Chapter 2.2 : Impact of the forest on the hydrological cycle and chemical balance in tropical humid watershed (Mule Hole, India)

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CHAPTER 2.2

IMPACT OF THE FOREST ON THE HYDROLOGICAL CYCLE AND CHEMICAL BALANCE IN A TROPICAL HUMID WATERSHED (MULE HOLE, INDIA)

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Abstract

It is commonly accepted that forest plays a role in modifying the water cycle at the watershed scale. However, the impact of forest on aquifer recharge is still under discussion: some studies indicate that infiltration is facilitated under forest while other studies suggest a decrease in recharge. This paper presents an estimate of recharge rates to groundwater in a humid forested watershed in India. Recharge estimates are based on the joint use of several methods: chloride mass balance, water table fluctuation, geophysics, groundwater chemistry and flow analysis. Two components of the recharge (direct and indirect) are estimated over a 3-years monitoring period (2003-2006). The direct and localised recharge resulting from rainfall over the entire watershed surface area is estimated at 45 mm/yr while the indirect recharge occurring from infiltration of the stream during flood events is estimated at 30 mm/yr for a two km-long stream. Calculated recharge rates, rainfall and runoff measurements are then combined in a water budget to estimate yearly evapotranspiration which ranges from 80 to 90% of the rainfall, and corresponds to 1100 mm/yr on average. This unexpectedly high value for a deciduous forest is nevertheless in agreement with the forest worldwide relationship between rainfall and evapotranspiration. The water yield of the watershed is low with an average of 180 mm/yr and does not increase much if rainfall increases, due to the high demand for water from the forest. The high level of evapotranspiration from the forest cover contributes to a decrease in the recharge rate; consequently the water table is low. This is the reason why the stream is highly ephemeral there. Another impact is that the water table is recharge-controlled and not topography-controlled. As a result, the water table is not a subdued replica of the local topography. It results in groundwater flows below the existing weir which contributes for more than 90 % of the total chemical outputs, and for less than 10 % of surface runoff. More generally, the impact of forest cover on the water yield (based on empirical worldwide curves, and compared with grass cover) appears to range between 210 and 340 mm/yr according to the rainfall rate, and other sites specific parametres.

Impact de la Forêt sur le Cycle Hydrologique et le Bilan de Matière dans un Bassin Versant Tropical Humide (Mule Hole, Inde)

Résumé

Il est généralement admis que la forêt joue un rôle en modifiant le cycle de l'eau à l'échelle d'un bassin versant. Cependant, l'impact de la forêt sur la recharge est encore source de débats : certaines études indiquent que l'infiltration est facilitée sous couvert forestier tandis que d'autres suggèrent une diminution de la recharge à la nappe aquifère. Cet article décrit l'estimation de taux de recharge vers les eaux souterraines dans un bassin forestier en zone humide en Inde. Les estimations de recharge sont basées sur l'utilisation conjointe de plusieurs méthodes : le bilan de masse en chlorures, les fluctuations de niveaux de nappe, la géophysique, la chimie de l'eau souterraine et l'analyse de flux souterrains. Deux composantes de la recharge (directe et indirecte) sont estimées durant trois années d'étude (2003-2006). Les taux de recharges directe et localisée sont estimés à environ 45 mm/an à l'échelle du bassin versant tandis que la recharge indirecte résultant de l'infiltration à partir du lit du ruisseau durant les épisodes de crue est estimée à 30 mm/an pour un cours d'eau long de deux kilomètres. Les flux de recharge, pluie et ruissellement sont ensuite combinés dans un bilan hydrologique pour estimer l'évapotranspiration réelle qui oscille entre 80 et 90% de la pluie totale, soit 1100 mm/an en moyenne. Cette valeur élevée pour une forêt décidue est néanmoins en accord avec la relation empirique mondiale entre pluie et évapotranspiration dans les bassins forestiers. La pluie efficace est faible avec une moyenne de 180 mm/an à l'échelle du bassin versant et n'a pas un potentiel d'augmentation très élevé en cas d'augmentation de la pluie compte tenu de la forte demande en eau de la forêt. La forte évapotranspiration du couvert forestier contribue à diminuer le taux de recharge ; par conséquent, le niveau de la nappe est bas. C'est la raison pour laquelle le ruisseau est très éphémère. Un autre impact est que le niveau de la nappe est contrôlé par la recharge et non pas par la topographie. Il en résulte que le niveau piézométrique n'est pas une réplique de la topographie et que les écoulements souterrains sous l'exutoire du bassin versant topographique contribuent à plus de 90% des exports de solutés, tandis que les exports via le ruissellement de surface sont inférieurs à 10%. Plus généralement, l'impact forestier sur la pluie efficace, estimé sur la base des courbes empiriques mondiales, varie entre 210 et 340 mm/an selon le taux de précipitations.

Introduction

The impact of forests on the water yield (runoff + recharge) of a watershed is a question of interest and is still under debate. The vast majority of studies indicate decreased runoff from areas under forests compared to areas under shorter land covers (Jewitt, 2005). References on the effects of forestation on natural recharge are less abundant, especially in fractured crystalline aquifers. Preliminary studies in Australia have shown that clearing of natural woodland or forest has resulted in the increase of recharge and rise of the water tables (Williamson, 1990). Sharma et al. (1987) arrived at the same conclusion after forest clearing in a lateritic environment. On the other hand, some studies show there is usually less soil compaction in forests and, depending on soil type, soil structure may improve causing more rainfall to infiltrate (Bruijnzeel, 2004; Ilstedt et al., 2007).

The total recharge to groundwater is made up of three main components. Direct recharge refers to diffuse recharge, such as from precipitation, whereas indirect recharge results from the percolation of a portion of runoff water through the stream beds, and localised recharge refers to concentrated recharge by preferential flow through cracks, joints and fissures (Lerner et al., 1990). The relative proportions of these components fluctuate according to climatic conditions, geomorphology and geology.

The aims of this study are (i) to identify significant recharge processes at the watershed scale in a forested area and (ii) to assess the direct, localised and indirect transient recharge rates. In order to reach these objectives, a combination of several approaches focusing on the saturated zone - chloride mass balance, water table fluctuation, geophysical investigations, pumping tests and groundwater flow modelling - is used. Then, a water budget and a chemical balance are calculated at the watershed scale. This study has been conducted on an experimental watershed in humid conditions located on the Indian crystalline basement. The watershed is the object of an integrated study including soil dynamics and erosion processes (Barbiero et al., 2007), geophysical investigations (Descloitres et al., 2008) and hydrological and biogeochemical cycles (Braun et al., 2008).

Field Setting and Methodology

The Mule Hole experimental watershed is situated in the Western Ghats, in southern India (Figure 1), at 11° 44' N and 76° 27' E (Karnataka state, Chamrajnagar district). The watershed area (4.1 km^2) is mostly undulating with gentle slopes and the elevation of the watershed ranges from 820 to 910 m above mean sea level. The watershed is covered by a dry deciduous forest composed of three main types of vegetation. (1) Anogeissus latifolia, Terminalia crenulata and Tectona grandis and referred here as “ATT community”; this is the dominant association and covers approximately 70% of the watershed; silicarich Themeda triandra (elephant grass) are associated with the ATT community. (2) The “Shorea community” is characterised by the presence of Shorea roxburghii and Lagerstroemia microcarpa; mainly present on shallow red soils, it covers 15% of the watershed. (3) The “Swamp community” consists of grass-covered glades with scattered trees (Ceristoides turgida and Diospyros melanoxylon) and is developed on black soils only and covers 5% of the watershed.

The study site is located in the climatic semi-humid transition area and the mean annual rainfall (n = 20 years) is 1120 mm. The mean yearly temperature is 27°C. On the basis of the aridity index (AI) defined as the ratio of mean annual precipitation (P) to potential evapotranspiration for grass reference (ET_0): $AI = P/ET_0 = 1.2$, the climate regime can be classified as humid (UNESCO, 1979). Nevertheless, the climate is characterised by recurrent but non-periodic droughts, depending on monsoon rainfalls.

The substratum belongs to the Precambrian Dharwar supergroup (Moyen et al., 2001) and consists of gneiss with amphibolite and quartz dykes. The mean strike value is N80°, with a dip angle ranging from 75° to the vertical. Because of the dominant erosion process in such a head catchment, no fine deposit is present in the stream bed.

Geophysical measurements were carried out to estimate the main geological units at the watershed scale (Descloitres et al., 2008) (Figure 1). 2D Electrical Resistivity Tomography (ERT) interpretation showed that (i) fresh bedrock is strongly affected by sub-vertical fractures, (ii) bedrock is overlaid by a thin weathered-fissured zone (see (Dewandel et al., 2006) for its definition) of a few metres, and then by weathered material ranging from clayey to sandy rocks, and (iii) the depth to the bedrocks is highly heterogeneous, i.e. from a few metres to several tens of metres at the watershed outlet area. Magnetic Resonance Soundings (MRS) were also implemented on the watershed (Legchenko et al., 2006). MRS interpretation indicated that the main groundwater reservoir is located in the weathered-fissured zone of the bedrock. Further, time-lapse geophysical measurements of both ERT and MRS conducted at the outlet of the watershed indicated seasonal infiltration of water under the stream (Descloitres et al., 2008).

A set of 13 observation wells (named P1 - P13) were drilled in the area. In this National Park which is home to protected wildlife, authorisation was obtained for drilling along the existing tracks only. Three wells (P3, P5 and P6 on Figure 1) were drilled at the boundaries of the watershed in order to obtain the background characteristics of the aquifer far away from the main flowing stream network. The remaining ten wells (including collapsed well P4, not shown on the map) were drilled along a straight line crossing the stream axis at its outlet (Figure 1) in order to monitor the effects of water seepage from the stream. The water levels have been monitored in all the wells from their drilling time up to now, either manually at a monthly time step or automatically at a hourly time-step.

Runoff at the outlet has been measured by a stream gauging station composed by a water level indicator coupled with a Doppler-flowmeter located in a specially designed section of the stream (the error on discharge rate is assumed to be less than 20 %). Rainfall has been measured at a meteorological station located at Mule Hole checkpost (Figure 1) with an estimated error of 6%. During three and half hydrological years of monitoring from 1st July 2003 to 31st December 2006, the stream only flowed for an equivalent period of 147 days, i.e. 11% of the monitoring period or about 42 days a year. The stream is therefore highly ephemeral with very fast recession periods and flows only in response to specific heavy rainfall events.

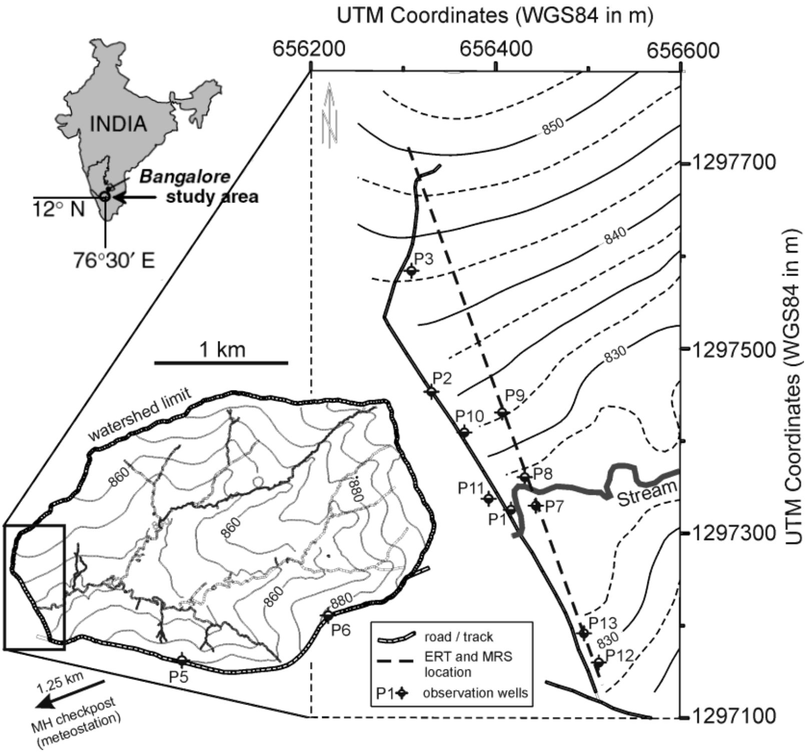


Fig. 2.2-1 - Location map of the experimental site at the outlet of the Mule Hole watershed (modified from Descloitres et al., 2008). Collapsed observation well P4 is not represented (elevation in MASL).

Apart from rainfall, the meteorological station provides an hourly time series of humidity, temperature and wind speed. As the watershed is mainly forested, the reference evapotranspiration (ET_{ref}) has been calculated at an hourly time step using the Penman-Montheit equation using alfalfa (tall: 0.5 meter) coefficients C_n = 66, C_d = 0.25 for daytime and C_d = 1.7 for night-time (Allen et al., 2006).

Bulk rain samplers continuously open to the atmosphere have been used to sample bulk deposition close to the meteorological station (checkpost, Figure 1). Chloride was also measured in the ephemeral stream, the automatic sampling frequency during floods depending on the water level. Chloride concentrations were measured using an Ion Chromatograph Dionex 600, with a detection limit lower than 1 µmol/L and an analytical error less than 5%.

The recharge has been determined using the Chloride Mass Balance (CMB) method in the observation wells. This method is based on the assumption of conservation of mass between the input of atmospheric chloride and the chloride flux in the subsurface (Dettinger, 1989).

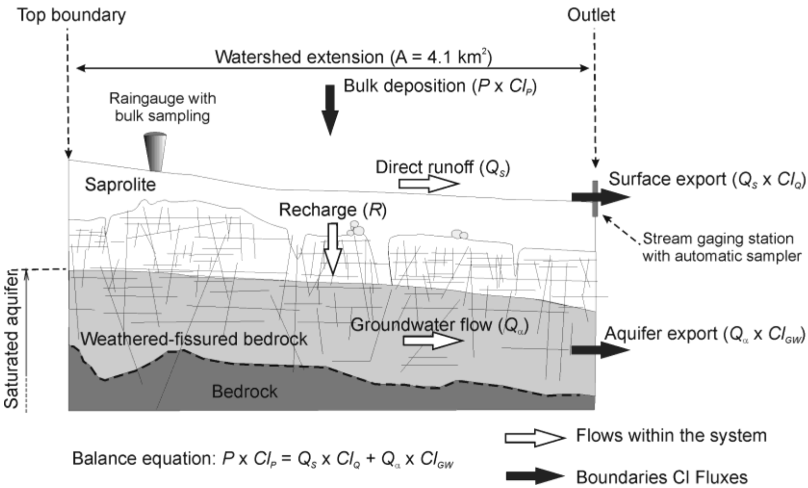


Fig. 2.2-2 - chloride mass balance in a small head watershed with water table disconnected from the ground and runoff measurements at the chosen outlet.

Considering (i) an unsaturated zone where evapotranspiration and mixing of rainfall and pore water takes place, (ii) measurements of runoff at the outlet, (iii) a recharged aquifer which flows below the outlet of a small head watershed and does not contribute to the direct runoff at the outlet (Figure 2), the equation of chloride mass balance is:

$$P\,Cl_P = Q_S\,Cl_Q + Q_\alpha Cl_{GW} \tag{1}$$

with Cl_P: weight-average chloride concentration in bulk precipitation (µmol/l), Cl_Q: weight-average chloride concentration in runoff (µmol/l), Cl_{GW}: average chloride concentration in groundwater (µmol/l), P: average precipitation rate (l/yr), Q_S: average runoff (l/yr) and Q_α(l/yr): average groundwater outflow (l/yr), all these values being averages over the sampling period. Assuming that the annual groundwater outflow is equal to the annual recharge rate ($Q_\alpha = R$) which corresponds to the absence of long-term systematic increase or depletion of the water table (see Figure 5), the recharge rate (R) through the unsaturated zone can be calculated by:

$$R = \frac{P\,Cl_P - Q_S\,Cl_Q}{Cl_{GW}} \tag{2}$$

In considering the runoff component, Equation (1) becomes an extended version of the equation commonly used in the literature for recharge estimation (Bazuhair and Wood, 1996; Wood and Sanford, 1995). This equation can be used in transient conditions provided regular, long-term (several years) monitoring of chloride is conducted on rainfall, surface and groundwater. The weight-average chloride in precipitation and runoff are calculated using the same equations as (Wood and Sanford, 1995). A simple average is calculated for chloride in groundwater using the waters regularly sampled in the observation wells.

Some recent studies have shown that chloride can be temporarily retained in the soils due to interaction with organic matter (Öberg and Sandén, 2005). It appears that the durations of the processes involved are, at most, a few months. This is irrelevant for long-term aquifer recharge estimation using several years' data (Alcalá and Custodio, 2008; Scanlon et al., 2006) and equation (2) can be used for recharge estimation.

In the Mule Hole watershed, the substratum is free of evaporites and very poor in chloride bearing minerals. Moreover, the watershed has been preserved from human activity since the 17th Century at least, because it belonged to the Maharaja of Mysore and was later incorporated into the Bandipur National Park. Then, it may be reasonably considered that chloride comes from atmospheric deposition only and that the CMB technique method can be applied with accuracy.

Assuming an analytical error of 5% on chloride content measurements in water, an instrumental error of 6% on rainfall and 20% on runoff gauging, a total error of 20% is obtained on recharge estimates using the classical uncertainty propagation theory.

Results

Groundwater Flows and Water Level Time Series

During the dry season, decreasing water levels from P6 to P5 and to the western wells-line suggest a regional groundwater system flowing roughly from east to west. Far away from the stream (wells P3, P5 and P6), the water table remains deep below the stream level even during the monsoon. The water table is then always disconnected from the stream (see a schematic cross section in Figure 2). This explains the absence of springs in the watershed and the absence of baseflow measured at the outlet. Below the stream, water seepage recharges the aquifer as shown by time-lapse geophysical investigations (Descloitres et al., 2008). At the experimental site, the stream has incised the soils profiles to a depth of 2 metres (Barbiero et al., 2007), therefore the case study corresponds to a disconnected losing stream without a clogged streambed and with a shallow water table (Peterson and Wilson, 1988).

At the experimental site crossing the stream axis, the water table fluctuates according to the monsoon regime. The water levels are located at an average depth of 8 metres below the stream bed during the dry season. During the monsoon, the water levels rise close to the stream (at monitoring wells P1 and P7) up to a level of about one meter below the stream bed.

The water table fluctuations (Δh) between pre- and post-monsoon seasons are highly variable over space. The influence of the distance to the stream network on the water table level is analysed in Figure 3 where three groups of wells can be identified. Group 1 (wells P3, P5 and P6) is characterised by small water table fluctuations (Δh < 2.5 m) observed at wells far away (> 200 m) from the nearest stream. Wells in group 2 (P8, P9, P10, P12, P13), located in the vicinity of the watershed flowing streams (distance comprised between 40 and 200m), show medium water level fluctuations (3 < Δh < 5 m). Group 3 (P1, P7, P11) is made up of wells located in the close vicinity of a surface stream (< 40 m) and characterised by high (Δh > 7 m) water level fluctuations. This relationship between water level fluctuations and distance to the surface stream confirms the existence of a heterogeneity in the recharge process, mainly caused by the losing stream.

The analysis of water level fluctuations in wells from group 3 provides information on the dynamics of indirect recharge from the stream. For instance, the response of this group (wells P1 and P7) to the first rain events of April – June 2005 was very marked (Figure 4). During the first event identified (57 mm rainfall on 10th April 2005), the water levels rose by 3.7 m in 46 hours at P1 and by 2.2 m in 56 hours at P7. At the same time, the water level rose gently by 0.61 m in 17 days at P13 (group 2). Wells P3 (group 1) and P10 (group 2) do not show any response to these events. During the second event (17 mm rainfall on 27th April 2005), the water levels rose by 3.37m in 63 hours at P1 and by 2.9 m in 68 hours at P7 while the rise is only 0.56 m in 11 days at P13. After that, despite several rainy events (i.e. 38 mm on 25th May and 29 mm on 20th June 2005), the water levels in all the wells decrease in the absence of any runoff in the stream. Large water level fluctuations are correlated with runoff occurrence in the stream. Over the whole observation period, the water levels are higher below the stream-axis than elsewhere which results in a groundwater mound. The resulting hydraulic gradient potentially induces groundwater flows from the stream axis towards North and South.

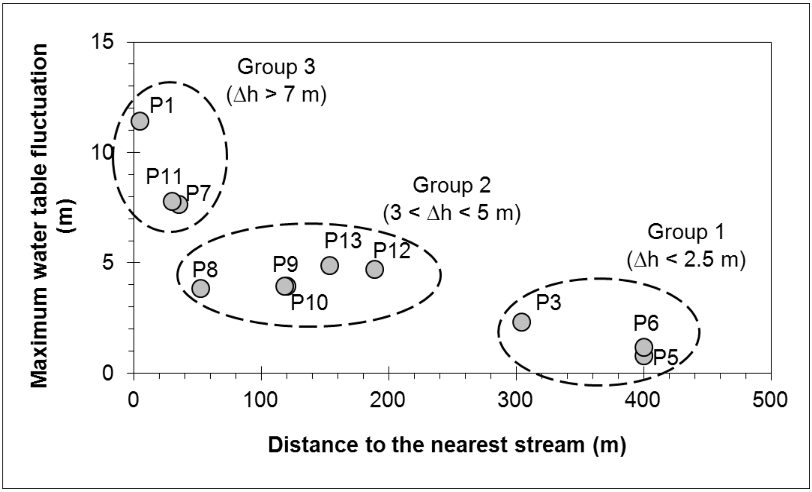


Fig. 2.2-3 - Maximum yearly water level fluctuation observed during the period 2003-2006 as a function of the distance from the nearest surface streams (P2 is not mentioned as water table fluctuations are disturbed by rainfall infiltration into the well tube)

The water table fluctuations in wells P1 and P7 (close to the stream) are linked to direct recharge from rainfall and indirect recharge from the stream. The relationships between rainfall and stream head (as input functions), and water levels below the stream (as output) have been evaluated using a cross-correlation analysis in order to estimate the relative importance of both processes. Although it is low in both cases, the correlation coefficient is clearly higher for the stream/aquifer relationship than for rainfall/aquifer relationship (Maréchal et al., 2009). The influence of the seepage from the stream is preponderant for the water level signal at wells P1 and P7 from group 3. In other words, the water table fluctuations at wells P1 and P7 are more dependent on the stream head than on rainfall. The low average correlation coefficient is induced by the difference in the duration of input and output events: short rainfall and storm events are not well correlated to long-duration water table fluctuations.

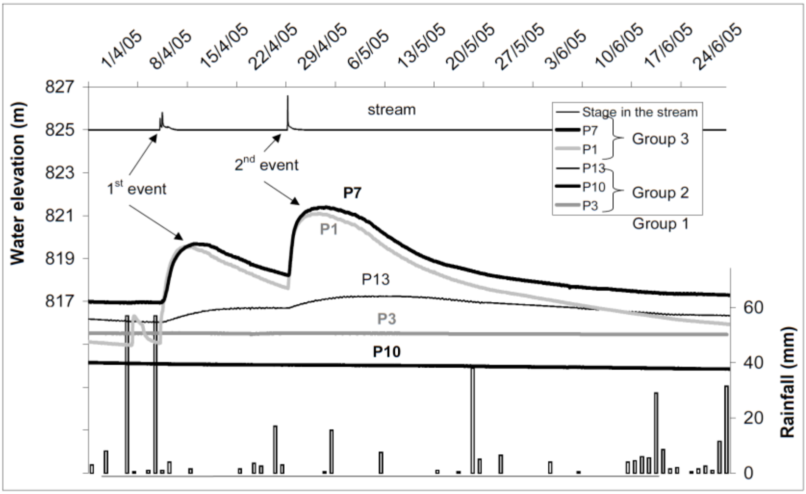


Fig. 2.2-4 - Hourly water levels in the stream and observation wells; April – June 2005

Chemical Content of Waters

The chloride concentrations in rainfall during the survey from 2003 to 2006 range from 1 to 129 µmol/l (*n* = 202). The weight-average concentration is 24.6 ± 1.2 µmol/l, which corresponds to an input flux of chloride of 275 ± 30 mol/ha/yr. The Na/Cl molar ratio is 0.71, within the range 0.70 – 0.85 classically found for seawater derived rainfall (Möller, 1990).

The chloride concentrations in surface water at the outlet during the survey from 2003 to 2006 range from 16 to 164 µmol/l (*n* = 198). The weighted chloride concentration in surface water is 38.1 ± 1.9 µmol/l. The mean discharge rate on 2003 - 2006 period at the outlet is Q_S = 3.24 x 10^5 m^3/yr (79 ± 16 mm/yr on average) and the resulting total output flux of chloride is 30 ± 7.5 mol/ha/yr.

The partition of groundwater into three groups based on water level fluctuations and distance from the main flowing stream (see above) is globally consistent with the chemical compositions of the groundwater and their variations through time. All sample compositions are dominated by alkalinity for anions and by calcium-magnesium-sodium for cations.

Waters from group 1 (P3, P5 and P6) are relatively concentrated, with mean electrical conductivity (EC) ranging from 670 µS/cm in P5 to 935 µS/cm in P6, and mean Cl concentrations of 913 µmol/l in P3, 528 µmol/l in P5 and 450 µmol/l in P6. The chemical composition of well P5 is stable over the monitoring period, as illustrated in Figure 5a.

Waters from group 2 are, on average, less concentrated than those from group 1. Mean EC ranges from 274 µS/cm in P12 to 670 µS/cm in P8. Mean Cl concentrations are 303 µmol/l in P8, 327 µmol/l in P9, 477 µmol/l in P10 and 164 µmol/l in P13. Chloride concentrations are sensitive to season, with dilution during monsoon and concentration during dry seasons (Figure 5b).

Waters from group 3 (P1, P7 and P11) are the most diluted, with mean EC ranging from 244 to 573 µS/cm and mean Cl concentrations from 114 to 137 µmol/l in P1 and P7, respectively. The overall chemical composition of these wells is also variable throughout the year, with well-marked concentration during dry seasons in 2004 and 2006 (Figure 5c). Chloride concentrations vary from 62 to 360 µmol/l in P1, from 33 to 251 in P7 and from 29 to 450 in P11, EC from 151 to 461 µS/cm in P1, from 161 to 716 µS/cm in P7 and from 100 to 783 µS/cm in P11.

Globally, calcium and magnesium represent between 60 and 95% of the cationic balance, with no strong differences in proportion between the three groups.

Discussion

Direct, Localised and Indirect Recharges

The chloride fluxes are used in equation (2) from the CMB method to calculate the recharge rates using the average chloride content in each observation well and weighted averages chloride content in precipitation and runoff. The values obtained at the various monitoring wells range from 27 to 250 mm/year (Table 1).

According to the most probable changes in groundwater flow directions, the chloride content fluctuates with time and characterises mixed groundwater from various recharge areas. Therefore, in wells of group 3 and a part of group 2 (wells P8, P9 and P12), the estimated recharge corresponds to a yearly average on groundwater flowing at the sampled point. The values of total recharge estimated by CMB method at these sites do not correspond necessarily to recharge which happened through the unsaturated zone above the sampled water table.

The calculated recharge increases from group 1 to group 3. The wells P3, P5 and P6 (group 1) are located sufficiently far away from the stream to be considered as not influenced by it: it is assumed that indirect recharge is negligible ($R_i = 0$) at these wells. In such a fractured system, the localised recharge through cracks and fractures is the main recharge mechanism, especially during storm events (Shivanna et al., 2004). Therefore, the recharge estimated from group 1 can be considered as the reference for direct and localised recharge $R_{d,l} = R_d + R_l$ in the watershed: it ranges from 27 to 55 mm/year with an average of about 45 mm/yr. The ratio $R_{d,l}/P$ of (direct and localised) recharge to rainfall at Mule Hole watershed is very low compared to values obtained for direct recharge only from tritium injection experiments in several geologically similar catchments located in non-forested areas of India (Rangarajan and Athavale, 2000). The application of the relationship proposed by these authors to the rainfall quantity observed on Mule Hole watershed (P = 1120 mm/yr) leads to an expected direct recharge value of about 150 mm/yr, much more than that observed at wells from group 1 (Figure 6). The most probable cause of the observed deficit of direct recharge in this watershed is the high evapotranspiration from forest cover. This process occurs in the vadose zone as observed in many forested watersheds throughout the World (Zhang et al., 2001). It means that the forest could consume an excess of about 100 mm/yr of actual evapotranspiration compared to a non-forested case. This low rate of direct recharge at watershed scale partially explains the great depth of the water

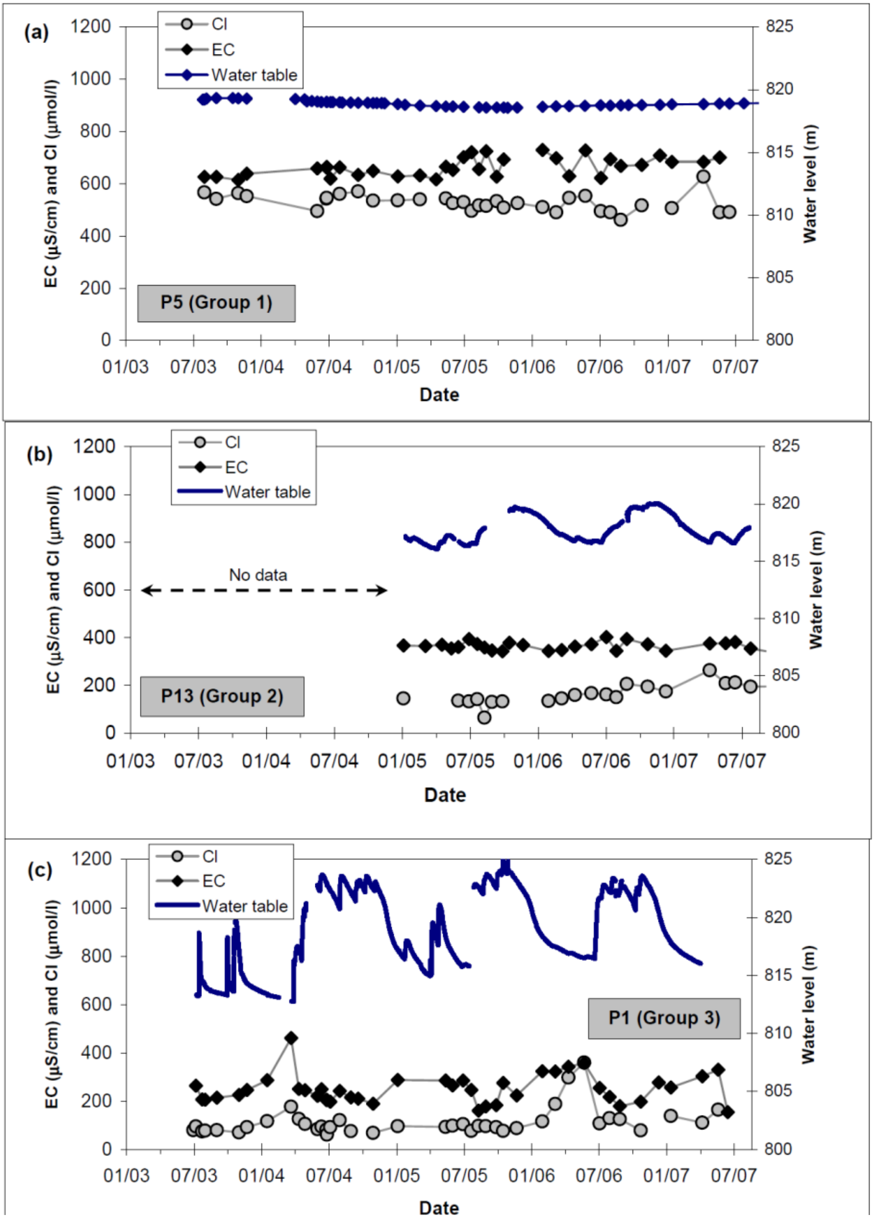


Fig. 2.2-5 - electrical conductivity (EC), chloride content and water table elevation in monitoring wells P5 (a), P13 (b) and P1 (c). *Well P13 was drilled in Jan. 2005.

Group	Well	Cl_{GW} (µmol/l)	R (mm/year)	R_i[1] (mm/year)
Group 1	P3 [2]	913 ± 46	27 ± 5	0 ± 14
	P5	528 ± 26	47 ± 9	2 ± 18
	P6 [2]	450 ± 22	55 ± 11	10 ± 20
Group 2	P2 [2]	533 ± 27	46 ± 9	1 ± 18
	P10	477 ± 24	52 ± 10	7 ± 19
	P9	327 ± 16	75 ± 15	30 ± 24
	P8	303 ± 15	81 ± 16	36 ± 40
	P12	202 ± 10	122 ± 24	77 ± 33
	P13	164 ± 8	150 ± 30	105 ± 39
Group 3	P7	137 ± 7	178 ± 35	133 ± 44
	P1	114 ± 6	215 ± 43	170 ± 52
	P11	98 ± 5	250 ± 50	205 ± 59

Table 2.2-1 - Average chloride content in groundwater and calculated recharge rate using CMB method (with P = 1120 mm/yr; Cl_P = 24.6 µmol/l; Cl_Q = 38 µmol/l) sorted according to increasing recharge rate. [1] Indirect recharge component R_i is estimated using the difference between total recharge and assumed direct and localised recharge $R_{d,l}$ = 45 ± 9 mm/year. [2] Chloride content in wells P2, P3 and P6 has been averaged over the first months after drilling in order to consider the pristine composition of groundwater not perturbed by vertical flows induced in the well.

table (40.3 m at P5 and 38.4 m at P6) in this watershed, reversing stream/aquifer interaction with losses from the stream. Lowering of the water table in this way, through a reduction in the recharge of groundwater by forest cover, has been already reported in the Karnataka region under Eucalyptus (Shiva and Bandyopadhyay, 1983).

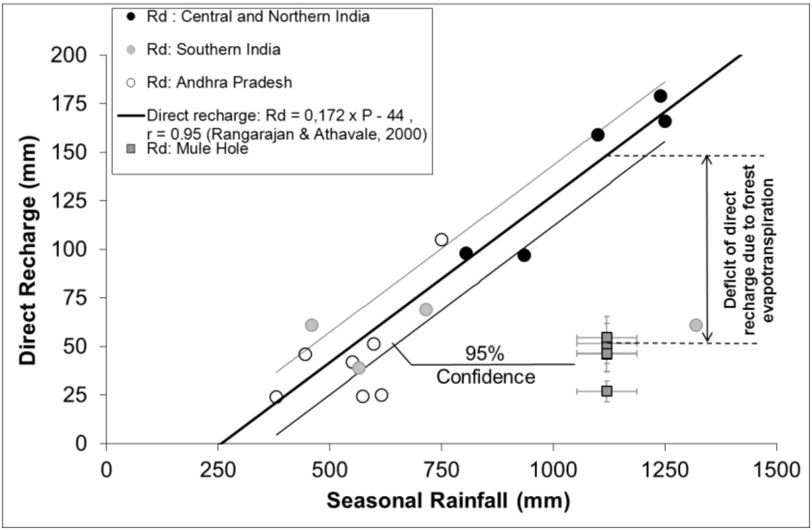


Fig. 2.2-6 - Rainfall - direct recharge in granite and gneiss. Andhra-Pradesh, Southern, Central and Northern India direct recharge estimated using tritium injection (Rangarajan and Athavale, 2000; Sukhija et al., 1996). Comparison with direct recharge measured in Mule Hole watershed. The straight line corresponds to the general trend from observation sites apart from Mule Hole and without significant forest cover.

The total recharge to the groundwater in areas very close to the stream (P1, P7 and P11, Figure 7) ranges from 180 to 250 mm/yr in monitoring wells from group 3. This total recharge corresponds to the addition of direct recharge from rainfall diffuse percolation, localised recharge from percolation through cracks and joints and indirect recharge from the stream. Intermediary values (120-150 mm/yr) at P12 and P13 in the vicinity of the stream (200 m) show an influence of indirect recharge (group 2). Wells located on the Northern bank of the stream are characterised by direct and localised recharge (P2, P3 and P10) or very low indirect recharge (P8 and P9). This is consistent with low water table fluctuations at these points (Figure 7) and suggests the existence of an asymmetry in groundwater flows from the stream axis. The water received from indirect recharge should mostly flow to the South as suggested by Figure 7 and by chloride concentrations.

Descloitres et al. (2008) compared MRS and ERT measurements carried out at the outlet of the watershed before and during the 2004 monsoon (Figure 1). They concluded that a seasonal indirect recharge occurs under the stream. Moreover, a cross section of MRS water content obtained from the interpretation of 15 soundings located at the outlet of the watershed (Figure 7) has been proposed by Legchenko et al. (2006). The MRS were carried out in November 2004 during the monsoon. The water content distribution indicates that the saturated water reservoir is strongly dissymmetric and mainly located on the south bank of the stream. MRS water content ranges between 0.5% (threshold value for the used device) and 2%.

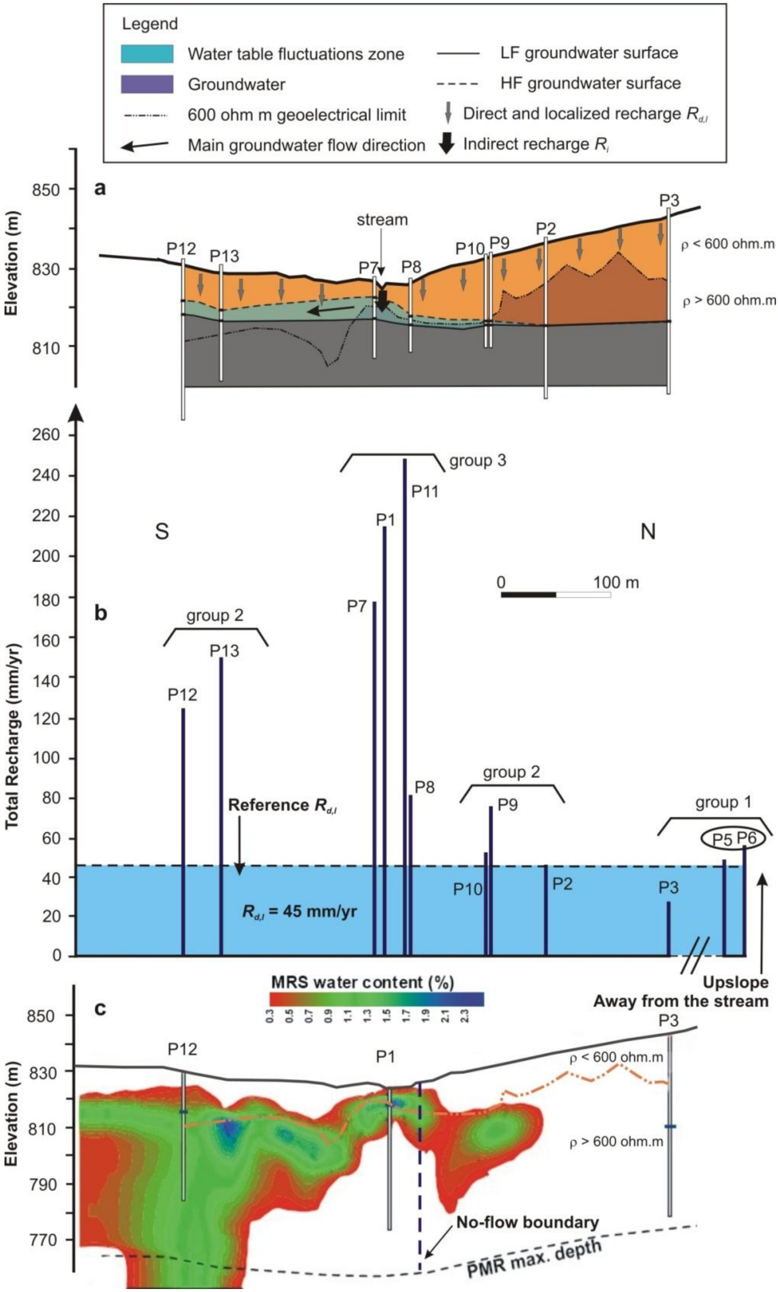
Hydraulic tests were carried out in several boreholes of the watershed in 2006. The interpretation of the 18 hours pumping tests conducted in P1 and monitored in P1, P7 and P8 indicated a no-flow boundary (Figure 7). This boundary is located north of the stream and is orientated parallel to the geological structures, i.e. N70°. It acts as an underground barrier and is probably caused by an uplift of the bedrock. This is the reason why indirect recharge under the stream cannot flow easily to the north. Such underground no-flow boundary is not clearly identified when interpreting the pumping test of 50 hours conducted in the south of the stream (pumping in P12 and monitored in P12 and P13).

Assuming $R_{d,l} = 45 \pm 9$ mm/year, the indirect recharge due to infiltration from the stream may be estimated (Table 1) and it fluctuates between 0 and 200 mm/yr. These values are representative of local areas close to the sampled wells and the stream; they cannot be scaled up to the entire watershed area, at which scale the indirect recharge is much less.

Groundwater Mound and Flows

To better understand how the aquifer responds to indirect recharge from the Mule Hole stream, the water level fluctuations at monitoring wells were modelled using an analytical solution. To this end, the analytical solution of Kresic (1997) was implemented in superposition to take into consideration the complex water level change close to the stream due to indirect recharge (Maréchal et al., 2009).

Fig. 2.2-7(next page) - Combination of the approaches. (a) Cross section of the experimental site with water table fluctuations (LF: pre-monsoon low flow in June 2006; HF: post-monsoon high flow in December 2005), Observation wells P1 and P11 are not figured for clarity reason. (b) Recharge estimated using CMB method ("S" and "N" mean "South" and "North", and is valid for figures a, b and c). (c) Water content from geophysical measurements (MRS) and no-flow boundary from hydraulic tests.



Legend
Water table fluctuations zone
Groundwater
600 ohm m geoelectrical limit
Main groundwater flow direction
LF groundwater surface
HF groundwater surface
Direct and localized recharge $R_{d,l}$
Indirect recharge R_i
a
Elevation (m)
850
830
810
stream
P12
P13
P7
P8
P10
P9
P2
P3
ρ < 600 ohm.m
ρ > 600 ohm.m
b
Total Recharge (mm/yr)
260
240
220
200
180
160
140
120
100
80
60
40
20
0
S
N
0
100 m
group 3
group 2
group 1
P11
P1
P5
P6
Reference $R_{d,l}$
$R_{d,l}$ = 45 mm/yr
Upslope
Away from the stream
c
MRS water content (%)
0.3
0.5
0.7
0.9
1.1
1.3
1.5
1.7
1.9
2.1
2.3
790
770
No-flow boundary
PMR max. depth

The groundwater mound south of the stream was analysed. The Kresic equation was applied in superposition to the pairing of wells P7-P13 in April-May 2005 over a long period (N = 35 days). The model was matched to measurements using a trial/error process. The best fit for the modelled head changed at P13 (x = 153 m) over 3 days leading to hydraulic diffusivity of $D = 0.018\ m^2/s$. The relatively good match of the model with the measurements at P7 using reasonable hydrodynamic parameters, consistent with pumping tests (Maréchal et al., 2009), shows that water table fluctuations at P13 can be explained by the groundwater mound propagation from P7 through the aquifer. This is consistent with the intermediary chemistry of the aquifer at P13.

One way to assess the water seepage from the stream is to consider that this quantity of water is flowing in the aquifer laterally towards the South. The flow in the aquifer can be obtained applying Darcy's law between wells P7 and P13:

$$q(t) = T\,\frac{h_{P7}(t) - h_{P13}(t)}{x} \tag{2}$$

with $q(t)$ the linear flow perpendicular to the stream (m^2/s) and $\frac{h_{P7}(t) - h_{P13}(t)}{x}$ the hydraulic gradient (-) between P7 and P13.

The average transmissivity $T = 1.6 \times 10^{-4}\ m^2/s$ was obtained by pumping tests in wells P7 and P13; this value is quite typical for the weathered-fissured part of a hard-rock aquifer (Maréchal et al., 2004). The hydraulic gradient and the flow rate between P7 and P13 were calculated hourly during 2006, applying equation (2). The total amount of linear flow is about 3 l/s/m, which corresponds to about 15 mm/year at the watershed scale, for a reference surface area of 4.1 km^2 and for a 1,000 m-long infiltrating stream.

The low-lying area, occupying the lower part of the slope and the flat valley bottoms are mainly covered by black soils to which 2:1 clays (smectite - montmorillonite) provide vertic properties (swelling clay) and low permeability (Barbiero et al., 2007). Therefore, water losses will take place where the stream flows through red soils or on the bedrock itself. Considering about 2,000 m of stream length on that kind of permeable material, the total rate of indirect recharge at the watershed scale is about 30 mm/yr. This amount is small but it must be related to the small number of flowing days in this ephemeral stream.

The low total recharge rate induces: (i) a lowering of the water table and (ii) a water table which is recharge-controlled and not topography-controlled (Haitjema and Mitchell-Bruker, 2005). As a result, the water table is most probably not a subdued replica of the local topography and the groundwater basin boundaries do not coincide with the gauged surface watershed as confirmed by groundwater outflow below its outlet.

Water Budget and Forest Evapotranspiration

Actual evapotranspiration (ET) is a major component of the water cycle which is very difficult to measure, especially in forested watershed. Where other components of the water cycle are known, actual evapotranspiration can be estimated as the remainder of the catchment water balance (Bosch and Hewlett, 1982):

$$ET = P - R_{d,l} - R_i - Q - \Delta S \tag{3}$$

where ET is evapotranspiration, P is precipitation, Q is surface runoff measured as stream flow, $R_{d,l}$ is direct and localised recharge to groundwater, R_i is indirect recharge and ΔS is the change in soil water storage.

Usually, this technique for evapotranspiration estimate is confronted with a lack of information on the recharge component. Using the recharges estimated here above and neglecting the yearly soil water storage fluctuations at the watershed scale, equation (3) is used to estimate the annual actual evapotranspiration in the Mule Hole watershed (Table 2).

The annual evapotranspiration is directly related to the available water on the watershed and therefore limited by rainfall. The average ET on years characterised by a normal rainfall (2004 and 2006) is about 1,050mm/yr. The comparison of evapotranspiration with other evergreen forests from the same region (Nilgiris, (Sharda et al., 1998)) and with the forested worldwide curve (Zhang et al., 2001) shows good compliance (Figure 8), confirming the role of vegetation on the water budget. In this forested watershed, evapotranspiration represents about 87% of rainfall: this very high actual evapotranspiration, close to the reference evapotranspiration for alfalfa (ET_{ref} = 1205 mm/yr), was unexpected in such a deciduous forest. It would require high crop coefficient throughout the year: despite the absence of rainfall during a long six-month period, the great depth and extent of the roots of the trees (Anogeissus latifolia, Terminalia crenulata and Tectona grandis with deep roots observed in observation boreholes of Mule Hole) would contribute to an increase in transpiration by forest in deep soils during dry season (Calder, 1990; Nepstad et al., 1994).

Year	P (mm/yr)	$R_{d,l}$ (mm/yr)	R_i (mm/yr)	Q (mm/yr)	ET (mm/yr)
2004	1216 ± 73	45± 9	30 ± 15	66 ± 13	1075 ± 110
2005	1434 ± 86			196 ± 39	1163 ± 150
2006	1170 ± 70			52 ± 10	1043 ± 105

Table. 2.2-2 - Values of the components of the water balance of the Mule Hole watershed from 2004 to 2006, *ET* is calculated using equation (3). The year 2003 is not considered because rainfall and runoff time series are not complete. *ET* during 2005 is most probably overestimated as soil water storage fluctuation could have been non negligible during this rainy year.

The error obtained on evapotranspiration is about 10% (Table 2), which is acceptable for a yearly water budget. The errors on recharge and evapotranspiration are mainly due to analytical errors on chloride content in waters (rainfall and groundwater). Additional uncertainties due to variable chloride content in rainfall would require a modelling approach to be quantified. Extensive monitoring of rainfall chloride content is necessary in order to minimize these uncertainties.

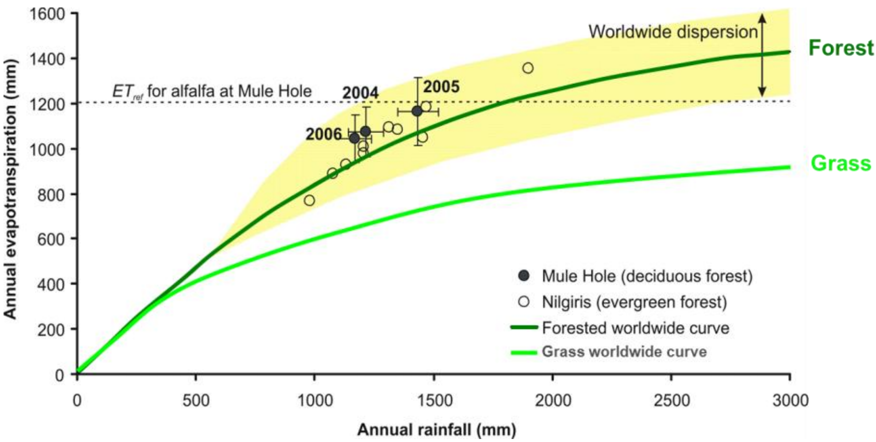


Fig. 2.2-8 - Relationship between annual rainfall and evapotranspiration at Mule Hole watershed compared to Nilgiris watersheds and worldwide curves from (Zhang et al., 2001). The yellow area represents the dispersion around the worldwide curve.

Inversely, the water yield constitutes the amount of water available in the watershed after the process of evapotranspiration from the forest. The relationship between annual rainfall and water yield is represented in Figure 9. If we consider the average worldwide curve, in the range of rainfall rates observed in Mule Hole area (1,000 to 1,500 mm/yr), the theoretical decrease of water availability due to the forest cover (difference between worldwide grass and forest curves) ranges from 210 to 340 mm/yr. In other words, in such a forested watershed, the increase of the water yield linked to an increase of rainfall of 500 mm/yr (for instance from 1,000 to 1,500 mm/yr) would only be about 250 mm/yr, whereas it would be about 400 mm/yr for a grassy watershed (the difference being 150 mm/yr, i.e. 30% of the additional rainfall in this range of rainfall).

Chemical Balance

Some chemical balance studies consider that the subsurface flow is negligible and that water flows out the basin mainly as surface runoff. Therefore, it is assumed that the measurements of the discharge rate and elements concentration in surface runoff at the watershed outlet weir are sufficient to estimate the total output fluxes of elements. The output flux due to inter-basin groundwater flow corresponds to water loss from the watershed via groundwater flow below the weir and is therefore not measured at the outlet. This component of the water balance is usually poorly defined (Nichols and Verry, 2001) or even “frequently ignored altogether in catchment studies in the quiet hope that it is, in fact, zero” (Penman, 1963).

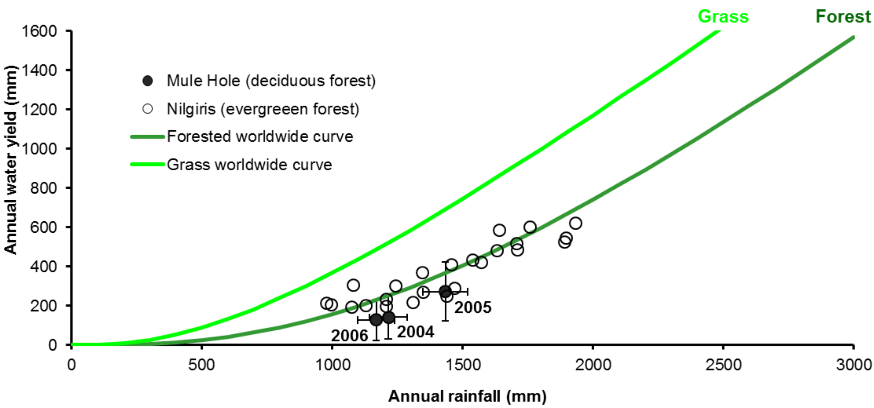


Fig. 2.2-9 - Relationship between annual rainfall and water yield at Mule Hole watershed compared to Nilgiris watersheds and worldwide curves

As stated above for Mule Hole, because of the low recharge rate and the induced lowering of the water table, the latter is not topography-controlled and results in groundwater flow below the weir. Moreover, groundwater flow is constituted by a hierarchical pattern of flow systems into the shallow "active zone" of a crystalline aquifer (Maréchal, 2010) that could be local, intermediate, and regional according to Toth theory (Toth, 1963).

The study of the mass balance of a small drainage watershed located at the head of the system (Figure 10) requires us to consider an extra flux corresponding to the output flux by groundwater flow $Q_{i,gw}$. Even if the flow velocities in local flow systems are often much faster than the flow velocity in a regional system, the flux of elements, which is highly dependent on the concentration in groundwater, may be non-negligible (depending on the considered chemicals).

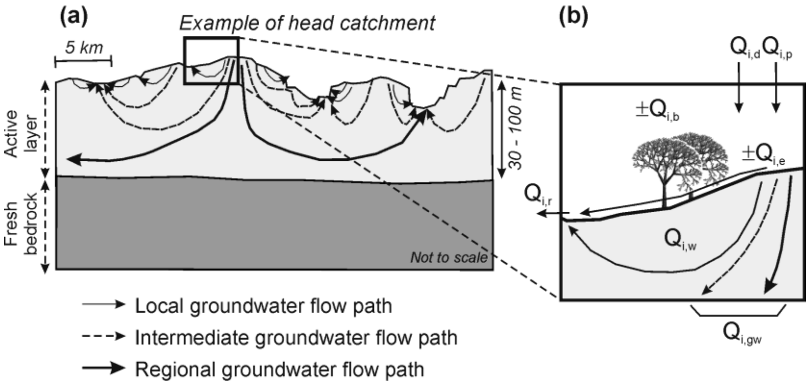


Fig. 2.2-10 - (a) Crystalline aquifer system showing flow paths associated with local, intermediate, and regional flow systems (modified from Toth, 1963). (b) Major fluxes of element i; symbols refer to the mass balance equation.

In such a case, the mass balance used to determine the chemical weathering fluxes for a species i must be corrected to:

$$Q_{i,w} + Q_{i,p} + Q_{i,d} = Q_{i,r} + Q_{i,gw} \mu Q_{i,b} \mu Q_{i,e} \quad (4)$$

where $Q_{i,r}$ = runoff flux

$Q_{i,w}$ = net chemical weathering flux from bedrock, saprolite and soil layers

$Q_{i,p}$ = input flux due to atmospheric wet deposition
$Q_{i,d}$ = input flux due to atmospheric dry deposition
$Q_{i,b}$ = source/sink flux due to biomass activity
$Q_{i,e}$ = source/sink flux due to exchange by sorption/desorption in weathering profiles
$Q_{i,gw}$ = output flux due to inter-basin groundwater flow

At Mule Hole watershed, given that the water table is disconnected and there is no long-term water level fluctuation, we assume that the groundwater baseflow is equal to natural recharge. Then, the chemical groundwater output can be determined at each observation well by multiplying element concentrations by recharge rate from Table 1. We compared these values to surface water chemical output calculated while multiplying discharge rate by element concentrations at the outlet (Table 3).

The results show the high variability of chemical groundwater outputs according to well location on the various groundwater flow paths (groups 1 to 3). As an average (Maréchal et al., 2011), the chemical groundwater output is much higher than the chemical output through surface runoff at the outlet: export through groundwater accounts for more than 95% of sodium, 93% of calcium, 95% of magnesium, and 84% of silica. Potassium is the only tracer for which the surface export is significant, approximately 35 % of the total output.

	Group	**Well**	**Chemical output $Q_{i,gw}$ (mol/hectare/year)**				
			Na	**K**	**Ca**	**Mg**	**SiO2**
Groundwater	Group 1	P5	582	100	726	482	351
	Group 2	P8	942	82	1716	843	317
	Group 3	P1	613	152	1195	913	1364
Surface water	-	Outlet	30	79	74	52	109

Table. 2.2-3 - mass balance of the Mule Hole watershed

On this basis, we can estimate that most of the output fluxes of chemicals are due to groundwater, but that these fluxes are strongly site dependant, which must be taken into account when estimating the global chemicals fluxes from the watershed.

Conclusion

A multidisciplinary approach consisting of chloride mass balance technique coupled with water table fluctuations study and groundwater flow analytical modelling leads to a rigorous estimate of direct, localised ($R_{d,l}$ = 45 mm/yr) and indirect recharge (R_i = 30 mm/yr) rates to groundwater in a forested watershed. Complementary geophysical measurements and hydraulic tests respectively confirm and explain the dissymmetry of recharge around the stream axis. The low values of recharge and runoff rates imply a very high evapotranspiration rate (about 87% of annual rainfall) from the deciduous forest cover in such a humid watershed, where evapotranspiration is controlled by both water availability and atmospheric demand.

The water yield of the watershed is low with an average of 180 mm/yr (105 mm/yr as runoff and 75 mm/yr as recharge) and its increase with the rainfall increase is reduced by the water demand from the forest. The impact of the forest on water yield (based on empirical worldwide curves) ranges between 210 and 340 mm/yr according to the rainfall rate.

The forest can impact the rate of recharge of percolating water to the water table. When deep tree roots are present, they can abstract water from the unsaturated zone during the transpiration process. In the case of Mule Hole, the resulting low recharge rate leads to a low water table which is disconnected from the surface stream. Therefore, the stream is highly ephemeral. While flowing, the stream infiltrates to the ground and indirect recharge takes place several dozen days per year. In the absence of groundwater flow to the stream, the total recharge to groundwater flows as outflow under the outlet of the watershed. This implies that vegetation abstraction has modified the groundwater basin, which is now different from the surface watershed. It results in groundwater flow below the weir which contributes for more than 90% of the main chemical outputs, and for less than 10% through surface runoff. As a consequence, the forest can impact both the hydrological cycle and the chemical balance of the watershed.

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