1.13 (100)

Seasonal contribution as percentage of total in parenthesis.
Uncertainties are propagated 1σ errors.
[a] Base flow index (BFI) is the ratio of the total volume of baseflow divided by the total volume of discharge following the method outlined in Gustard et al. (1992).
[b] Catchment-wide rainfall is reported for February 2010 to January 2011 to coincide with the study period.

Table 3. Breakdown of streamflow into its sources.

	n	Fraction wet season rainfall[a]	Fraction dry season rainfall[a]	Fraction cloud water[a]	Wet season rain as a source (mm month^{-1})[b]	Dry season rain as a source (mm month^{-1})[b]	Cloud water as a source (mm month^{-1})[b]	Total stream runoff (mm month^{-1})
10 Feb	28	0.62 ± 0.04	0.24 ± 0.04	0.11 ± 0.03	231 ± 16	91 ± 17	41 ± 13	372 ± 38
10 Mar	2	0.62 ± 0.15	0.25 ± 0.16	0.10 ± 0.12	261 ± 70	105 ± 72	44 ± 56	420 ± 41
10 Apr	2	0.65 ± 0.14	0.21 ± 0.14	0.11 ± 0.12	200 ± 49	66 ± 49	35 ± 42	309 ± 41
10 May	3	0.61 ± 0.12	0.25 ± 0.13	0.11 ± 0.10	143 ± 34	59 ± 35	27 ± 28	235 ± 40
10 Jun	1	0.63 ± 0.20	0.22 ± 0.20	0.12 ± 0.17	103 ± 40	36 ± 40	19 ± 35	163 ± 35
10 Jul	2	0.58 ± 0.13	0.28 ± 0.14	0.12 ± 0.11	61 ± 17	28 ± 18	13 ± 14	105 ± 30
10 Aug	3	0.58 ± 0.13	0.27 ± 0.14	0.12 ± 0.10	51 ± 15	24 ± 16	10 ± 12	87 ± 26
10 Sep	3	0.59 ± 0.13	0.26 ± 0.23	0.12 ± 0.11	42 ± 12	19 ± 13	8 ± 10	70 ± 23
10 Oct	1	0.64 ± 0.22	0.26 ± 0.23	0.08 ± 0.16	127 ± 53	51 ± 55	16 ± 39	199 ± 36
10 Nov	3	0.52 ± 0.14	0.30 ± 0.15	0.14 ± 0.12	86 ± 28	50 ± 31	23 ± 25	164 ± 34
10 Dec	4	0.56 ± 0.12	0.30 ± 0.13	0.10 ± 0.09	188 ± 44	98 ± 50	35 ± 33	333 ± 42
11 Jan	2	0.55 ± 0.16	0.29 ± 0.18	0.14 ± 0.14	186 ± 62	99 ± 69	46 ± 54	339 ± 43
Fractional contributions by season[c]								
Wet		0.59 ± 0.07	0.27 ± 0.08	0.11 ± 0.06				
Wet–dry		0.65 ± 0.14	0.21 ± 0.14	0.11 ± 0.12				
Dry		0.61 ± 0.09	0.25 ± 0.09	0.12 ± 0.07				
Dry–wet		0.59 ± 0.16	0.28 ± 0.17	0.11 ± 0.13				
Annual		0.60 ± 0.05	0.26 ± 0.05	0.11 ± 0.04				

[a] Calculated from monthly average values of mixing model results. Reported errors are propagated uncertainty (1σ) from individual samples per month, accounting for uncertainties from the Monte Carlo mixing modelling (Table S6 in the Supplement).
[b] Calculated from monthly fractional contributions and monthly runoff. Reported errors are propagated uncertainty (1σ) from the mixing modelling and from the variation in stream runoff.
[c] Calculated based on runoff totals for each month, from each source, summed for a given season. Reported errors are propagated uncertainty from monthly runoff estimates from each source.
n = number of samples.

Figure 1. (a) The Kosñipata Valley, Eastern Andes of Peru, showing the Kosñipata River catchment measured at the San Pedro (SP) river gauging station and the nested sub-catchment at the Wayqecha (WQ) river gauging station, overlaid on 90 m × 90 m digital elevation model (Shuttle Radar Topography Mission) (Farr et al., 2007). Black box indicates the extent of the TRMM 3B43 tile used in this study (cf. Fig. 2a). The meteorological stations used for rainfall data are numbered 1 to 9 (Table S2 in the Supplement). **(b)** The Kosñipata River flows into the Alto Madre de Dios (AMdD) and then into the Madre de Dios River, a major tributary of the Amazon River **(c)**. The river network was produced from HydroSHEDS (hydrological data and maps based on shuttle elevation derivatives at multiple scales) (Lehner et al., 2008).

Figure 2. Mean monthly rainfall data for the 9 meteorological stations in the Kosñipata Valley study area (from ~ 900 to ~ 3500 m a.s.l.) (dark dashes and light error bars) and estimated mean monthly rainfall (grey circles and dark error bars) covering the months of the 1-year study period (February 2010 to January 2011) determined using the linear regression equations for each met station derived from tropical rainfall measuring mission (TRMM) data (Table S7 in the Supplement). The grey line is the linear fit with elevation for the estimated mean monthly rainfall (mm month^{-1} = $-0.1216 \pm 0.0187 \times$ elevation + 593.16 ± 44.94, $R^2 = 0.86$; $P = 0.0003$). The error bars are 2 × standard error of monthly data. Mean monthly rainfall for 5 meteorological stations outside of the study area but within the larger Madre de Dios Basin are also shown as triangles (Rapp and Silman, 2012). The shaded box shows the TRMM 3B43 v7a monthly mean rainfall for the 1-year study period with 2 × standard error. Elevation range is shown for the 34.5 km × 34.5 km TRMM tile. The altitudinal range of the study area is represented by the dashed lines at 1350 and 4000 m a.s.l. **(b)** Linear regressions of estimated catchment wide rainfall by month from February 2010 to January 2011, colour-coded by season. The distribution of annual rainfall with elevation by season for the Kosñipata River are shown for the San Pedro (SP) gauging station **(c)** and at the Wayqecha (WQ) gauging station **(d)** at 100 m intervals using the monthly linear regressions **(b)**.

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Figure 3. Runoff for the Kosñipata River at the San Pedro (SP) and Wayqecha (WQ) gauging stations. Rainfall (top axis) from the Wayqecha meteorological station is on the secondary axis. The Kosñipata River runoff at San Pedro and baseflow were measured nearly continuously through the year, with a 31-day gap partly in July and August that is covered by three manual measurements and the gap filled using linear interpolation. The Kosñipata River runoff at Wayqecha was measured throughout the year from a daily to a monthly interval and is discussed in the Supplement.
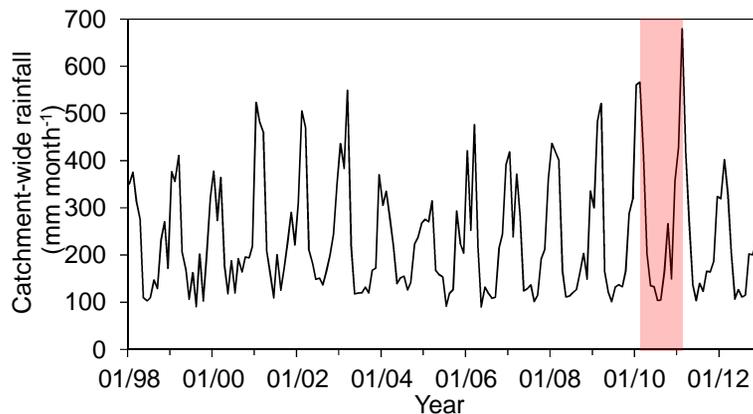
Figure 4. Catchment wide TRMM calibrated rainfall for the Kosñipata Valley from 1998 to 2012. The red shaded area represents the 1-year study period.

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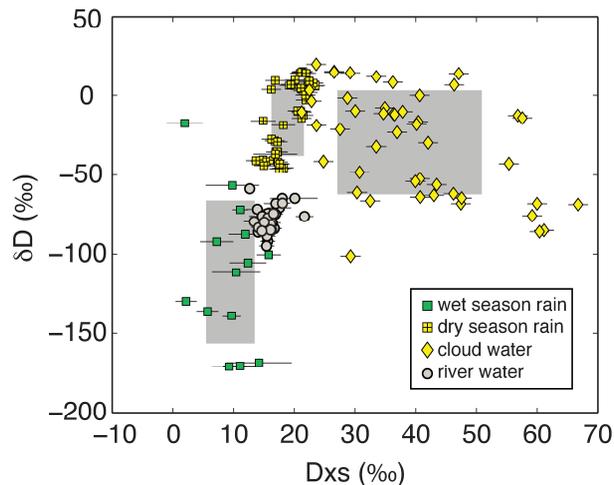
Figure 5. Hydrogen isotope ratio (δD, ‰) and deuterium excess (Dxs, ‰) of dry season cloud water vapour (yellow diamonds), and river water (grey circles) from the Kosñipata Valley. Rainwater samples (squares) are from the dry season (May to August, yellow) and from the wet season (December to March, green). All error bars correspond to two standard deviations. The grey shaded regions encompass the mean δD and Dxs values and one standard deviation for each end-member (i.e. wet season rainfall, dry season rainfall, and dry season cloud water vapour). The ranges defined by these grey boxes were used to generate random sets of end-member compositions for the three end-member mixing model.

Figure 6. Time series of river water hydrogen isotopes (δD_{river}) and river water deuterium excess (Dxs_{river}), and the calculated mixing proportions of different sources for the Kosñipata River. **(a)** The time series of δD_{river} values with error bars signifying 2 standard deviations. **(b)** The time series of Dxs_{river} values with error bars signifying 2 standard deviations. Time series of the 5th (lower error bar), 50th (open circle), and 95th percentile (upper error bar) values of the distribution of fractional contributions to river discharge, calculated using the three end member mixing model, of: **(c)** wet season rain; **(d)** dry season rain; and **(e)** cloud water vapour. **(f)** Time series of the mean contributions of wet season rain (circle), dry season rain (square), and cloud water vapour (diamond) to river discharge.
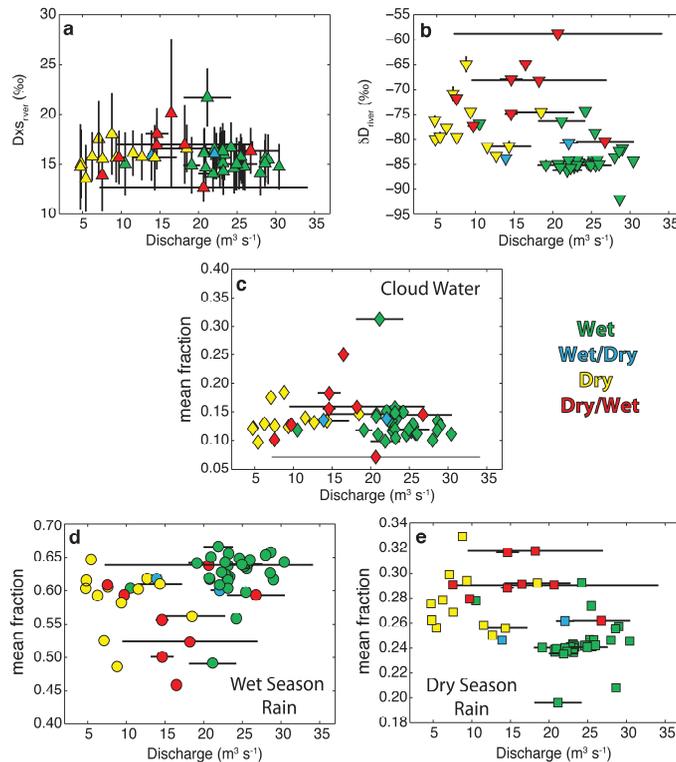
Figure 7. Variation in the isotopic composition of river water **(a)** deuterium excess (Dxs_{river}, %) and **(b)** hydrogen isotope ratio (δD_{river}, ‰) plotted vs. discharge ($m^3 s^{-1}$). Variation in the mean contributions to river flow as a function of water discharge for cloud water vapour **(c)**, wet season rain **(d)**, and dry season rain **(e)** as calculated by the end member mixing analysis, also plotted vs. discharge.
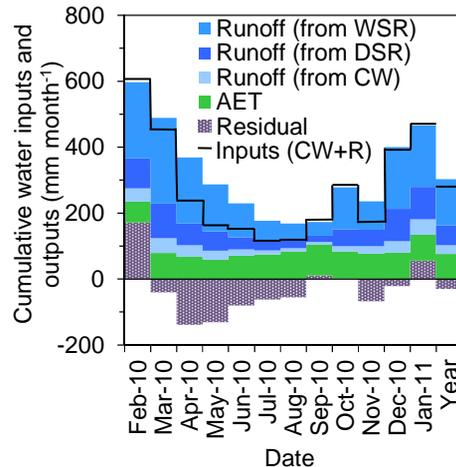
Figure 8. Cumulative water inputs (rainfall and cloud water interception) are represented by the black line. Cumulative water outputs (river runoff and actual evapotranspiration) and the residual are separated out into cumulative coloured stacked bars. Runoff is separated into its 3 sources: wet season rainfall (WSR), dry season rainfall (DSR), and cloudwater (CW) (Table 3). The study period is separated by month and the monthly balance is determined for the study year, February 2010 to January 2011.
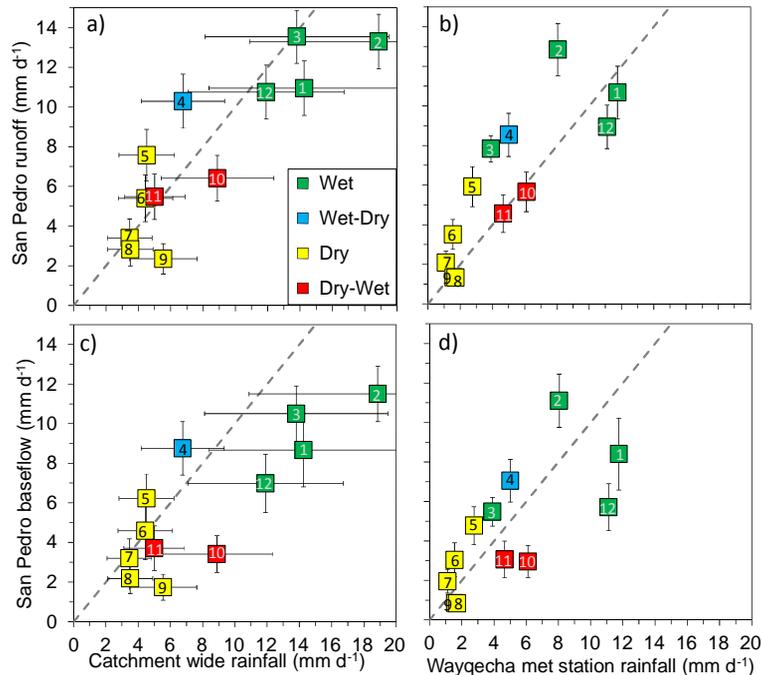
Figure 9. Mean monthly rainfall vs. river runoff ($mm\,d^{-1}$) for the Kosñipata Valley, showing anticlockwise hysteresis throughout the year, with months numbered chronologically and colour coded by season (see Figs. 2 and 7). Plots show stream runoff vs. **(a)** catchment wide rainfall and **(b)** meteorological station rainfall for the Wayqecha met station (2900 m a.s.l.), and baseflow vs. **(c)** catchment wide rainfall and **(d)** meteorological station rainfall at Wayqecha. Error bars represent one standard deviation. The one-to-one line for rainfall to river runoff is represented by the grey dashed line. Note: in **(b)** and **(d)** days with zero rainfall were excluded as per the approach used by Andermann et al. (2012).

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