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▶ **To cite this version:**

Jean-Christophe Maréchal, Benoît Dewandel, Shakeel Ahmed, Laurent Galeazzi, Faisal K. Zaidi. Combined estimation of specific yield and natural recharge in a semi-arid groundwater basin with irrigated agriculture. Journal of Hydrology, 2006, 329 (1-2), pp.281-293. 10.1016/j.jhydrol.2006.02.022 . hal-00458378

HAL Id: hal-00458378
https://hal.science/hal-00458378

Submitted on 21 Feb 2010

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

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Keywords: Water balance; Recharge; Semi-arid environment, India, GIS

Abstract

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

2. Study area

The Maheshwaram pilot watershed (Figure 1a), 53 km^2 in area, is located 35 km south of Hyderabad (Andhra Pradesh State, India). The area is characterized by a relatively flat topography 590 to 670 m above sea level and the absence of perennial streams. The

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

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

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

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

Figure 1. (a) Maheshwaram watershed (53 km^2), location of farmers pumping borewells in July 2002 (MS: meteorological station; IFP7: observation well whose hydrograph is used on Figure 3); (b) weathering profile of the hard-rock aquifer with mean altitude of layer limits

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

3. Methodology

3.1. Principle.

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

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

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:

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

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:

$$\Delta S = S_y . \Delta h \qquad (3)$$

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:

184 $$S_y = \frac{RF^{dry} + Q_{on}^{dry} - E^{dry} - PG^{dry} - Q_{off}^{dry}}{\Delta h^{dry}} \quad (7)$$

185 $$R = \Delta h^{wet}.S_y - RF^{wet} - Q_{on}^{wet} + E^{wet} + PG^{wet} + Q_{off}^{wet} \quad (8)$$

186 Equation (7) known as the "water-budget method" for estimating S_y (Healy and Cook,
187 2002), was initially proposed by Walton (1970) and was afterwards used namely by Hall
188 and Risser (1993) and Gburek and Folmar (1999). The water-budget method is the most
189 widely used technique for estimating specific yield in fractured-rock systems, probably
190 because it does not require any assumptions concerning flow processes (Healy and Cook,
191 2002).

192 Various authors (Sokolov and Chapman, 1974; Sophocleous, 1991) distinguish the
193 terms "specific yield" and "fillable porosity"–specific yield being the volume of water
194 released from a unit volume of saturated aquifer material drained by a falling water table,
195 whereas fillable porosity is the amount of water that an unconfined aquifer can store per
196 unit rise in water table and per unit area. Because of hysteresis, under some conditions, the
197 fillable porosity can be smaller than the specific yield (Kayane, 1983). The difference
198 between specific yield during water table decline and fillable porosity during water table
199 rise is due to the presence of air trapped in pore space below the water table when it rises
200 rapidly (Kayane, 1983). Since entrapped air disappears with time by diffusion, the fillable
201 porosity is a function of time and increases towards the value of specific yield. Therefore,
202 maximum water levels should be measured at least one month after the rise in order to
203 obtain the true water table fluctuation for a storage corresponding to the specific yield
204 value. Therefore, in the study area, measurements were done in mid-November, more than
205 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 m^3 (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 m^3) and electrical power consumption (9.0 million m^3) (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

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

311 With a 15mm/day water requirement during the Rabi 2003, 209 ha of paddy fields
312 required about 4.2 million m^3 (80 mm), confirming the value of about 4.4 million m^3 (83
313 mm, Table 2) estimated using the well inventory. This means that the relative error on
314 groundwater abstraction for rice can be considered to be about 5 %.

315

316 **3.4. Return flow from irrigation**

317 Since most of the water pumped in the basin is used for irrigation, a large part of it
318 can return to the aquifer by direct infiltration. This may lead to high irrigation return flow.
319 In some cases, e.g. in paddy fields, more than 50 % of the pumped water returns to the
320 aquifer (Jalota and Arora, 2002). Therefore, a water budget method has been applied in
321 order to determine irrigation return flow from the irrigated crops at the watershed and
322 seasonal scale, i.e. for rice, vegetable and flower fields. However, for fruit and grapes,
323 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:

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

or

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

Where z [m] is the depth, h [m] the pressure head, Ks [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 $$\frac{\theta}{\theta s} = \Theta = \left(\frac{h_{bc}}{h} \right)^{\lambda} \quad (11)$$

349 where Θ [-] is the saturation index, θ [m^3/m^3] the moisture content, θs [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 $$\frac{K(\theta)}{Ks} = \Theta^{\eta} \quad (12)$$

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

357 Values of τ depend on the chosen capillary model, in this case, the Burdine model (τ=1,
358 $\eta = \frac{2}{\lambda} + 3$).

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

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

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

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

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

372 All calculations are done at a daily time step.

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

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

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

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

389 comparable to values found by previous studies in various regions of Southeast Asia: 51 %
390 in Northern India (Jalota and Arora, 2002) and 59 % in Taiwan (Chen et al., 2002). The
391 estimated return flow coefficient is also consistent with the one evaluated by APGWD
392 (1977) for paddies on granitic rocks (60 %). For vegetables and flowers, the mean C_{RF} is
393 about 17 %, a value similar to the one proposed by CGWB (1998) (20 %). No data are
394 available for domestic and chicken poultries, but since return flow probably exists, a value
395 of 20 % was assumed for the coefficient.

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

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

404

405 **3.5. Horizontal flow across the boundaries of the watershed**
406 Flow was computed using a finite-differences model (Modflow) with hydrodynamic
407 and geometry properties acquired on the basin, in order to obtain a spatial distribution of
408 flows on the grid of square cells (Figure 5c and Table 4).

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

411 Low in-flow occurs mainly across the southern border of the watershed due to the
412 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 (~ 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

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

442 It is generally assumed that specific yield varies with depth–especially in hard-rock
443 aquifers where fracture density and porosity change with depth, namely between the
444 different layers constituting the aquifer (Maréchal et al., 2004; Dewandel et al., 2006).
445 Water budget results in 2002 and 2003 seem to indicate that the specific yield does not
446 vary. In fact, the water table is located mainly in the fissured layer of the aquifer (Figure
447 1b) and water table fluctuations are small enough so that the water table remains in the
448 same portion of the aquifer, characterized by a constant specific yield.

449

450 **4.2. Natural recharge estimates**

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

455

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

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

462 (0.13-0.24, Rao et al., 2004), an alluvial aquifer in Uttar Pradesh (0.06-0.19, Kumar and
463 Seethapathi 2002) and the value assumed by CGWB (1998) for hard-rock aquifer (0.12).
464 Its fluctuation, year to year, depends mainly on the intensity and temporal distribution of
465 rainfall events during the monsoon. Notice that the recharge coefficient increases with the
466 number of rainy days during the monsoon (Table 6).

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

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

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

498

499 **Figure 6. Rainfall – recharge (R_d: direct recharge; R_{il}: indirect and localized recharge; R: total**
500 **recharge) relationship in granite and gneiss. Andhra-Pradesh, Southern, Central and Northern India**
501 **direct recharge estimated using tritium injection (Rangarajan and Athavale, 2000; Sukhija et al., 1996)**

502 Both black triangles in Figure 6 corresponding to the estimated total recharge at the
503 Maheshwaram basin scale are higher (compared to the 95 % confidence interval of the
504 linear regression) than the recharge expected from the linear regression. This is really
505 significant for 2003 because the discrepancy in 2002 is almost in the range of the error.
506 This difference could be due to the contribution of indirect and localized recharge
507 ($R_{il} = R_i + R_l$) to the total recharge. This contribution can be estimated by subtracting
508 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 ([1]Direct recharge is estimated using the relation R = 0.172 x 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.

Acknowledgments

This study was carried out at the Indo-French Center for Ground water Research (BRGM-NGRI). The authors wish to thank the French Ministry of Foreign Affairs and the Embassy in India for their support. The Indo-French Center for Ground water Research has also benefited from CNRS funding within the framework of the ACI Program "Water and Environment" and from the Indo-French Center for the Promotion of Advanced Research. This paper benefited from detailed comments provided by Patrick Lachassagne and research assistance provided by Géraud Bournet. The authors thank B. Bourgine who contributed to error estimation and two anonymous reviewers whose comments have contribute to improve the manuscript.

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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	
			Δh^{wet} = 1.2 ± 0.27 = 1.2 ± 22.5 %
11-22 November 02	107	614.7	
			Δh^{dry}= -4.4 ± 0.35 -4.4 ± 0.08 %
2-11 June 03	114	610.3	
			Δh^{wet} = 8.3 ± 0.32 =8.3 ± 0.04 %
10-21 November 03	155	618.6	
			Δh^{dry}= -5.1 ± 0.23 = -5.1 ± 0.04 %
14-25 June 04	134	613.5	

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.**

680
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 ± 3.8	83.4 ± 4.2	62.5 ± 3.1	108.7 ± 5.4
Vegetables & flowers (0.35 km^2)	1.3 ± 0.1	1.7 ± 0.1	1.2 ± 0.1	1.6 ± 0.1
Fruits & grapes (1.02 km^2)	4.1 ± 0.2	10.0 ± 0.5	3.8 ± 0.2	9.4 ± 0.5
Domestic & chicken poultries (-)	3.1 ± 0.2	4.2 ± 0.2	3.3 ± 0.2	4.0 ± 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.**

684

685

	C_{RF} inRice (%)		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 ± 3.6 %	44 ± 1.7 %	15 ± 2.1 %	10 ± 0.8 %	37 ± 3.3 %	38 ± 1.4 %
June 03 - June 04	51 ± 4.6 %	48 ± 1.8 %	18 ± 2.5 %	15 ± 1.1 %	46 ± 4.1 %	42 ± 1.6 %

686 **Table 11. Seasonal irrigation return flow coefficients (C_{RF}=*RF/PG*) and absolute errors for paddy fields**

687 **(rice) and vegetable and flower fields from June 2002 to June 2004. C_{RF}_Total: for all ground water**

688 **abstraction, i.e.: rice, vegetables, flowers, fruit, grapes, domestic and poultries.**

689

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

690 **Table 12. Seasonal horizontal flow across the boundaries of the watershed (Q_{on}: horizontal in-flow,**

691 **Q_{off}: horizontal out-flow). Maximum relative error of 100% (i.e. 1 mm) is assumed.**

692
693

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

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

695 **errors**

696

697

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**

700
701

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 ([1]Direct recharge is estimated using the relation R = 0.172 x P –**

703 **44; [2] Indirect and localized recharge are estimated by difference between total and direct recharge)**

704

705

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 ± 8	+3.2 ±0.55

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

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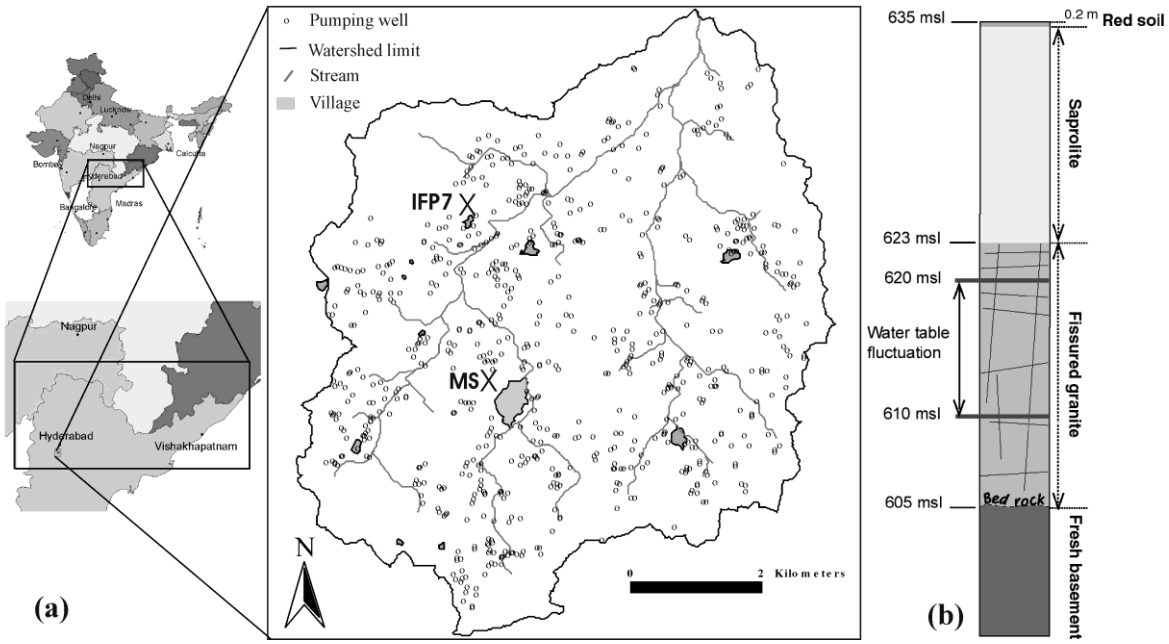


Figure 1

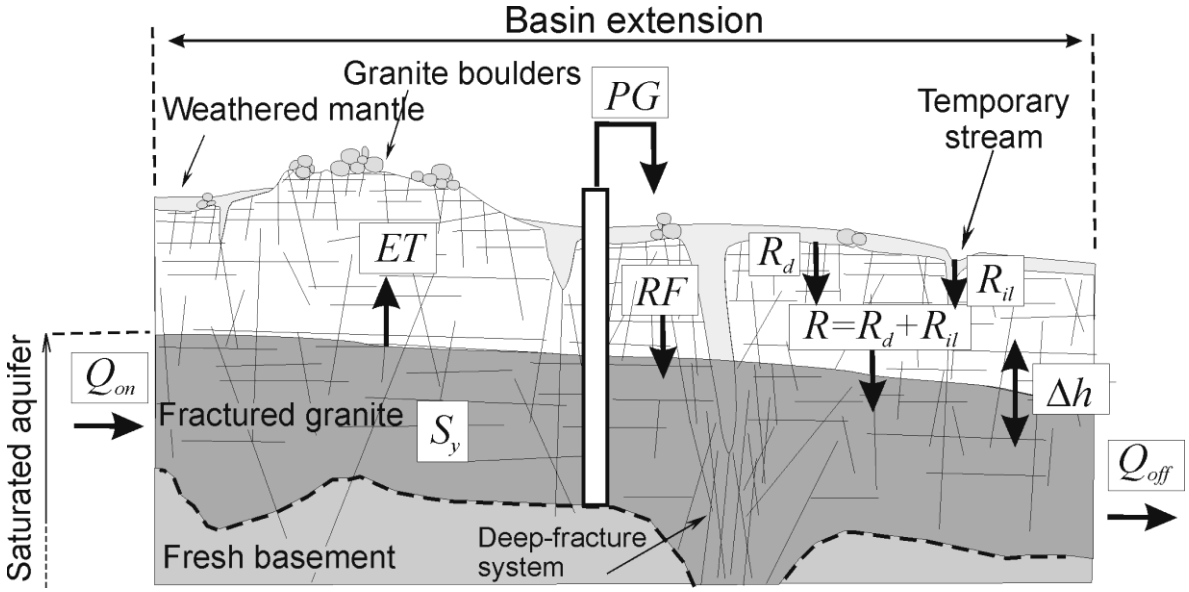
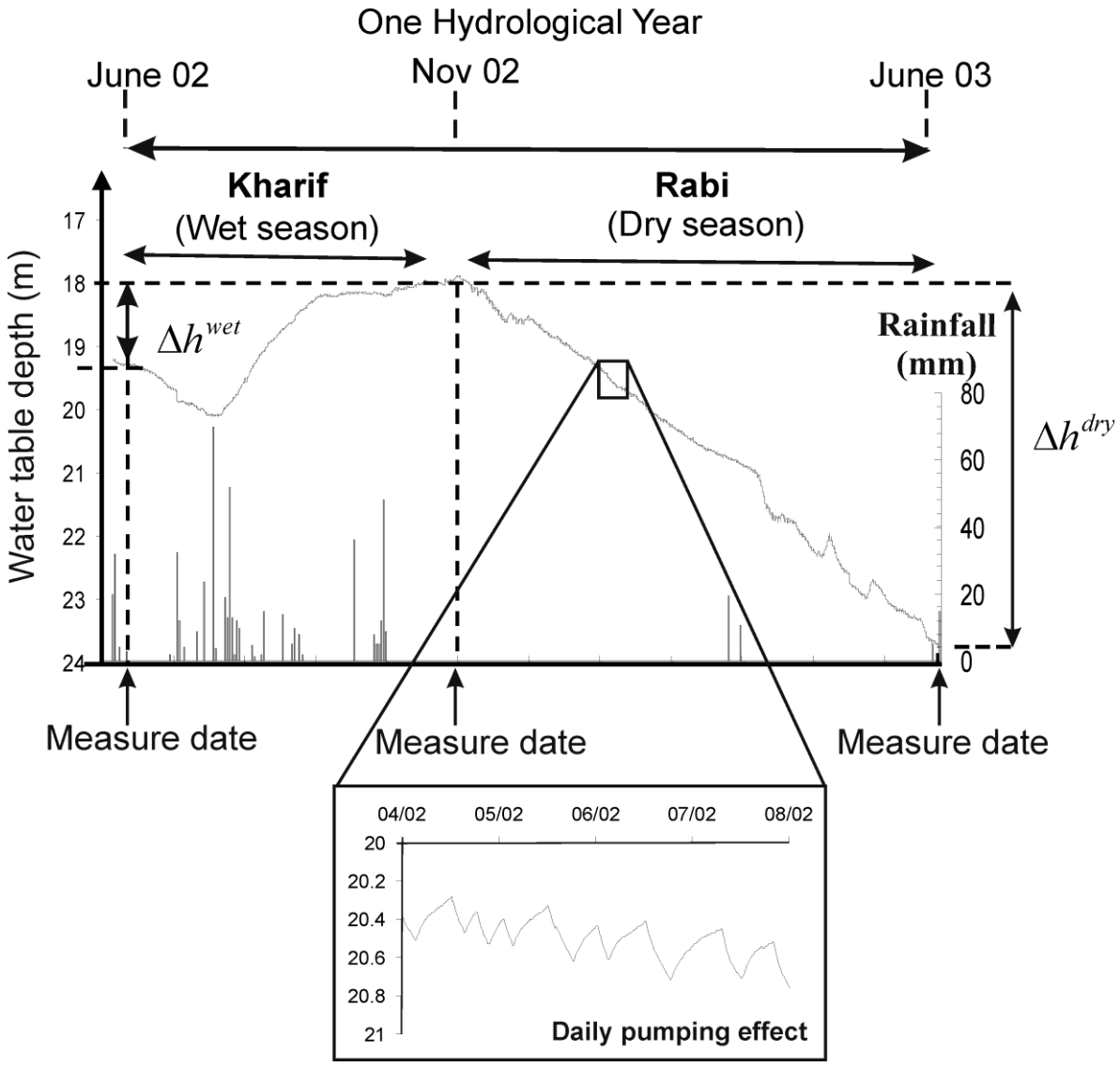


Figure 2



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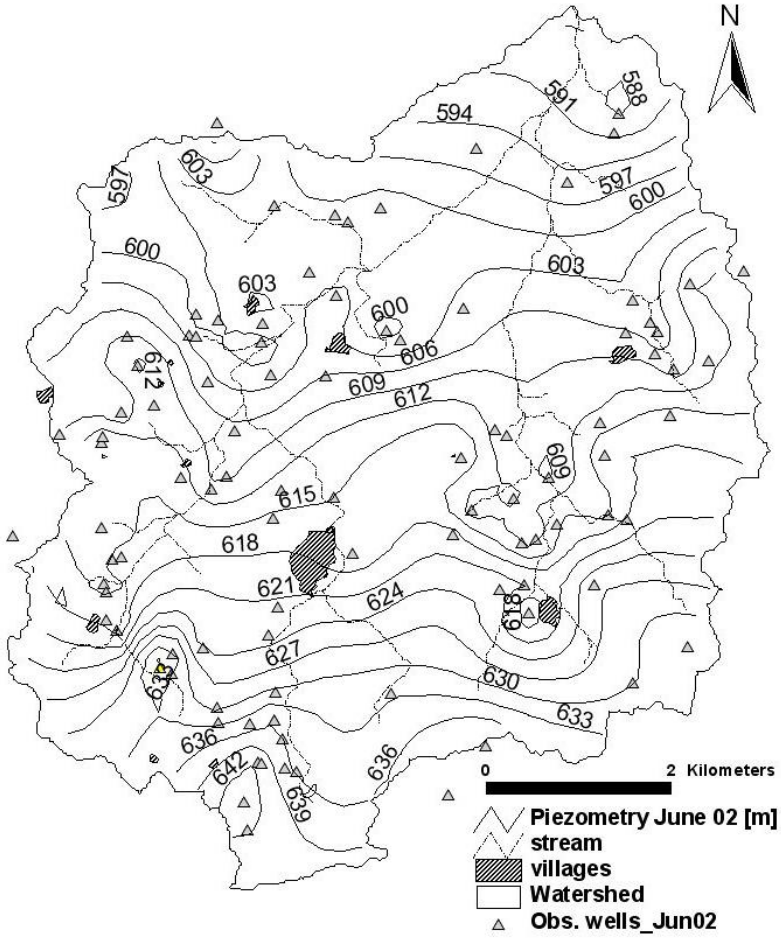

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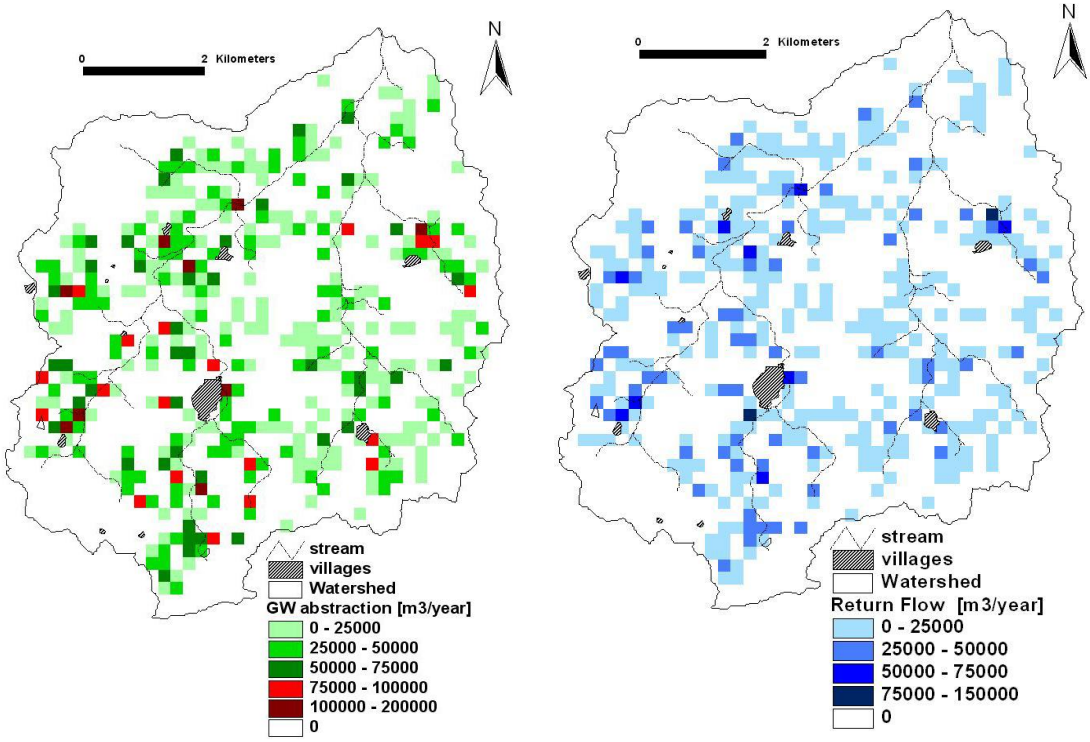

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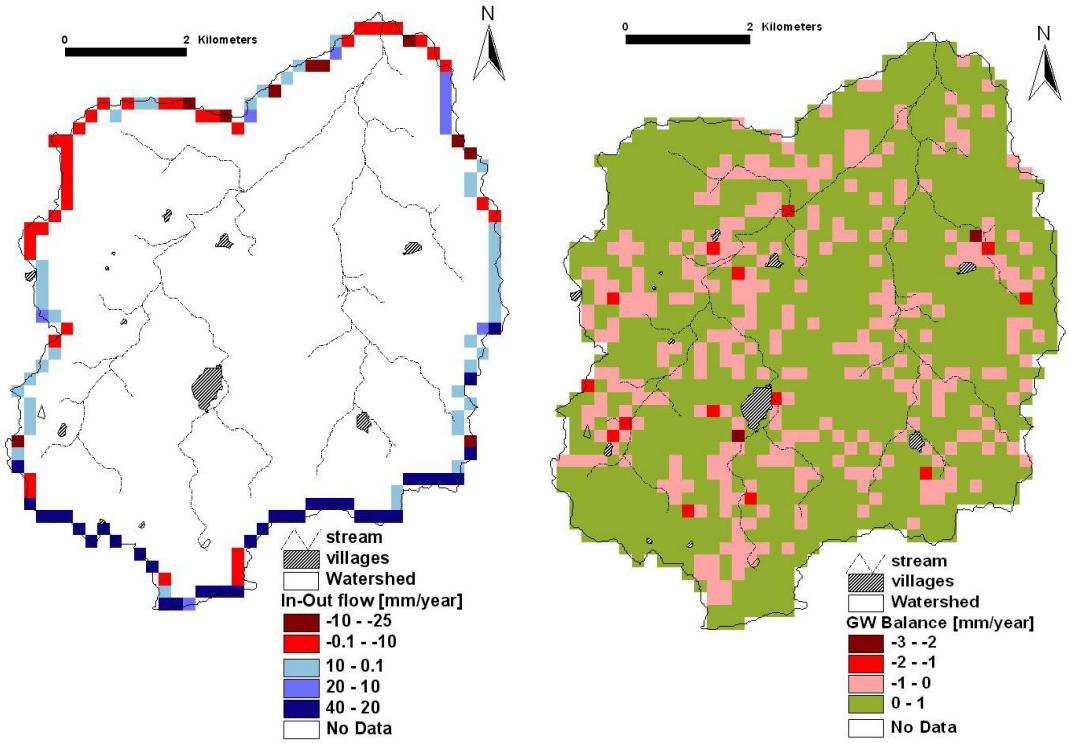

732 Figures 5a to d

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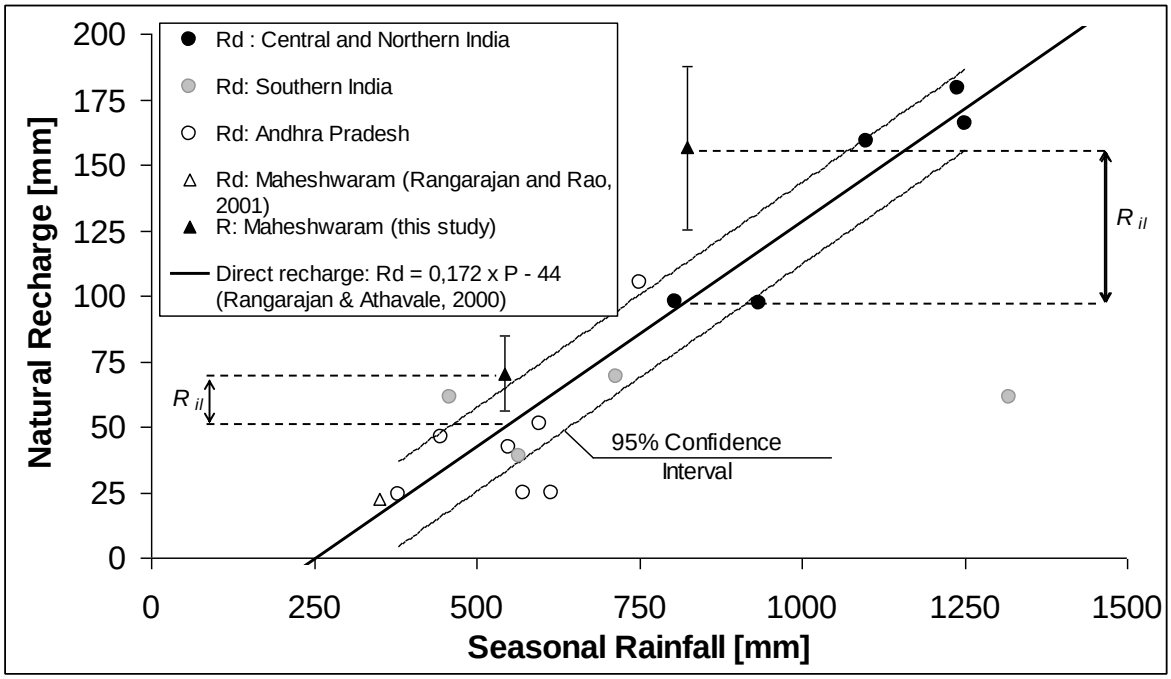
Natural Recharge [mm]
Seasonal Rainfall [mm]
Rd : Central and Northern India
Rd: Southern India
Rd: Andhra Pradesh
Rd: Maheshwaram (Rangarajan and Rao, 2001)
R: Maheshwaram (this study)
Direct recharge: Rd = 0,172 x P - 44 (Rangarajan & Athavale, 2000)
95% Confidence Interval
R il
R il
0
25
50
75
100
125
150
175
200
0
250
500
750
1000
1250
1500

Figure 6