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Pluri-annual recharge assessment using vertical soil temperature profiles: Example of the Seine river basin (1984–2001)

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Abstract

Among the various methods used to determine vertical water seepage in unsaturated soils, thermal convection presents significant advantages: temperature measurements are simple to perform and record, and a wide range of time scales can be considered. The authors analysed the data recorded by the meteorological stations of the Seine river basin, at three different depths: 20, 50 and 100 cm. As the measurement sensitivity was limited to 0.1 K, long series of data needed to be stacked in order to obtain sufficient precision to quantify the convective component of heat transfer, in a predominantly conductive context. For the period from 1984 to 2001, it was possible to determine the average recharge at each station, and the recharge variation between groups of three-year periods. By interpolating these data over the whole basin, a global assessment has been made and compared to the exported flow rate at the river mouth: the resulting value of 94 mm yr^{-1} lies between the lowest annual rate, 52 mm yr^{-1}, and the mean total exported value of 252 mm yr^{-1}. ***To cite this article: A. Tabbagh et al., C. R. Geoscience 341 (2009).***

Résumé

Estimation de la recharge pluriannuelle à partir de suivis de profils verticaux de la température du sol : exemple du bassin de la Seine (1984–2001). La détermination de l'infiltration verticale à partir de l'analyse de l'évolution du profil de température du sol est facile à mettre en œuvre et permet d'aborder une très large gamme d'échelles de temps. Nous avons utilisé les données acquises à 20, 50 et 100 cm de profondeur, aux stations météorologiques du bassin de la Seine. Leur précision étant limitée à 0,1 K, il est nécessaire d'accumuler une quantité suffisante de données pour pouvoir mettre en évidence l'effet de la convection dans un processus de transfert où la conduction domine ; mais du fait de l'intégration sur plusieurs cycles annuels, l'infiltration correspond à la recharge. Les résultats obtenus sur une période de 18 ans interpolés sur tout le bassin donnent une recharge de 94 mm par an, valeur très vraisemblable, puisque comprise entre le flux à l'étiage qui correspond à 52 mm par an et l'export total à 252 mm par an. La variation de la recharge entre périodes de trois ans est aussi une information précieuse sur la variation du stock d'eau souterrain. ***Pour citer cet article : A. Tabbagh et al., C. R. Geoscience 341 (2009).***

Keywords: Recharge; Vertical soil temperature profile; Seine river basin; France

Mots clés : Recharge ; Profil vertical de température du sol ; Bassin de la Seine ; France

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1. Introduction

In the absence of run-off, the water balance on the land's surface is comprised of rainfall, real evapotranspiration (RET) and infiltration. Although rainfall can be measured directly, the direct determination of RET is impossible due to the adaptability of the vegetation cover. Only its upper bound can be restricted by the potential evapotranspiration (PET). It is very important to determine the level of infiltration, and, better still, net infiltration [10] that corresponds to the net gain in underground water [1,6] and will be equated to the recharge in the present study. However, the choice of spatial and time scales, as well as that of the measurement technique, can pose serious problems. Water, which penetrates into the first centimetres, or even metres, of the ground may be taken up by plant roots, without contributing to any increase in the underground storage. Although piezometer measurements indicate the water table depth, they do not provide any information concerning the amount of water stored in the unsaturated zone.

A rather large panel of geophysical methods can be used to assess the amount of water stored in the ground. Some of them, such as the gravity method, have a strongly integral (damped) effect [7,13], whereas others are more locally sensitive [5]. Even those capable of directly determining the liquid water content (MRS, Magnetic Resonance Sounding) are not sensitive to water seepage. However, at least one method, based on temperature distribution, is sensitive to the water flow rate, i.e. Darcy's velocity, and can be used to determine it by a very simple principle: a certain quantity of convective heat transfer can be associated with a measured volume of fluid seepage, such that analysis of the temperature distribution (temporal variations at different depths) can be used to evaluate the flow rate. This principle is well understood and has been applied for several decades [11,12,14,15].

The theoretical temperature variation with depth in a layered 1D ground from which water is exiting by vertical seepage has already been described [3]. Its behaviour is governed by the following equation:

$$k_i \frac{\partial^2 T}{\partial z^2} - u C_{\mathrm{w}} \frac{\partial T}{\partial z} - C_{\mathrm{vi}} \frac{\partial T}{\partial t} = 0 \qquad (1)$$

where k_i is the thermal conductivity of the i^{th} layer, u is the vertical component of Darcy's velocity, C_{w} is the volume heat capacity of water, and C_{vi} is the bulk volume heat capacity of the i^{th} layer. By defining the thermal diffusivity $\Gamma_i = k_i / C_{\mathrm{vi}}$ and the heat advection velocity $\nu_i = u(C_{\mathrm{w}}/C_{\mathrm{vi}})$, one can write:

$$\Gamma_i \frac{\partial^2 T}{\partial z^2} - \nu_i \frac{\partial T}{\partial z} - \frac{\partial T}{\partial t} = 0 \qquad (2)$$

By applying to this equation the Fourier Transform, the time variable is eliminated and a solution can be easily proposed for the temperature spectrum $\hat{T}(z, \omega)$ in each layer. The unknown constants of this solution are determined due to the continuity of the temperature and of the vertical flux at each interface.

It is thus possible by fitting a theoretical solution to the experimental frequency variations of the temperature at different depths to determine Γ_i, υ_i, the volume water content θ and Darcy's velocity [14]. However, two difficulties, arising from the fact that conduction is the dominant mode of heat transfer, must be overcome:

1) a possible lack of measurement sensitivity (which can be partly compensated for by stacking) [2];
2) prior determination of variations in thermal properties versus depth [3].

2. Inversion method from temperature profile to water flux

One must first define where, at which depth, with which sampling step, and when, temperature should be measured. If the sensors are placed beyond the root zone, this option is quite difficult to implement and will prevent the monitoring of rapid variations. Positioning of the sensors within the first metre is more straightforward; it allows both short time and long time variations to be recorded, and when the results are integrated over one or more annual cycles the vegetation uptake is cancelled.

Ground temperature measurements are recorded by meteorological stations at several depths, and over periods of several decades. It thus makes sense to consider the relevance of these data, even if more sensitive and more closely spaced measurements need to be made later. We thus chose to make initial recharge determinations over the entire hydrographic basin of the Seine River, using the data collected by meteorological stations.

There are 15 stations which have similar measuring conditions: they are situated on a horizontal surface (absence of run-off), have a grassy cover, provide daily samples taken at depths of 0.1, 0.2, 0.5 and 1 m, and have a temperature sensitivity of 0.1 K, resulting from the Least Significant Bit (LSB), the sensor noise being

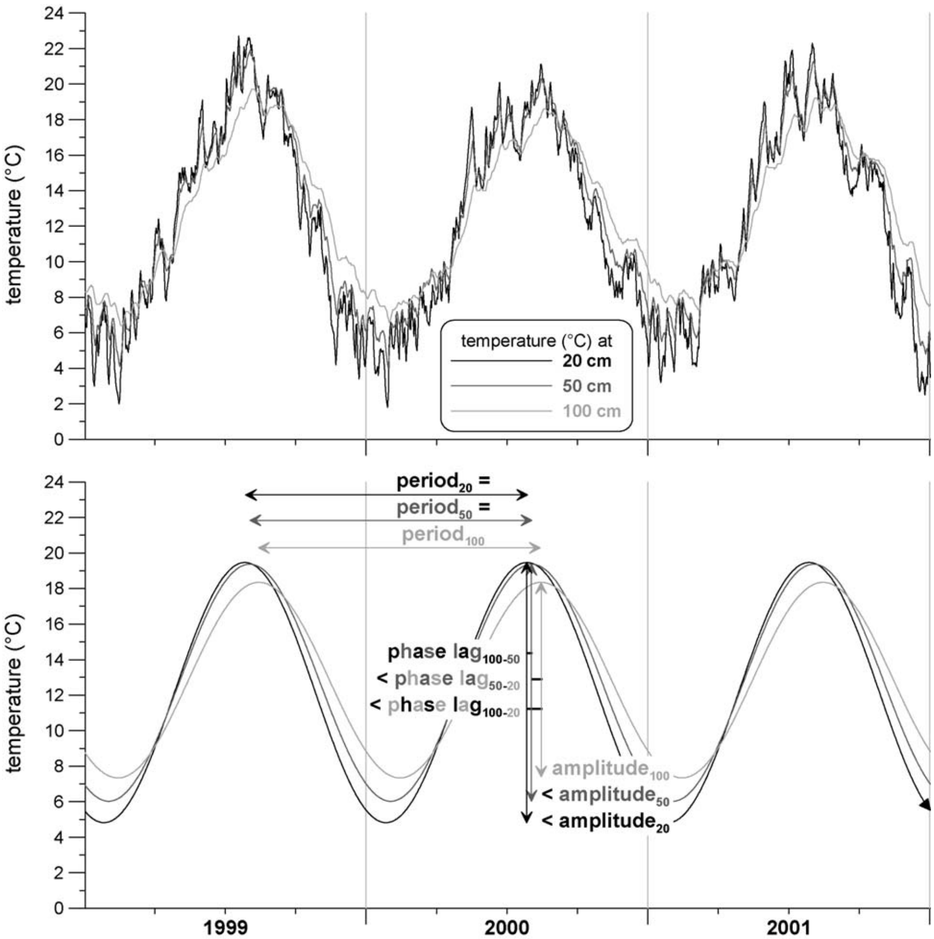

Fig. 1. Temperature record at 0.2, 0.5 and 1 m, Vélizy-Villacoublay station (Yvelines, France) for the [1999–2001] period and corresponding sine annual variations.

Fig. 1. Température à 0,2, 0,5 et 1 m, station de Vélizy-Villacoublay (Yvelines, France) pour la période [1999–2001] et variations sinusoïdales correspondantes.

lower. The 0.1 m data were not used, to avoid spectral aliasing. In order to compensate for the lack of sensitivity and the limited quantity of data, we analysed records dating back 18 years, from all 15 stations. The temperature records and the correspondent sine annual variation are presented in Fig. 1 for the [1999–2001] period at Vélizy-Villacoublay (Yvelines, France).

At each station, electrical soundings were made [3] at the exact location of the temperature sensors, in order to identify the vertical position of the ground structure interfaces (Table 1 shows the interpretation of the electrical sounding at the Vélizy-Villacoublay station). The results were then interpreted using a three-layer model, which has a sufficient number of interfaces to fit any variation of the soil properties as a function of depth. The thermal properties of each of the three layers depend on several parameters: time, volume water content, θ, and time–independent variables i.e. porosity, n, and thermal conductivity of the solid fraction, k_{si}. To simplify the model, we assumed the porosity to be uniform and equal to 0.45, and assumed the volume water content to be independent of depth (because long periods are considered). Consequently, only the solid fraction conductivity varies with depth, and thermal properties can be determined using the following equations [4]:

$$k_i = (0.8908 - 1.0959n)k_{\mathrm{si}} + (1.2236 - 0.3485n)\theta \tag{3}$$

and

$$C_{\mathrm{v}} = C_{\mathrm{w}}\theta + (1 - n)C_{\mathrm{s}}, \tag{4}$$

Table 1
Three layer structure, Vélizy-Villacoublay station (Yvelines, France).

Tableau 1
Structure du terrain, station de Vélizy-Villacoublay (Yvelines, France).

	Resistivity (ς m)	Thickness (m)	Thermal conductivity of the solid fraction k_s (W K^{-1} m^{-1})	Thermal conductivity k (W K^{-1} m^{-1})
1st layer	260	0.22	2.38	1.28
2nd layer	73	0.49	3.27	1.63
Substratum	16		2.07	1.16

where the solid fraction volume heat capacity, C_s, is assumed to be uniform and equal to 2×10^6J m^{-3} K^{-1}, and the water volume heat capacity, C_w, is 4.18×10^6J m^{-3} K^{-1}.

In order to determine the four unknowns k_{s1}, k_{s2}, k_{s3} and θ, one proceeds in two steps:

1) The full set of recordings is divided into N equal periods: in the present case, the 18 years recording was divided into six three-year periods. For each period and for each depth, a least squares regression was used to fit a sinusoidal function to the data, using the expression:

$$T(z,t) = T_0 + Az + B\cos\omega t + D\sin\omega t \tag{5}$$

where ω is the angular frequency corresponding to the annual cycle, and A, B and D depend on the considered depth.

2) For each pair of depths, (0.2, 0.5) and (0.5, 1.0), the values of B and D are used to calculate two apparent thermal diffusivities: Γ_{ph} is derived from the phase lag, and Γ_{amp} is derived from the amplitude ratio [3,14]. The apparent diffusivities are defined as the diffusivities of a homogeneous stretch of ground which, for the same pairs of depths, would lead to the same values of phase lag and amplitude ratio: if one has B_1 and D_1 at depth z_1 and B_2 and D_2 at depth z_2, one would have:

$$\Gamma_{amp} = \frac{\omega}{2} \frac{(z_1 - z_2)^2}{\left(\ln\left(\sqrt{B_1^2 + D_1^2}/\sqrt{B_2^2 + D_2^2}\right)\right)^2} \tag{6}$$

and

$$\Gamma_{ph} = \frac{\omega}{2} \frac{(z_1 - z_2)^2}{\left(\mathrm{arctg}\frac{D_1}{B_1} - \mathrm{arctg}\frac{D_2}{B_2}\right)^2} \tag{7}$$

In the absence of seepage in homogeneous ground, these quantities are equal. As Γ_{ph} is practically independent of seepage, its values are used to determine the thermal properties of the layers, from two independent pairs of depths and N periods. This leads to $2N$ relationships that need to be solved for four unknowns.

Once the thermal properties have been determined, a period is chosen for which the value of u is needed. The two values of Γ_{amp} corresponding to this period are then computed, and used to calculate the best fit for the value of u.

3. Groundwater recharge interpretation of Seine catchment

From this data, we first determined Darcy's velocity, which is equivalent to the recharge, for the whole period. The numerical results are presented in Table 2 together with rainfall, potential evapotranspiration, potential excess precipitation (P-PET), and real excess precipitation (P-RET) calculated monthly using Thornthwaite's simplified algorithm [8,16] and then averaged over the 18 years. The potential excess precipitation constitutes a lower bound for the recharge and the real excess precipitation normally constitutes an upper bound (it takes into account the run-off). However, the determination of this second term is very sensitive to the time interval over which it is calculated: for long intervals (in the present case the month) it is underestimated. We then considered, Table 3, the recharge differences, for the [1996–1998] and [1999–2001] three-year periods, which correspond to periods during which the average rainfall increased significantly.

The recharge map is shown in Fig. 2. The values are interpolated by kriging using a spherical variogram model. The rainfall was the highest towards the north-west (near the sea) and the south-east (higher altitude), with an average value of 691 mm yr^{-1} over the catchment. The recharge variations are higher in the north than in the south, with an average value of 94 mm yr^{-1} and a standard deviation around 10 mm yr^{-1} as established in [3] for each station. These values are not far from the simple means of 101 for recharge and 676 mm yr^{-1} for rain and can be compared to Thiessen's polygons surface weighted

Table 2
Rainfall, potential evapotranspiration, potential excess precipitation (P-PET), real excess precipitation (P-RET) calculated by Thornwaite's algorithm and recharge during the [1984–2001] period, in millimetres per year. The X corresponds to an absence of data.

Tableau 2
Pluie, ETP, pluie-ETP, pluie-ETR calculée par l'algorithme de Thornwaite et recharge sur la période [1984–2001]. Le X correspond à une absence de données.

Station	Rainfall	PET	P-PET	P-RET (Thornthwaite)	Recharge
Abbeville	794.0	684.5	109.5	245.1	151.3
Auxerre	723.1	767.6	−44.5	160.9	163.4
Beauvais	695.3	733.6	−38.4	166.6	225.2
Bonneuil (Le Bourget)	646.3	771.1	−124.8	108.4	52.0
Boos (Rouen)	857.1	696.3	160.8	281.3	X
Brétigny	629.3	633.7	−4.4	301.8	36.3
Bricy (Orléans)	637.5	802.2	−164.7	119.3	−32.5
Chartres	599.4	730.3	−130.9	104.3	36.9
Courcy (Reims)	635.8	725.9	−90.2	105.3	197.7
Fontaine (St Quentin)	707.2	697.0	10.2	171.0	142.2
Huest (Evreux)	607.8	722.4	−114.6	106.5	179.0
Nevers	787.0	725.4	61.7	218.5	−46.8
Paris-Montsouris	656.5	794.0	−137.5	100.2	103.3
Troyes	648.6	743.6	−95.0	124.2	193.0
Vélizy-Villlacoublay	694.4	782.5	−88.1	143.7	16.0

Table 3
Increases in millimetres per year between [1996–1998] and [1999–2001] periods in rainfall intensity, in potential evapotranspiration, and in recharge rate. The X corresponds to an absence of inversion convergence.

Tableau 3
Augmentation en millimètre par an entre [1996–1998] et [1999–2001] de la pluie, de l'évapotranspiration potentielle, et de la recharge. Les X correspondent à une absence de convergence du processus d'inversion.

Increase between the [1996–1998] and [1999–2001] periods			
Station	Rainfall intensity	PET	Recharge
Abbeville	348.0	3.8	188
Auxerre	179.6	−24.3	X
Beauvais	289.2	−3.1	223
Bonneuil (Le Bourget)	277.3	−12.2	339
Boos (Rouen)	284.4	−3.3	117
Brétigny	211.5	−40.9	X
Bricy (Orléans)	282.5	−27.6	X
Chartres	209.3	6.6	X
Courcy (Reims)	162.7	30.4	160
Fontaine (St Quentin)	196.0	25.6	330
Huest (Evreux)	221.7	−3.5	377
Nevers	117.9	−6.7	318
Paris-Montsouris	231.4	−0.7	X
Troyes	207.0	3.9	X
Vélizy-Villlacoublay	277.7	−33.9	310

results (presented in Fig. 3), which, respectively, deliver 111 and 676 mm yr^{-1}. One must also note that the application of Thiessen's polygons to the Thornthwaite's excess (P-RET) leads to 112 mm yr^{-1} for the basin.

Although such a recharge value is plausible, it merits comparison with other sources of data. The river flow rate at the mouth of the Seine (Caudebec en Caux gauge station) is 586.6 $m^3 s^{-1}$, which corresponds to 252 mm yr^{-1} for an upstream catchment area of 73 300 km^2. The difference between these two recharge rates is significant (by a factor of 2.7), but could be attributed to run-off that feeds the river. In order to assess the likelihood of this result, one can consider the

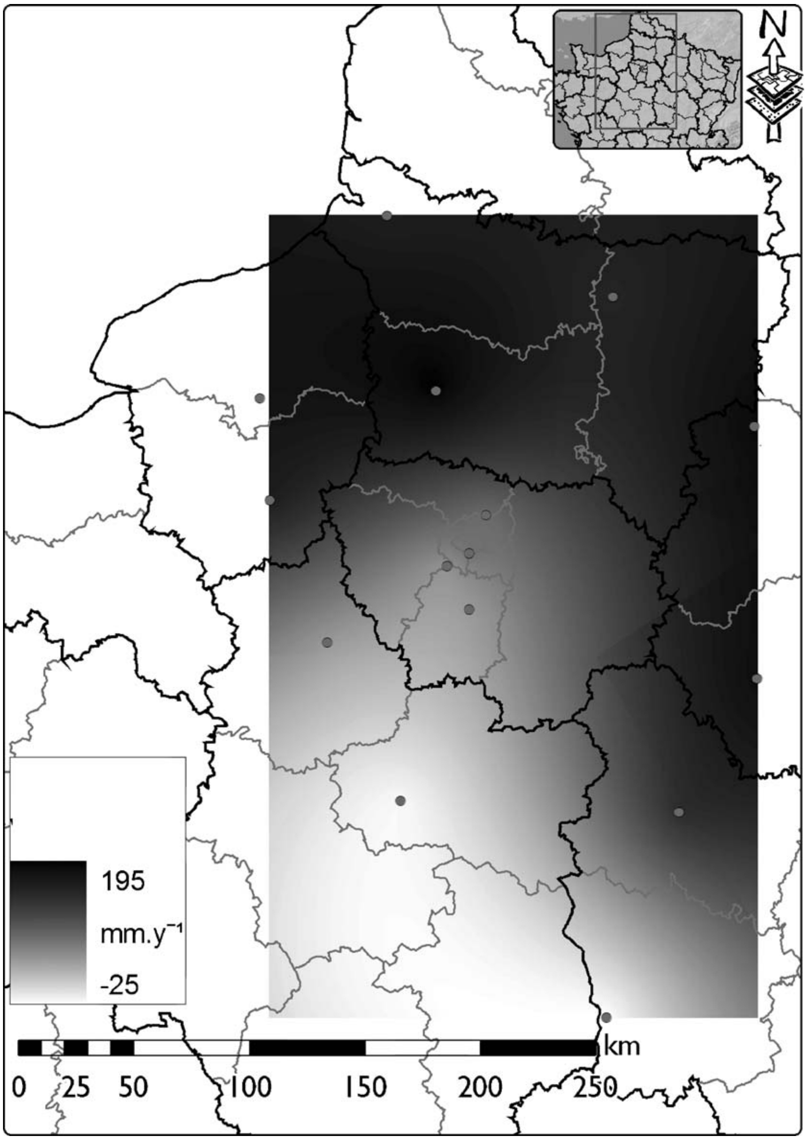

Fig. 2. Map of the average recharge for the [1984–2001] period.

Fig. 2. Carte de la recharge entre 1984 et 2001.

lowest (summer) water level, which corresponds to a lower bound for the recharge. Although the lowest flow rate at the river mouth is around 200 $m^3 s^{-1}$ [9], corresponding to 86 mm yr^{-1}, this term includes the water delivered to the river by the lakes, which is estimated to represent 40% of the total flow. The lowest bound for the recharge would thus be 52 mm yr^{-1}. It is likely that the average value over the full period of 18 yearly cycles is 1.8 times greater.

The rainfall intensity increased significantly during the [1996–1998] and the [1999–2001] three-year periods, with an average value of 226 mm yr^{-1}. It was even higher in the north-western part of the basin where the Somme River flood occurred during the spring of 2001. The calculation of the difference in recharge can be achieved at only nine of the 15 stations (Table 3). The average difference in recharge rate is 262 mm yr^{-1}, which is slightly higher than that of the rain. Two explanations can be proposed for the significant increase in recharge rate: wetter soils have a greater hydraulic conductivity, and rainy weather leads to a reduction in RET. Between the two three-year

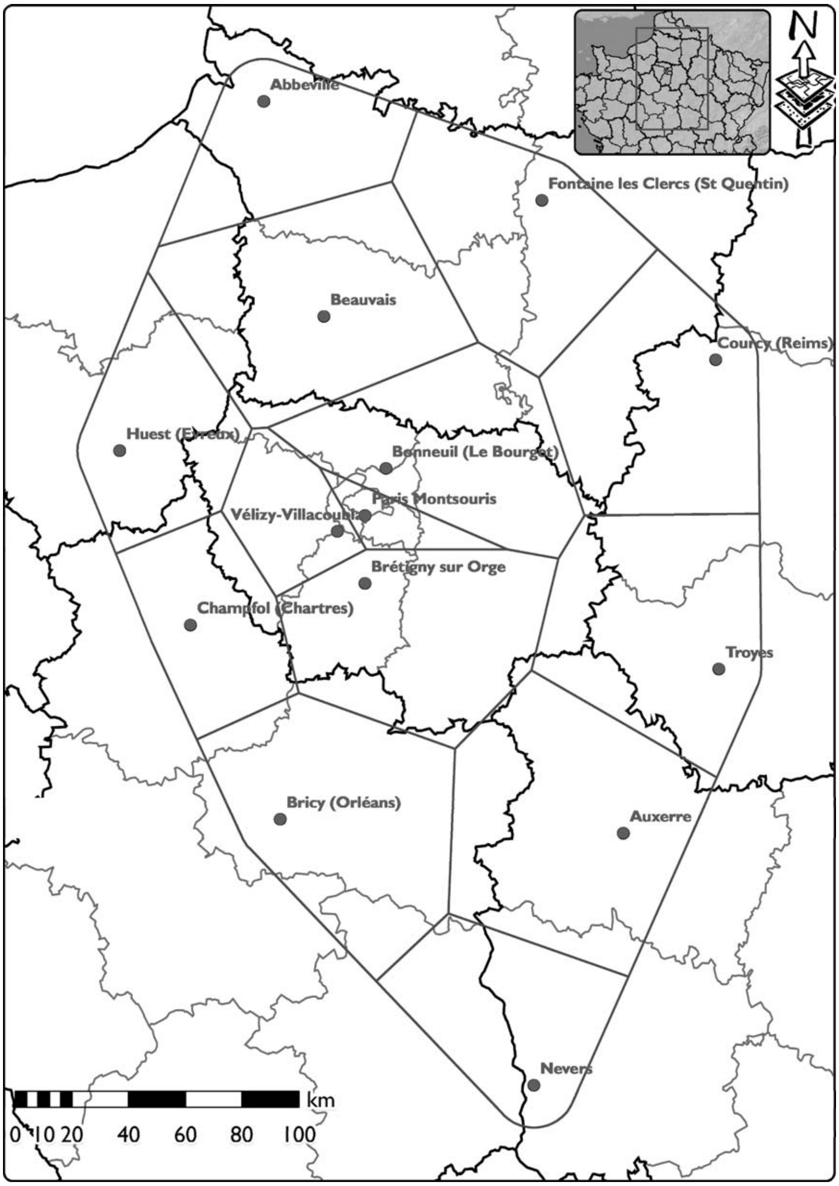

Fig. 3. Influence of each station determined by Thiessen's polygons areas.

Fig. 3. Aire d'influence de chaque station, déterminée par la méthode des polygones de Thiessen.

periods, the flow at the Seine mouth increased from 453 to 821 $m^3 s^{-1}$, corresponding to an annual difference of 158 $mm\,yr^{-1}$, which suggests that underground water storage increased during the second three-year period.

4. Conclusion

The calculation of recharge rates using existing values of ground temperature is difficult to implement, due to the large time intervals in the recorded data at meteorological stations, and to the low sensitivity of the measurements. Nevertheless, by using data recorded over long periods of time, it is possible to obtain plausible results that can be combined with other types of hydrological data to characterise the long-term water balance.

The possibilities offered by modern technology, to significantly enhance both the time step (a quarter of an hour can easily be achieved) and the sensitivity (0.001 K is possible) of recordings open up very wide perspectives for the assessment of infiltration and recharge rates, over a large range of spatial and temporal scales.

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