1 **Combined estimation of specific yield and natural recharge in a semi-arid**
2 **groundwater basin with irrigated agriculture**

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15
16 **Abstract**

17 A water budget approach is developed to jointly estimate specific yield and natural
18 recharge in an unconfined aquifer with significant seasonal water table fluctuations. Water
19 table fluctuations are due to distinct seasonality in groundwater recharge. The separation of
20 the hydrologic year into two (or more) extended seasons of recharge (wet season) and no-
21 recharge (dry season) with accompanying changes in water table allows for a split use of
22 the water table fluctuation (WTF) method, first to estimate specific yield from the water
23 table drop during the dry season (no recharge) and, second, to estimate recharge from the

water table rise during the wet season, after considering all other water budget components explicitly. The latter includes explicit computation of groundwater storage with the WTF method. The application of the WTF method requires a large number of water level measurements throughout the unconfined aquifer before and after each season. The advantage of the method is that specific yield and recharge are estimated at the scale of interest to basin hydrologic studies and that the method requires no extensive in situ instrumentation network. Here, the method is demonstrated through a case study in a fractured hard-rock aquifer subject to intensive groundwater pumping for irrigation purposes.

1. Introduction

Quantification of the rate of ground water recharge is a basic prerequisite for efficient ground water resource management (Sophocleous, 1991). This constitutes a major issue in regions with large demands for ground water supplies, such as in semiarid areas, where such resources are the key to agricultural development. However, the rate of aquifer recharge is one of the most difficult components to measure when evaluating ground water resources (Sophocleous, 1991). Its determination in arid and semiarid areas is neither straightforward nor easy. This is a consequence of the time variability of precipitation in arid and semiarid climates, and spatial variability in soil characteristics, topography, vegetation and land use (Lerner et al., 1990). Moreover, recharge amounts are usually small in comparison with the resolution of investigation methods. The more arid the climate, the smaller and potentially more variable is the recharge flux (Allison et al., 1994).

According to Sophocleous (1991), the main techniques used to estimate ground water recharge rates can be divided into physical methods and chemical methods (Allison,

1988; Foster, 1988). Among the physical methods, the water table fluctuation technique (WTF) links the change in ground water storage with resulting water table fluctuations through the storage parameter (specific yield in unconfined aquifer). This method is considered to be one of the most promising and attractive due to its accuracy, ease of use and low cost of application in semiarid areas (Beekman and Xu, 2003). The WTF method was first used to estimate ground water recharge and has since been used in numerous studies for the same purpose (Leduc et al., 1997; Moon et al., 2004) or groundwater storage changes estimation (Ruud et al., 2004).

The main limitations of the WTF technique are (1) the need to know the specific yield of the saturated aquifer at a suitable scale and (2) the fact that its accuracy depends on both the knowledge and representativeness of water table fluctuations (Beekman and Xu, 2003). In order to determine the specific yield at a suitable scale, and consequently the recharge, a double water table fluctuation method (DWTF) that is a combination of the ground water budget and water table fluctuation procedures, is employed. It is illustrated by its application to a case study in an overexploited hard-rock aquifer in India where numerous observation wells enable an accurate knowledge of water table fluctuations in such a heterogeneous environment. Special attention has been paid, in this paper, to accurately estimate all the components of the ground water budget.

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66 **2. Study area**

67 The Maheshwaram pilot watershed (Figure 1a), 53 km² in area, is located 35 km
68 south of Hyderabad (Andhra Pradesh State, India). The area is characterized by a relatively
69 flat topography 590 to 670 m above sea level and the absence of perennial streams. The

70 region has a semiarid climate controlled by the periodicity of the Monsoon (rainy or
71 "Kharif" season from June to October). Mean annual precipitation (P) is about 750 mm, of
72 which more than 90 % falls during the Monsoon season. The mean annual temperature is
73 about 26 °C, although in summer ("Rabi" season from March to May) the maximum
74 temperature can reach 45 °C. The resulting potential evaporation from soil plus
75 transpiration by plants (PET) is 1,800 mm/year. Therefore, the aridity index ($AI = P/PET =$
76 0.42) is $0.2 < AI < 0.5$, typical of semiarid areas (UNEP, 1992). Surface streams are dry
77 most of the time, except a few days a year after very heavy rainfalls during the monsoon.
78 The geology of the watershed is relatively homogeneous and mainly composed of the
79 Archean granite commonly found in the region characterized by remains of ancient and
80 more recent weathering profiles. Recent results (Dewandel et al., 2006) describe a typical
81 weathering profile (Figure 1b) comprised of the following layers having specific
82 hydrodynamic properties. From top to bottom:

83 - Saprolite (or alterite or regolith), a clay-rich material, derived from prolonged *in*
84 *situ* decomposition of bedrock, a few tens of meters thick (where the layer is not eroded).
85 Because of its clayey-sandy composition, the saprolite layer has a high porosity, and a low
86 permeability. When it is saturated, this layer constitutes part of the storage capacity of the
87 aquifer.

88 - A fissured layer, generally characterized by dense horizontal fissuring (Maréchal
89 et al., 2003) in the first few meters and a depth-decreasing density of subhorizontal and
90 subvertical fissures (Maréchal et al., 2004). This layer mainly assumes the transmissive
91 function of the aquifer and is tapped by most of the wells drilled in hard-rock areas.

92 - The fresh basement is permeable only locally, where tectonic fractures are present.

93

94 **Figure 1. (a) Maheshwaram watershed (53 km²), location of farmers pumping borewells in July 2002**
 95 **(MS: meteorological station; IFP7: observation well whose hydrograph is used on Figure 3); (b)**
 96 **weathering profile of the hard-rock aquifer with mean altitude of layer limits**

97 The Maheshwaram watershed is a representative Southern India catchment in terms
 98 of overexploitation of its hard-rock aquifer (more than 700 borewells in use), its cropping
 99 pattern (rice fields dominating), rural socio-economy (based mainly on traditional
 100 agriculture) and agricultural practices. Ground water resources face a chronic depletion
 101 that is observable by the drying-up of springs and streams and a declining water table.
 102 Water table is now 15 to 25 m deep and is disconnected from surface water : no spring, no
 103 baseflow, no regular infiltration from surface streams beds is observed.

104

105 **3. Methodology**

106 **3.1. Principle.**

107

108 The employed methodology is based on applying the water table fluctuation (WTF)
 109 method in conjunction with the groundwater basin water budget method. The water budget
 110 method focuses on the various components contributing to groundwater flow and
 111 groundwater storage changes (Figure 2). Changes in ground water storage can be attributed
 112 to recharge, irrigation return flow and ground water inflow to the basin minus baseflow
 113 (ground water discharge to streams or springs), evapotranspiration from ground water,
 114 pumping, and ground water outflow from the basin according to the following equation
 115 adapted from Schicht and Walton (1961):

$$116 \quad R + RF + Q_{on} = ET + PG + Q_{off} + Q_{bf} + \Delta S \quad (1)$$

where R is total ground water recharge (sum of direct recharge R_d through unsaturated zone and indirect and localized recharge R_{il} respectively from surface bodies and through local pathways like fractures, this point is discussed in details at § 4.2.), RF is irrigation return flow, Q_{on} and Q_{off} are ground water flow onto and off the basin, ET is evaporation from water table, PG is the abstraction of ground water by pumping, Q_{bf} is baseflow (ground water discharge to streams or springs) and ΔS is change in ground water storage.

Figure 2: schematisation of flow components of the groundwater budget in a depleted unconfined aquifer (modified after Maréchal et al., 2004)

Due to the significant thickness of the unsaturated zone overlying the unconfined aquifer in the Maheshwaram basin - on average more than 17 m - the following simplifications can be made to the water budget:

- Groundwater discharge to surface water, Q_{bf} , via stream discharge or springs does not exist ($Q_{bf} = 0$). All groundwater discharge is via groundwater pumping.
- Transpiration from the water table is negligible due to large depth to groundwater higher than the depth of trees roots evaluated to maximum 10 m in this area from borewells and dugwells observation. Therefore, this flow can be neglected and the evaporation (E) from the water table has been estimated according to the water table depth using the relation proposed by Coudrain-Ribstein et al. (1998) for semi-arid areas,

Equation (1) can be rewritten:

$$R + RF + Q_{on} = PG + E + Q_{off} + \Delta S \quad (2)$$

The main advantage of the ground water budget method compared to the classical hydrologic budget is that evapotranspiration from the root zone of soils - already included in the natural recharge - which usually constitutes a major component with large associated uncertainties is not present in Eq (2).

The methodology used to determine the unknown ground water storage is the Water Table Fluctuations method (WTF), which links the change in ground water storage ΔS with resulting water table fluctuations Δh :

$$\Delta S = S_y \cdot \Delta h \quad (3)$$

where S_y is the specific yield (storage) or the fillable porosity of the unconfined aquifer.

Because the water level measured in an observation well is representative of an area of at least several tens of square meters, the WTF method can be viewed as an integrated approach and less a point measurement than methods based on very local data in the unsaturated zone for example. Techniques based on ground water levels are among the most widely applied methods for estimating recharge rates (Healy and Cook, 2002). This is likely due to the abundance of available ground water-level data and the simplicity of estimating recharge rates from temporal fluctuations or spatial patterns of ground water levels.

The WTF method, applicable only to unconfined aquifers, is best applied to shallow water tables that display sharp water-level rises and declines. Deep aquifers may not display sharp rises because wetting fronts tend to disperse over long distances (Healy and Cook, 2002). In the study area, the monitoring of water table between 2000 and 2003 using ten automatic water level recorders shows that the aquifer displays well-identified large seasonal water-level fluctuations due to percolation of water during monsoon period

through a rather thick unsaturated zone and small daily fluctuations due to pumping cycles (Figure 3). The Kharif season, during which the water table level rises several meters due to rainfall recharge, is followed by the Rabi season during which the water level drops due mainly to ground water pumping (Figure 3). Therefore, the hydrological year can be divided into two distinct seasons each with a distinct water level rise or decline. To each of these seasons, the WTF method can be applied separately.

Figure 3. Well hydrograph observed (IFP7; Fig.1a) in the study area with seasonal water table fluctuations. The rise of water table during the Kharif season is general on the whole basin at the same time (a small delay of several days is observable according to wells local context).

Combining the water budget equation (2) with the WTF method expressed in (3), we obtain:

$$R + RF + Q_{on} = PG + E + Q_{off} + S_y \Delta h \quad (4)$$

As is typical for semi-arid basins with irrigated agriculture, two terms that cannot be evaluated independently without extensive in situ instrumentation are the basin-average natural recharge rate, R and the basin-average, effective specific yield, S_y . By applying (4) separately to the dry season, during which $R = 0$, and to the wet season, we obtain two equations with two unknown parameters:

$$RF^{dry} + Q_{on}^{dry} = PG^{dry} + E^{dry} + Q_{off}^{dry} + S_y \Delta h^{dry} \quad (5)$$

$$R + RF^{wet} + Q_{on}^{wet} = PG^{wet} + E^{wet} + Q_{off}^{wet} + S_y \Delta h^{wet} \quad (6)$$

which can be solved sequentially, first by obtaining S_y by solving (5), then by solving (6) for R , given the season-specific values for the known parameters:

$$S_y = \frac{RF^{dry} + Q_{on}^{dry} - E^{dry} - PG^{dry} - Q_{off}^{dry}}{\Delta h^{dry}} \quad (7)$$

$$R = \Delta h^{wet} \cdot S_y - RF^{wet} - Q_{on}^{wet} + E^{wet} + PG^{wet} + Q_{off}^{wet} \quad (8)$$

Equation (7) known as the “water-budget method” for estimating S_y (Healy and Cook, 2002), was initially proposed by Walton (1970) and was afterwards used namely by Hall and Risser (1993) and Gburek and Folmar (1999). The water-budget method is the most widely used technique for estimating specific yield in fractured-rock systems, probably because it does not require any assumptions concerning flow processes (Healy and Cook, 2002).

Various authors (Sokolov and Chapman, 1974; Sophocleous, 1991) distinguish the terms “specific yield” and “fillable porosity”—specific yield being the volume of water released from a unit volume of saturated aquifer material drained by a falling water table, whereas fillable porosity is the amount of water that an unconfined aquifer can store per unit rise in water table and per unit area. Because of hysteresis, under some conditions, the fillable porosity can be smaller than the specific yield (Kayane, 1983). The difference between specific yield during water table decline and fillable porosity during water table rise is due to the presence of air trapped in pore space below the water table when it rises rapidly (Kayane, 1983). Since entrapped air disappears with time by diffusion, the fillable porosity is a function of time and increases towards the value of specific yield. Therefore, maximum water levels should be measured at least one month after the rise in order to obtain the true water table fluctuation for a storage corresponding to the specific yield value. Therefore, in the study area, measurements were done in mid-November, more than one month after the average water level peak had been reached (Figure 3). It is assumed

that this time interval is sufficient to allow entrapped air to be evacuated, especially in a pumped aquifer where induced flow increases air diffusion. Therefore, the specific yield determined using (7) can be introduced in (8).

In the following sections, it is described the methods used for obtaining the "known" parameters in equations (7) and (8), which are needed for the estimation of S_y and R . The flow components are considered to be spatially distributed throughout the groundwater basin on a 200 x 200 meter cells-length grid with measurements taken from June 2002 until June 2004 (Figures 5a – 5d). A Geographical Information System is used to compute all parameters in (7) and (8) cell by cell which are then aggregated at the groundwater basin scale. Q_{on} and Q_{off} are reliably determined only at the larger basin scale through the basin boundaries, hence S_y and R can only be computed at groundwater basin scale.

3.2. Water table fluctuation (Δh)

The WTF method requires a very good knowledge of the piezometric level throughout the entire basin. This could be achieved owing to a very dense observation network (99 to 155 wells, Table 1) provided mainly by defunct or abandoned agricultural borewells. Sophocleous (1991) pointed out that the WTF method can be misleading if the water-level fluctuations are confused with those resulting from pumping, barometric, or other causes. Continuous (15 minutes of recording time interval) monitoring of the water table using ten automatic water level recorders has shown, however, that barometric and earth tides do not affect this unconfined aquifer and care was taken to avoid any interference from pumping wells. No measurements were done in pumped wells and the rare cases of observed drawdown in the monitored wells due to interference by nearby pumping wells are never

more than 10-20 cm, which is little compared to water table fluctuations at the seasonal scale (several meters). At the same time, the continuous monitoring of the water table contributes to determine the relevant time for piezometric campaigns. Standard deviation of the error on the water table fluctuation measurement has been calculated by geostatistics (Table 1). Admitting a Gaussian statistical distribution of errors, it defines the 66% confidence interval of the error. The relative error on water table fluctuation logically decreases with the increasing number of measurements (Table 1). Water table elevations are computed by difference between ground elevation from a Digital Elevation Model obtained by a couple of satellite images stereoscopy treatment (grid resolution : 30 m ; accuracy: 1 meter) and water depth obtained from piezometric measurements. The water table maps were then interpolated using the kriging technique. The map was then critically evaluated. The automatic interpolation technique gave satisfactory results owing to the very dense observation network and to the fact that there is no surface water capable of locally modifying the water table. The map for June 2002 (Figure 4) shows that the water table roughly follows the topographic slope, as is usually observed in flat hard-rock areas. However, local water table depletion is observed in highly pumped areas where natural flow paths are modified by ground water abstraction.

Table 1: Number of piezometric observations, mean water table elevation for pre-Monsoon (June) and post-Monsoon (November) periods from 2002 to 2004 (mean value of the kriged grid), water table fluctuations with absolute and relative errors.

Water table levels are fluctuating between 610 and 619 m, which indicates that the water table is always in the fissured aquifer layer (Figure 1b).

Figure 4: water table map in June 2002

Figure 5: spatially distributed flow component maps; (a) volume (m^3/year) pumped from the aquifer during Rabi 2003 (Nov 02-Jun 03); (b) irrigation return flow (m^3/year) during Rabi 2003 (Nov 02-Jun 03); (c) horizontal flows (mm/year) across the limits of the watershed during Rabi 2003 (Nov 02-Jun 03); (d) annual ground water balance expressed as water table fluctuation (m/year) between June 2002 and June 2003

3.3. Pumping flow

Paddy fields (rice) and fields of vegetables (tomatoes, brinjals, ladies' fingers (okra), chilies, etc.), flowers and fruits (mangoes, goya and grapes) are irrigated with ground water due to the absence of perennial surface water, the low cost of drilling and free electricity for farmers (according to implemented regulation policies), the possibility of getting water near the crops, etc. These crops are irrigated throughout the year, even during the monsoon season, albeit at a lower rates.

The annual pumping rate was estimated using two methods: an inventory of borewells and a land use map using remote sensing technique.

A database of the borewells existing in the watershed from June 2002 to September 2002 was created. Nine hundred and twenty-nine wells were located using portable GPS and the discharge rate of the 707 in use was measured (rates between 5 L/min and 700 L/min with an average of 130 L/min). Information about daily duration of pumping, annual

number of pumping days and use (rice, vegetables, flowers, fruit, grapes, domestic, chicken factories) was gathered in order to estimate the annual abstracted volume.

The daily duration of pumping depends mainly on electrical power availability and automatic water level recorders installed in five observation wells enable daily observation of pumping phases. Observations (6.5, 7.1, 7.4, 6.7, and 6.6 hours of pumping per day) are consistent with information collected from the farmers. Computation of monthly pumping rates at the watershed scale (Table 2) is based on the average daily pumping duration in five observation borewells and on the discharge rates of the 707 borewells in use.

Table 2. Ground water abstraction according to use. Value in mm per season (and absolute error) at the basin scale, from June 2002 to June 2004.

During the studied period (June 02 to June 04), the mean total annual ground water abstraction estimated using the well inventory is about 8.8 million m³ (or 165 mm). This value is in accordance with those evaluated in 1999 using techniques based on census data, agricultural uses of water (9.1 million m³) and electrical power consumption (9.0 million m³) (Engerrand, 2002). Most of the abstracted ground water is used for paddy fields (87 %), whereas domestic consumption, estimated using inventory wells, represents only about 2 % in this rural area. Geographically, pumping is concentrated in lower elevation zones, on flat areas allowing agriculture and close to the villages (Figure 5a).

A land use map has been made from a infra-red satellite image (image-resolution: 20 x 20 m) acquired in January 2002 during the Rabi season 2002. Since paddy fields consume, by far, most of the ground water abstracted in the area, special attention was paid to accurately evaluating their surface area. A total area of about 209 ha was found for this period. In order to convert the total paddy field area into ground water abstraction, it was

necessary to estimate the mean daily pumped water need per square meter of paddy field during the same period. Therefore 11 paddy fields were surveyed in order to measure the water requirements during this period and during the Kharif season 2003. For both periods, relatively good linear relationships were found between the irrigated paddy surface and the daily pumped water. This means that farmers size their paddy fields according to their borewell yields. Requirements differ with seasons: during the Rabi season, about 15 mm of pumped water is required daily for the field while only about 10 mm is needed during the Kharif season because of the additional contribution of monsoon rainfall. Given the moderate decrease of ground water abstraction from Rabi to Kharif periods (Table 2), the contribution of rainfall allows farmers to extend the size of their paddy fields during the Kharif season.

With a 15mm/day water requirement during the Rabi 2003, 209 ha of paddy fields required about 4.2 million m³ (80 mm), confirming the value of about 4.4 million m³ (83 mm, Table 2) estimated using the well inventory. This means that the relative error on groundwater abstraction for rice can be considered to be about 5 %.

3.4. Return flow from irrigation

Since most of the water pumped in the basin is used for irrigation, a large part of it can return to the aquifer by direct infiltration. This may lead to high irrigation return flow. In some cases, e.g. in paddy fields, more than 50 % of the pumped water returns to the aquifer (Jalota and Arora, 2002). Therefore, a water budget method has been applied in order to determine irrigation return flow from the irrigated crops at the watershed and seasonal scale, i.e. for rice, vegetable and flower fields. However, for fruit and grapes, irrigation return flow was not calculated since these crops use drip irrigation techniques that

eliminate irrigation losses. No return flow was thus assumed. The principle of the method is here briefly described.

The model is based on the daily variations of water stock present in the field. The water balance is (Chen et al., 2002):

$$PG + P = ETR + RF + D + dw \quad (9)$$

where PG is the pumping flow, P the rainfall, ETR the evapotranspiration of irrigated crops, RF the irrigation return flow, D the overflow (runoff) and dw the change in ponded water depth or water storage in the soil profile; all in mm/day. Lateral seepages across the field edges are assumed to be nil.

Runoff (D) was assumed to occur when surface storage exceeds a water depth that corresponds to the mean field edge.

Irrigation return flows (q) are computed using the Darcy-Buckingham equations (Buckingham, 1907) for one-dimensional flow that consider the flow theory in non-saturated and saturated media:

$$q = -K(\theta) \cdot \left(\frac{dh}{dz} - 1 \right) \text{ for unsaturated profile} \quad (10a)$$

or

$$q = -K_s \cdot \left(\frac{dh}{dz} - 1 \right) \text{ for saturated profile} \quad (10b)$$

Where z [m] is the depth, h [m] the pressure head, K_s [m/s] the soil hydraulic conductivity at saturation and $K(\theta)$ [m/s] the unsaturated hydraulic conductivity of the soil.

Water-retention, h - θ , and the k - θ curves for the different soil types are estimated using the power law models of Brooks and Corey (1964).

347 The pressure head, h , is further a function of moisture content (θ):

$$348 \quad \frac{\theta}{\theta_s} = \Theta = \left(\frac{h_{bc}}{h} \right)^\lambda \quad (11)$$

349 where Θ [-] is the saturation index, θ [m^3/m^3] the moisture content, θ_s [m^3/m^3] the moisture
 350 content at saturation, h_{bc} [m] is the air entry suction, and λ [-] a texture-dependent
 351 dimensionless soil parameter that depends on the pore-size distribution.

352 The unsaturated hydraulic conductivity is a function of saturation index:

$$353 \quad \frac{K(\theta)}{K_s} = \Theta^\eta \quad (12)$$

354 where η [-] is the pore-disconnect-edness index, a dimensionless parameter function of λ
 355 and a parameter function of the soil tortuosity, τ .

$$356 \quad \eta = \frac{2}{\lambda} + 2 + \tau \quad (13)$$

357 Values of τ depend on the chosen capillary model, in this case, the Burdine model ($\tau=1$,

$$358 \quad \eta = \frac{2}{\lambda} + 3).$$

359 Calculation of θ is done at daily time-step using the continuity equation:

$$360 \quad \frac{\partial \theta}{\partial t} = \frac{\partial q}{\partial z} \quad (14)$$

361 For saturated profile the left-hand side of the above equation is zero. For unsaturated layers,
 362 the rate of change of θ is calculated from a linearized form of this equation. After each
 363 time-step, the new θ is calculated by subtracting the outflow from the inflow during that
 364 time-step, dividing the difference by layer thickness, integrating the resulting rate of change
 365 over time-step, and adding the change to the previous θ value. For the next time-step, the

pressure head h corresponding to the new moisture content is assessed again, and the whole procedure is repeated.

The hydraulic properties of the different soil types (e.g. K_s , θ_s and h_{bc}) have been assessed by field measurements (De Condappa, 2005). As an average, for rice soils are sandy clay loam with K_s : 2.5×10^{-7} m/s, θ_s : $0.40 \text{ m}^3/\text{m}^3$, h_{bc} : 0.14 m and λ : 0.148; for the other crops soils are sandy loam soil with K_s : 4.2×10^{-6} m/s, θ_s : $0.37 \text{ m}^3/\text{m}^3$, h_{bc} : 0.03 m and λ : 0.09.

All calculations are done at a daily time step.

Computation of daily PG at the watershed scale is based on the daily duration of pumping (see section 3.3) and on the seasonal water requirements of the field assessed during a field survey (see section 3.3 for rice, 7.7 mm/d for the vegetables and 4.9 mm/d for the flowers). Therefore it is assumed that for each season the mean seasonal PG does not vary significantly (e.g. for rice all Rabi seasons have a mean PG of 15 mm/d).

Daily evapotranspiration of irrigated crops (ETR) has been computed according to the FAO method (Allen et al., 1990).

The error on irrigation return flow coefficients ($C_{RF} = RF/PG$) has been evaluated according to the error introduced by PG (5%, see section 3.3) and to the variability of the soil saturated hydraulic conductivity (e.g. for rice soil: 10^{-7} to 4×10^{-7} m/s), error on C_{RF} due to other hydraulic parameters being negligible when compared to the error introduced by the uncertainty on soil saturated hydraulic conductivity.

Table 3 gives the average value of irrigation return flow coefficients for the different seasons from June 2002 to June 2004 with their absolute errors. Since climate conditions and pumping flow fluctuate, the return flow coefficient is variable with seasons. The mean value of the rice irrigation return flow coefficient is about 48 %, which is

comparable to values found by previous studies in various regions of Southeast Asia: 51 % in Northern India (Jalota and Arora, 2002) and 59 % in Taiwan (Chen et al., 2002). The estimated return flow coefficient is also consistent with the one evaluated by APGWD (1977) for paddies on granitic rocks (60 %). For vegetables and flowers, the mean C_{RF} is about 17 %, a value similar to the one proposed by CGWB (1998) (20 %). No data are available for domestic and chicken poultries, but since return flow probably exists, a value of 20 % was assumed for the coefficient.

Table 3. Seasonal irrigation return flow coefficients ($C_{RF}=RF/PG$) and absolute errors for paddy fields (rice) and vegetable and flower fields from June 2002 to June 2004. C_{RF_Total} : for all ground water abstraction, i.e.: rice, vegetables, flowers, fruit, grapes, domestic and poultries.

Therefore, a large proportion of the water pumped (~40 %, C_{RF} Total; Table 3) returns to the aquifer. The only water that does not return to the aquifer is that which evapotranspires from crops and soils. The map of spatially distributed irrigation return flow was calculated applying the estimated C_{RF} to each of the pumping rates according to their uses (Figure 5b).

3.5. Horizontal flow across the boundaries of the watershed

Flow was computed using a finite-differences model (Modflow) with hydrodynamic and geometry properties acquired on the basin, in order to obtain a spatial distribution of flows on the grid of square cells (Figure 5c and Table 4).

Table 4. Seasonal horizontal flow across the boundaries of the watershed (Q_{on} : horizontal in-flow, Q_{off} : horizontal out-flow). Maximum relative error of 100% (i.e. 1 mm) is assumed.

Low in-flow occurs mainly across the southern border of the watershed due to the regional south-north gradient linked to the regional topographical slope (Figure 4 and

Figure 5c). In-flow from the west and east is due to water table depletion near the boundaries, induced by pumping wells. The balance between horizontal in- and out-flow is close to nil. As expected, in this flat hard-rock aquifer, the regional water table being sub-parallel to regional topography, ground water flow through the boundaries of the surface watershed are negligible. Disturbance of natural flow by pumping does not significantly affect this context due to the fact that effects statistically nullify each other, in the case of regular distribution.

4. Results and discussion

4.1. Specific yield estimates

Basin-wide effective specific yields obtained from (7) were 0.014 ± 0.003 for both dry seasons (Table 5). Because these values reflect an effective basin-wide process, they are insensitive to local heterogeneities in the fractured rock aquifer system, in comparison with locally obtained values using lab samples or local aquifer testing, which are highly variable and relatively unreliable (Bardenhagen, 2000). Therefore, for water resource assessment at the watershed scale, this methodology for specific yield estimation is much more sound than the aforementioned punctual techniques. Error on specific yield ($\sim 20\%$, Table 5) has been computed cumulating all the sources of errors described above.

Table 5. Ground water budget during the Rabi seasons, estimation of specific yield and absolute errors

The specific yield obtained is realistic for fissured granite and is of the same order of magnitude as values estimated at the sub-basin scale through global modeling (one value: 0.01, Engerrand, 2002) and at the well scale using pumping data in the fissured layer itself (six values with an average of 6.3×10^{-3} , Maréchal et al., 2004). Higher values obtained with the water budget method can be explained by the fact that the upper part of

the weathering cover (saprolite with specific yield much higher than in the fissured zone, Chilton and Foster, 1995) can be partially saturated in some areas after Monsoon, which increases the global storage at the watershed scale. Heterogeneity effects can also explain this apparent increase of S_y with scale.

It is generally assumed that specific yield varies with depth—especially in hard-rock aquifers where fracture density and porosity change with depth, namely between the different layers constituting the aquifer (Maréchal et al., 2004; Dewandel et al., 2006). Water budget results in 2002 and 2003 seem to indicate that the specific yield does not vary. In fact, the water table is located mainly in the fissured layer of the aquifer (Figure 1b) and water table fluctuations are small enough so that the water table remains in the same portion of the aquifer, characterized by a constant specific yield.

4.2. Natural recharge estimates

Equation (8) was used to estimate natural recharge (Table 6). Natural recharge is determined at the watershed scale, not cell by cell like other budget components, and is therefore not spatially distributed. Relative error on natural recharge (22 - 24 %, Table 6) has been computed cumulating all the sources of errors described above.

Table 6. Ground water balance during monsoon seasons, estimation of natural recharge and absolute errors

At Table 6, the recharge is compared to precipitation during the monsoon (seasonal rainfall) between June and November. During both hydrological years of monitoring, the recharge coefficient R/p varies between 0.13 and 0.19. This is similar to recent results obtained in India under the same climate conditions for a coastal aquifer in Karnataka

(0.13-0.24, Rao et al., 2004), an alluvial aquifer in Uttar Pradesh (0.06-0.19, Kumar and Seethapathi 2002) and the value assumed by CGWB (1998) for hard-rock aquifer (0.12). Its fluctuation, year to year, depends mainly on the intensity and temporal distribution of rainfall events during the monsoon. Notice that the recharge coefficient increases with the number of rainy days during the monsoon (Table 6).

Total recharge can be divided into three main components (Lerner et al., 1990): *direct* recharge R_d (by direct vertical percolation through the vadose zone – saprolite, Figure 1b), *indirect* recharge R_i (percolation to the water table through the beds of surface-water courses, close to nil in the study area due to absence of water in surface streams) and *localized* recharge R_l (various-scales pathways such as those formed by shrinkage cracks, roots, and burrowing animals, trenches, dugwells, brick factories and caused by major landscape features, Gee and Hillel, 1988).

In the WTF method for recharge evaluation, no assumptions are made concerning the mechanisms by which water travels through the unsaturated zone. Hence, the presence of preferential flow paths (indirect or localized recharge as defined above) within the vadose zone in no way restricts its application to evaluation of total recharge. The estimated recharge flow includes all recharge types. This point is illustrated in Figure 6 where the total recharge R calculated using the WTF technique is compared to estimates of recharge using tritium injection tests on the same type of lithology (granite and gneiss) in semiarid regions of India (Rangarajan and Athavale, 2000; Sukhija et al., 1996). Tritium injection tests enable an estimation of only one part (direct recharge R_d) of the total recharge R by interpretation of artificial tracer transfer through the soils after an injection of tracer before the monsoon. Rangarajan and Athavale (2000) have shown a linear relationship between

direct recharge and seasonal rainfall in hard-rock regions of India. The regression line suggests that a certain minimum seasonal rainfall (about 250 mm) is required for initiating deep percolation and recharge to the phreatic aquifer system. As a comparison, in various lithological and morphological contexts in South Africa, Botswana and Zimbabwe, the regional recharge is very low where rainfall is less than 400 mm/year (Selaolo, 1998 cited in de Vries and Simmers, 2002). This can be considered as the minimum rainfall required for recouping the soil moisture deficit in the vadose zone (Rangarajan and Athavale, 2000). Recharge does not vary a lot for the same seasonal rainfall (Figure 6). This means that significant recharge does not result from infrequent large events and that describing mean annual recharge as a proportion of seasonal rainfall is valid in such a context. Inversely, such a statement cannot be made in a similar climatic context in South Africa, Botswana and Zimbabwe where recharge varies by a factor of up to 100 for the same seasonal rainfall (Selaolo, 1998).

Figure 6. Rainfall – recharge (R_d : direct recharge; R_{il} : indirect and localized recharge; R : total recharge) relationship 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)

Both black triangles in Figure 6 corresponding to the estimated total recharge at the Maheshwaram basin scale are higher (compared to the 95 % confidence interval of the linear regression) than the recharge expected from the linear regression. This is really significant for 2003 because the discrepancy in 2002 is almost in the range of the error. This difference could be due to the contribution of indirect and localized recharge ($R_{il} = R_i + R_l$) to the total recharge. This contribution can be estimated by subtracting direct recharge (roughly estimated using the linear relationship with the observed seasonal

rainfall) from total recharge (obtained with the WTF technique). For both years of available data, indirect and localized recharge accounts for about 30-40 % of total recharge (Table 7). The indirect recharge R_i should be small in the watershed as stated above. Consequently, most of the additional recharge probably corresponds to localized recharge at various scales ($R_{il} \approx R_l$).

Table 7. Estimation of recharge types (¹Direct recharge is estimated using the relation $R = 0.172 \times P - 44$; ² Indirect and localized recharge are estimated by difference between total and direct recharge)

4.3. Annual ground water budget

The “double water table fluctuation method” consists in aggregating dry and rainy seasons water budgets. The annual ground water balance was calculated from June 2002 to June 2004 (Table 8) and we see a respective deficit and excess of water due to discrepancies between annual rainfall and an average rainfall of about 740 mm/year (average in Maheshwaram since 1985). Considering the uncertainty on the components of the budget, this suggests that the balance should be lightly negative for an average rainfall. Historical water level data shows a global depletion of the aquifer at a rate of about one meter per year in pumped areas, confirming that the overexploitation threshold has been reached in such areas. Moreover, given the abstraction rate in the basin, any deficient monsoon (the 2002 monsoon, for example) causes a significantly negative balance followed by a drop in the water table, which can be fully or only partially replenished by the next heavy monsoon. In spite of the fact that the pumping areas represent only 25 % of the 1324 cells of the basin (Figure 4d), the entire balance is negative.

Table 8. Ground water balance during two hydrological cycles

The importance of irrigation return flow (*RF*) justifies the need for accurate techniques for its determination. Its relative importance will guide ground water sustainability solutions because any reduction in pumping triggers a corresponding reduction in ground water recharge from irrigation drainage. Regarding cropping pattern changes, choices should be guided by the same constraint: to halt water table decline beneath these ground water-irrigated areas, evapotranspiration must decrease. Therefore, sustainability (defined as stabilizing ground water levels) begins not with reducing irrigation pumping per acre, but rather with reducing the total acreage of irrigated land (Kendy, 2003) or changing the cropping pattern in order to decrease the total amount of evapotranspiration at the watershed scale.

5. Conclusions

The advantage of the proposed method is that specific yield and recharge are estimated at the scale of interest to basin hydrologic studies and that the method requires no extensive in situ instrumentation network. This methodology enables to overcome the main limitation of the classical WTF technique, i.e. unknown specific yield, by determining it at the suitable watershed scale and within an acceptable range of uncertainty according to the available observation network. Obviously, the accuracy of the technique increases with the number of measurements on the water table. Therefore, this technique is well suited to developing countries and semiarid areas, where the presence of many agricultural dugwells and borewells throughout a basin provides a high-density observation network. For economic reasons, it is important to optimize the amount of piezometric data needed to guarantee an acceptable accuracy in the application of this methodology. Therefore, a geostatistical approach combined with hydrogeological information must be used in order

to assess the impact of observation well density reduction on water budget calculations and therefore optimize the density and observation well distribution. This will be the subject of a future publication.

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676 TABLES

Date	Number of piezometric measurements	Mean elevation of water table [m]	Water table fluctuation [m]
10-21 June 02	99	613.5	$\Delta h^{wet} = 1.2 \pm 0.27$ $= 1.2 \pm 22.5 \%$
11-22 November 02	107	614.7	
2-11 June 03	114	610.3	$\Delta h^{dry} = -4.4 \pm 0.35$ $-4.4 \pm 0.08 \%$
10-21 November 03	155	618.6	$\Delta h^{wet} = 8.3 \pm 0.32$ $= 8.3 \pm 0.04 \%$
14-25 June 04	134	613.5	$\Delta h^{dry} = -5.1 \pm 0.23$ $= -5.1 \pm 0.04 \%$

677 Table 9: Number of piezometric observations, mean water table elevation for pre-Monsoon (June) and
678 post-Monsoon (November) periods from 2002 to 2004 (mean value of the kriged grid), water table
679 fluctuations with absolute and relative errors.

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681

Hydrological year	June 02 to June 03		June 03 to June 04	
Usage (area)	Kharif [mm]	Rabi [mm]	Kharif [mm]	Rabi [mm]
Rice (2.1 km ² in Rabi)	75.8 $\pm$ 3.8	83.4 $\pm$ 4.2	62.5 $\pm$ 3.1	108.7 $\pm$ 5.4
Vegetables & flowers (0.35 km ²)	1.3 $\pm$ 0.1	1.7 $\pm$ 0.1	1.2 $\pm$ 0.1	1.6 $\pm$ 0.1
Fruits & grapes (1.02 km ²)	4.1 $\pm$ 0.2	10.0 $\pm$ 0.5	3.8 $\pm$ 0.2	9.4 $\pm$ 0.5
Domestic & chicken poultries (-)	3.1 $\pm$ 0.2	4.2 $\pm$ 0.2	3.3 $\pm$ 0.2	4.0 $\pm$ 0.2

682 Table 10. Ground water abstraction according to use. Value in mm per season (and absolute error) at
683 the basin scale, from June 2002 to June 2004.

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	C_{RF} in Rice (%)		C_{RF} in Vegetables+Flowers (%)		C_{RF} Total (%)	
Period	Return flow Kharif	Return flow Rabi	Return flow Kharif	Return flow Rabi	Return flow Kharif	Return flow Rabi
June 02 - June 03	$40 \pm 3.6 \%$	$44 \pm 1.7 \%$	$15 \pm 2.1 \%$	$10 \pm 0.8 \%$	$37 \pm 3.3 \%$	$38 \pm 1.4 \%$
June 03 - June 04	$51 \pm 4.6 \%$	$48 \pm 1.8 \%$	$18 \pm 2.5 \%$	$15 \pm 1.1 \%$	$46 \pm 4.1 \%$	$42 \pm 1.6 \%$

Table 11. Seasonal irrigation return flow coefficients ($C_{RF}=RF/PG$) and absolute errors for paddy fields (rice) and vegetable and flower fields from June 2002 to June 2004. C_{RF_Total} : for all ground water abstraction, i.e.: rice, vegetables, flowers, fruit, grapes, domestic and poultries.

Hydrological year	June 02 to June 03		June 03 to –June 04	
Season	Kharif [mm]	Rabi [mm]	Kharif [mm]	Rabi [mm]
Q_{on}	1.1	1.5	0.7	1.6
Q_{off}	1.1	1.8	1.8	0.5
$Q_{on} - Q_{off}$	0.0 ± 1	-0.3 ± 1	-1.2 ± 1	1.1 ± 1

Table 12. Seasonal horizontal flow across the boundaries of the watershed (Q_{on} : horizontal in-flow, Q_{off} : horizontal out-flow). Maximum relative error of 100% (i.e. 1 mm) is assumed.

Season	Date	RF^{dry} [mm]	PG^{dry} [mm]	E^{dry} [mm]	$Q_{on}^{dry} - Q_{off}^{dry}$ [mm]	Δh^{dry} [m]	S_y [-]
Rabi 2003	Nov02-Jun03	37.9 ± 3.2	99.3 ± 5	0.6 ± 1	-0.3 ± 1	-4.4 ± 0.35	0.0140 ± 0.0029
Rabi 2004	Nov03-Jun04	53.7 ± 3	123.8 ± 6.2	1.3 ± 1	1.0 ± 1	-5.1 ± 0.23	0.0138 ± 0.0027

Table 13. Ground water budget during the Rabi seasons, estimation of specific yield and absolute errors

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Season	Date	RF^{wet} [mm]	PG^{wet} [mm]	E^{wet} [mm]	$Q_{on}^{wet} - Q_{off}^{wet}$ [mm]	Δh^{wet} [m]	R [mm]	Seasonal rainfall [mm]	Rainy days	R/P (-)
Kharif 2002	Jun02- Nov02	31.0 ± 4.6	84.2 ± 4.2	0.5 ± 1	0.0 ± 1	1.2 ± 0.27	70.5 ± 15.8	543	43	0.13 ± 0.03
Kharif 2003	Jun03- Nov03	32.6 ± 4.6	70.8 ± 3.5	1.0 ± 1	-1.2 ± 1	8.3 ± 0.32	156.5 ± 37.5	824	54	0.19 ± 0.05

698 **Table 14. Ground water balance during monsoon seasons, estimation of natural recharge and absolute**699 **errors**

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Year	Annual rainfall	Seasonal rainfall	Total recharge	Direct ¹ recharge	Indirect ² and localized recharge	Indirect and localized recharge
	[mm]	[mm]	[mm]	[mm]	[mm]	[%]
2002	613	543	70.5 ± 15.8	~49	~21	~30 %
2003	889	824	156.5 ± 37.5	~98	~59	~38 %

702 **Table 15. Estimation of recharge types (¹Direct recharge is estimated using the relation $R = 0.172 \times P -$** 703 **44; ² Indirect and localized recharge are estimated by difference between total and direct recharge)**

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Year	Annual Rainfall	R (mm/yr)	RF_{TOT} (mm/yr)	PG_{TOT} (mm/yr)	E_{TOT} (mm/yr)	$Q_{onTOT} - Q_{offTOT}$ (mm/yr)	BAL (mm/yr)	Δh_{TOT} (m)
2002- 2003	613	70.5 ± 15.8	68.9 ± 7.8	183.5 ± 9.2	1.1 ± 2	-0.3 ± 2	-45.5 ± 9	-3.2 ± 0.62
2003- 2004	889	156.5 ± 37.5	86.3 ± 7.6	194.6 ± 9.7	2.30 ± 2	-0.1 ± 2	$+45.8 \pm 8$	$+3.2 \pm 0.55$

706 **Table 16. Ground water balance during two hydrological cycles**

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Figure 1

Figure 2

Figure 3

Figure 4

Figures 5a to d

Figure 6