Available online at www.sciencedirect.com

SCIENCE DIRECT®

Forest Ecology and Management 224 (2006) 95–103

Forest Ecology and Management

www.elsevier.com/locate/foreco

Reversibility of forest conversion impacts on water budgets in tropical karst terrain

David G. Chandler *

Department of Plants, Soils and Biometeorology, Utah State University, 4820 Old Main Hill, Logan, UT 84322-4820, USA

Abstract

A conceptual model of the control of tropical land use and vegetative cover on bedrock recharge is developed for highly permeable geologic substrates. A case study of water budgets is then developed from field data and simple modeling for upland sites with three different vegetative covers (cropland, intensively grazed pasture and forest regrowth) in Leyte, Philippines. Water budget model results show that annual precipitation is divided primarily between evapotranspiration and overland flow for the pasture, but apportioned more to evapotranspiration and inputs to bedrock storage for the crop and forest sites. Modeled evapotranspiration from the forest (1906 mm) was not sufficiently greater than that for either the crop (1661 mm) or pasture (1476 mm) sites to offset the greater overland flow from those sites. The differences in overland flow are related to depth profiles of soil bulk density, which decreased between crop and forest and increased between crop and pasture, and drainable porosity, which increased between crop and forest and decreased between crop and pasture. Dry season streamflow is assumed to be primarily base flow and dependent on wet season bedrock recharge, which was dramatically lower for the pasture (106 mm) than for the crop (1134 mm) or forest covers (1320 mm), for 2946 mm of rainfall. The results support the premise that for landscapes with adequate storage in bedrock fractures, forest regrowth can increase recharge to perched aquifers, and hence dry season baseflow, relative to cropping and that dramatic reductions in overland flow and increases in dry season baseflow may be achieved by reforestation of compacted pastures.

Keywords: Hydrological balance; Bedrock; Baseflow; Bulk density; Hydrologic pathways

1. Introduction

Increases in seasonal flooding and upland water shortages have been a persistent problem in many areas of the humid tropics where forests have been converted to agriculture and pasture (Hamilton and King, 1983), particularly in regions with thin soils and karstic geomorphology (Urich, 1993; McDonald et al., 2002). Numerous studies have been conducted to clarify the impact of forest harvesting and regrowth on annual water yield (Bosch and Hewlett, 1982) as well as on low flows (Johnson, 1998). Changes in hydrologic response induced by land cover change may result from a shift in the water balance between evapotranspiration and runoff processes. These changes can be dramatic in the humid tropics due to extremes in rainfall intensity, soil hydraulic conductivity and topography. In a recent review of the influence of forest cover on annual water yield and seasonal flows in Southeast Asia, Bruijnzeel (2004) found evidence that annual water yield increased primarily in proportion to the extent of biomass removal and secondarily to the extent of surface disturbance, supporting the premise that the degree to which forest cover increases storage depends on the extent to which infiltration capacity is increased beyond the greater transpiration by trees during forest regrowth (Bruijnzeel, 1990). However, Bruijnzeel (2004) found conflicting evidence concerning whether baseflow will increase with forest regrowth following low disturbance conversions, raising the question of the importance of site geology on the response. Smakhtin (2001) noted that contradictory results are theoretically possible within a single site, depending on the time since conversion. Initially following forest conversion, reduced transpiration, interception and infiltration capacity may increase both surface runoff and soil moisture, which could presumably increase deep drainage in the short term. Eventually, the reduced recharge to groundwater caused by increased interception and transpiration of the regenerating forest could lead to a decline in deep drainage at the same site. This observation correctly identifies that base flows may depend on drainage from soil moisture storage, groundwater or both; that both may be impacted by changes in evaporation due

* Tel.: +1 435 797 7326; fax: +1 435 797 3376.
E-mail address: david.chandler@usu.edu.

0378-1127/$ – see front matter
doi:10.1016/j.foreco.2005.12.010

to land cover change; and that the timing of the base flow response to forest conversion depends on the affected hydrologic pathways.

Relatively few studies have been conducted on the hydrologic impact of reforestation of intensively farmed lands. This literature is dominated by studies of commercial eucalyptus plantations, which are renowned for their high water use (e.g. Cornish and Vertessy, 2001). These results support the hypothesis that the decrease in low flows following high disturbance conversions may not be reversible through reforestation if the change in hydrologic response results in a loss of soil water storage capacity prior to the establishment of forest cover (Bruijnzeel, 2004). Whereas, this view is likely correct for low flows derived from soil drainage (Anderson and Burt, 1980), it is less likely to be the case for baseflow derived from groundwater drainage via fracture flow or from perched aquifers above a hydrologic discontinuity (Smakhtin, 2001), as in karst terrain, or if the soil depth in the catchment is inadequate for trees to develop deeper rooting systems than grasses (Trimble et al., 1963; Andreassian, 2004).

This paper presents a case study of water budgets for three land covers after forest conversion in karst terrain. First, a conceptual model for the difference between catchments underlain by high permeability bedrock and those underlain by low permeability bedrock, typically used in hydrologic studies, is presented. Second, water budgets are presented for three zero-order catchments: cropped (*Zea mais*), heavily grazed pasture (*Paspalum conjugatum*) and secondary forest (*Leucaena leucocephala*). The water budgets are developed by building on the rainfall–runoff relationships previously presented by Chandler and Walter (1998) for a karst catchment in Leyte, Philippines with supplemental data and modeling. The model results are then used to clarify the relationships among runoff, evapotranspiration, soil moisture storage and bedrock recharge (assumed to control base flow) for each of the land use/land cover types.

2. Conceptual model

Annual water yield is typically represented in terms of a simplified water balance equation:

$$P = R + \mathrm{ET} + \Delta S \tag{1}$$

where P is precipitation (mm), R the runoff (mm), ET the evaporation and transpiration (mm), ΔS the change in water storage (mm).

In order to balance the *annual* water budget (Eq. (1)), hydrologists often avoid ‘‘leaky catchments’’ by selecting experimental sites with low permeability bedrock, assuming that the change in annual storage (defined for the depth of the soil) and groundwater recharge (often defined as a loss from the system) are negligible. If these assumptions are well justified for the defined conditions, the annual water budget is simply a division of annual precipitation between runoff and evapotranspiration. If groundwater recharge is not negligible, the annual water budget equation gains a term:

$$P = R + \mathrm{ET} + \Delta S + \mathrm{GW} \tag{2}$$

where GW is groundwater recharge (mm).

This representation of the water budget presents a ‘dilemma of residuals’ for achieving closure: either evapotranspiration or recharge to groundwater, which are both difficult to measure, must be quantified, and the residual term will include the error in all other terms. Whereas, current methods of estimating evapotranspiration have an associated error of up to 20%, the annual water balance approach can be problematic for calculating groundwater recharge (Bigelow, 2001). This error decreases at event-scale time steps for large rainfall amounts because ET is small relative to rainfall, which is partitioned among changes in storage, surface runoff (SRO) and deep drainage. Confidence in ET, and therefore in the water balance approach, is improved if the temporal record of change in storage can be replicated at intermediate time steps in a modeling exercise, since change in storage below field capacity is driven primarily by evapotranspiration (Kendy et al., 2003).

A more detailed conceptual model of the water balance components and fluxes which control hydrologic response at the event time scale are depicted in Fig. 1. The greater detail may be considered as an expansion of the runoff and storage terms in Eq. (1).

Runoff is divided into surface runoff, which occurs at the *event* scale, interflow which varies over periods somewhat longer than the *event* scale and baseflow and responds to large events and varies *seasonally to annually*, depending on the geology, soils vegetation and climate:

$$R = \mathrm{SRO} + \mathrm{IF} + \mathrm{BF} \tag{3}$$

where SRO is surface storm flow (mm), IF the subsurface storm flow, or interflow (mm), BF the base flow (mm).

For event water budgets, soil moisture storage and canopy storage or interception become important (Kendy et al., 2003; Bigelow, 2001). For landscapes developed in highly permeable substrates, such as karst, storage in perched aquifers may be important to consider at the event, seasonal and annual time scales. The conceptual model presented here divides the storage component into canopy storage of intercepted precipitation, soil moisture storage, and bedrock storage.

$$\Delta S = \Delta S_{\mathrm{c}} + \Delta S_{\mathrm{rz}} + \Delta S_{\mathrm{ds}} + \Delta S_{\mathrm{br}} \tag{4}$$

where ΔS_{c} is the change in canopy storage (mm), ΔS_{rz} the change in soil moisture storage within root zone (mm), ΔS_{ds} the change in soil moisture storage below root zone (mm), ΔS_{br} the change in bedrock storage in perched aquifer (mm).

Substitution of Eqs. (2) and (3) into Eq. (1) and rearrangement yields:

$$P - \mathrm{SRO} - \mathrm{IF} - \mathrm{ET} - \Delta S_{\mathrm{c}} - \Delta S_{\mathrm{rz}} - \Delta S_{\mathrm{ds}} = \mathrm{BF} + \Delta S_{\mathrm{br}} + \mathrm{GW} \tag{5}$$

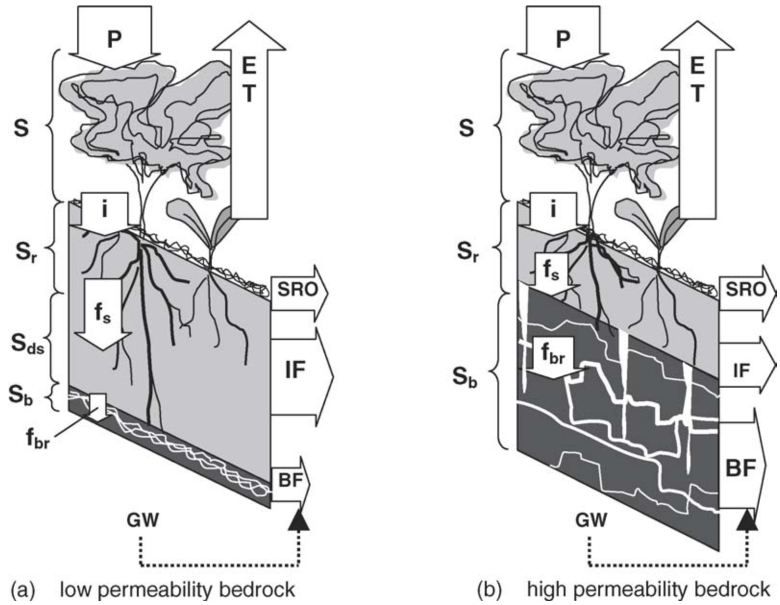

Fig. 1. Schematic representation of the stores and fluxes in low permeability bedrock and high permeability bedrock catchments. Fluxes include precipitation (P), infiltration (i), vertical flux of water through the soil (f_s) and bedrock (f_{br}), evapotranspiration (ET), surface runoff (SRO), interflow (IF), baseflow from perched aquifers and bedrock fractures (BF) and groundwater flow from below the regional phreatic surface (GW).

The conceptual model of the expanded water balance (Eq. (5)) is illustrated for two cases: (1) a typical study catchment characterized by low permeability bedrock with deep soil (Fig. 1a); and (2) a leaky catchment characterized by high permeability bedrock with thin soil (Fig. 1b). For the low permeability bedrock catchment, bedrock storage and losses to groundwater are assumed to be negligible and soil moisture storage is divided between shallow (root zone) storage and deep soil storage. For periods with little rain, no SRO or IF is assumed and

$$P - \Delta S_{rz} = \mathrm{ET} + \Delta S_c \tag{6}$$

By substitution, Eq. (5) is reduced to:

$$\mathrm{BF} = -\Delta S_{ds}, \tag{7}$$

indicating than dry season baseflow is dependent on deep soil drainage.

For the high permeability bedrock catchment, the bedrock storage and losses to groundwater are important components of the water balance and for thin soils the root zone may be assumed to occupy the complete depth of the soil. For periods with little rain, Eq. (5) is reduced to:

$$\mathrm{BF} = -\Delta S_{br} - \mathrm{GW}, \tag{8}$$

indicating that, in this geologic setting, dry season baseflow is dependent on drainage from bedrock storage, as depleted by losses to groundwater. Although this is highly site specific, it is assumed that at the small catchment scale, perched aquifers dominate the baseflow response ($\Delta S_{br} \ll \mathrm{GW}$, note sign convention) with an increasing proportion of groundwater contribution to baseflow at larger scales until baseflow is dominated by groundwater contributions ($\Delta S_{br} \gg \mathrm{GW}$).

The fluxes among the stores in both types of catchments are controlled largely by the attributes associated with the vegetation and soils. In catchments with a well-developed forest cover, the canopy buffers short duration rainfall intensity and reduces the effective precipitation through canopy storage (Keim and Skaugset, 2003). Surface cover and biological activity enhance infiltration rate (i), soil hydraulic conductivity, and the vertical flux through the soil (f_s). As effective precipitation seldom exceeds infiltration capacity, surface runoff rarely occurs. Subsurface storm flow requires an infiltration rate greater than the vertical hydraulic conductivity of the soil or of the bedrock (k_{br}) and soil moisture storage greater than field capacity. If the input during the event is greater than the soil storage capacity available at the beginning of the event, surface runoff is also generated.

It is important to recognize the different roles that the sequence of system stores and permeabilities play in the balance between runoff and baseflow generation for the different geologic and edaphic settings. For the deep soil, low permeability bedrock case (Fig. 1a), event runoff generation is largely dependent on the available storage in the canopy and soil and ground surface, and partitioning among vertical (i, f_s) and lateral (SRO, IF) flows (Elsenbeer et al., 1999, Elsenbeer and Vertessey, 2000). As such, runoff response is largely dependent on antecedent soil moisture storage conditions and greatly influenced by ET. Since flux to groundwater is limited by the soil moisture storage capacity and the low vertical conductivity of the bedrock, baseflow is dependent on drainage from the deep soil storage (and bedrock interface), which may be reduced by evapotranspiration if rooting depth is adequate. On the other hand, runoff from the shallow soil, high permeability bedrock case (Fig. 1b) is also dependent on infiltration capacity and soil hydraulic conductivity; but is less dependent on soil moisture storage because the permeability of the bedrock is much less limiting. In this case, bedrock storage accounts for water stored in perched aquifers and unsaturated bedrock above the phreatic surface and can be a large store with seasonal to annual residence times. Base flow is generated primarily by drainage from bedrock storage, which is influenced less by evapotranspiration because of the (assumed) limited access to roots. If the annual input to bedrock storage is large, it is reasonable to assume that a portion of wet season bedrock recharge will contribute to dry season base flows.

3. Study area

The study sites are situated at 250–350 m above sea level in the hills above Matalom (10.3°N, 124.8°E) in Leyte, Philippines. The regional geomorphology is characterized by well-dissected limestone hills along the orogenic belt at the eastern perimeter of the Visayan Basin and hosts a variety of karst landforms. In the hilly areas above 200 masl, outcrops of limestone are common as the soils are generally shallow. The soils of the area are Lithic Troporthents and Rendolls derived from limestone and calcareous shale (Chandler, 1998).

The uplands near Matalom were predominantly forested until the 1940s, when refugees from the coastal plains settled the hillsides. Since that time, most of the primary forest has been cleared. The predominant land cover types in hilly areas are now secondary forests, pasture and cropping of corn (*zea mais*) and sweet potato (*Ipomea batatas*). Land is initially prepared for cropping by slashing and burning, which may either be repeated for several cropping cycles or followed by tillage, depending on the slope, cover and time since forest clearance. Much of the current forest is secondary growth, as a result of farmland abandonment and episodic reforestation campaigns to promote soil conservation or rural economic growth (Pretty and Shah, 1997). At the time of the study, the upland farmers' primary concern was the decreasing availability of water, both for household and agricultural use, which they attributed to the removal of forest cover. However, they were also faced with decreasing crop yields on the intensively farmed landholdings due to erosion and depletion of nutrients in the topsoil (Agus et al., 1999). At the time, both alley farming between contour hedgerows (Fujisaka, 1993) and reforestation by community tree planting were actively promoted for soil and water conservation in the region (Garrity et al., 2002).

Southern Leyte receives in excess of 2200 mm of rainfall annually, with >5 mm of rainfall on nearly half of the days in a given year (IRRI Climate Unit, 1995, 1996). Rainfall is relatively evenly distributed for the months June through January, with a lower probability of rain from February through May. Significant episodes of high intensity rainfall often accompany tropical depressions and typhoons (Fornis et al., 2005). Typhoons are most frequent in October and November (Paningbatan et al., 1995), but may occur anytime from September to February.

4. Methods

Water budgets are developed for the 1995–1996 water year for three land uses, cropped (*Zea mais*), heavily grazed pasture (*Paspalum conjugatum*) and secondary forest (*Leucaena leucocephala*) (sites PL, PF and FS, respectively, in the original study of Chandler and Walter, 1998). Both the forest and pasture sites succeeded cropland and were the primary alternatives for fallowing depleted cropland. The sites were selected from the original five treatments to best represent the range in hydrological response as well as for their joint proximity to a recording rain gage. Each treatment was a zero-order catchment, concave both up and across the slope. Catchment sizes for cropped field, the forest and the pasture (0.25, 0.20 and 0.13 ha, respectively) are typical of the patch size of land cover types within the heavily populated uplands, as well as typical of the size of zero-order basins (e.g. Tsuboyama et al., 2000). Streamflow was not measured, since the upland streams were intermittent, integrated the response from several land uses, and were generated by undefined source areas in the karst terrain. Detailed descriptions of each treatment are presented in Chandler and Walter (1998).

Rainfall was recorded with a tipping-bucket rain gage (Texas Electronics, Dallas, TX, USA) near the sites. Runoff was estimated from application of rainfall–runoff relationships previously developed by Chandler and Walter (1998) based on the daily precipitation record (Table 1).

Periodic soil moisture measurements were made at three locations at each site with a TRIME-FM (Ettlingen, Germany) time domain reflectometry (TDR) soil moisture measurement device. Within each site, 1 m access tubes for the TRIME T3 probe were installed to allow four measurements at 20 cm depth increments. Soil moisture storage was calculated for each site as the average of the depth-integrated measurements. Although the TRIME-FM is a TDR device, and was calibrated for the soils in the study site, the accuracy of dielectric techniques for measuring soil water content is limited in clay soil (Robinson et al., 2003). Therefore, the soil moisture storage records calculated from these data are better viewed as a record of the soil wetting and drying history at each site, rather than as a precise measurement of storage changes.

Daily reference evapotranspiration (ET_0) values were calculated by the Penman–Monteith equation using the procedure presented by Allen et al. (1998) from data

Table 1
Rainfall–runoff relationships for hillslope experiments (reproduced from Chandler, 1998)

	Antecedent conditions	Surface runoff		Interflow	
		Runoff ratio (runoff/rainfall)	Threshold rainfall (mm)	Runoff ratio (runoff/rainfall)	Threshold rainfall (mm)
Pasture	Wet	0.8	4	0.2	30
	Dry	0.7	28	0.2	56
Crop	Wet	0.4	33	0.2	40
	Dry	0.4	50	0.2	72
Forest	Wet	0.1	95	0.1	62

collected at the Matalom weather station (IRRI Climate Unit, 1995, 1996).

$$\mathrm{ET_o} = \frac{0.408\Delta(R_n - G) + \gamma(900/(T+273))u_2(e_s - e_a)}{\Delta + \gamma(1 + 0.34u_2)} \quad (9)$$

where $\mathrm{ET_o}$ is the reference evapotranspiration (mm day^{-1}), R_n the net radiation at the crop surface (MJ m^{-2} day^{-1}), G the soil heat flux density (MJ m^{-2} day^{-1}), T the air temperature at 2 m height (°C), u_2 the wind speed at 2 m height (m s^{-1}), e_s the saturation vapour pressure (kPa), e_a the actual vapour pressure (kPa), $e_s - e_a$ the saturation vapour pressure deficit (kPa), Δ the slope vapour pressure curve (kPa °C^{-1}), γ the psychrometric constant (kPa °C^{-1}).

The rainfall, soil moisture and $\mathrm{ET_o}$ records were compared to determine wet and dry seasons during the 1995–1996 water year. The $\mathrm{ET_o}$ values were adjusted by season and cover type to represent the "crop" ET ($\mathrm{ET_c}$), by multiplying the daily $\mathrm{ET_o}$ values by crop coefficient (K_c) values:

$$\mathrm{ET_c} = K_c\,\mathrm{ET_o} \quad (10)$$

Water budgets were calculated at 4-day time steps to be consistent with the antecedent condition period previously used by Chandler and Walter (1998) to discriminate between wet and dry conditions for their rainfall–runoff relationships (Table 1). Initially, K_c values ranging from 0.5 to 1.2 were selected for similar cover types from established tables (Allen et al., 1998). Optimal K_c values were subsequently selected for each cover type by fitting the modeled soil water storage records to those calculated from the periodic field measurements. Finally, the sensitivity of the water balance components to K_c was tested by comparing the water balance model results for each land use over the initially selected range of $\mathrm{ET_c}$ values.

The modeling approach allows most terms in Eq. (5) (P, SRO, IF, ET, ΔS_{rz}) to be accounted for at an event scale time step, but certain assumptions remain. Canopy storage (ΔS_c) is neither measured nor calculated, but is implicit in other water balance components. The impact of canopy storage on runoff is first accounted for, along with soil water storage changes, in the increasing thresholds to runoff among cover types and seasons (Table 1). Secondly, canopy storage is accounted for by the procedure used to adjust ET using the soil moisture storage record, since the intercepted water would likely be partitioned among evaporation, transpiration and soil moisture storage in the 4-day time step. Deep soil moisture storage (ΔS_{ds}) is not accounted for in the water balance, as the root zone is assumed to reach the bottom of the soil profile in all treatments. Measurements of bedrock storage (ΔS_{br}) and outflows to base flow and groundwater are beyond the scope of this work, and may not be possible. Therefore, positive changes in bedrock storage (ΔS_{br}) are calculated as the residual of the water budget,

$$P - \mathrm{SRO} - \mathrm{IF} - \mathrm{ET} - \Delta S_{rz} = \Delta S_{br} \quad (11)$$

and reported as sums, S_{br} in the results. It is assumed that greater bedrock storage will result in increased base flows and groundwater input as modified by the local geologic setting.

Average soil bulk density was calculated by depth from a set of soil cores taken from each site with a 50 mm slide hammer (Blake, 1965). Samples were collected at five depths between the surface and 45 cm at four locations in the forest and 10 locations each at the crop and pasture sites. All samples were collected 3 days after a rain event to allow drainable porosity to be calculated from accompanying soil moisture content measurements.

5. Results

The wet and dry seasons were determined from the precipitation, $\mathrm{ET_o}$ and field soil moisture storage records. Rainfall was frequent throughout the year, but the occurrence of storms >50 mm decreased dramatically from March 8 until June 1, 1996, hereafter referred to as the dry season. Of the total rainfall (3651 mm), 2946 mm fell during the wet season and 705 mm fell during the dry season. While the daily variability in $\mathrm{ET_o}$ is similar to the seasonal fluctuation (Fig. 2a), the seasonal differences in $\mathrm{ET_o}$ relative to precipitation depth are quite dramatic. The wet season $\mathrm{ET_o}$ (1247 mm) was less than half the wet season rainfall, whereas the dry season $\mathrm{ET_o}$ (867 mm) exceeded dry season rainfall. Comparison of the 4 day sums for $\mathrm{ET_o}$ and rainfall (Fig. 2b) further demonstrates the seasonal shift from large rainfall events, which generate runoff and are greatly in excess of $\mathrm{ET_o}$ in the wet season, to a balance between rainfall and $\mathrm{ET_o}$ in the dry season. The soil moisture storage records also respond to the seasonal balance between rainfall and $\mathrm{ET_o}$ (Fig. 3), drying for the period from March through May and then wetting again in June, when 4-day rainfall once again exceeds 4-day $\mathrm{ET_o}$.

Following precipitation, evapotranspiration was the largest component of the water balance at all sites. The best agreement between modeled and measured soil moisture storage was obtained by individually optimizing K_c values for the wet and

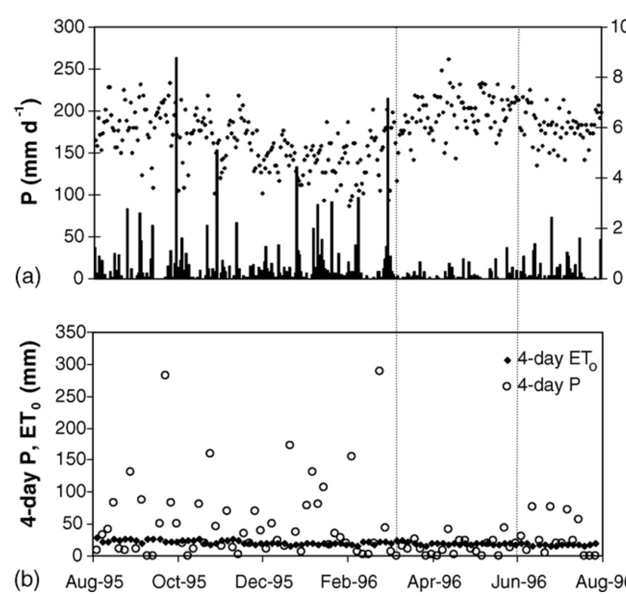

Fig. 2. Daily and 4-day summations of precipitation (P), and reference evapotranspiration ($\mathrm{ET_o}$) for Matalom, Leyte. The dry season is indicated by the dashed lines.

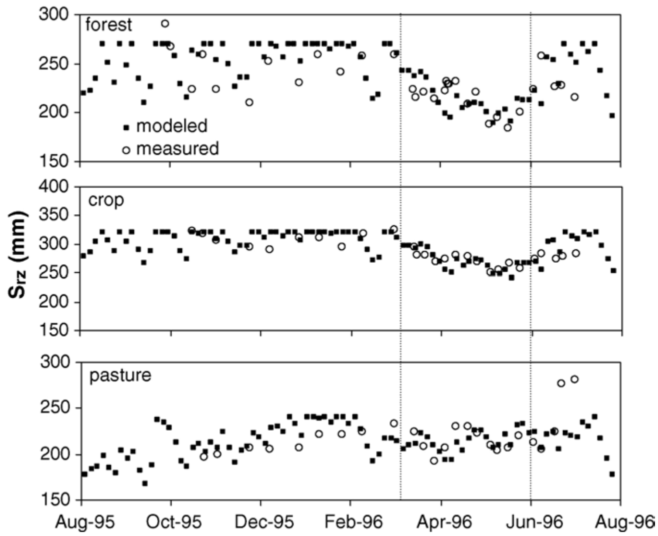

Fig. 3. Soil moisture storage (S) calculated from soil moisture measurements at the forest, crop and pasture sites and soil moisture storage calculated at 4-day time steps from the water budget model.

dry season for each land use (Fig. 3). The seasonal difference in soil moisture storage at the forest and crop sites was approximately double that at the pasture site, however these differences among sites (12–29 mm) are relatively small in comparison to the annual values of precipitation, evapotranspiration, runoff or bedrock storage (Table 3). The fitted K_c values and the resultant ET_c values decreased from forest (1906 mm) to crop (1661 mm) to pasture (1476 mm) for the wet season (Table 3). However, dry season ET_c was similar for the crop (304 mm) and forest (302 mm) sites and somewhat less for the pasture (222 mm), reflecting the limitations of soil moisture storage in thin soils.

During the wet season, runoff was an important water budget component for the non-forest sites, and occurred primarily as surface runoff at the crop (377 mm) and pasture (1630 mm) treatments. Interflow was similar to surface runoff (77 mm and 50 mm, respectively) from the forest, and increased in amount but decreased in proportion of total runoff at the crop (172 mm interflow of 549 mm total) and pasture (185 mm interflow of 1815 mm total) sites. Similarly, the occurrence of runoff generation was almost exclusively during the wet season, when the daily rainfall exceeded the threshold depth for both interflow and surface runoff generation (Table 1) for all sites. The number of surface runoff events increased as surface cover decreased: forest cover—2 events; crop cover—10 events; and pasture cover—38 events. The number of interflow events from the forest (3) was half of that from the crop and pasture (6) treatments. Dry season runoff was limited to surface runoff (34 mm) in the pasture treatment. The difference in annual bedrock storage input values among the sites is primarily due to differences in runoff. The forest had the least runoff and the most bedrock recharge (1320 mm), drainage under crop cover was somewhat less (1134 mm), and the bedrock input from the pasture was minimal (106 mm) (Table 3).

The partitioning of rainfall among runoff, evapotranspiration and bedrock recharge varied by event and was largely dependent on the event frequency and individual event depths. Cumulative values for the major components of the water budget for the year of record are presented in Fig. 4 to demonstrate the episodic nature of bedrock recharge. The moderate increase in diversion to runoff in the crop treatment and dramatic increase in runoff from the pasture treatment is attended by commensurate reductions in bedrock recharge, as compared to the forest (Table 3). This effect is especially apparent during the very wet December to March period. For example, of the 288 mm of rain which fell in the 4-day period from 25–28 February 1996, 175 mm was recharged to bedrock, 52 mm supplemented soil moisture storage, and 33 mm ran off the forest site as storm flow (Fig. 4, point a). At the crop site, the effective precipitation was divided primarily between 114 mm of bedrock recharge, 45 mm of soil moisture storage and 106 mm of runoff (Fig. 4, points b, c); at the pasture site, 246 mm of runoff was generated (Fig. 4, point d). The differential influence of runoff on groundwater recharge among the treatments is partially offset by the differences in wet season ET among the sites, but is apparently not affected by the reduced ET during the dry season.

The consequence of varying K_c among the sites was tested by comparing the major water budget component responses across the range of tested values. The water balance component (other than ET_c) most sensitive to changes in K_c was bedrock recharge, which decreases with increasing ET_c. With the exception of the pasture, runoff was relatively insensitive to

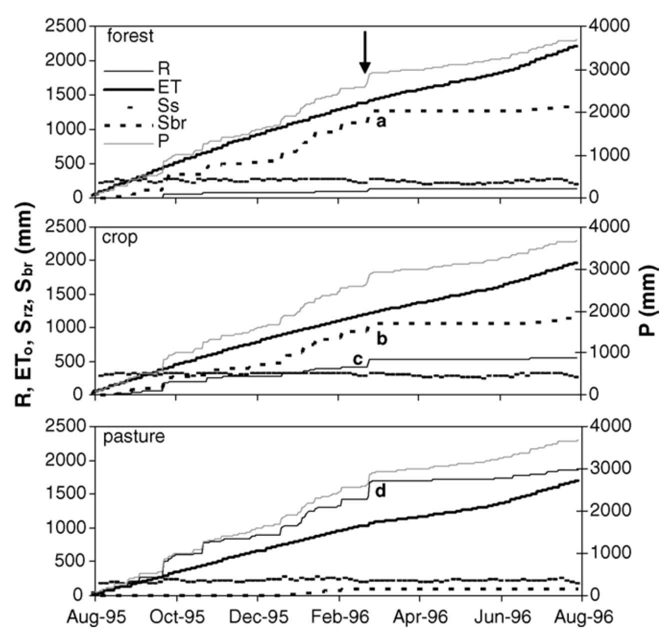

Fig. 4. Major water budget model components for the 1995–1996 water year. Cumulative precipitation (P) is plotted on the right axis. Interflow and surface runoff are summed as total stormflow (R) as calculated from Table 1. Crop evapotranspiration (ET_c) is calculated for each site from individual crop coefficients (Table 3). Soil moisture storage is calculated for the limiting values in Table 2. Cumulative R, ET_c, and bedrock storage (S_{br}), as well as 4-day S_s values, are plotted on the left axis. The arrow in the upper plot demarcates a 288 mm rain event ending on 28 February, 1996. The bedrock recharge under forest (a), to bedrock storage (b) and runoff (c) under crop, to runoff (d) from the pasture are identified as an example of the shift in hydrologic response among the land cover types.

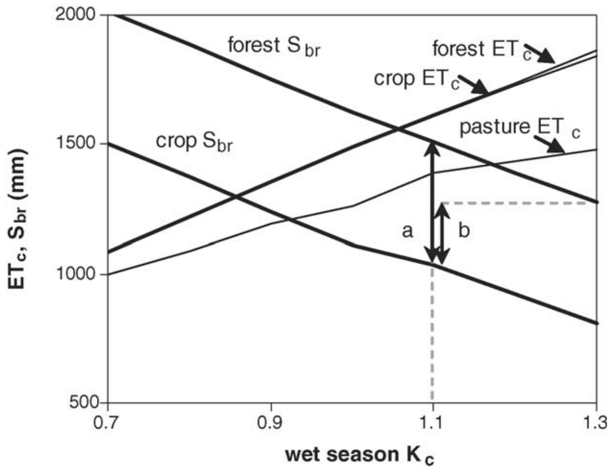

Fig. 5. Influence of crop coefficient value (K_c) on modeled evapotranspiration (ET_c) at all sites and on bedrock recharge (S_{br}) at the forest and crop sites. The difference in S_{br} between the forest and crop sites doubles when a wet season K_c value of 1.1 is used for both sites (point a), rather than the different K_c values used in the water budget (forest K_c = 1.3, point b).

Table 2
Soil depths, minimum and maximum volumetric soil moisture content and storage limits used in the model

	Soil depth (mm)	θ_{min} ($m^3\ m^{-3}$)	θ_{max} ($m^3\ m^{-3}$)	$S_{s\ min}$ (mm)	$S_{s\ max}$ (mm)
Pasture	600	0.20	0.40	120	240
Crop	800	0.15	0.40	120	320
Forest	800	0.15	0.40	120	320

Table 3
Annual values for water balance component calculated with K_c optimized by fitting modeled soil moisture storage to measured soil moisture storage

	Season	K_c	SRO (mm)	IF (mm)	ET_c (mm)	ΔS_s (mm)	S_{br} (mm)
Forest	Wet	1.3	50	77	1906	20	1320
	Dry	0.8	0	0	302	−24	0
Crop	Wet	1.1	377	172	1661	29	1134
	Dry	0.7	0	0	304	−27	0
Pasture	Wet	1.1	1630	185	1476	−13	106
	Dry	0.5	34	0	222	12	0

modeled changes in ET_c. Hereafter, the implications of varying K_c will be presented only for the wet season since ET_c balances P throughout the dry season across the range of tested K_c values. Wet season ET_c for forest and crop sites are nearly identical for a given K_c (Fig. 5) because other soil properties are also the same (Table 2). The resulting bedrock recharge for those treatments are complementary to ET_c and differ for a given K_c value by runoff, as indicated by Eq. (1). The pasture ET (Fig. 5) exhibited a complementary response to surface runoff (Fig. 4) and was independent of groundwater storage, which remained near 100 mm, regardless of changes in K_c (Table 3).

The forest and pasture soil properties associated with hydraulic conductivity differed markedly from those of the cropped site, underscoring the impact of land use on properties controlling infiltration and soil water fluxes. Soil bulk density was greater at all depths for the crop than the forest site and again greater for the pasture then the crop treatment (Fig. 6). Whereas, the soil bulk density of the crop site was relatively uniform with depth, the pasture soil was more compacted near the soil surface and bulk density in the surface soils of the forest site was lower than in the underlying soil layers, similar to the results of Krishnaswamy and Richter (2002). Drainable porosity (Fig. 6) increases from crop to forest and decreases from crop to pasture at all depths, and generally declines with increasing depth. It is noteworthy that the drainable porosity at the surface is dramatically different between the pasture (0.07) and the crop and forest sites (0.22 and 0.24, respectively), and that the drainable porosity at 40 cm depth increases markedly among the pasture (0.05), crop (0.12) and forest (0.16) sites.

6. Discussion

Whereas, the episodic diversion of precipitation to runoff controls the flux *to* storage at the event time scale, base flows depend on fluxes *from* storage, which occur over a much longer time scale. Any change in hydrological response within a given climate and geologic setting then depends on changes in flux or

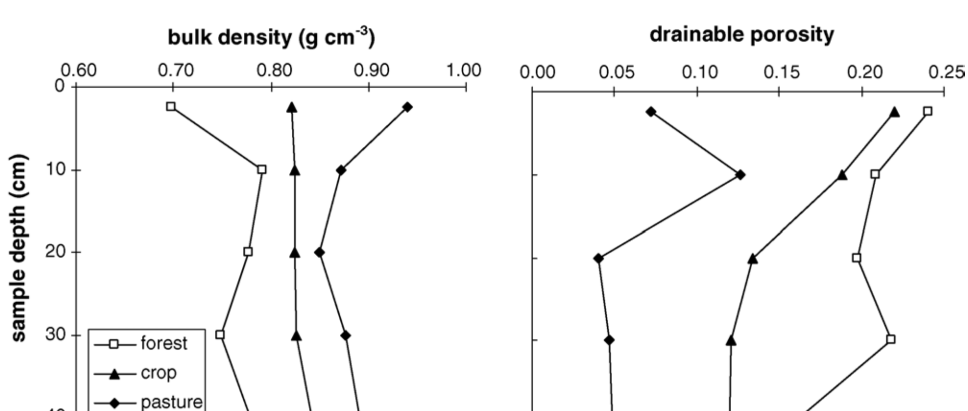

Fig. 6. Depth profiles of soil bulk density and drainable porosity for the forest, crop and pasture sites.

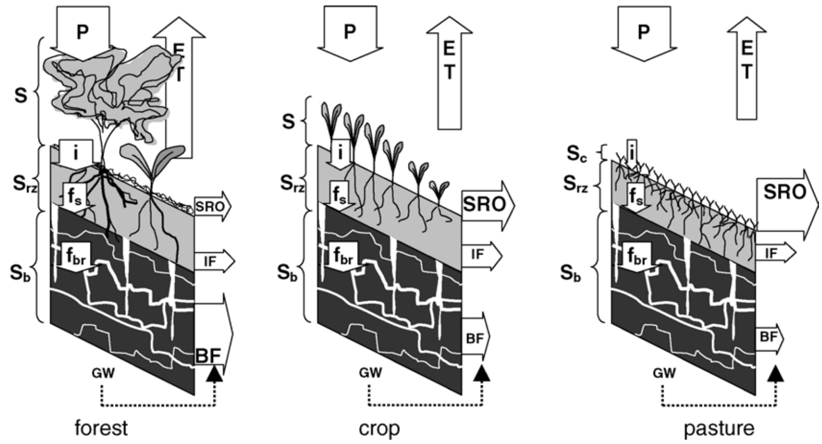

Fig. 7. Schematic representation of the stores and fluxes in the study treatments. Fluxes include precipitation (P), infiltration (i), vertical flux of water through the soil (f_s) and bedrock (f_{br}), evapotranspiration (ET), surface runoff (SRO), interflow (IF), baseflow from perched aquifers and bedrock fractures (BF), and groundwater flow from below the regional phreatic surface (GW).

storage limitations associated with land use change. The bedrock recharge results presented in this paper rely first on the difference between precipitation depth and wet season runoff, which dominates the response at the pasture, and secondly on the annual losses to evapotranspiration. The conceptual model for the high permeability bedrock catchment has been expanded to depict the flux and storage limitations in the studied treatments (Fig. 7).

In three tropical zero-order basins with drastically different land cover, above ground biomass and soil infiltration capacity are inexorably linked to the land use/land cover types, exerting a synergistic effect on multiple hydrologic processes. From forest to crop to pasture, reduced canopy storage increases the flux of effective precipitation to the ground. Along the same progression of treatments, infiltration capacity, hydraulic conductivity and soil storage capacity also decrease. Infiltration capacity and soil hydraulic conductivity are expected to decrease with increased bulk density (Fig. 6) and soil moisture storage capacity can be diminished for event or seasonal time scales from decreased evapotranspiration and/or reduced soil depth following erosion, or both in the case of the pasture (Table 2). This is consistent with the presented wet season threshold rainfall depths for surface runoff, which decrease dramatically with biomass removal from forest (95 mm) to crop (33 mm) to pasture (4 mm) cover types (Table 1) and the resulting increases in surface runoff and interflow (Table 3). Whereas, the runoff threshold values are influenced by storage and infiltration capacity, the greater dry season threshold values (threshold not reached for forest) indicate that the depth of storage available in the canopy and soil is the first-order control over surface runoff for rainfall events <100 mm. Most bedrock recharge is highly episodic and occurs during the wet season when available storage in the canopy and soil is low relative to event rainfall (Fig. 4). Transmission of precipitation to the bedrock was greatest and most frequent at the forest, less frequent and a lesser total at the cropped site and negligible at the pasture site, as reflected in the complementary bedrock recharge or runoff responses. The variance between the results presented in this paper and previous results reported for primary forest, forest regrowth and pasture in the Amazon (Godsey and Elsenbeer, 2002) underscores the importance of investigating the relationships among the hydrologic fluxes and stores in converted tropical forest systems under different geologic, edaphic, disturbance and post-disturbance vegetation settings.

In the water budgets for the three zero-order basins, the difference between the bedrock recharge component at the forest site and the cropped site is largely influenced by the selection of the wet season K_c value, which was considerably greater for the forest (1.3) than for the crop site (1.1). It is worthwhile to point out that the difference in recharge between the forest and the crop treatments would have doubled (Fig. 5, point a) had a K_c value of 1.1 been used for both sites (Fig. 5, point b). This being the case, the potential recharge values for the forest are considered conservative and are unlikely to be offset by further differences in evapotranspiration, barring significant abstraction of water in the bedrock fractures by tree roots.

7. Conclusion

The results support the premise that if the bedrock storage capacity is large relative to soil moisture storage, and if low season flow depends on wet season recharge of the bedrock storage, then the critical factor in maintaining or restoring low season flows is the maintenance of high vertical hydraulic conductivity rates between the soil surface and the bedrock reservoir. In this regard, forest cover may have some specific advantages: the soil environment under forest is more conducive to macropore development, both by the flora (large roots) and fauna (Noguchi et al., 1997, 1999) and the effective precipitation input fluxes are buffered through interception and transient canopy storage (Keim and Skaugset, 2003).

Low flows depend on a succession of diversions and abstractions of annual precipitation to stores and fluxes within any landscape. If precipitation is less than evaporation during the dry season, base flows depend on outflow from soil and bedrock storage. In the case of karst terrain with thin soils, soil moisture storage contributions to dry season flows are likely to be small compared to contributions from bedrock storage. As such, the differences in losses to evapotranspiration among the forest, crop, and pasture were found to be less important to dry season baseflow generation than was the increased recharge to bedrock storage with improved soil condition among the pasture, crop and forest treatments. This study underscores the importance of recognizing the different times scales over which deep percolation and surface runoff contribute to base flow and how the division between these two runoff components may vary with both vegetative cover and geologic substrate.

Acknowledgments

This work was supported through funding by the Cornell International Institute for Food and Agricultural Development and the Utah Agricultural Experiment Station, Utah State University Logan, Utah; approved as journal paper number 7657. Technical support was provided by the International Rice Research Institute. Logistical and field support was graciously provided in the Philippines by the farmers in Matalom, Leyte, the Farm and Resource Management Institute and the Visayas State College of Agriculture, Baybay, Leyte. Thanks to Mark Seyfried, Thijs Kelleners, Jim McNamara, Roy Sidle and Alan Ziegler for their helpful comments during the preparation of this manuscript.

References

Allen, R.G., Pereira, L.S., Raes, D., Smith, M., 1998. Crop evapotranspiration (guidelines for computing crop water requirements). FAO Irrigation and Drainage Paper No. 56, ISBN 92-5-104219-5.

Agus, F., Garrity, D.P., Cassel, D.K., 1999. Soil fertility in contour hedgerow systems on sloping oxisols in Mindanao Philippines. Soil Till. Res. 50 (2), 159–167.

Anderson, M.G., Burt, T.P., 1980. Interpretation of recession flow. J. Hydrol. 46, 89–101.

Andreassian, V., 2004. Waters and forests: from historical controversy to scientific debate. J. Hydrol. 291, 1–27.

Bigelow, S., 2001. Evapotranspiration modeled from stands of three broad-leaved tropical trees in Costa Rica. Hydrol. Process. 15 (14), 2779–2796.

Blake, G.R. 1965. Bulk density. In: Black, C.A. (Ed.), Methods of Soil Analysis, Part 1, first ed., vol. 9. Agron. Monogr., ASA, Madison, WI, pp. 374–390.

Bosch, J.M., Hewlett, J.D., 1982. A review of catchment experiments to determine the effect of vegetation changes on water yield and evapotranspiration. J. Hydrol. 55, 3–23.

Bruijnzeel, L.A., 1990. Hydrology of Moist Tropical Forests and Effects of Conversion: A State of Knowledge Review. Humid Tropics Programme, UNESCO IHP, Paris.

Bruijnzeel, L.A., 2004. Hydrological functions of tropical forests: not seeing the soil for the trees? Agric. Ecosys. Environ. 104 (1), 185–228.

Chandler, D.G. 1998. Athropogenic Effects On Hydrology and Soil Movement in Tropical Calcareous Uplands. Ph.D. Dissertation. Cornell University, Ithaca.

Chandler, D.G., Walter, M., 1998. Runoff response among common land uses in the uplands of Matalom, Leyte Philippines. Trans. ASAE. 41 (6), 1635–1641.

Cornish, P.M., Vertessy, R.A., 2001. Forest age-induced changes in evapotranspiration and water yield in a eucalypt forest. J. Hydrol. 242 (1–2), 43–63.

Elsenbeer, H., Newton, B.E., Dunne, T., de Moraes, J.M., 1999. Soil hydraulic conductivities under pasture, forest and teak in Rondonia Brazil. Hydrol. Proc. 13, 1417–1422.

Elsenbeer, H., Vertessey, R.A., 2000. Stormflow generation and flowpath characteristics in an Amazonian rainforest catchment. Hydrol. Proc. 14, 2367–2381.

Fujisaka, S., 1993. A case of farmer adaptation and adoption of contour hedgerows for soil conservation. Exp. Agric. 29 (1), 97–105.

Fornis, R.L., Vermeulen, H.R., Nieuwenhuis, J.D., 2005. Kinetic energy-rainfall intensity relationship for Central Cebu Philippines for soil erosion studies. J. Hydrol. 300 (1–4), 20–32.

Garrity, D.P., Amoroso, V.B., Koffa, S., Catacutan, D., Buenavista, G., Fay, P., Dar, W., 2002. Landcare on the poverty–protection interface in an Asian watershed. Conserv. Ecol. 6 (1) Art. No. 12.

Godsey, S., Elsenbeer, H., 2002. The soil hydrologic response to forest regrowth: a casestudy from southwestern Amazonia. Hydrol. Proc. 16, 1519–1522.

Hamilton, L.S., King, P.N., 1983. Tropical Forested Watersheds. Hydrologic and Soils Response to Major Uses or Conversions. Westview, Boulder.

IRRI Climate Unit, 1995. Compilation of Weather Data for 1995. International Rice Research Institute Publication, Los Banos.

IRRI Climate Unit, 1996. Compilation of Weather Data for 1996. International Rice Research Institute Publication, Los Banos.

Johnson, R., 1998. The forest cycle and low river flows: a review of UK and international studies. For. Ecol. Manag. 109 (1), 1–7.

Keim, R.F., Skaugset, A.E., 2003. Modelling effects of forest canopies on slope stability. Hydrol. Process. 17 (7), 1457–1467.

Kendy, E., Gérard-Marchant, P., Walter, M.T., Zhang, Y., Liu, C., Steenhuis, T.S., 2003. A soil–water-balance approach to quantify groundwater recharge from irrigated cropland in the North China Plain. Hydrol. Proc. 17 (10), 2011–2031.

Krishnaswamy, J., Richter, D.D., 2002. Properties of advanced weathering-stage soils in tropical forests and pastures. Soil Sci. Soc. Am. J. 66, 244–253.

McDonald, M.A., Healey, J.R., Stevens, P.A., 2002. The effects of secondary forest clearance and subsequent land-use on erosion losses and soil properties in the Blue Mountains of Jamaica Source. Agric. Ecosyst. Environ. 92 (1), 1–19.

Noguchi, S., Tsuboyama, Y., Sidle, R.C., Hosoda, I., 1997. Spatially distributed morphological characteristics of macropores in forest soils of Hitachi Ohta Experimental Watershed Japan. J. For. Res. 2, 207–215.

Noguchi, S., Tsuboyama, Y., Sidle, R.C., Hosoda, I., 1999. Morphological characteristics of macropores and the distribution of preferential flow pathways in a forested slope segment. Soil Sci. Am. J. 63, 1413–1432.

Paningbatan, E.P., Ciesiolka, C.A., Coughlan, K.J., Rose, C.W., 1995. Alley cropping for managing soil erosion of hilly lands in the Philippines. Soil Tech. 8, 193–204.

Pretty, J.N., Shah, P., 1997. Making soil and water conservation sustainable: from coercion and control to partnerships and participation. Land Degrad. Dev. 8 (1), 39–58.

Robinson, D.A., Jones, S.B., Wraith, J.M., Or, D., Friedman, S.P., 2003. A review of advances in dielectric and electrical conductivity measurement in soils using time domain reflectometry.

Smakhtin, V.U., 2001. Low flow hydrology: a review. J. Hydrol. 240, 147–186.

Tsuboyama, Y., Sidle, R.C.S., Noguchi, S., Murakami, Shimizu, T., 2000. A zero-order basin—its contribution to catchment hydrology and internal Hydrological Processes. Hydrol. Processes 14, 387–401.

Trimble, G.R.J., Reinhart, K.G., Webster, H.H., 1963. Cutting the forest to increase water yields. J. Forest 61, 635–640.

Urich, P.B., 1993. Stress on tropical karst cultivated with wet rice: Bohol Philippines. Environ. Geol. 21, 129–136.