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Estimation of groundwater recharge using the cumulative rainfall departure method for Bac Lieu province, Mekong Delta, Vietnam

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Abstract. Estimation of the groundwater (GW) recharge from rainfall is important for determining GW resources in water resources development and management. GW is currently extensively exploited and is an important source of freshwater for people in the Mekong Delta, Vietnam, especially during dry seasons. To achieve sustainable utilization of GW resources in the delta, it is essential to determine the annual renewable GW reserve from the rainfall recharge. The study provides evidence for the application of the cumulative rainfall departure (CRD) method for the GW recharge estimation for deep aquifers. The monitored rainfall data and GW levels of the aquifers in Bac Lieu province are used. The results of the analysis by the CRD method show that the fractions of cumulative rainfall departure for Holocene (qh), Upper Pleistocene (qp_3), Middle-Upper Pleistocene (qp_{2-3}), and Lower Pleistocene (qp_1) aquifers are 0.08 %, 0.18 %, 0.55 %, and 0.50 %, respectively, which only equals 1.31 % of the total rainfall. The Pearson correlation between the observed and model water levels is high, from 0.898 to 0.925. The total GW annual recharge from the rainfall over the province is estimated to be 74.07 million m^3, equivalent to 203 000 m^3/day, i.e., which is 16 % lower than the current water abstraction of 23 600 m^3/day. The obtained results are important for subsequent comparison with the Red River basin in northern Vietnam, where it is necessary to keep track of the groundwater inflow along with its volume/resource, including the inflow from the geothermal system of the rift zone of the Red River.

Keywords: Mekong Delta, groundwater monitoring, Pleistocene, Holocene, Pearson correlation, net recharge

Оценка пополнения запасов подземных вод, с использованием метода кумулятивного оттока осадков, для провинции Бак Льеу, дельта реки Меконг, Вьетнам

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Резюме. Оценка подпитки грунтовых вод (ГВ) за счет атмосферных осадков важна для определения ресурсов ГВ при освоении и управлении водными ресурсами. В настоящее время ГВ широко эксплуатируются и являются важным источником пресной воды для населения во вьетнамской дельте р. Меконг, особенно в засушливые сезоны. Для достижения устойчивого использования ресурсов ГВ важно определить ежегодный возобновляемый резерв ГВ благодаря подпитке дождевой водой. В работе приведена аргументация применимости метода кумулятивного притока осадков (cumulative rainfall departure, CRD) для оценки подпитки ГВ для глубоких водоносных горизонтов. Используются данные о количестве осадков и уровнях ГВ водоносных горизонтов в провинции Бакльеу. По результатам анализа методом CRD, доли кумулятивного оттока осадков для водоносных горизонтов голоцена (*qh*), верхнего плейстоцена (qp_3), среднего–верхнего плейстоцена (qp_{2-3}) и нижнего плейстоцена (qp_1) соответственно составляют 0.08 %, 0.18, 0.55 и 0.50 %, то есть всего 1.31 % от количества осадков. Корреляция Пирсона между наблюдаемыми и модельными уровнями воды высокая, от 0.898 до 0.925. Общий годовой запас ГВ от осадков над провинцией оценивается в 74.07 млн м^3, что эквивалентно 203 000 м^3/день, т.е. на 16 % ниже текущего забора в 23 600 м^3/день. Полученные результаты важны для последующего сравнения с бассейном р. Красная на севере Вьетнама, где кроме оценки ресурсов грунтовых вод необходимо вести учет поступления подземных вод, в том числе из геотермальной системы рифтовой зоны р. Красная.

Ключевые слова: дельта Меконга, мониторинг подземных вод, плейстоцен, голоцен, корреляция Пирсона, возобновляемые водные ресурсы

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Introduction

Groundwater (GW) is an important freshwater source for people in the Mekong Delta, Vietnam, especially during dry seasons when the quality of surface water is deteriorated in most parts of the delta [1–4]. Groundwater is abstracted via dug wells, small-scale household tube wells, or medium- and large-scale central supply wells that were dug as part of the Rural Clean Water Supply Program [5–7]. The Division for Water Resources Planning and Investigation found that 60 percent of wells access the Pleistocene aquifers of the delta (qp_{2-3} and qp_1 in Fig. 1 a), and that most water supply projects for domestic and industrial water supply use this aquifer. The GW exploitation, especially over-exploitation, causes land subsidence, which in coastal areas poses a flood inundation hazard in the Delta. The average annual rate of GW decline is about 0.3 m and the land subsidence is at an average annual rate of 1.6 cm [8]. From 1991–2015, the Delta sank on average about 18 cm and in 2016 within 1.1 cm/yr – 2.5 cm/yr as a consequence of GW withdrawal [9, 10]. Minderhoud et al. [11] presented projections of extraction-induced subsidence and consequent delta elevation loss for this century using a 3D hydrogeological model with a coupled geotechnical module. The results show that if the GW extraction continues to increase continuously as in the past decades, extraction-induced subsidence would drown the Mekong Delta before the end of the century. The combined effect of global sea-level rise [11] and groundwater abstraction and its induced subsidence result in further saltwater intrusion [8, 12].

To have a future sustainable utilization of GW resources in the Mekong Delta, the determination of the annual GW renewable component, including the recharge of GW in flood periods, is an important issue [13]. One of the components of GW renewable reserves is the dynamic reserve thanks to the rainwater recharge. Although the annual rainfall in the Mekong Delta is from 1300 mm to 2500 mm [14], the wide distribution of the top surface soil from medium (sand, silty sand) to weak (silt, semi-permeable clay) formations limits the GW recharge from rainfall. The authors Jan et al. [15] pointed out that, the fresh GW volume in the Mekong Delta is huge in comparison with the present yearly extraction rate. However, the annual replenishment is very limited, the extraction rate can cause rapid upcoming of brackish and saline GW, which is a serious salinization threat to the fresh GW volume.

Within this work, a study on the GW recharge via the rainfall infiltration using the revised cumulative rainfall departure (CRD) method [16] for GW levels obtained in the monitoring GW boreholes in Bac Lieu province in the Mekong Delta is carried out and presented.

About recharge estimation methods

Groundwater recharge estimation is a difficult, sensitive, and delicate problem and varies very much in accuracy and uncertainty. The authors Kinzelbach et al. [17] in their survey work on the most common methods of recharge estimation have classified into the following groups with accuracy ratings in three classes, according to regional recharge estimates: 1) class 1: factor of 2 (two times larger or two times smaller than the true value); 2) class 2: factor of 5 (of the same order of magnitude); and 3) class 3: factor of 10 or more (with large errors likely). Following Kinzelbach et al. [17], the CRD method proposed by Bredenkamp et al. [18] has advantages in simplicity and error stabilization thanks to the long time series, and the disadvantage in requirement of storage coefficient, of known discharge (including abstractions), and the accuracy class 2 to 3.

GW recharge may occur thanks to different mechanisms, four types of which can be distinguished by Xu and Beekman [19]: 1) vertical water flow through the unsaturated zone reaching the water table; 2) lateral and/or vertical inter-aquifer flow; 3) artificial recharge, such as from infiltration ponds or injection water wells; and 4) induced recharge from nearby surface water bodies (rivers, streams, lakes), which results from the groundwater abstraction.

The CRD methods have been applied in several studies, mainly for shallow aquifers. The authors Xu and Van Tonder [16] demonstrated the application of the CRD method to several shallow fractured aquifers with low storativity. Adams et al. [20] used the chloride mass balance (CMB), the saturated volume fluctuation (SVF), and the CRD methods to quantify GW recharge of the aquifers in the Central Namaqualand of South Africa and obtained that the SVF and CRD results are of a good agreement and that the GW recharge is higher in the mountainous areas than in the lower lying areas, and that the GW recharge mainly occurs through the alluvial aquifers with significant soil cover.

The author Baalousha [21] applied the CRD method for quantification of GW recharge in the Gaza Strip (Palestine), a transitional zone between the semi-humid zone and the semi-arid loess plains. The aquifer system in the area of study. The aquifer is mainly phreatic with thickness from a few meters at the east to 170 m and consists of Pleistocene calcareous sandstone and gravel and Holocene sands interbedded with some silts, clay, and conglomerate. The aquifer storativity values range between 0.0005 and 0.03. The average GW recharge had been determined to be from 31.64 % to 41.10 %, on average 36.74 %.

The authors Rasoulzadeh and Moosavi [22] utilized CRD and revised CRD methods to estimate the GW recharge from rainfall to the Quaternary aquifer, with thickness from 30 m to more than 300 m in the Tashk Lake area, Iran. The aquifer consists of rubble stone, gravel, sand, and silt with a small amount of clay and has a storativity of 0.0342 as determined by the pumping test data. The results showed that the rainfall recharge is 35 % with a lag time of 1 month between the rainfall and GW level change.

The authors Sun et al. [23] applied the rainfall infiltration breakthrough (RIB), a modified revised CRD model for the estimation of GW recharge by the rainfall percolating through the unsaturated zone to the water table in two research sites characterized by two extremely different types of geology in west coastal South Africa: the coastal plain sand aquifer with a diffuse recharge in Riverlands Nature Reserve (Western Cape, South Africa), and the mountain group aquifer (TMG) with a localized recharge in Oudebosch catchment in the Kogelberg Nature Reserve (Western Cape, South Africa). The storativity ranges from 0.05 to 0.15 for fine to medium sand cover, and the water table depth ranges from 1.6 m to 3.5 m. The GW recharge estimated is 16–47 % at the daily scale and 9.3–27.8 % at the monthly scale at Riverlands, while at Oudebosch is 51.5 % of MAP at the daily scale and and 15.7 % at the monthly scale. Similarly, Nguyen Duc Roi [24] applied the RIB method to the shallow Holocene aquifer in Hung Yen province, Northern Vietnam.

Some issues remain with the CRD method, such as the applicability of the method for deep aquifers [16], for different climatic areas under different hydrogeological conditions [23]. The present study attempts to deal with the climatic conditions different from the arid and semiarid, i.e., the tropical climatic conditions, and for deep aquifers with vertical inter-aquifer flow through semi-pervious aquitards and rainfall percolation through the ground surface aquitard in the Vietnamese Mekong delta in the southernmost region of Vietnam.

Cumulative rainfall departure method (CRD)

Method description

The revised CRD method utilizes the relationship between water level fluctuations and the departure of rainfall from the mean rainfall of a preceding time [16] which is defined as:

$$ {}_{t}^{1}CRD_i = \sum_{n=1}^{i} R_n - \left(2 - \frac{1}{R_{av} i}\sum_{n=1}^{i} R_n\right)\sum_{n=1}^{i} R_t; $$

$$ (n = 1,\ 2,\ 3 \ldots N), \quad (1) $$

where:

N is the total length of rainfall series;

R_i is the rainfall amount at i^{th} time scale (daily, monthly, or annually);

R_{av} is the mean precipitation of the whole time series;

P_t is a threshold value representing the boundary conditions.

A linear relationship between the CRD_i value and water level change during the i^{th} time duration is assumed:

$$ \Delta h_i = \frac{r}{S}\left({}_{t}^{1}CRD_i\right); \quad (i = 0,\ 1,\ 2,\ 3 \ldots N), \quad (2) $$

where:

Δh_i is the water level change during the i^{th} time duration;

r is the fraction of cumulative rainfall departure which results in recharge;

S is the aquifer storativity (i.e., the aquifer storage coefficient S* for confined aquifer or the specific yield μ for the unconfined aquifer).

Therefore, Eq. (2) may be used to estimate the ratio of the aquifer recharge to the aquifer storativity through a regression analysis between CRD_i and Δh_i.

In case the monitored groundwater level fluctuations are effective by the natural aquifer discharge and/or pumping out, the following is used instead of Eq. (2):

$$ \Delta h_i = \frac{r}{S}\left({}_{t}^{1}CRD_i\right) - \frac{Q_{outi}}{AS}; $$

$$ (i = 0,\ 1,\ 2,\ 3 \ldots N) \quad (3) $$

where:

Q_{outi} is the total natural aquifer discharge and pumping out during the i^{th} time duration;

A is the area under the effect of the aquifer discharge, pumping out, or pumping in.

If the rainfall departure is positive, the water level definitely will rise and vice versa. However, as long as there is a surplus of recharge over the discharge of an aquifer, the natural water level may continue to rise even though the departure is negative.

About Rt values

Following the authors Xu and Van Tonder [16], the value of $R_t = 0$ represents a closed aquifer system, which means that the recharge at i[th] time scale only depends on preceding rainfall events, while a value of $R_t = R_{av}$ represents an open system, which perhaps is regulated by spring flow. Both r and R_t values are determined during the simulation process. The value of R_t slightly greater than R_{av} had been revealed in the current work and shall be addressed and discussed as follows.

For accessing the R_t value, let us consider the case that there is no recharge other than the rainfall recharge and any discharge, and rewrite (1) and (2) as follows:

$$r = \frac{S\Delta h_i}{\sum_{n=1}^{i} R_n - \left(2 - \frac{1}{R_{av} i}\sum_{n=1}^{i} R_n\right)\sum_{n=1}^{i} R_t};$$

$$(n = 1,\ 2,\ 3\ldots N) \tag{4}$$

For $R_t = 0$: $\Delta h_i = \frac{r}{S}\sum_{n=1}^{i} R_n$ (n = 1, 2, 3...N)

which shows that Δh_i is increasing with time as $\sum_{n=1}^{i} R_n$; (n = 1, 2, 3...N) increases with n.

For $R_t = R_{av}$, the rainfall events R_i do not have a trend, and the cumulative rainfall average would conform to R_{av} [16]:

$$\Delta h_i = \frac{r}{S}\left[\sum_{n=1}^{i} R_n - \left(2 - \frac{1}{R_{av} i}\sum_{n=1}^{i} R_n\right)\sum_{n=1}^{i} R_{av}\right] =$$

$$= \frac{r}{S}\left(\sum_{n=1}^{i} R_n - \sum_{n=1}^{i} R_{av}\right);\quad (n = 1,\ 2,\ 3\ldots N) \tag{5}$$

Eq. (5) is the CRD formula proposed by Bredenkamp et al. [18] with $k = 1$, i.e., there is no pumping and/or natural discharge:

$$_t^iCRD_i = \sum_{n=1}^{i} R_n - k\sum_{n=1}^{i} R_{av};$$

$$(n = 1,\ 2,\ 3\ldots N). \tag{6}$$

Let us consider the case of using $R_t > R_{av}$ for coping with the net effect of pumping and/or natural outflow Q_{outi} in Eq. (3), as the actual *CRD* analysis has shown an increasing water level fluctuation tendency with the use of $R_t = R_{av}$.

$$_t^1CRD_i(R_{av}) - {}_t^1CRD_i(R_t) =$$

$$= \left[\sum_{n=1}^{i} R_n - \left(2 - \frac{1}{R_{av} i}\sum_{n=1}^{i} R_n\right)\sum_{n=1}^{i} R_{av}\right] -$$

$$- \left[\sum_{n=1}^{i} R_n - \left(2 - \frac{1}{R_{av} i}\sum_{n=1}^{i} R_n\right)\sum_{n=1}^{i} R_t\right] =$$

$$= \left(2 - \frac{1}{R_{av} i}\sum_{n=1}^{i} R_n\right)\left(\sum_{n=1}^{i} R_t - \sum_{n=1}^{i} R_{av}\right);$$

$$(n = 1,\ 2,\ 3\ldots N) \tag{7}$$

where: $_t^1CRD_i(R_{av})$, $_t^1CRD_i(R_t)$ are *CRD* values determined with $R_t = R_{av}$ and $R_t > R_{av}$, respectively.

Therefore, the groundwater level drawdown (Δs_i) due to the net effect of pumping and/or natural discharge is:

$$\Delta s_i = \frac{r}{S}\left(2 - \frac{1}{R_{av} i}\sum_{n=1}^{i} R_n\right)\left(\sum_{n=1}^{i} R_t - \sum_{n=1}^{i} R_{av}\right);$$

$$(n = 1,\ 2,\ 3\ldots N). \tag{8}$$

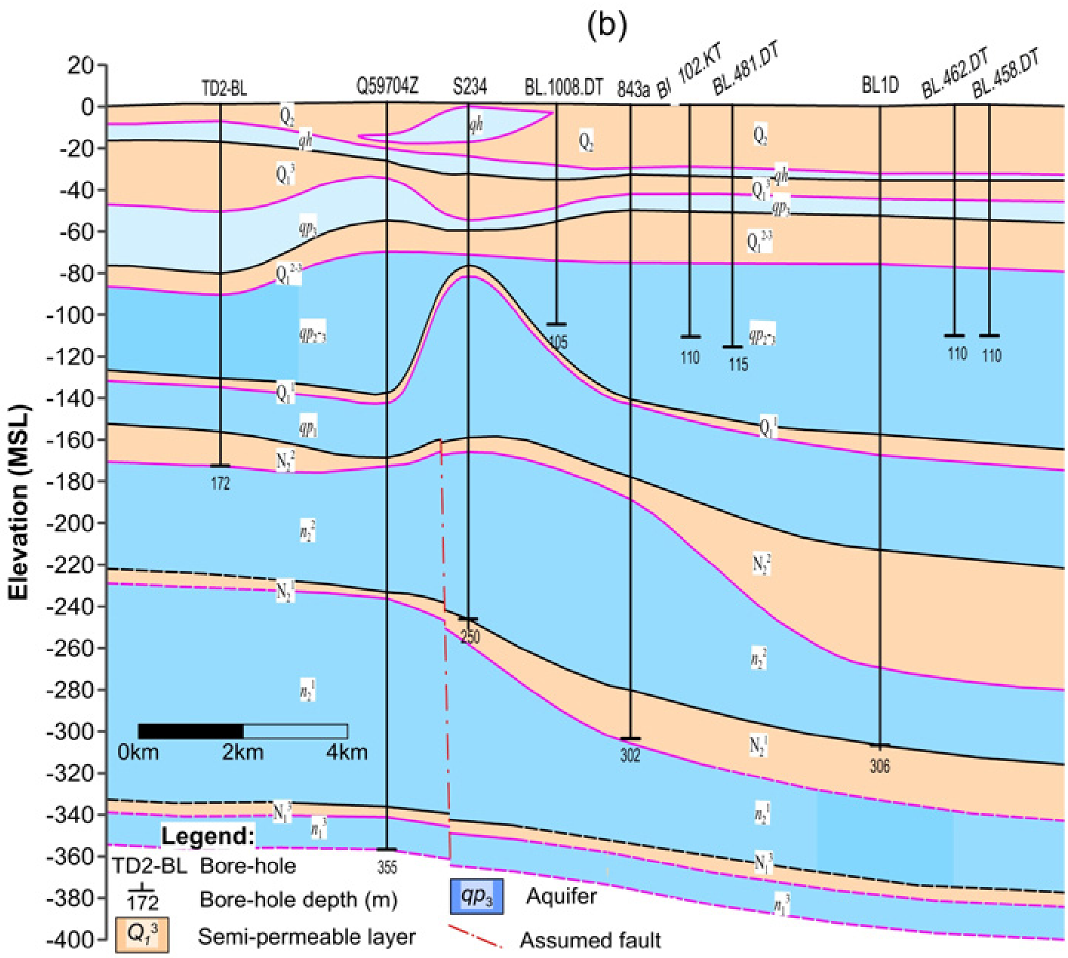
Application of CRD method to Bac Lieu province, South Vietnam

Hydrological conditions of the study area

The study area is Bac Lieu province with an area of 2669 km^2, located in the southernmost region of Vietnam in the Vietnam Mekong Delta, the most downstream area of the Mekong River basin Delta (Fig. 1 a). The Vietnam Mekong Delta region includes Can Tho city and 12 provinces (Long An, Tien Giang, Ben Tre, Vinh Long, Tra Vinh, Hau Giang, Soc Trang, Dong Thap, An Giang, Kien Giang, Bac Lieu and Ca Mau). The Mekong Delta has a total area of 40816 km^2 and a total population of nearly 18 million [25].

The hydrogeological conditions of the Mekong Delta in general and Bac Lieu province in particular, are characterized by the following 7 aquifers and 7 semi-permeable layers (Fig. 1 b) [26].

Fig. 1. Map of Southern Vietnam showing the study area (Bac Lieu province) within the Mekong delta including groundwater monitoring wells and hydrogeological cross-section AB location illustrated (at the top). Hydrogeological cross-section along line AB (below). The bore-holes (wells), their ID and depth indicated as well as semi-permeable sediment layer (quaternary, yellow colored), aquifers (blue colored) and fault position provided.



Holocene porous aquifer (qh)

The Holocene aquifer (*qh*) consists of layers of ash-gray, blue-gray, yellow-gray fine sands to medium sands with grits or gravels, in some places interbedded with layers of silty sands and silt. The Holocene aquifer is exposed to the ground surface in some places but is mostly underlain by the Holocene semi-pervious layer (Q_2) and lying above the upper Pleistocene layer (Q_1^3). The Holocene aquifer thickness varies from 4 m to 32 m, on average 18 m.

Aquifer storativity (*S*) which is equal to the multiplication of the specific storativity of the aquifer material (S_s) and aquifer thickness is an important required parameter in the determination of the rainwater recharge in the aquifer. The aquifer-specific storativity is determined by the water unit weight, water, and aquifer material compressibilities, and total porosity. The water at 25 °C has a compressibility of 4.6×10^{-10} m^2/N. For fine sands to medium sands with grits or gravels, the compressibility would be 1.3×10^{-8} m^2/N [27]. With the porosity of the Holocene and Pleistocene aquifers of around 0.25 [26], the specific storativity of the aquifers' materials is around 0.00013/m. Therefore, the aquifer storativity is 0.00234.

There is only a groundwater monitoring well (Q17701T) in the *qh* aquifer which is located near the study province Bac Lieu in the west of province Ca Mau (Fig. 1).

Upper Pleistocene porous aquifer (qp_3)

The Upper Pleistocene aquifer (qp_3) consists of layers of ash-gray, blue-gray, yellow-gray fine sands to medium sands with grits or gravels, in some places interbedded with layers of gray, blue-gray, gray-yellow, light gray silty sands, silt, and silty clay. The Upper Pleistocene aquifer is underlain by the upper Pleistocene semi-pervious layer (Q_1^3) and is lying above the Middle–Upper Pleistocene semi-pervious layer (Q_1^{2-3}). The qp_3 aquifer thickness varies very much from 2 m to 65 m, on average 17 m. Similarly to aquifer *qh*, the aquifer storativity is 0.00221.

The groundwater monitoring well (Q5970) in the qp_3, qp_{2-3}, and qp_1 aquifers is located in the east of province Bac Lieu in the west of Ca Mau province (Fig. 1).

Upper Middle–Upper Pleistocene porous aquifer (qp_{2-3})

The Middle–Upper Pleistocene aquifer (qp_{2-3}) consists of layers of ash-gray, blue-gray, yellow-gray fine sands to medium sands with grits or gravels, in some places interbedded with layers of gray, blue-gray, gray-yellow, light gray silty sands, silt, and silty clay. The Upper Pleistocene aquifer is underlain by the Upper Pleistocene (Long Toan formation) semi-pervious layer (Q_1^{2-3}) and is lying above the Lower Pleistocene (Ca Mau formation) semi-pervious layer (Q_1^1). The qp_{2-3} aquifer thickness varies very much from 5 m to 100 m, on average 46 m. Similarly to aquifers *qh* and qp_3, the aquifer qp_{2-3} storativity is 0.00598. The groundwater monitoring well in the qp_{2-3} aquifer is Q5970 (Fig. 1).

Lower Pleistocene porous aquifer (qp_1)

The Lower Pleistocene aquifer (qp_1) consists of layers of dark-gray, ash-gray fine sands to medium-coarse sands, in some places interbedded with layers of gray, dark-gray silty sands, and silt. The Upper Pleistocene aquifer is underlain by the Lower Pleistocene (Ca Mau formation – Q_1^1cm) semi-pervious layer (Q_1^1) and is lying above the Middle Pliocene (Nam Can Formation – N_2^2nc) semi-pervious layer (N_2^{2-}). The qp_1 aquifer thickness varies from 25 m to 53 m, on average 37 m. Similarly to aquifers *qh* and qp_3 and $qp_{2-3,}$ the aquifer qp_1 storativity is 0.00481. The groundwater monitoring well in the qp_1 aquifer is Q5970 (Fig. 1).

Middle Pliocene porous aquifer (n_2^2)

The Middle Pliocene aquifer (n_2^2) consists of layers of blue-gray, light yellow fine sands to coarse sands, interbedded with layers of silty sands, and clayey sands. The Middle Pliocene aquifer is underlain by the Middle Pliocene semi-pervious layer (N_2^2) and is lying above the Lower Pliocene (Can Tho Formation – N_2^1ct) semi-pervious layer (N_2^1). The n_2^2 aquifer thickness varies from 18 m to 123 m, on average 78 m. Similarly, to the above aquifers, the aquifer n_2^2 storativity is 0.01014.

Lower Pliocene porous aquifer (n_2^1)

The Lower Pliocene aquifer (n_2^1) consists of layers of ash-gray, blue-gray, light gray fine sands to coarse sands mixed with some clay, in some places with gravels, interbedded with layers of gray, ash-gray silty sands, silts. The Lower Pliocene aquifer is underlain by the Lower Pliocene semi-pervious layer (N_2^1) and is lying above the Upper Miocene (Phung Hiep Formation – N_1^3ph) semi-pervious layer (N_1^3). The n_2^1 aquifer thickness varies from 33 m to 52 m, on average 43 m. Similarly, to the above aquifers, the aquifer n_2^2 storativity estimate is 0.00559.

Upper Miocene porous aquifer (n_1^3)

The Upper Miocene aquifer (n_1^3) consists of layers of fine sands to coarse sands, in some places interbedded with layers of silts. The Upper Miocene aquifer is underlain by the Upper Miocene semi-pervious layer (N_1^3). The n_1^3 aquifer thickness varies from 15 m to 122 m, on average 51 m. The aquifer n_2^2 storativity estimate is 0.01586.

Semi-pervious layers

The semi-pervious layers that have been mentioned above have the lithological contents and the thickness as follows.

– The semi-pervious *Holocene layer* Q_2 has a thickness from 11 m to 25 m (20 m on average) and consists of ash-gray, dark-gray, yellow clayey silts, and silts interbedded with lenses of fine sands.

– The semi-pervious *Upper Pleistocene layer* Q_1^3 has a thickness from 5 m to 43 m (16 m on average) and consists of ash-gray, dark-gray, blue-gray silty clay, silts, silty sands, in some places interbedded with lenses of fine sands.

– The semi-pervious *Upper-Middle Pleistocene* layer $Q_1^{2–3}$ has a thickness from 7 m to 65 m (26 m on average) and consists of ash-gray, blue-gray laterite clay, silts, and clayey silts.

– The semi-pervious *Lower Pleistocene layer* Q_1^1 has a thickness from 3 m to 60 m (15 m on average) and consists of blue-gray, dark-gray, red-gray laterite clay, silty clay, silts, in some places interbedded with fine sands mixed with many silts.

– The semi-pervious *Middle Pliocene layer* N_2^2 has a thickness from 3 m to 28 m (11 m on average) and consists of blue-gray, light-gray, and red-gray clay.

– The semi-pervious *Lower Pliocene layer* N_2^1 has a thickness from 3 m to 20 m (13 m on average) and consists of blue-gray, light-gray, dark-gray, yellow-gray clay, silty clay, clayey silts with laterite grits.

– The semi-pervious *Upper Miocene layer* N_1^3 consists of mainly tight silts and clay. Its thickness is not well determined as there is only one bore-hole drilled through the layer in 14 m of length.

The semi-pervious layers have low hydraulic conductivity from $10^{–7}$ m/day to $10^{–4}$ m/day [26].

The above-described aquifers and aquitards are all distributed within Vietnam territory, under the East Sea, and expanded to the Kingdom of Cambodia and the Lao People's Democratic Republic. The distributed recharge to the aquifers is exclusively from the rainfall: the rainwater infiltrates to recharge the Holocene aquifer *qh*, which in turn recharges the Upper aquifer qp_3 and so on [26]. The water level elevations are in the lowering sequence from the topmost aquifer (Holocene aquifer *qh*) to the lowest Quaternary aquifer (Lower Pleistocene porous aquifer qp_1) as shown in Fig. 2.

On the capability of the CRD method of dealing with deep vertical inter-aquifer flow aquifers in the study area

The groundwater level of the aquifer qp_3 and $qp_{2–3}$ have been being monitored since 1995, during which very insignificant groundwater abstraction took place in the context of the country's extremely low economic activity. The quality of the groundwater level monitoring is not so good during the first monitoring year. Analyzing the monitored data, reasonable and seemingly good monitoring data in the monitoring well Q5970 (shown in Fig. 1) is from the 27th of February 1996 to the 27th of March 1997 [28] (Fig. 2 b). The Pearson correlation analysis shows a high relationship (the correlation coefficient is 0.930) between the daily water level of aquifers qp_3 and

$qp_{2–3}$ (Fig. 3). This would allow us to conclude that the increase of qp_3 water level causes vertical groundwater flow from qp_3 to $qp_{2–3}$ through the semi-pervious layer $Q_1^{2–3}$. The component of the vertical flow rate which recharges aquifer $qp_{2–3}$ may be estimated using Darcy law and that vertical flow rate may be calibrated with the assumption that the rate is determined by the difference in water level of aquifers qp_3 and $qp_{2–3}$, the thickness and the hydraulic conductivity of the semi-pervious layer $Q_1^{2–3}$.

Therefore, the fluctuations of the water levels of the aquifers in the multi-aquifer system in the study area are synchronized with each other. This would also be supported by the high Pearson correlations between CRD simulated GW levels and observed GW levels for all aquifers under consideration.

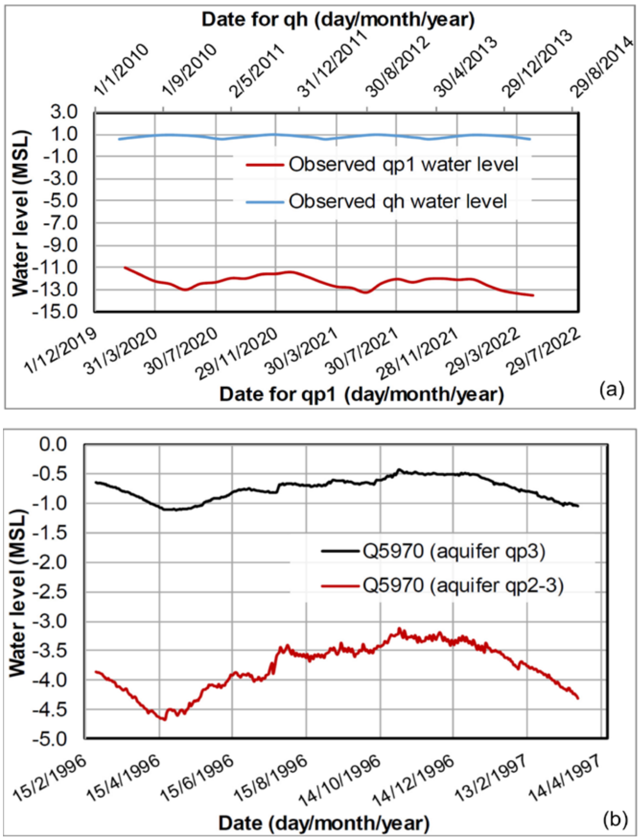


Fig. 2. Groundwater levels time series monitoring data: (a) qh (monitoring well Q17701T) and qp_1 (monitoring well Q5970) during 01.12.2019-29.07.2022; (b) qp_3 and $qp_{2–3}$ from 27.02.1996 to 27.03.1997 in monitoring well Q5970. On the (a) and (b) variations of groundwater levels observed, fluctuated seasonally and annually. Clearly exposed by monitoring data that even during dry season's groundwater level stay still keep enough high values. This is optimistic observed phenomena making rainfall method helpful to use data for groundwater management in the area.

Let us now consider the relationship between rainfall and groundwater level of the uppermost aquifer *qh*. The aquifer *qh* receives recharge from the rainfall via the rainwater infiltration neither through the unsaturated or saturated zone above the uppermost aquifer. The water level fluctuation of the uppermost aquifer is dependent upon that rainwater infiltration rate. In our case, the rainwater infiltration takes place in the Holocene semi-pervious silty layer Q_2 of thickness of 20 m on average. The lack time of for the water to respond to the rainfall is the time the wetting front reaches the semi-pervious layer bottom. Roi [24] showed that the lag time would be 5.5 days for the Holocene silty unsaturated zone thickness of 3 m in Hung Yen province (North of Vietnam), based on the assumption that the lag time is linearly proportional to the thickness of a semi-pervious layer, the lack time of the water level response of the Holocene aquifer *qh* to the rainfall in this study area would be around one month. Yet, the CRD method is also capable of analyzing the lag time through the time series of CRD simulated water level fluctuations and the time series of monitored groundwater levels.

The lag time of one month is initially used in the CRD model: the data series of the monthly rainfalls and the groundwater levels at the month's end are used.

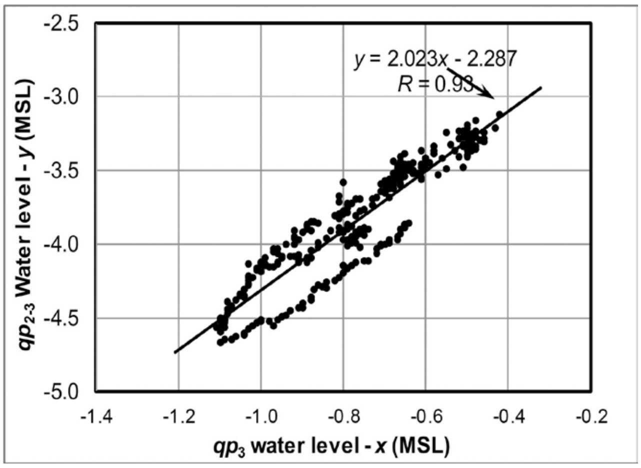


Fig. 3. Correlation between water level of qp_3 and $qp_{2–3}$ in monitoring well Q5970. Here and on the Figures 4–7, *R* indicates the correlation coefficient. Good correlation of water levels in the well on different depth reflects high permeable and dynamic groundwater systems entirely reflecting variations under the certain local and regional factors.

An illustration with long-term semi-hypothetical repeated rainfall and GW level

The data series of the monthly rainfalls and the groundwater levels of the aquifer qp_3 at the months' ends from the 31st Mar. 1995 to the 29th Feb. 1996 are used for the next 7-year duration (Fig. 4 a). The total rainfall during the 12 months from the 31st Mar. 1995 to the 29th Feb. 1996 is 1.989 m. With the storage coefficient of the aquifer equal to 0.003, the fraction of the rainfall that recharged the aquifer is equal to 0.12 %. Therefore, the aquifer qp_3 annual recharge from the rainfall is 0.0024 m. The Pearson correlation coefficients for $R_t = R_{av}$ and $R_t = 1.023R_{av}$ are 0.965 and 0.952, respectively (Fig. 4 b). The observed and CRD simulated water level fluctuations are presented in Fig. 4 a along with the month rainfalls. It seems that the lag time of one month is appropriate for the aquifer water level. The CRD simulated monthly water level (WL) fluctuations for $R_t = 1.023R_{av}$ seem to be of a 12-month cycle.

The application results

As it was mentioned before, the groundwater level monitoring has been not so well implemented, i.e., the monitoring wells in the aquifers had not been installed at the same time, and the monitoring data are most likely not corrected for some periods. Therefore, some separated periods for each of the aquifers have been selected for the analysis. Each couple of figures, Figures 5–7

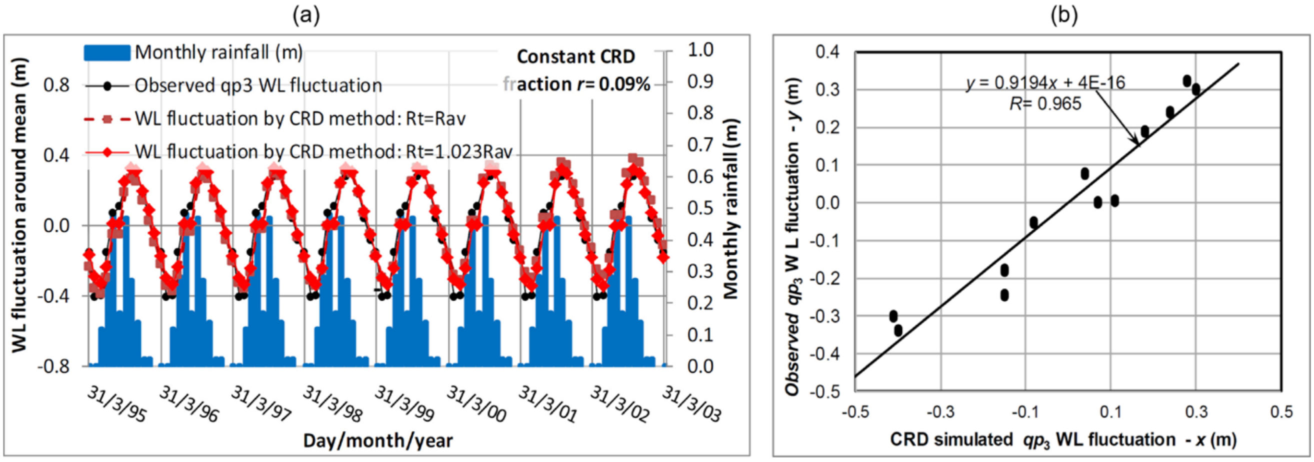


Fig. 4. Observed and CRD simulated WL fluctuations of qp_3 in monitoring well Q5970 for long-term semi-hypothetical repeated rainfall (a) and the correlation between them (b). This figure clearly illustrates approval of CRD method verified by experimental observations and simulated computing during 31.03.1995 – 31.03.2003 confirmed by well correlated values for demonstrated sites of *qp*3.

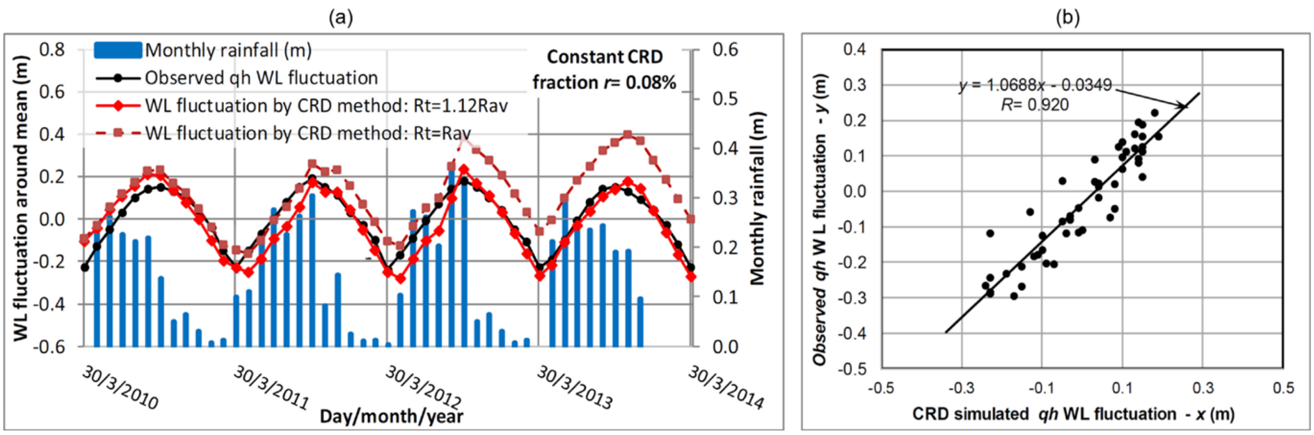


Fig. 5. Observed and CRD simulated WL fluctuations of *qh* during Mar. 2010 – Mar. 2014 (a) and the correlation between them (b). This figure approves by the CRD method obtained data verified by experimental observations and simulated computing confirmed by well correlated values for demonstrated sites of *qh* during 30.03.2010 – 30.03.2014.

present the CRD simulated monthly water level fluctuations and Pearson correlation between observed and simulated water level fluctuations for *qh*, qp_3, qp_{2-3} and qp_1 aquifers, respectively. The Pearson correlation coefficient is relatively high, from 0.894 to 0.925. Table summarizes the analysis results.

Meanwhile, the present GW abstraction from the Quaternary aquifers is 236 000 m^3/day, which is much higher than the net recharge from the rainfall.

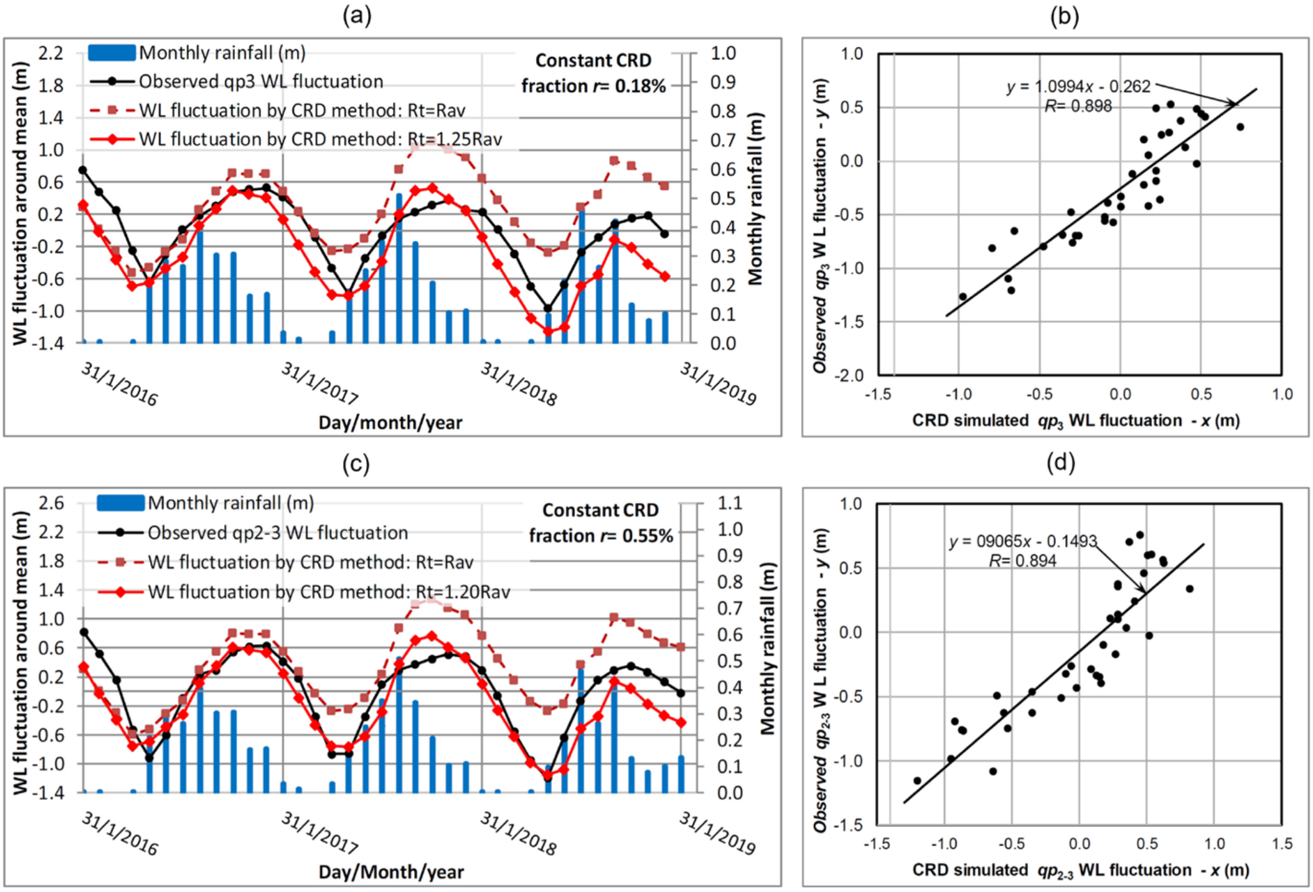


Fig. 6. Observed and CRD simulated WL fluctuations during Jan. 2016 – Jan. 2019: qp_3 (a) and the correlation between them (b); qp_{2-3} (c) and the correlation between them (d). This figure clearly illustrates approval of CRD method verified by experimental observations and simulated computing confirmed by well correlated values for demonstrated sites of qp_3 and qp_{2-3} during 31.01.2016 – 31.01.2019.

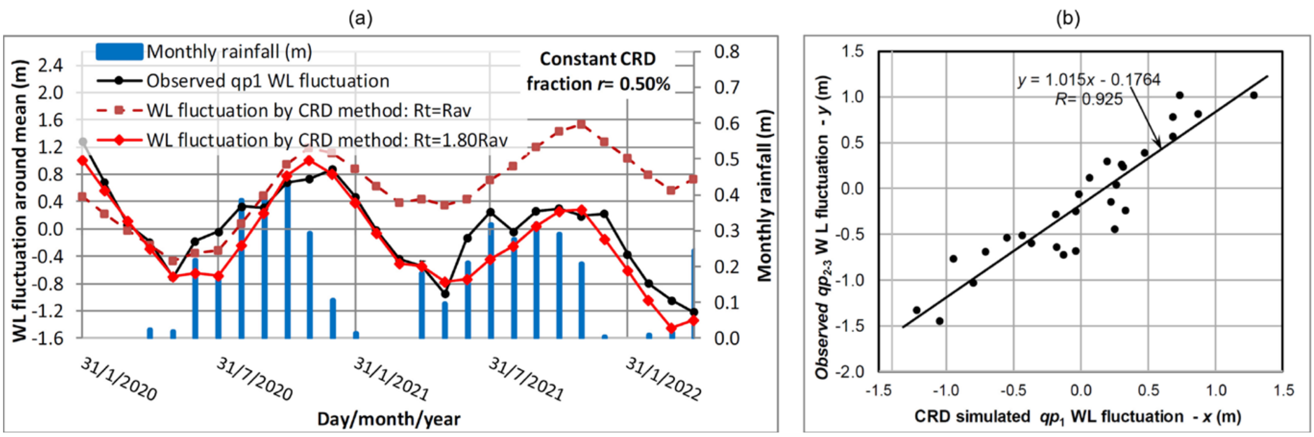


Fig. 7. Observed and CRD simulated WL fluctuations of qp_1 during Jan. 2020 – Jan. 2022 (a) and the correlation between them (b). This figure illustrates approval of CRD method verified by experimental observations and simulated computing confirmed by well correlated values for demonstrated sites of qp_1 during 31.01.2020 – 31.01.2022.

Table. Determined not to recharge values by the CRD method

Aquifer	R_t	Pearson correlation coefficient (R)	Annual rainfall (mm)	Percentage of rainfall (%)	Annual net recharge (mm)
qh	1.12	0.920	2207	0.08	1.77
qp_3	1.25	0.898	2000	0.18	3.60
qp_{2-3}	1.20	0.894	2056	0.55	11.31
qp_1	1.80	0.925	2216	0.50	11.08
			Average:	Total	
			2120	1.31	27.75

Note. From the whole Bac Lieu province: with area of 2669 km^2, annual net recharge is 74.07´10^6 m^3/year, or 203 000 m^3/day.

Discussions

The CRD method's authors suggested that for an open aquifer system, the R_t is equal to the average rainfall R_{av}. The present study shows that R_t is equal to 1.12 R_{av} for the shallowest aquifer *qh* from which there is no GW exploitation, (1.20–1.25*)* R_{av} for aquifers qp_3 and qp_{2-3}, and to a very high value (R_t = 1.80R_{av}) for aquifer qp_1. This issue would be significant matter to be further investigated.

Finally, the fractions of cumulative rainfall departure for aquifers *qh*, qp_3, $qp_{2-3,}$ and qp_1 are 0.08 %, 0.18 %, 0.55 %, and 0.50 %, respectively, which is equal to 1.31 % of the rainfall. The fractions are rather small in comparison to 34.1 % – 38.1 % obtained by Roi [24] for Hung Yen province, Northern Vietnam. The total fraction of 1.31 % would provide the total GW annual recharge of 74.07 million m^3 from the rainfall over the whole Bac Lieu province, equivalent to 203 000 m^3/day. The current GW abstraction of 236 000 m^3/day in the province is 16 % higher than the estimated GW recharge from the rainfall, which needs to be paid attention to in the water planning in general and reconsideration of the GW exploitation plan in particular for Bac Lieu province.

Comparable research should be performed between the Southern (Mekong River basin) and Northern (Red River basin) Vietnam areas to get clear pattern of groundwater recharge not only by rainfall but also by underlying deep water saturated strata also influenced by hydrogeological geothermal processes.

Summary

The CRD method originally developed for arid and semi-arid Southern Africa has been proven to apply to Northern Vietnam [24] and Southern Vietnam (the present study) with tropical climates. The applicability of the CRD method for deep aquifers would be judged through the Darcy law describing a vertical inter-aquifer flow which is determined by the difference in water level of upper and lower aquifers with an in-between semi-pervious layer having a certain hydraulic conductivity and the results of the present study.

The present study shows that the CRD simulated GW levels of all four Quaternary aquifers in the study area have a tight correlation with the monthly rainfall, which means that the preceding monthly rainfall results in the water level fluctuation at the end of the month.

For the further research authors will continuing these studies using additionally ecological and landscape methods, and performing hydrochemical, microbiological and gasgeochemical and geophysical research in the study areas, which will allow us to specify the hydrogeological situation and clarify the dynamics of groundwater resources and their quality during seasonal and annual fluctuations.

Credit authorship contribution statement. All authors contributed to the study conception and design, material preparation, data collection, and analysis. Trinh Hoai Thu: Material preparation, Data collection, Writing-review & editing. Renat Shakirov: Methodology, Writing-original draft, and final version.

Natalia Lee & Elena Maltceva: Formal analysis, Methodology, Writing-original draft. Nguyen Van Hoang: Investigation, Methodology, Writing-review & editing. Tran Thi Thuy Huong: Investigation, Methodology, Visualization. Nguyen The Chuyen: Investigation, Data collection, Visualization. Anna Venikova: Formal analysis.

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Data availability: Data will be made available on request.

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