Journal of Hydrology 380 (2010) 460–472

Contents lists available at ScienceDirect

Journal of Hydrology

journal homepage: www.elsevier.com/locate/jhydrol

Water balance modelling in a tropical watershed under deciduous forest (Mule Hole, India): Regolith matric storage buffers the groundwater recharge process

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ARTICLE INFO

Article history:
Received 27 February 2009
Received in revised form 17 September 2009
Accepted 16 November 2009

This manuscript was handled by P. Baveye, Editor-in-Chief, with the assistance of Hans-Jörg Vogel, Associate Editor

Keywords:
Matric porosity
Deep tree roots
Groundwater recharge
Lumped model
Semi-arid tropics
Evapotranspiration

SUMMARY

Accurate estimations of water balance are needed in semi-arid and sub-humid tropical regions, where water resources are scarce compared to water demand. Evapotranspiration plays a major role in this context, and the difficulty to quantify it precisely leads to major uncertainties in the groundwater recharge assessment, especially in forested catchments. In this paper, we propose to assess the importance of deep unsaturated regolith and water uptake by deep tree roots on the groundwater recharge process by using a lumped conceptual model (COMFORT). The model is calibrated using a 5 year hydrological monitoring of an experimental watershed under dry deciduous forest in South India (Mule Hole watershed).

The model was able to simulate the stream discharge as well as the contrasted behaviour of groundwater table along the hillslope. Water balance simulated for a 32 year climatic time series displayed a large year-to-year variability, with alternance of dry and wet phases with a time period of approximately 14 years. On an average, input by the rainfall was 1090 mm year^{-1} and the evapotranspiration was about 900 mm year^{-1} out of which 100 mm year^{-1} was uptake from the deep saprolite horizons. The stream flow was 100 mm year^{-1} while the groundwater underflow was 80 mm year^{-1}.

The simulation results suggest that (i) deciduous trees can uptake a significant amount of water from the deep regolith, (ii) this uptake, combined with the spatial variability of regolith depth, can account for the variable lag time between drainage events and groundwater rise observed for the different piezometers and (iii) water table response to recharge is buffered due to the long vertical travel time through the deep vadose zone, which constitutes a major water reservoir. This study stresses the importance of long term observations for the understanding of hydrological processes in tropical forested ecosystems.

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Introduction

Accurate assessment of water balance at the watershed scale is of major importance in a context of a global dramatic increase of human demand for water, either for urban or agricultural requirements. This balance results from the interaction of climate, geology, morphology, soil and vegetation (De Vries and Simmers, 2002). Its assessment is complex especially in arid and semi-arid regions, where evapotranspiration equals or surpasses average precipitation, because groundwater recharge and stream discharge fluxes are weak and variable (Scanlon et al., 2002, 2006; Sekhar et al., 2004; Anuraga et al., 2006). Moreover, despite the large amount of work dedicated to assess evapotranspiration at the watershed scale, this flux remains a large source of uncertainty in water budgeting, especially in the case of forested watersheds (Zhang et al., 2001, 2004).

A classical way of modelling watershed hydrology is to assess groundwater recharge as the residual of soil water balance equation (Beven, 2001; Beaujouan et al., 2001; Scanlon et al., 2002; Wagener et al., 2004). In that case, water draining below the soil zone is considered to be immediately transferred to the groundwater as recharge. However, significant delays between rainfall events and groundwater responses are sometimes observed, in particular

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doi:10.1016/j.jhydrol.2009.11.020

where deep unsaturated regoliths are present (O'Reilly, 1998; Ghauri, 2004). Some authors have addressed this problem by introducing a transfer function to simulate the traveltime of water from the bottom of the soil to the groundwater (O'Reilly, 2004). This approach assumes that the water content of the deep regolith does not vary significantly. However, this simplification could be inadequate in the case of forested catchments, because deep tree roots are able to uptake water in the deep regolith zone (Nepstad et al., 1994; Canadell et al., 1996; Collins and Bras, 2007).

The objective of this paper is to test the hypothesis that water uptake by deep tree roots in the unsaturated regolith zone can play a major role in controlling water balance in semi-arid forested watersheds. The experimental setting used in this study is a small forested watershed in South India, monitored for 5 years as part of the project "Observatoire de Recherche en Environnement – Bassin Versant Expérimentaux Tropicaux" (http://www.ore.fr/) (Braun et al., 2005).

The hypothesis was tested by using a conceptual model (COMFORT) introduced in this paper, which was designed to account explicitly for the water uptake by deep roots in the deep unsaturated regolith, while being as parsimonious as possible (only nine parameters). This was done because most of the few existing models accounting for deep unsaturated regolith usually require the calibration of many parameters: for example the model WAVES (Zhang et al., 1996) requires the calibration of at least 29 parameters.

Our results show that the deep vadose zone plays a major role in buffering the groundwater response to the water percolation.

Site description

The study site is situated in South India (Fig. 1), at 11°44′N and 76°27′E (Karnataka state, Chamrajnagar district).

It is located in the transition zone of a steep climatic and geomorphologic gradient at the edge of the rifted continental passive margin of the Karnataka Plateau, which was the focus of extensive geomorphologic studies (Gunnell and Bourgeon, 1997). This plateau, developed on the high-grade metamorphic silicate rocks of the West Dharwar craton (Moyen et al., 2001), is limited westward by the Western Ghâts, a first order mountain range. This mountain forms an orographic barrier, inducing an steep climatic gradient, with annual rainfall decreasing from west to east from about 6000 mm to 500 mm within a distance of about 80 km (Pascal, 1982). These Ghâts are of critical ecological and economical importance and also an important source of all major South Indian rivers, flowing eastward towards the Gulf of Bengal. The climatic transition zone is mainly covered by dry deciduous forests, belonging to the wildlife sanctuaries of Mudumalai, Waynad, Bandipur and Nagarahole (Prasad and Hedge, 1986). Such a tropical climosequence is comparable, although much steeper (Gunnell, 2000), to the well documented monsoonal West African and the Northeast Brazilian climosequences (Gunnell, 1998).

The Mule Hole experimental watershed (4.1 km^2) is located in the climatic semi-humid transition area and the mean annual rainfall (n = 25 years) is 1120 mm. The mean yearly temperature is 27 °C. On the basis of the aridity index defined as the ratio of mean annual precipitation to potential evapotranspiration, the climate regime can be classified as humid (UNESCO, 1979). Nevertheless, the climate is characterized by the occurrence of a marked dry season (around 5 months from December to April) and by recurrent droughts, depending on the monsoon rainfalls. The rainfall pattern is bimodal, as it is affected by both the South West Monsoon (June–September) and the North East Monsoon (October–December) (Gunnell and Bourgeon, 1997). Streams are ephemeral, and their flow duration ranges from a few hours to a few days after the storm events.

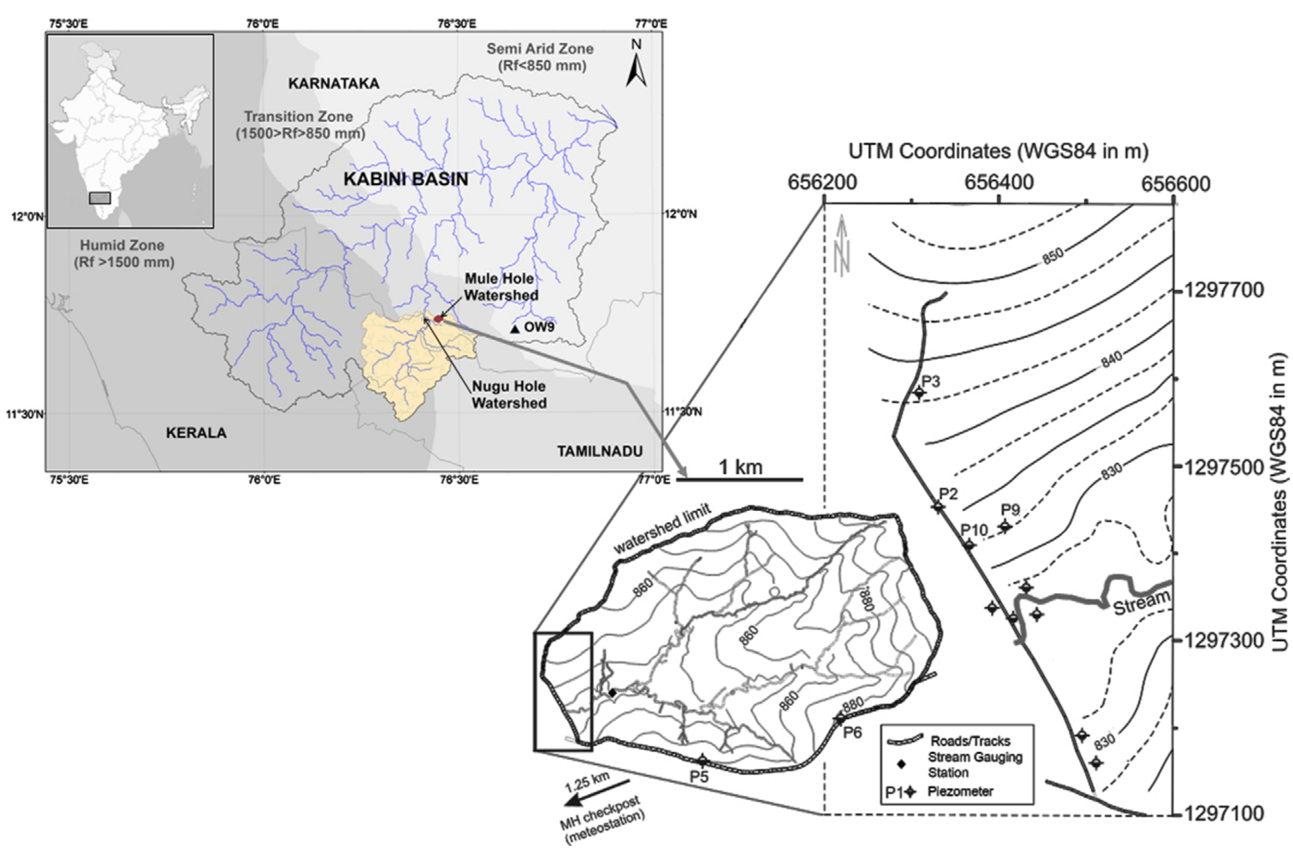

Fig. 1. Location map of the experimental site.

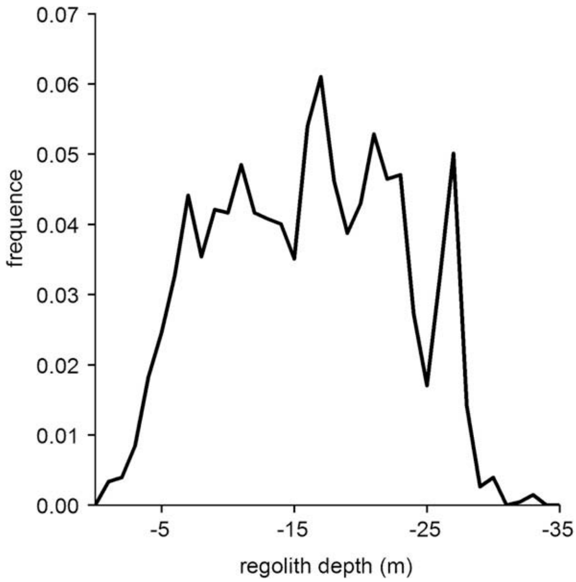

Fig. 2. Distribution of regolith depth across the Mule Hole watershed (from a geophysical and geochemical survey by Braun et al. (2008)).

The watershed is mostly undulating with gentle slopes and the elevation of the watershed ranges from 820 to 910 m above sea level (Fig. 1). The morphology of the watershed is convexo-concave highly incised by the temporary stream network. The lithology, representative of the West Dharwar craton (Naqvi and Rogers, 1987), is dominated by complexly folded, heterogeneous Precambrian peninsular gneiss intermingled with mafic and ultramafic rocks of the volcano-sedimentary Sargur series (Shadakshara Swamy et al., 1995). The Peninsular gneiss represents at least 85% of the watershed basement and the average strike value is N80°, with a dip angle ranging from 75° to the vertical (Descloitres et al., 2008). In such hard-rock context, the aquifer can generally be divided into two parts: one upper part is the porous clayey to loamy regolith with an apparent density lower than the rock bulk density, the other is in the fractured–fissured protolith with an apparent density close to the bulk density of the rock and a network of fractures of a density decreasing with depth (Sekhar et al., 1994; Maréchal et al., 2004; Wyns et al., 2004; Dewandel et al., 2006). In the Mule Hole watershed, the average depth of the regolith is 17 m, as estimated through an extensive geophysical and geochemical survey (Braun et al., 2006, 2008). The distribution of the regolith depth (Fig. 2) shows that the range of variation is 5–27 m. No correlation with the position on the hillslope was found. Average total porosity of the regolith is around 12%. Magnetic Resonance Sounding (MRS) performed on the watershed (Legchenko et al., 2006) showed that drainage porosity is around 1% in the regolith and below the detection level (<0.5%) in the fractured rock. Finally, timelapse geophysical measurements of both electrical resistivity tomography (ERT) and MRS conducted at the outlet of the watershed indicated a seasonal infiltration of water under the stream (Descloitres et al., 2008). A hydrological investigations allowed estimating the indirect recharge from the stream at around 30 mm year^{-1} at the watershed scale (Maréchal et al., 2009).

The soil distribution in the watershed was determined by Barbiéro et al. (2007). The gneissic saprolite, cohesive to loose sandy, crops out both in the streambed and at the mid-slope in approximately 22% of the watershed area. The lower part of the slope and the flat valley bottoms (12% of the area) are covered by black soils (Vertisols and Vertic intergrades), which are 2 m deep on an average. Shallow red soils (Ferralsols and Chromic Luvisols), which are of 1–2 m deep, cover 66% of the entire watershed area. The watershed is covered by a dry deciduous forest with different facies linked to the soil distribution (Barbiéro et al., 2007). Minimal human activity is present as it belongs to the Bandipur National Park, dedicated to wildlife and biodiversity preservation. The predominant tree component of the vegetation consists of *Anogeissus–Terminalia–Tectona* association (ATT facies) forming a relatively open canopy not exceeding 20 m (Prasad and Hedge, 1986; Pascal, 1986). Phenology is marked by a strong seasonality, with leaf senescence starting in December and leaf flushing occurring in early April, 1 or 2 month before the first significant monsoon rains. This surprising behaviour, leading to a deciduous period of only 2–3 months, much shorter than the dry season, is a general feature of the Asian forests (Singh and Kushwaha, 2005), and was quoted as the "paradox of Asian monsoon forest" by Elliot et al. (2006).

Model description

The model COMFORT (**CO**nceptual **M**odel for hydrological balance in **FOR**es**T**ed catchments) proposed in this article allows simulating the daily water budget of a forested watershed, through a simple and widely accepted conceptual description of the hydrological processes. It includes a lumped model for the soil moisture and the evapotranspiration in the forest (Granier et al., 1999), a surface runoff model based on the variable source area theory (Moore et al., 1983; Beven, 2001) and linear reservoirs for the recharge and groundwater discharge (Beven, 2001; Putty and Prasad, 2000). Its originality relies on the introduction of an additional water reservoir located in the weathered vadose zone (saprolite) below the soil, accessible to tree roots but not to understorey vegetation roots.

The model includes two modules, calibrated and run successively (Fig. 3): the first one is a slightly modified and simplified version of the lumped water balance model presented by Granier et al. (1999), which simulates the daily water balance for the forested soil and the surface runoff Q_s, while the second one simulates the flow of water through the deep vadose zone and the groundwater flow. Forcing variables are the daily rainfall (Rf), the Penman–Montieth potential evapotranspiration (PET), and the forest leaf area index (LAI). The details of each of these modules is given below.

Module 1: soil moisture

This module computes the daily variations in the soil moisture deficit (SMD, in mm) as:

$$\Delta\mathrm{SMD} = E_{\mathrm{In}} + T + E_{\mathrm{u}} + \mathrm{PR} + Q_{\mathrm{s}} - \mathrm{Rf}$$

and

$$0 < \mathrm{SMD} < \mathrm{SMD}_{\max}$$

with E_{In} (mm day^{-1}) being the evaporation of rainfall intercepted by the forest canopy, T (mm day^{-1}) the tree transpiration, E_{u} (mm day^{-1}) the evapotranspiration of the understorey layer, PR (mm day^{-1}) is the percolation below the soil layer, also called "potential recharge" by De Vries and Simmers (2002), Q_s (mm day^{-1}) the surface runoff, Rf (mm day^{-1}) the rainfall and $\mathrm{SMD}_{\max}$ (mm) the maximum soil water deficit, equivalent to the soil water holding capacity.

The throughfall ($\mathrm{Rf} - \mathrm{E_{In}}$) on the saturated area of the watershed (SA in %) reaches the stream as surface runoff (Q_s in mm day^{-1}):

$$Q_{\mathrm{s}} = (\mathrm{Rf} - E_{\mathrm{In}}) \times \mathrm{SA}/100$$

Saturated area is usually computed as a function of water storage in the soil and the groundwater (see for example Putty and Prasad, 2000). In our context, groundwater level is far below the ground level, thus SA is modelled as an exponential function of the soil moisture deficit:

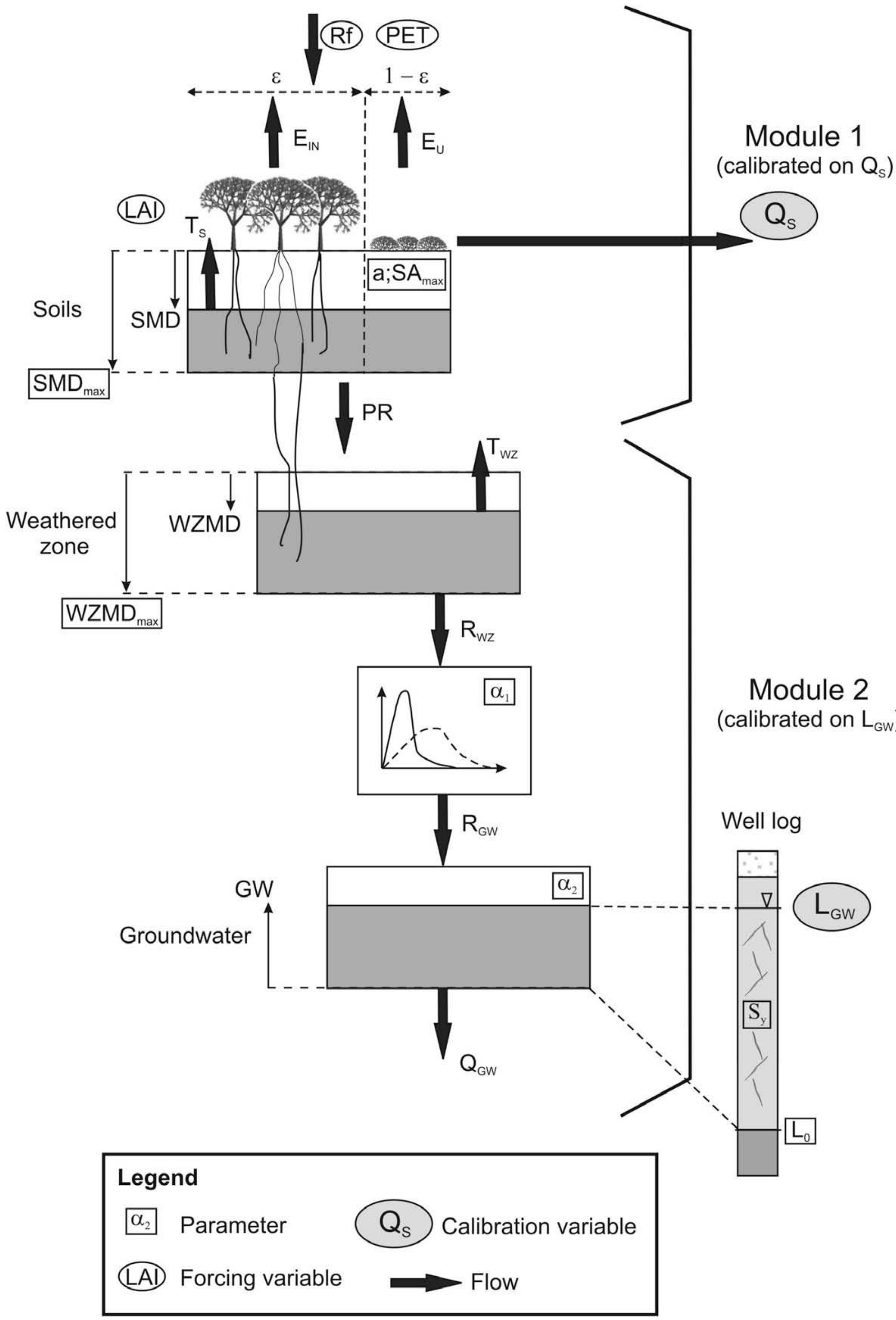

Fig. 3. A schematic representation of the model COMFORT.

$$SA = SA_{max} \times \exp(-a \times SMD)$$

with SA_{max} the maximal extension of the saturated area (%), and a being the exponent constant. The proportion of the soil surface covered by tree leaves (ε) is calculated from the forest LAI with the Beer–Lambert function assuming a light coefficient of extinction of 0.5 (Granier et al., 1999).

$$\varepsilon = 1 - e^{-0.5\times \mathrm{LAI}}$$

The evaporation of rainfall intercepted by the forest canopy, or interception losses, E_{In} (mm day^{-1}) is computed as:

$$E_{\mathrm{In}} = \mathrm{minimum}(\varepsilon \times \mathrm{Rf}; \varepsilon \times \mathrm{PET}; \varepsilon \times \mathrm{In})$$

with In (mm) the canopy storage capacity. This equation implies that the canopy cannot store water for more than 1 day, and thus interception losses are nil during non rainy days. This simplification is justified by the fact that In is generally smaller than PET at a daily time step.

As proposed by Granier et al. (1999), to account for the fact that the rate of evaporation of intercepted water is approximately four time greater than transpiration rate (Rutter, 1967), PET is reduced by 20% of the amount of intercepted water and the total actual evapotranspiration (AET) is limited to 1.2 × PET. The tree transpiration from the soil layer (T_s in mm day^{-1}) is then calculated as:

$$T_s = \mathrm{minimum}(\mathrm{SMD}_{\max} - \mathrm{SMD}; (\varepsilon \times \mathrm{PET}) - (0.2 \times E_{\mathrm{In}}); 1.2 \times \mathrm{PET} - E_{\mathrm{In}})$$

with $\mathrm{SMD}_{\max}$ the maximum soil water deficit (mm), equivalent to the soil water holding capacity. In their model, Granier et al. (1999) propose that T/PET ratio decreases linearly when soil moisture reaches a critical level of 40% of the water holding capacity of the soil. In our model, for the sake of simplicity, transpiration is only limited by the amount of water present in the soil reservoir.

Evapotranspiration from understorey vegetation is calculated as:

$$E_u = \mathrm{minimum}(\mathrm{SMD}_{\max} - \mathrm{SMD} - T_s; (1-\varepsilon) \times \mathrm{PET}; 1.2 \times \mathrm{PET} - T_s - E_{\mathrm{In}})$$

Actual evapotranspiration from the soil layer and the canopy (AETs in mm day^{-1}) is then:

$$\mathrm{AET}_s = E_{\mathrm{In}} + T_s + E_u$$

Eventually, the water in excess in the soil reservoir (when SMD < 0) percolates below the soil layer, as "potential recharge" (PR in mm day^{-1}). This flow is an input to the weathered zone reservoir (module 2).

Module 2: weathered zone and groundwater

This module simulates the daily variations of the moisture deficit in the weathered zone (WZMD in mm) as:

$$\Delta \mathrm{WZMD} = T_{\mathrm{WZ}} + R_{\mathrm{WZ}} - \mathrm{PR}$$

where T_{WZ} is the tree transpiration from the weathered zone below the soil (mm day^{-1}), PR the potential recharge calculated in the module 1 (mm day^{-1}) and R_{WZ} (mm day^{-1}) the recharge. The potential transpiration from the weathered zone ($\mathrm{PT}_{\mathrm{WZ}}$ in mm day^{-1}) is the residual of the potential transpiration of trees minus the actual tree transpiration in the soil zone:

$$\mathrm{PT}_{\mathrm{WZ}} = \varepsilon \times \mathrm{PET} - (0.2 \times E_{\mathrm{In}}) - T_s$$

The actual transpiration from the weathered zone T_{WZ} (mm day^{-1}) is then:

$$T_{\mathrm{WZ}} = \mathrm{minimum}(\mathrm{WZMD}_{\max} - \mathrm{WZMD}; \mathrm{PT}_{\mathrm{WZ}})$$

with $\mathrm{WZMD}_{\max}$ (mm) the maximum water deficit of the weathered zone.

Recharge (R_{WZ} in mm day^{-1}) is the water in excess in the weathered zone (when WZMD < 0). A linear reservoir model was used to account for the observed smoothness of the recharge process. R_{WZ} is directed to a recharge reservoir R (mm), and the effective recharge (R_{GW} in mm day^{-1}) reaching the groundwater reservoir (GW in mm) is calculated as:

$$R_{\mathrm{GW}} = \alpha_1 \times R$$

with α_1 (day^{-1}) the recession coefficient of the recharge reservoir R.

Eventually, the daily variation of water content in the groundwater reservoir (GW in mm) is calculated as:

$$\Delta \mathrm{GW} = R_{\mathrm{GW}} - Q_{\mathrm{GW}}$$

with the Q_{GW} the groundwater flow (mm day^{-1}) calculated using a second linear reservoir model as:

$$Q_{\mathrm{GW}} = \alpha_2 \times GW$$

with α_2 (day^{-1}) the recession coefficient of the groundwater reservoir GW. As the groundwater level is always deeper than the stream bed at the outlet, this flow is considered as an underflow.

The groundwater table level (L_{GW} in meters above sea level) is then calculated as:

$$L_{\mathrm{GW}} = L_0 + \mathrm{GW}/(\mathrm{Sy} \times 1000)$$

with L_0 (masl) the altitude of the base of the aquifer and Sy its specific yield.

Data acquisition and model calibration procedure

The climatic data (daily rainfall and PET) necessary to run the model were available for the period 2003–2007 from the automatic weather station (CIMEL, type ENERCO 407 AVKP) installed at the Mulehole forest check post, which is 1.5 km West to the watershed outlet (Fig. 1). Daily rainfall data were available at the same location from 1976 to 1995 from Indian Meteorological Department. Daily rainfall data were also available from the Ambalavayal weather station, located 20 km west of the study site, for the years 1979–2004. As statistical analysis showed a strong correlation between the two stations, the seven missing years (1996–2002) in the Mule Hole were inferred from the Ambalavayal data. Granier et al. (1999) have shown that soil water content can be equally simulated in forest stands under different climates either with models based on Penman–Montieth potential evapotranspiration or with mechanistic approaches tacking into account the canopy structure. Although many studies have shown that evapotranspiration is greater from forest than for the short size vegetation (Zhang et al., 2001) this difference is mainly attributed to better access to soil water at depth. Thus Penman–Montieth potential evapotranspiration (Allen et al., 1998) was used as a forcing variable to the model.

As year to year variations in PET were little during the years 2003–2007, an average daily PET series was calculated and applied to the period 1976–2002. Using an average annual curve of PET does not affect much the rainfall–runoff models (Burnash, 1995; Oudin et al., 2005) and even less groundwater models (O'Reilly, 2007). The simulations were run on the reconstructed time series 1976–2007. Stream discharge (Q_s) was measured since August 2003 at a 6 min time step using a flume built at the outlet of the watershed. Due to technical problems, level recording was not available from 15th April 2007 to 7th August 2007. A set of 13 observation wells were drilled in the area in 2003 (P1–P6) and 2004 (P7–P13) (Fig. 1). Most of these wells were dedicated to the monitoring of the effects of water seepage from the stream. Wells P2, P3, P5, P6, P9 and P10 were not influenced by the indirect recharge from the stream (Maréchal et al., 2009), and can be used to assess the direct recharge. The water levels were monitored in all the wells either manually at a monthly time step or automatically at an hourly time-step. Due to technical problems (including elephant attacks), P2 and P9 did not give reliable records and were not included in the analysis. Additional data from an observation well (OW9) monitored by the Department of Mines and Geology (Karnataka State) since 1975, and located 20 km east of the watershed, which is close to the forest border, was used in the analysis.

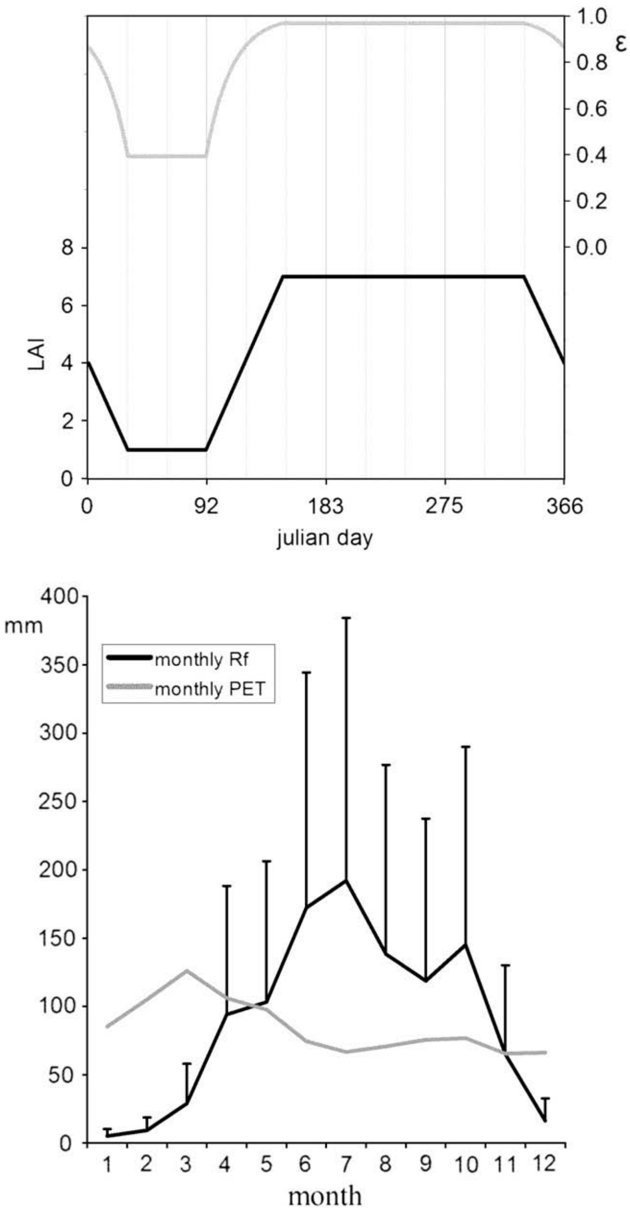

Fig. 4. (a) Daily forest LAI and coefficient of extinction ε and (b) average monthly rainfall (Rf) (1976–2007) and PET (2003–2007). Vertical bars indicate standard deviation of Rf.

As no measurement of forest leaf surface were carried out in the watershed, the evolution of LAI was hypothesised from qualitative observations from the site and references from literature concerning local tree phenology (Prasad and Hedge, 1986; Sundarapandian et al., 2005) and NDVI records for Indian forests (Prasad et al., 2005). Seasonality of leaf flushing and senescence is mostly driven by photoperiod, and therefore can be taken as a constant from one year to the other (Elliot et al., 2006; Singh and Kushwaha, 2005). The proposed LAI pattern is presented in Fig. 4a along with ε variations. The comparison with the average monthly rainfall and PET over 32 years period (Fig. 4b) shows that maximum PET is reached during the deciduous period and that on average, rainfall exceeds PET during 5 months.

The parametrization and the uncertainties in the parameters were estimated successively for the two modules using the Generalized Likelihood Uncertainty Estimate (GLUE) approach (Beven and Binley, 1992). The Nash and Sutcliffe (1970) efficiency criterion was used as an objective function. The GLUE approach is based on Monte Carlo simulations, in which simulations are performed for all the parameter sets. The parameter sets are sampled from a prior range of the parameters using the Latin Hypercube Sampling (McKay et al., 1979). The behavioural (best) parameters set are those which gives the best fit between measured and simulated variables. These behavioural parameters set were used to get the different properties of parameters like their uncertainties, correlation among parameters, etc.

Module 1 was calibrated against the observed stream discharge values. This module has three forcing variables (Rf, PET and LAI) and four parameters (In, a, SA_{max} and SMD_{max}). As the sensitivity analysis showed that the model had little sensitivity to the In value, it was set at 1 mm and automatic calibration was performed on the three remaining parameters. The time series obtained for the potential evapotranspiration from the weathered zone (PT_{WZ}) and the potential recharge (PR) were then used as forcing variables for module 2, which included five parameters ($WZMD_{max}$, α_1, α_2, Sy and L_0). Module 2 was calibrated for each piezometer against observed water table levels. To minimize the risk of nonunique solutions, the following procedure was adopted: because $WZMD_{max}$ determines the date of initial water table rise, it was first adjusted by trial and error; then the four remaining parameters were automatically calibrated, with Sy values constrained smaller than 0.01, according to the conclusions of MRS survey (Legchenko et al., 2006). For each calibration, the solver was run with contrasting sets of initial parameter values, which later converged towards the same solution.

Results

Fig. 5 compares the observed and the simulated surface runoff (Q_s) at the Mule Hole watershed outlet for the 5 years of monitoring. The Nash–Sutcliffe parameter calculated using the monthly values is 74. Despite this relatively low value, the model was able to reproduce the general trend of observed runoff, in particular the delay between the first monsoon rains and the first observed stream runoff. On the other hand, it was unable to reproduce the runoff observed after summer storm events in 2005, because they are due to Hortonian flow, a mechanism that was not accounted for in the model. However, Hortonian flow is of marginal importance in this pedoclimatic context. Considering that rainfall is measured in only one weather station located outside the watershed and the high spatial variability of rainfall, especially during strong individual storms, a perfect fit was not expected. In particular in 2005, two events, on 22nd of October and 4th of November, produced 34 mm day^{-1} and 20 mm day^{-1} of runoff for rains of 52 mm day^{-1} and 28 mm day^{-1} respectively. For these events, (in which no Hortonian flow was observed) the actual rainfall in the watershed was probably much higher than the measured one. This is likely to have affected the assessment of water balance for the year 2005, as these two storms represent about one quarter of the total yearly runoff. However, this kind of event remains very rare: during the five monitoring years, only these two events produced more than 20 mm of runoff, and only five other individual events produced more than 10 mm of runoff.

The calibrated parameter values are: a = 0.1, SA_{max} = 33.3% and SMD_{max} = 173 mm, and their corresponding uncertainties are 0.003, 0.56%, and 19.6 mm respectively. The value of maximal saturated area is probably overestimated, considering that the flat valley bottom overlaid by black soil occupies about 12% of the watershed area. The two exceptional storm events mentioned above played an important role in this overestimation. The calibrated value of SMD_{max} is consistent with the soil moisture monitoring carried out in 2004 and 2005 in the site, showing a maximum variation of volumetric water content of 7% in red and black soils

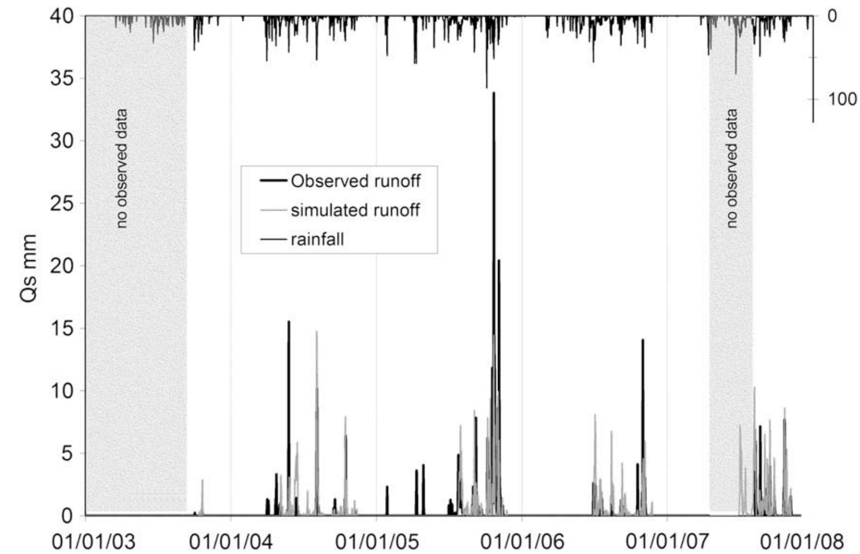

Fig. 5. Observed (black line) and simulated (grey line) daily surface runoff (Q_s in mm day^{-1}) at the Mule Hole watershed outlet for the 5 years of monitoring. Secondary axis is daily rainfall (Rf) in mm.

Table 1
Soil water balance (mm year^{-1}) for the monitored years and yearly average for the monitored period and the whole 32 year simulated period. Signification of terms is in the text. cv (%) is the standard deviation of the respective quantity for the respective time period divided by the mean × 100.

	2003	2004	2005	2006	2007	2003–2007	cv (%)	1976–2007	cv (%)
Rf	706	1216	1434	1170	1252	1155	(*23*)	1091	(*32*)
PET	1101	1074	1017	1012	963	1034	(*5*)	1067	–
E_{In}	78	130	133	123	136	120	(*20*)	98	(*23*)
E_u	53	79	111	127	72	89	(*34*)	82	(*33*)
T_s	555	627	624	649	546	600	(*8*)	624	(*11*)
AET_S	686	837	868	900	754	809	(*11*)	803	(*14*)
PR	6	225	372	183	318	221	(*64*)	197	(*87*)
PT_{wz}	268	144	97	87	118	143	(*51*)	158	(*58*)
Obs Qs	1	66	196	52	–	79[a]	(*105*)	–	–
sim Q_s	5	117	154	92	162	93[a]	(*68*)	94	(*79*)

[a] 2003–2006 Average.

(Barbiéro et al., 2007) and an average 2 m depth (Braun et al., 2006).

Annual soil water balances as well as averages for the monitoring period and the 32 years simulation period are presented in Table 1. Evapotranspiration is the most important sink for water, accounting for about 70% of rainfall. Interception is about 10% of rainfall, which is in the range of references values given in the literature for broad leaved forests (Ward and Robinson, 2000). Potential recharge is large (197 mm year^{-1} on the whole period), and displays an important year to year variability. It is larger than the estimate based on the regression equation proposed by Rangarajan and Athavale (2000) from tritium injection experiments in granitic areas in India, which gives a value of about 150 mm year^{-1} for the conditions of Mule Hole. This difference might be due to specificities of the study site, in particular the relatively low PET, mainly due to low temperatures linked with the altitude and to the forested environment which contributes to decrease soil compaction and increase infiltration potential (Bruijnzeel, 2004; Ilstedt et al., 2007). Results also show that the monitoring period is quite representative of the whole 32 years period with respect to the soil water balance.

Fig. 6 compares the simulated daily potential recharge with observed relative variations in the water table level in hillslope piezometers. Although all piezometers showed a consistent global tendency to water level rise, they displayed significant contrasted behaviour. The relatively shallow piezometer (P10) responded each year to recharge, and the water level rise was almost simultaneous with the first occurrence of the potential recharge. Then the water level variation pattern was smooth, and the yearly maximum level was reached each year about 2 months after the end of the potential recharge period. For P3, the water table level did not increase in 2004, however showed a steep increase at the end of the 2005 rainy season, followed by a gentle continuous increase. In this well, ephemeral water table variations suggest the occurrence of some preferential flow (from August to October in 2006 and 2007). Finally, the deep piezometers P5 and P6 displayed a declining tendency in 2004 and 2005, a stabilisation or a very gentle rise in 2006, and a marked rise at the end of the rainy season 2007. These contrasted patterns of water table variation among the different hillslope piezometers suggest that they are linked with local processes and not by a regional aquifer dynamics.

Comparison of the observed and the simulated water level variations in piezometers for the monitoring period show a very good agreement (Fig. 7), with Nash criteria values of 96.7, 94.1 and 95.0 for P10, P3 and P5 respectively. Calibrated parameters (Table 2) are relatively similar for the three piezometers. The most contrasted parameter is $WZMD_{max}$, the maximum water deficit of the weathered zone, because it accounts for the observed very long lag time

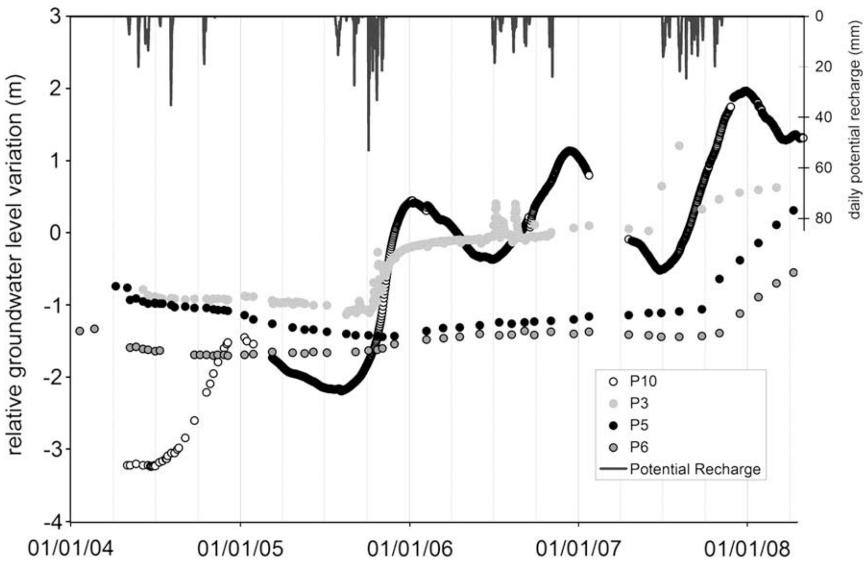

Fig. 6. Observed relative variation of water table level in hillslope piezometer, compared to simulated potential recharge. Reference values for water table depth are 16.8 m, 27.8 m, 39.1 m and 37.4 m for P10, P3, P5 and P6 respectively.

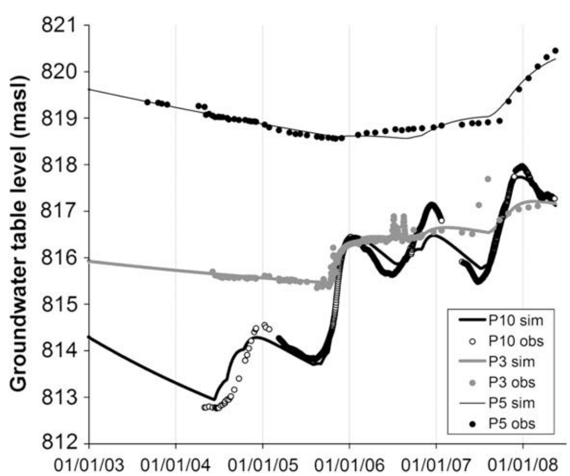

Fig. 7. Observed (dots) and simulated (lines) water level (in meter above sea level) in piezometer P3, P5 and P10 for the monitoring period.

Table 2
Model parameters for the three simulated piezometers. 95% confidence limits are given in the brackets, except for $WZMD_{max}$ which was adjusted by trial and error.

	P10	P3	P5
Ground level (masl)	832.82	844.24	859.14
Initial watertable depth (m)	16.82	27.75	39.14
Parameters			
$WZMD_{max}$	50	250	470
L_0	809.93 (1.691)	814.73 (0.394)	814.85 (0.101)
α_1	6.96×10^{-4} (4.99×10^{-4})	4.89×10^{-4} (2.84×10^{-4})	2.32×10^{-4} (4.82×10^{-5})
Sy	2.10×10^{-3} (1.17×10^{-3})	6×10^{-3} (2.5×10^{-3})	5.11×10^{-3} (1.9×10^{-3})
α_2	3.03×10^{-2} (1.5×10^{-2})	1.71×10^{-2} (7.1×10^{-3})	4.39×10^{-3} (3.2×10^{-3})

between water table rise observed in P10 in comparison with P3 (more than 1 year) and with P5 (more than 3 years). The calibrated value of the specific yield is consistent with the values obtained in similar fractured rock context in the region (Sekhar et al., 2004; Sekhar and Ruiz, 2006; Maréchal et al., 2006). The correlations among the parameters are summarized in Table 3a and b for the module 1 and module 2 respectively.

Table 3
Correlation matrix among the model parameters (a) module 1 and (b) module 2. For module 2, the correlation matrix displays the range of variation of the correlation coefficients obtained for the three piezometers.

	In	*a*	SA_{max}	SMD_{max}
a				
In	1	−0.03	−0.01	0.09
a		1	−0.2	−0.24
SA_{max}			1	0.18
SMD_{max}				1
b				
	L_0	α_1	α_2	Sy
L_0	1	0.39–0.83	−0.07 to 0.64	0.42–0.61
α_1		1	−0.09 to 0.55	0.20–0.64
α_2			1	−0.88 to −0.32
Sy				1

The most surprising result is the small value of the recession coefficient of the recharge reservoir (α_1). The recharge flow reaching the groundwater table is then very smooth, and it is mostly compensated by the groundwater discharge. The consequence is that the variations in the water content of the deep vadose zone (regolith and recharge reservoirs) are very large, with a maximum range of variation for the entire 32 year period of 650 mm, 745 mm and 851 mm for P10, P3 and P5 respectively. Although very large, these variations are compatible with the material porosity, considering the depth of this vadose zone (from 15 m to 40 m).

Average water balances components for each piezometer during the monitoring period and the 32 years simulation period are presented in Table 4. Groundwater underflow was lower during the monitoring period compared to the whole simulation period, even though rainfall and potential recharge were slightly higher. This is due to the fact that the effects of the drought period from 2001 to 2003 persisted longer at depth, which is apparent from the late rising of the deepest piezometers. The simulated long term variations of piezometers (Fig. 8) reveals an alternance of wet and dry phases, of 12–15 years duration period. Maximum level in P5 is

Table 4
Watershed balance components (in mm/year) for the piezometers P10, P3 and P5, for the monitoring period and the 31 year simulation. *cv % is the standard deviation of the respective quantity for the respective time period divided by the mean × 100.*

	2003–2007	cv (%)	1976–2007	cv (%)
Rf	**1155**	*(23)*	**1091**	*(32)*
AET_S	809	*(11)*	803	*(14)*
sim Q_s	107	*(59)*	94	*(79)*
Local balance				
P10 T_{WZ}	31	*(83)*	43	*(52)*
P10 Q_{GW}	115	*(28)*	141	*(26)*
P3 T_{wz}	62	*(89)*	114	*(53)*
P3 Q_{GW}	49	*(38)*	68	*(35)*
P5 T_{wz}	62	*(89)*	141	*(47)*
P5 Q_{GW}	34	*(8)*	42	*(22)*
Watershed balance				
T_{wz}	**53**	*(88)*	**104**	*(43)*
Q_{GW}	**62**	*(23)*	**78**	*(26)*
Water balance	**125**	*(103)*	**12**	*(1346)*

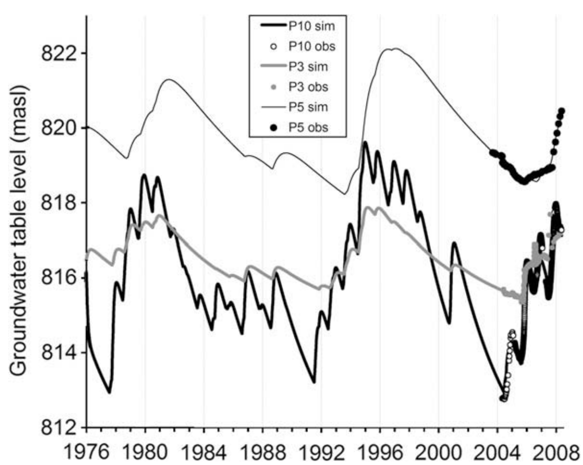

Fig. 8. Simulated variations of water table level (in masl) in piezometer P3, P5 and P10 for the 32 year simulation period (lines). Dots represent observed values.

reached 2–3 years later than in P10. This simulated pattern was compared to the data recorded by the Department of Mines and Geology (Karnataka State) from 1974 to 2007 in a shallow observation well located outside the forest zone 20 km east of the study site (Fig. 9). Although at the annual scale the observation well is much more reactive than P10, they display a very similar long-term trend. This observation suggests that the model described reasonably the global long-term behaviour of the groundwater.

Fig. 10 illustrates the simulated variations of evaporation and transpiration during 2 years. It shows that transpiration by trees from the soil layer is the dominant flux during most of the year, especially during rainy season. Soil evapotranspiration by the understorey vegetation can be significant during dry season, depending on the occurrence of isolated rainy events. Transpiration of water from the deep weathered zone occurs mainly during dry season, and during dry periods on the course of the monsoon season. Because the model computes transpiration successively from soil and then from the weathered zone, the latter can occur only when soil water is completely depleted, leading to abrupt alternances between the two fluxes (Fig. 10c), which are probably much smoother in reality.

With an objective to obtain an assessment of the water balance at the watershed scale, we need to assess the representativity of the monitored piezometers with respect to the entire area. The parameter $WZMD_{max}$, which is driving the most important part of the observed piezometer variability, is probably linked with the regolith depth in the vicinity of the piezometer. P10 is located in an area where the regolith depth is around 8 m, P3 around 15 m and P6 more than 20 m. A resistivity logging performed on P5 suggested a regolith depth of 22 m (Braun et al., 2008). As a first approach, we can consider that P10, P3 and P5 are representative of area with regolith depth of 0–12 m, 12–18 m and more than 18 m respectively. According to the regolith depth distribution in the watershed (Fig. 2), the proportion is 30%, 27% and 43% for P10, P3 and P5 respectively. With this hypothesis, the watershed balance (Table 4) appears roughly equilibrated during the 32 years simulation period, while the water gain was 125 mm year^{-1} during the monitoring period. The average water uptake by trees from the deep weathered zone is 104 mm, and average groundwater underflow is 78 mm year^{-1}. Groundwater recharge is equivalent to underflow on the long term. This value is close to the recharge that

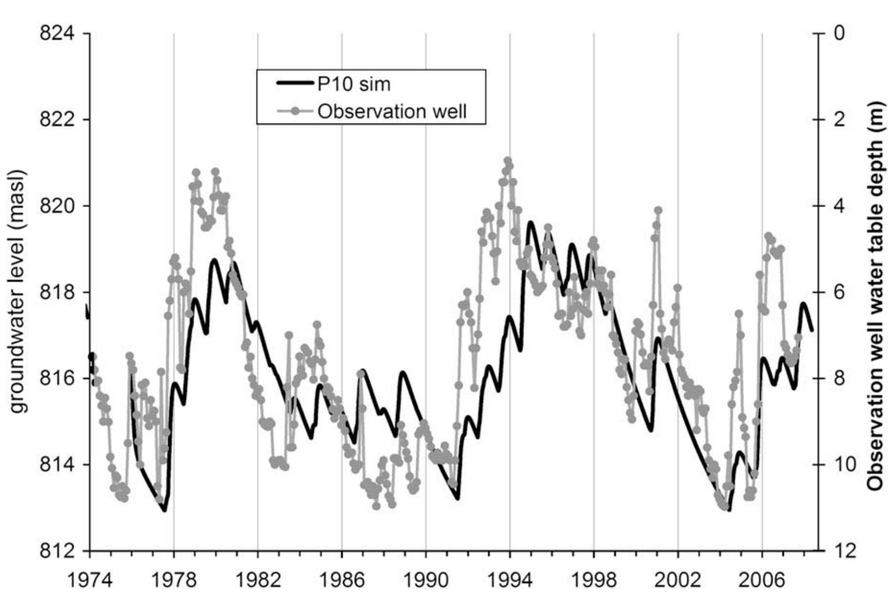

Fig. 9. Simulated variations of water table level (in masl) in piezometer P10 (black line, left *Y*-axis) and water level (depth to ground level in meters) recorded by Central Groundwater Board from 1974 to 2007 in a shallow observation well located in an agricultural zone 20 km east of the study site (grey line with dots, right *Y*-axis).

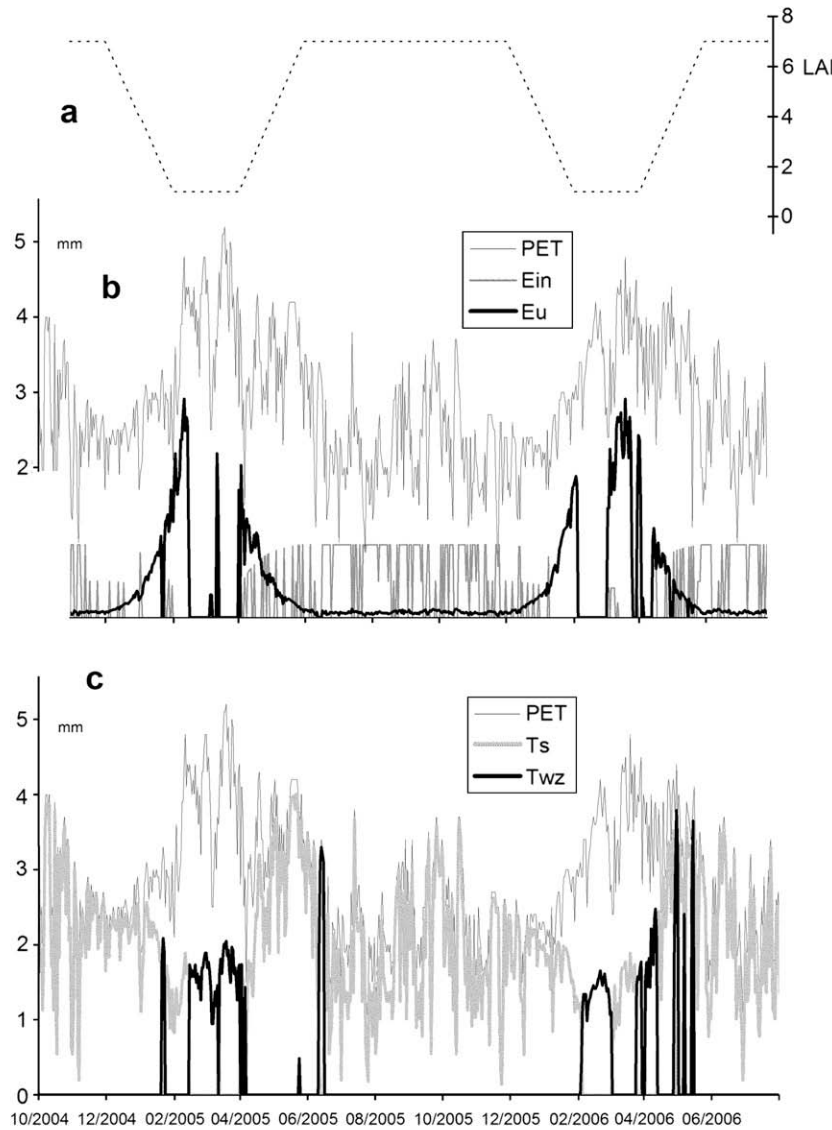

Fig. 10. Example of simulated daily evapotranspiration fluxes (in mm day^{-1}) compared to PET during 2 years (a) variations of LAI; (b) E_{In} and Eu; (c) T_s and T_{wz} (see text for signification of abbreviations).

was assessed in this area with Chloride mass balance method (45 mm year^{-1}, Maréchal et al., 2009).

Discussion

The 5 years monitoring allowed us to give a tentative assessment of the water balance of a small experimental watershed using a simple conceptual lumped model. However, the exercise has proven to be difficult, due to the climatic context and the presence of forest. The plausibility of the model results and the future work needed to validate them are discussed in this section.

One uncertainty is linked to the assessment of evapotranspiration in forest stands. This issue continues to generate a great deal of controversy in the literature (Andréassian, 2004; Bruijnzeel, 2004; Robinson et al., 2003). The difficulty is also greater for regions with the index of dryness (PET/Rf) close to 1.0 (Zhang et al., 2004), which is the case in our study site. However, despite its limitations, the Penman–Montieth PET approach has proven its ability to simulate soil water moisture in a broad range of climatic conditions and tree species (Granier et al., 1999). The average total evapotranspiration found in our long term simulations (around 900 mm year^{-1}, out of which about 100 mm are linked to extra transpiration by deep tree roots, see Table 4) is in very good agreement with the worldwide evapotranspiration curve proposed for forests and grasslands by Zhang et al. (2001). In our climatic context, water percolation towards the deep vadose zone is mainly concentrated during short rainy periods of the monsoon season, usually few days to 2 weeks, followed by drier periods. In this configuration, soil water budget is less sensitive to evapotranspiration assessment. However, direct measurements of forest evapotranspiration would allow a better calibration of the model.

According to our model, the spatial variability of the water reservoir located at depth in the weathered zone and accessible to deep roots of trees is a key parameter for water budgeting in semi-arid forested watersheds. Our calibration gave values ranging from 50 to 470 mm (Table 2). Few studies have been dedicated to measure the hydraulic properties of weathered granitic rocks (Jones and Graham, 1993; Katsura et al., 2006). They suggest that weathered rocks have the capacity to hold appreciable amount of water that is available to plants (Jones and Graham, 1993; Williamson et al., 2004).

The fact that tree roots are able to uptake water at considerable depth, especially in water limited ecosystems is widely accepted (Nepstad et al., 1994; Canadell et al., 1996; Collins and Bras, 2007). Importance of water storage in deep weathered rock in forested ecosystems has recently gained recognition, and large scale surveys to quantify deep water reservoirs have been attempted, for example in Cambodia (Ohnuki et al., 2008), leading to estimates as high as 1350 mm. The average total porosity of the saprolite in the Mule Hole watershed was estimated at 12% from geophysical and geochemical studies (Braun et al., 2008). Even assuming that the proportion of the porosity available to plants is only 5%, considering a 16 m deep saprolite would lead to an average water storage capacity of 800 mm, which is compatible with our findings. Considerable spatial variability of water stress and tree mortality during drought period is commonly reported in dry deciduous forests (Nath et al., 2006). A survey dedicated to check whether part of this variability can be explained by the regolith depth would constitute a validation of our hypothesis.

The most surprising consequence of the model calibration is the great importance of the recharge reservoir, and its low recession coefficient. In temperate regions, recharge process is usually considered to be very quick, and in most models water percolated below the soil zone is immediately transferred to the groundwater. This is acceptable because groundwater table is generally shallow, and the regolith matric porosity is filled every year by recharge. However, the fact that matric water can drain for a long time is well documented (Healy and Cook, 2002). In a regional study in Florida, O'Reilly (1998) found a increasing delay in groundwater response with aquifer depth. In a chalk aquifer, Price et al. (2000) demonstrated that the delay between recharge period and the smooth water table rise can be explained by matric water storage in the vadose zone, which is slowly released to groundwater during dry periods. By monitoring water content variations in a 21 m deep sandstone vadose zone, Rimon et al. (2007) found a variation of 660 mm of water content during a rainy season, and a slow decrease in water content during the dry season. In a hydrogeologic survey in Australia, in a granitic environment, Ghauri (2004) observes a long delay between rainfall and groundwater response, attributed to matric storage in the deep vadose zone. If this hypothesis is confirmed, it could be of considerable importance for water resource evaluation in hard rock aquifers, because in most cases recharge is assessed through water table level methods (Healy and Cook, 2002). Monitoring of water content variations in the deep vadose zone of our experimental watershed would allow validating this hypothesis.

The modelled watershed balance leads to an average water flow of about 180 mm year^{-1}, out of which 80 mm is groundwater underflow. This underflow might reach the streams of higher order rivers, like Nugu Hole or Kabini (Fig. 1). Indeed, it is very small compared to the estimates of the flows produced in the humid zone of the climatic gradient that range from 900 to 4700 mm year^{-1} (Putty and Prasad, 2000). However, these high flows are produced during the rainy season, and during the dry season rivers virtually dry up (Putty and Prasad, 2000). Because groundwater underflow from transition area produces a fairly constant load, it might be of significant importance in sustaining the baseflow in large rivers during the dry season. This will be assessed through a regional modelling in a future work.

Conclusions

This study is based on a 5 years monitoring of an experimental forested watershed in the South India, and using a conceptual model of the water balance over a 32 years period allowed us to draw the following conclusions:

(i) In tropical forest ecosystems, deciduous trees can uptake a significant amount of water from the deep regolith. This mechanism is particularly important at the end of the dry summer period, because the leaf flushing can precede monsoon rains by several weeks. More investigations are needed to check if variability in regolith depth can be linked to the variability of tree mortality during dry periods in monsoon season.

(ii) This water uptake, combined with the spatial variability of regolith depth, can account for the variable lag time between drainage events and groundwater rise observed for the different piezometers.

(iii) Water table response to the recharge is buffered due to the long vertical travel time through the deep vadose zone, which constitutes a major water reservoir. The 5 years monitoring period reveals that the watershed water balance is not equilibrated, mainly due to large variations in water content in the vadose zone. This observation is of great importance for water resource assessment, as water level fluctuation method is often used to estimate yearly groundwater recharge, especially in India (G.E.C., 1997). Our results show that this method can lead to an underestimation of recharge when vadose zone is large.

This study stresses the importance of long term observatories for the understanding of the hydrological processes in tropical forested ecosystems.

Acknowledgements

The Mule Hole basin is part of the ORE–BVET project (Observatoire de Recherche en Environnement – Bassin Versant Expérimentaux Tropicaux, www.orebvet.fr). Apart from the specific support from the French Institute of Research for Development (IRD), the Embassy of France in India and the Indian Institute of Science, our project benefited of funding from IRD and INSU/CNRS (Institut National des Sciences de l'Univers/Centre National de la Recherche Scientifique) through the French programmes ECCO–PNRH (Ecosphère Continentale: Processus et Modélisation – Programme National Recherche Hydrologique), EC2CO (Ecosphère Continentale et Côtière) and ACI-Eau. It is also funded by the Indo-French programme IFCPAR (Indo-French Center for the Promotion of Advanced Research W-3000). The multidisciplinary research carried on the Mule Hole watershed began in 2002 under the control of the IFCWS (Indo-French Cell for Water Sciences), joint laboratory IISc/IRD. We thank the Department of Mines and Geology of Karnataka for providing data and the Karnataka Forest Department and the staff of the Bandipur National Park for all the facilities and support they provided. Thanks to the Editor-in-Chief and the Associate Editor, as well as A.M. O'Reilly and one anonymous reviewer for their thoughtful reviews and advice.

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