Chapter

Characteristics and Assessment of Groundwater

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Abstract

Groundwater system is very vital to humanity and the ecosystem. Aquifers are determined based on the absence or presence of water table positioning, that is, confined, unconfined, leaky aquifers and fractured aquifers. The objective of this chapter is to discuss the characteristic and assessment of groundwater within the scope of vertical distribution of GW, types of the aquifer system, types of SW-GW interface, and SW-GW interaction at both local and regional scales. The properties of the aquifer depend on the physical characteristics of the materials (porosity, permeability, specific yield, specific storage, and hydraulic conductivities) which are determined by techniques like resistivity surveys and pumping tests followed by remote sensing and geographic information system for better information on the groundwater system. Furthermore, understanding the SW-GW interactions through available methods (seepage meter, heat tracer, and environmental tracer) is useful in watershed management, that is, risk management and assessment of the aquifer system.

Keywords: aquifer characteristics, GW distribution, SW-GW interaction, SW-GW methods, resistivity survey, pumping test, RS, GIS

1. Introduction

Groundwater (GW) belongs to all subsurface water, including saturated and unsaturated zones. More than 1.5 billion inhabitants around the globe depend on the groundwater for agriculture usage and industrialization consumption. However, pollutions were identified as one of the major challenges in hampering GW withdrawal (**Figure 1**) [1]. The exchange of the chemical and physical characteristics in water will affect the quality of groundwater resources, hence leading to the availability of humans in terms of quantity [2].

Groundwater is deposited between the pore spaces of rock/soils, cracks, joints, and fractures and various geological formations. The movement of groundwater in soils and rocks depends on the hydraulic characteristics of the shape and size of void spaces. Water can flow easily through certain rocks through the soil into the underground aquifer system, but water typically penetrates through fractures, cracks, and some other geological formations. Generally, there are three distinct types of geological formations of groundwater that determine the availability of groundwater resource, namely, aquifers, aquitard, and aquiclude.

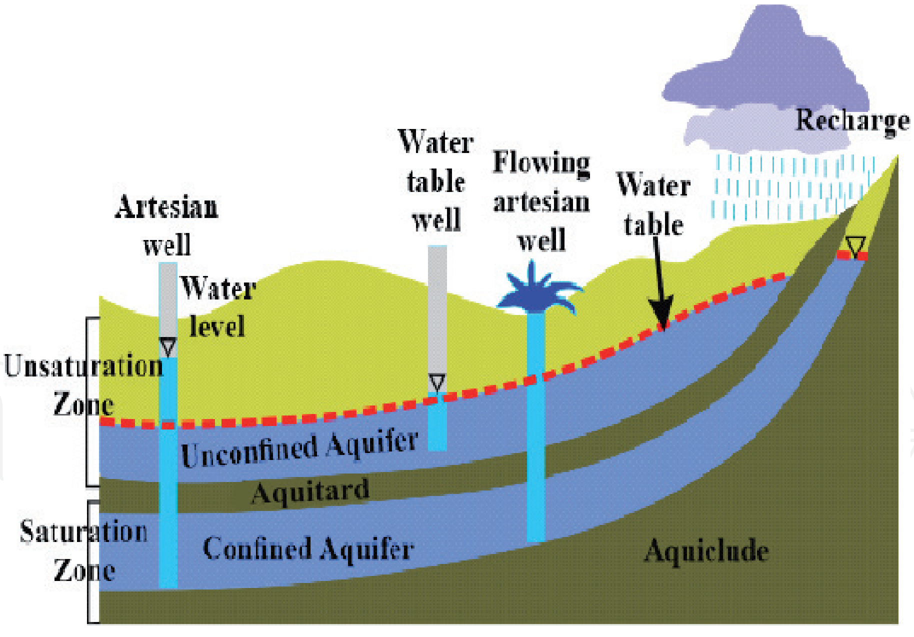

Figure 1.
The figure is showing unconfined aquifer and confined aquifer.

An aquifer is a highly permeable or porous saturated formation (conglomerate, sandstone, limestone, unconsolidated sand, gravels, fractured limestone, fractured basalt, etc.) that not only stores water but also provides adequate amounts of water and thus is considered as significant groundwater resources. The aquitard is a partially saturated formation (shale or clay) that allows water through it but does not provide enough available water than the aquifer. An aquiclude is an impermeable layer (clay) produces a considerable volume of water because of its high porosity but does not provide significant amount of water.

Groundwater passes from recharge zones to discharge zones along flow routes of variable lengths and comes in contact with surface water (SW) essentially at low elevated areas [3]. Surface water resources mostly depend on regional precipitation/rainfall and it may be lost by infiltration through the streambed, layer of soil-moisture, and cracks or fractures to interacts with the groundwater system and the area of mixing of both known as the hyporheic zone [4]. The interaction of surface water and groundwater takes several forms in which if surface water moves toward the groundwater system, it is referred to as a losing stream whiles the other way round is called gaining stream [5].

The surface water (SW)-groundwater (GW) interactions in the hyporheic zone take place within the close-streambed sediments at few scales, which depend on the hydraulic-potential strengths and bed geometry [6]. In SW-GW interactions and hyporheic exchange, earlier studies used three types of scales such as sediment scale (<1 m), local scale (1–1 km), and catchment scale (>1 km) [7]. However, Todd and Mays [8] classified only two scales such as local scale and regional scale which are associated with small watershed and large watershed, respectively. Here, SW-GW interaction is associated with the direction of streamflow, shallow GW aquifer property, and local GW flow system.

The scope of this chapter is to discuss the vertical distribution of GW, types of the aquifer system, types of SW-GW interface, and SW-GW interaction at both local and regional scales. This chapter has been divided into four sections; (i) groundwater distribution and aquifer characteristics, (ii) the SW-GW interactions at local and regional scale, (iii) types of SW-GW interface, and (iv) the methods for investigation of SW-GW interactions and aquifer system.

2. Groundwater distribution and aquifer characteristics

The distribution of groundwater is classified into two zones based on the water table, namely, unsaturation zone and saturation zone. The aquifer system is mainly divided into confined and unconfined aquifers, and its characteristics depend upon the main physical parameters such as porosity, permeability, transmissivity, specific yield, specific storage, and hydraulic conductivity.

2.1 Vertical distribution of groundwater

The groundwater occurrence is typically categorized into two major zones based on the water table namely unsaturation zone and saturation zone. The zone of unsaturation is also known as the zone of aeration (vadose zone), which is also subclassified into the soil moisture zone, intermediate vadose zone, and capillary zone. The unsaturation zone is comprised of interstices or void spaces that are partially filled with water and air. All interstices are fully saturated with water under hydrostatic pressure in the saturated zone under the water table.

2.1.1 Soil-moisture zone

The soil-moisture zone occurs across the main root zone beneath the earth's surface, but its thickness varies with the types of soil and vegetation. This zone plays a significant role in the recognition of hydrological processes [9] and also important for the interaction of the land-surface atmosphere [10]. The practices of agriculture and irrigation, particularly in arid and semiarid areas, primarily depend on the timely characterization of spatial and temporal soil moisture fluctuations in the root zone as a consequence of the soil moisture effect on health status and production of crops and salinization [11]. Several environmental factors such as physicochemical characteristics of water, surface slope and roughness, soil hydraulic conductivity, the porosity of the soil, and pre-existing soil pore moisture content are controlling the soil matrix's capacity to transfer water of which affecting the infiltration process [12].

2.1.2 Intermediate vadose zone

The intermediate vadose zone is located beneath the soil moisture zone and upper part of the capillary zone. Water that drops into this zone can be either drawn into the capillary interstices of the transition area through the molecular attraction or drawn downwards to the adjacent saturated zone.

2.1.3 Capillary zone

The zone is the lowest part of the aeration zone and directly above the water table where water as a component of the capillary action can be drawn back toward it. For a capillary zone of clay with a 0.0005 mm porous radius, the typical height may be 3 m, contrasted with fine sand of less than 10 cm with a 0.02 mm porous radius. Capillary water is the water stored above a surface of the water table in the capillary openings of unsaturated or saturated substances.

2.1.4 Saturation zone

The saturation zone is above the water table which is often referred to as the phreatic zone or aquifer system. Water that has profoundly infiltrated through the vadose zone enters the saturation zone and filled all pore spaces with water.

The thickness of the saturation zone varies from several meters beneath the earth's surface to numerous hundred meters. The factors to determine the thickness of this zone depend upon the local geology, accessibility of openings or pores in the rock formation, and water flow within the zone from recharging to discharge points. This saturation can take place range from several days or weeks to many months in duration. Moreover, groundwater is controlled by quantity and rainfall intensity, temperature, rock porosity, and permeability, dryness of the air, vaporization intensity during the rainy season, land slope, vegetative covering, and water absorption ability for soil. As well, significant volume of water can be contained within fractures and joints structures. The following are typical opening types contained in rock: (1) openings in gravel and sandstone formations with individual particles; (2) vugs, caverns, and solution channels in dolomite and limestone rock; and (3) joints, crevices, gas holes and faults in metamorphic rocks and igneous formations.

2.2 Types of aquifers

Aquifers are generally categorized into two major classifications, confined and unconfined aquifers; leaky and fractured aquifers are sometimes addressed in some other aquifers (**Figure 1**).

An unconfined aquifer is a layer of water-bearing formations or rocks that do not have a confining bed at the top of the groundwater which is referred to as the groundwater table where the pressure becomes equivalent to the atmospheric pressure. The variation of groundwater levels varies and depends on the pumping from the wells, permeability, area of recharge and discharge, in effect impacting the increasing or declining water rates in wells that are extracted from aquifers. The water table is free to rise or to fall which is often called the free or phreatic surface. Contour graphs and water table profiles of wells that use the water to determine water quantities available as well as water distributing, and movement may be prepared from elevations of wells. The perching water sources, as shown in (**Figure 1**) are a case of unconfined aquifers. Their high susceptibility to contamination is a major problem with non-confined aquifers. If something dumps on the surface, it will penetrate vertically and go down into the storage of groundwater.

2.2.1 Confined aquifer

The definition of the confined aquifer as "a formation in which the groundwater is isolated from the atmosphere at the point of discharge by impermeable geologic formations; confined groundwater is generally subject to a pressure greater than atmospheric" [13]. It is also known as "artesian or pressure aquifers" and it occurs mostly just above the base of confined rock bodies or layers which is mostly composed of clay that can protect it from surface pollution. Punctured wells from artesian aquifers are more prone to fluctuate with their depth of water because of changes in pressure than the amount in stored water. When such an aquifer is well penetrated, the water level should increase over the base of the confined layer, as illustrated by the flowing and artesian wells of (**Figure 1**). The water reaches a confined aquifer in a region in which the confining layer reaches the surface. The groundwater flow system into aquifers is frequently affected by gravity and geological formations in such areas either vertically or horizontally. A zone that provides water to a restricted area is considered a recharge area and water may even be leak into a restricted bed. Water ups and downs in confined aquifers penetrating wells mainly result from pressure changes instead of storage volume changes. Confined aquifers thus show only limited variations in storage and are predominantly used as conduits to move water to natural or artificial discharges from recharge areas.

2.2.2 Leaky aquifer

Aquifers that are fully unconfined or confined appear less often than aquifers that are leaky, or semi-confined. This is a common occurrence of plains, alluvial valleys, or former lake basins where a semi-pervious aquitard or semi-confining bed is underlain or overlain by a permeable layer. Pumping water from a well into a leaky aquifer eliminates water in two directions such as the vertical flow into the aquifer through the aquitard and the horizontal flow in the aquifer.

2.2.3 Fractured aquifer

The fractured rock aquifers vary from the subsurface water systems that are stored in the geological formation. Although sedimentary aquifers hold and move a significant amount of water between specific sedimentary granules through pore spaces, however, fractured rock aquifers hold and move water in an otherwise impermeable rock mass through as cracks, joints, and fractures (**Figure 2**).

Therefore, fractured rock aquifers have hydraulic characteristics that vary from those found in sedimentary aquifers with accessible water (common in terms to be described as bore yield) and are typically defined by nature (opening, size, and extent) and degree of interconnection between discontinuities in the rock mass. The long-term yield from well in fractured rock aquifers depends on the location of the degree of discontinuity and the relationship of discontinuities in the total mass of the rock instead of on the permeability of the geological substances near the extraction phase. The aquifers in fractured rock typically depend on the amount of precipitation that caused the surface water runoff of which considerably greater than in flat regions. Moreover, permeability fractured rock aquifers can also be dramatically decreased by the weight of the overlying rock mass as open spaces progressively decrease between fractures and cracks.

2.3 Characteristics of aquifer

Several properties that contribute to the identification and characterization of the aquifer are discussed.

2.3.1 Porosity

Porosity (n) is the intrinsic characteristic of a substance and refers to the amount of void or empty space in each material. The porosity (void space) occurs between

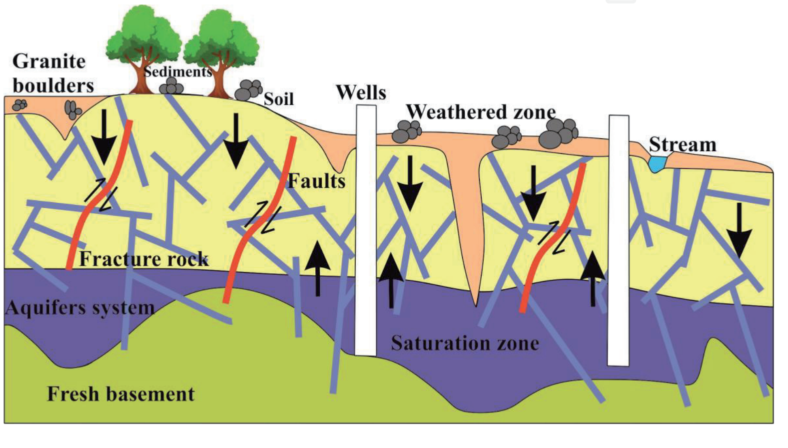

Figure 2.
The figure is showing the aquifer system in fractured rock formations.

the fragments of soil or rock. It is defined by the ratio between the volume of the void space and the volume of rocks/soils.

$$n = \frac{V_v}{V} * 100\% \tag{1}$$

where V_v is the volume of void space in a unit volume of earth material; and V is the unit volume of earth material (solids and voids).

2.3.2 Hydraulic conductivity and permeability

Permeability is defined as the ability of water movement through rock or soil which is directly related to porosity and it applies to the interconnected of pore spaces in rock or soil. Considering the relationship between driving and resisting forces on a microscopic scale during flow to porous media, hence, the permeability, k, is a function only of the area where the hydraulic conductivity K is defined:

$$k = \frac{K\mu}{\rho g} \tag{2}$$

where k is the permeability, K is the hydraulic conductivity, g is the acceleration due to gravity, ρ is the fluid density, and μ is the viscosity.

Hydraulic conductivity (K) is a physical characteristic that calculates the capacity of substance in the context of an applied hydraulic gradient to transfer water across the pore spaces and fractures of rock/soil [14]. It depends on various physical variables including porosity, the structure of the soil matrix, grain size distribution, type of soil fluid, particle arrangement, water contents, void ratio, and other factors [15, 16].

2.3.3 Transmissivity

The transmission (T) is the rate of discharge where the water is transferred under a hydraulic gradient over a unit width of an aquifer. It is calculated by a formula and expressed in m^2/s, or $m^3/day/m$ or l/day/m.

$$T = Kb(confinedaquifer) \tag{3}$$

$$T = Kh(unconfinedaquifer) \tag{4}$$

where K is the hydraulic conductivity, b is the aquifer thickness, and h is equivalent to the depth of confined aquifers.

2.3.4 Specific yield

Specific yield (S_y) as defined by Freeze and Cheery [14] is the storage term for unconfined aquifer where *the amount of water from the unconfined aquifer releases from the storage per unit surface area of aquifer per unit decline in the water table.* It is also known as unconfined storativity.

In other view, specific yield can be defined as the ratio of the volume of water that a saturated rock or soil will yield by gravity to the total volume of the rock or soil [15]. It is expressed in percentage.

$$S_y = \frac{V_w}{V} * 100\% \tag{5}$$

where V_w denotes the volume of water in a unit volume of earth materials; and V indicates the unit volume of earth material, including both voids and solids.

2.3.5 Specific storage

Specific storage (S_s) is the volume per unit amount of a saturated formation that is a deposit from the storage because of the compressibility of the mineral skeleton and the pore water per unit change in head. The specific storage is given by Jacob (1940) and is typically represented in cm^{-1} or m^{-1}.

$$S_s = \rho_\omega g(\alpha + n\beta) \tag{6}$$

where ρ_ω denotes water density, g is the acceleration of gravity, α shows compressibility of the aquifer skeleton, n indicate porosity, and β is the compressibility of water.

3. Groundwater and surface water interaction

Groundwater moves across flow paths arranged in space and develop a flow system. GW flow system is classified into local, intermediate, and regional flow systems (**Figure 3**) [17]. Water travels to the adjacent discharge area in a local flow system. One or more topographical low and high located between their discharge and recharge regions describe an intermediate flow system; however, contrary to the regional flux system, it does not occupy both the bottom of the basin and the major topographic high [18]. Water flows at a longer distance than the local flow system in a regional flow system and often discharge into large streams and lakes.

3.1 Characteristics of SW-GW interactions at the local scale

The range of groundwater at the local flow system is from 10 m to 10 km between the adjacent aquifers system and the stream reach. The recharge and discharge zones are associated with high and low areas respectively, associated with sub-watershed boundaries and local streams, respectively. The local GW flow system depends on the slope of topography and hydrogeology of the region (subsurface rock, streambed-sediment characteristics, and climatic conditions). At this scale, the seasonal effect on the hydrological response to recharge is high due to local flow systems, high water flux, and unsteady flow conditions. The fluctuation of the water table in local GW flow systems varies in different climatic conditions. For instance, the low water level in arid and semi-arid climate due to the low amount of precipitation and infiltration while the higher water level in a tropical environment is due to higher rainfall and infiltration. Therefore, SW-GW interaction is found more in a tropical and humid climate.

Generally, SW-GW can be introduced for homogeneous interaction of a stream and adjacent shallow aquifers system with hydrological processes, which is controlled by a SW-GW head and a streambed leakage coefficient. In hydrological processes, water moves with huge quantities of nutrients and streambed sediment and modifies the earth's surface through deposition and erosion. The hydrological processes give information about the drainage basin, small watershed, stream basin, evaporation, transpiration, evapotranspiration, runoff, and infiltrations rate (**Figure 4**). The hydrological exchange between GW and SW is through the downwelling and upwelling processes (**Figure 5**). Upwelling processes are those in which local GW flow moves toward SW and on the other hand, the situation is referred to as downwelling processes. During these processes, if the shape of the longitudinal streambed profile

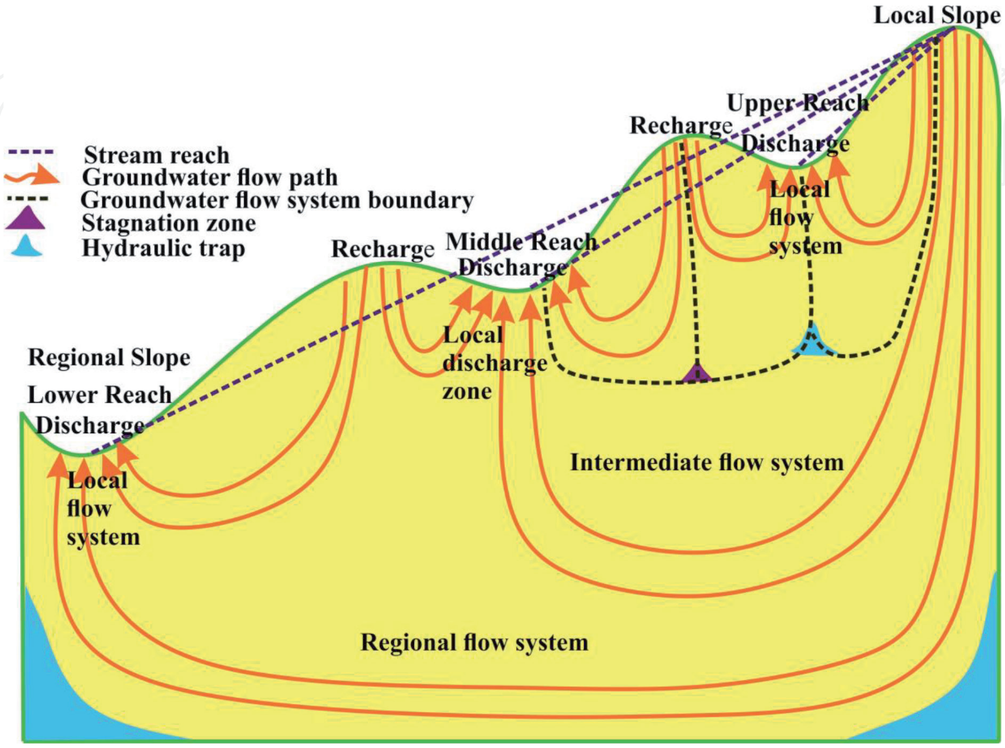

Figure 3.
GW flow system at the local and regional scale.

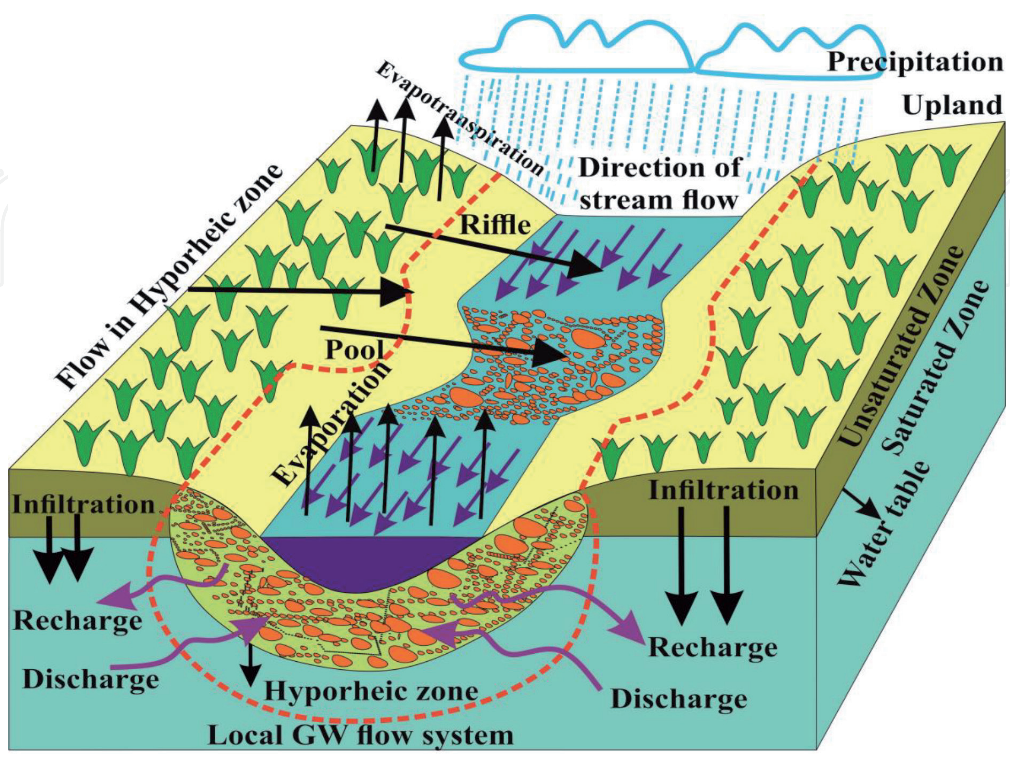

Figure 4.
River water and groundwater interactions in the hyporheic zone at the local scale.

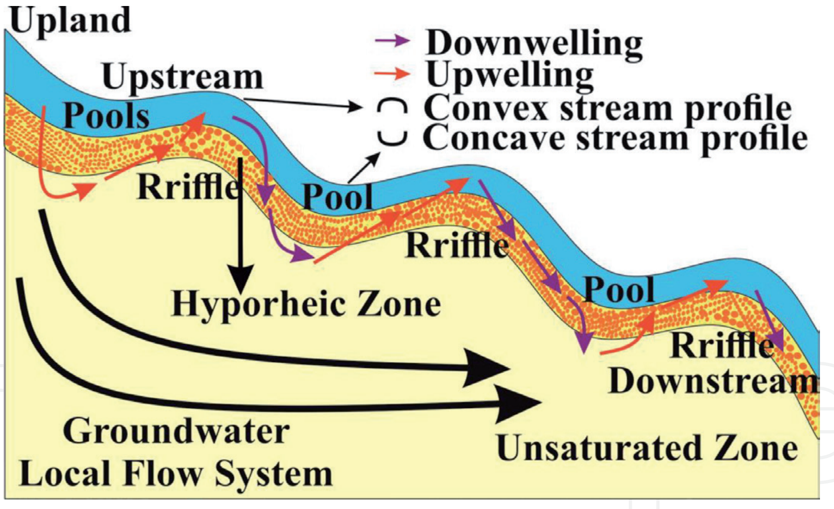

Figure 5.
Downwelling, upwelling, and hyporheic exchange processes.

is convex then the SW movement is through downwelling processes in the hyporheic zone whereas, if the shape of the longitudinal streambed profile of SW is concave then SW movement is through upwelling processes in the hyporheic zone [19]. The shapes of longitudinal streambed profiles are related to pool-riffle sequence and sediment bars, dunes, and ripples. The movement of stream water from riffles to pools is showing in (**Figure 5**) which is affected by the channel's sinuosity and bed load materials.

3.2 Characteristics of SW-GW interactions at the regional scale

The interaction of SW-GW is related to low topographical pathways varying from 10 to 100 km or more at a regional scale. The recharge or discharge

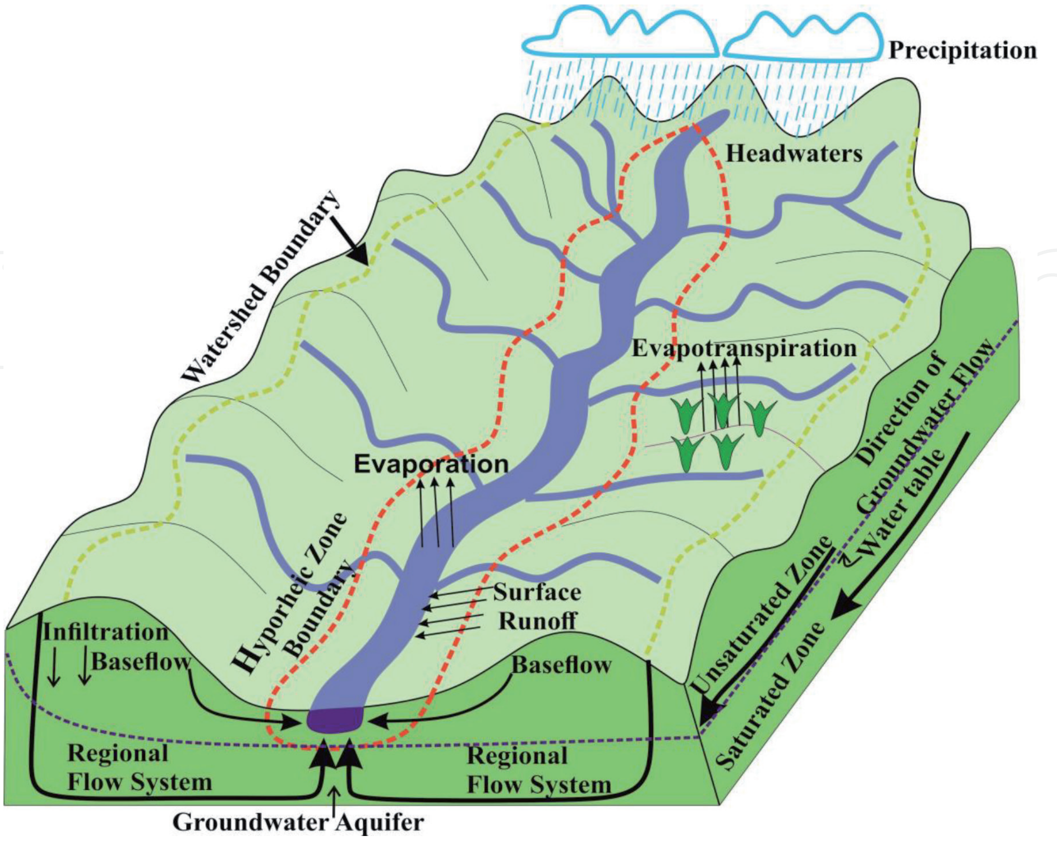

Figure 6.
River water and groundwater interactions at the regional scale.

trends of groundwater, regional topographic, hydrological conditions are mainly characterized by the regional groundwater flow system at a regional scale. The hydrological processes of the large watershed such as precipitation, surface run-off, infiltration, evapotranspiration, base flow, streamflow, and channel conditions are described in **Figure 6**. All hydrological processes can cover on the regional scale of a large watershed and a small stream reach conditions can cover local scales. Slow hydrological response to recharging areas will result in an insignificant seasonal impact on the regional GW flow system. The regional flow system is developed with long-distance SW flows and low charging levels on higher topographical slopes. It explains why recharging and discharge levels fluctuate more at the local level than at the regional level. The recharge or discharge rate investigation of the regional flow system can be analyzed using environmental tracer isotopes and hydro-chemical characteristics (major ions and heavy metals) [6].

4. Different types of SW-GW interactions

There are several types of interactions between GW and SW. A losing river does not lose water as it flows downstream by percolation, but it can also lose water through evaporation, use of plants, and consumption of human activities.

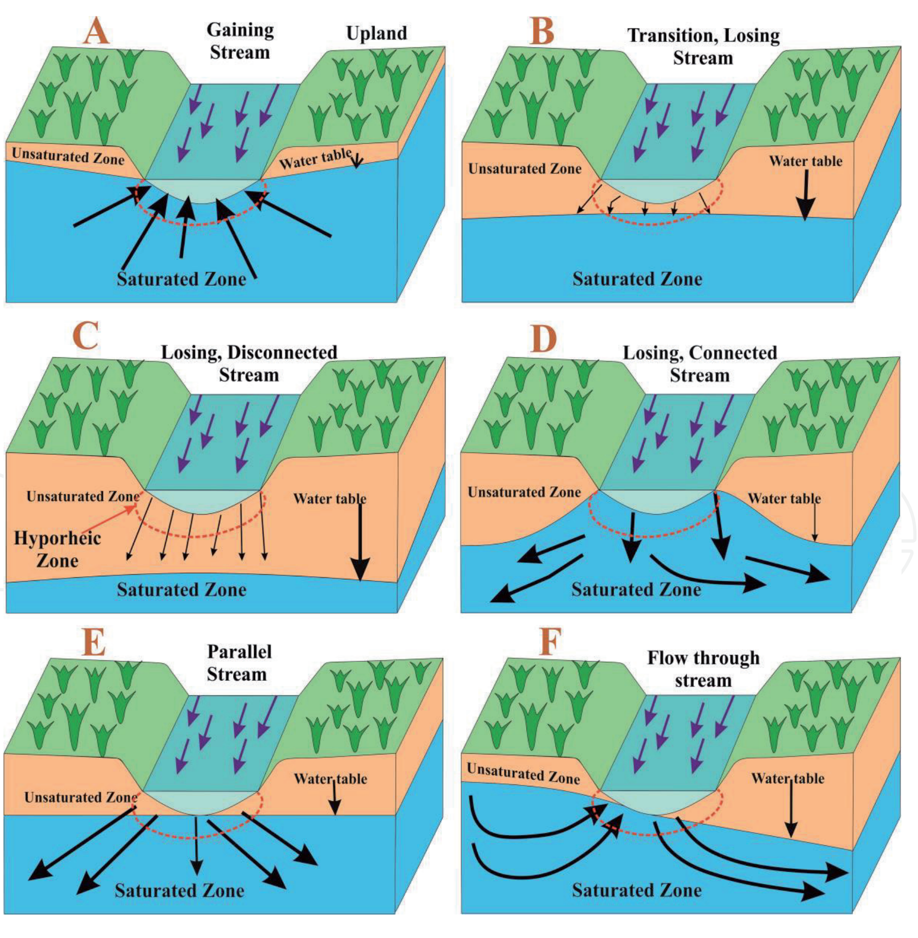

Figure 7.
Types of stream water and GW interactions, while (A) a stream water gains from GW, known as gaining stream, (B) a stream loses water to GW at shallow depth known as disconnected or losing or transition stream, (C) a stream loses water to GW at great depth known as a disconnected stream, (D) a stream loses water, but connected with GW known as a connected stream (E) a stream stage and the GW head are equally known as the parallel Stream, and (F) a stream shows the gain on one side and the loss on the other side to GW known as flow-through stream.

4.1 Gaining stream or effluent stream

In this connection, the level of GW is higher than the riverbed which recharges the river (**Figure 7A**). It can also be described as entering of GW into SW when SW reaches its base level which results in gaining stream connection.

4.2 Losing stream or influent stream

In this connection, given the level of GW is lower than the streambed, hence, SW recharges the GW. Losing streams connection divided into two types in which it is either connected or disconnected with the GW table. The term "transition" is used to define the condition of connected and disconnected streams. As shown in (**Figure 7B**), the unsaturation zone is presented in transition with a shallow GW table between riverbed. Note, there are distinguished interactions of the disconnected stream with shallow and deep water table [20].

4.3 Losing disconnected stream

In this connection, the unsaturation zone of sediments exists between the channel and regional water table hence it can be said that the system may be hydraulically disconnected (**Figure 7C**). The term disconnected has been criticized because it can suggest a system where there is no exchange of recharge and discharge of the GW system [21]. Therefore, the rate of infiltration of a disconnected system has been referred to as a "maximum losing condition"—stream discharge mechanism. Thus, the water table occurs at greater depth in the disconnected system and at shallow depth in the transition zone.

4.4 Losing connected stream

A stream is a stream that loses water while it flows downstream. The water penetrates the ground and recharges the local GW flow as the water table lies below the level of the channel with the absence of an unsaturation zone (**Figure 7D**).

4.5 Parallel stream

This interaction occurs when the stage of the stream and the head of the groundwater is equal (**Figure 7E**).

4.6 Flow-through stream

Where the stage of the channel is less than the head in the groundwater on one bank side and larger than the head in the groundwater on the opposite bank side, this process is seen as a flow-through reaches (**Figure 7F**). This interaction occurs most frequently when the stream cuts perpendicular to the regional GW flow, which in the case of fluvial plains is along their axis.

5. Methods

The methods for the investigation of aquifer systems such as remote sensing (RS) and geographic information system (GIS), resistivity test, and pumping tests will be discussed. Several related approaches also will be discussed such as seepage meter, Darcy's law, heat tracer method, and environmental tracer method for the investigation of SW-GW interactions.

5.1 Resistivity survey

In groundwater system, evaluation geophysical methods (geothermal gravity, electrical resistivity, etc.) have been well recognized. The electrical resistivity survey is one of the tools that is very effective to identify subsurface profiles without interfering with the structure of the soil [22]. The usage of this method enables the measurement of groundwater quantities and quality. This includes detailed knowledge concerning the geological and hydrological information of the GW system such as subsurface mapping to identify aquifer-protective structures, the analysis of infiltration of the vadose zone, measuring the extent of volume and internal aquifer structure, and groundwater contamination [23]. It is effectively used to estimate soil porosity and soil permeability as a non-destructive process. In addition, it is commonly utilized for the interaction of changes in the resistivity of the subsurface with the soil characteristics. The negligible porosity and permeability of the hard rock, as well as igneous and metamorphic rocks, in terms of soil exploitability, but the alteration processes taking place in the first 10–100 m of depth can significantly increase their fracture permeability. In the zone influenced by modification, this may create moderate secondary porosity aquifers.

These aquifers are very critical for irrigation and the availability of potable water in many parts of the world. Altered methods often influence the overall porosity of the rock such as water content which results in a varying spectrum of electric resistivity within the transition region [24]. Consequently, it is a good potential technique for the study of alteration zones in hard rocks, electrical resistivity in rocks influenced by differing weathering degrees. Schlumberger array system [25] was used to perform the resistivity survey. "ABEM SAS 1000 Terrameter" was the device used for performing Vertical Electrical Sounding (VES). For resistance measurement, four electrodes were selected at a certain time. Two existing electrodes situated on the outside of the potential electrodes were inserting currents into the field. The potentially different electrodes were quantified and the ground resistance was measured by Eq. (2).

$$\rho = K_g * resistanceofearth = K_g * \left(\frac{V}{I}\right) \tag{7}$$

$$K_g = \left\{\pi\left(\frac{AB}{2}\right)^2 * \left(\frac{MN}{2}\right)^2\right\} / MN \tag{8}$$

where ρ denotes the apparent resistivity (ohm-meter), K_g is the geometric constant, I indicate current (ampere), V is the voltage (volt), AB is spacing between the current electrode (m) and MN is spacing between the potential electrode (m). The geometric factor (K_g) is based on field observation calculation and by multiplying the geometric factor with data of resistivity the apparent resistivity values can be calculated. For instance, the transmissivity estimates were 0.588, 0.578, and 0.756 m^2/min, respectively, by the analysis of grain size distribution, the resistivity survey, and the pumping test [25]. The finding on the results indicated that the values of aquifer transmissivity have been found much similar to each other by measurement of grain size distribution, pumping test, and resistivity survey.

5.2 Pumping test

Pumping test is a field technique and it is used for the assessment of the aquifer characteristics such as hydraulic conductivity (K), storage coefficient (S), and transmissivity (T). Aquifer hydraulic parameters are spatially and temporarily influenced by their heterogeneity, complicated geologic conditions, as well as multipart boundaries but these characteristics in various aquifer areas, are challenging to describe efficiently [26]. The geological formation of the aquifer (confined, leaky, unconfined, and fractured aquifer) influences the hydraulic parameters to estimate; thus, various interpretive techniques are applied. Implementation of geophysical studies and pumping test techniques may be used to maximize the comprehension of hydrogeology models by accurately detecting such essential aquifer characteristics: permeability, thickness, porosity, transmissivity, hydraulic conductivity, etc. Various pumping test methods are used to determine aquifer hydraulic characteristics; but, long term, step pumping, and recovery tests are mostly utilized. Aquifer characteristics can be found by using easy methods such as the first analytical solutions proposed by [27] Thiem (1906) for a steady-state condition that gives an equation for the groundwater flow in aquifers subject to pumping. After this, [28] Theis (1935) and Cooper and Jacob (1946) [29] find extremely restrictive conditions in terms of a transient state that limit their implementation to aquifers that are uniform, homogeneous and isotropic, constant thickness, porous and permeable which produce pumping with a constant discharge in a completely penetrating well. The following formulas calculate the aquifer properties by Theis-Jacob method:

$$s(r,t) = \frac{Q}{4\pi T}\left[W\left(\frac{4^2 S}{4Tt_p}\right)\right] \tag{9}$$

$$T = KB = \frac{\gamma b_m^3}{12\mu} N_f \tag{10}$$

where Eq. (5) shows that the $s(r,t)$ is the drawdown, Q is the pumping rate, T is the hydraulic transmissivity, W is the well function for a confined aquifer, S is the aquifer storativity and t_p is the pumping period (t) for $t \le t_p$. The Eq. (6) is showing that K is the hydraulic conductivity, N_f is defined by imposing $N_f \times b_m^3 = n_e \times B$, and n_e is the effective porosity ($\cong 0.003 \pm 0.002$) of the studied fractured aquifer given by tracer tests [29]. For example, Alfy et al. [30] were used the pumping test and geophysical logging for the investigation of the hydraulic and petro-physical characteristics of the folded UmmerRadhuma (UeR) Formation, Saudi Arabia. The findings were obtained showing that, concerning efficient porosity, permeability, hydraulic conductivity, and transmissibility, the average values of 220%, >100 mD, 3.30×10^{-5}–1.34×10^{-3} m/s, and 1.49×10^{-3}–6.04×10^{-2} m^2/s.

5.3 Remote sensing (RS) and geographic information system (GIS)

Intensive performance applications of RS and GIS are spatial data analysis and monitoring methods for groundwater sources. RS data integration with the GIS environment seems to be very beneficial in considerably identifying the specific groundwater potential areas. In the short time available, RS and GIS cover a vast and unacceptable region of the earth's surface to assess areas of possible groundwater and to identify natural recharging locations [31]. RS and GIS information are valuable for many geological resources including mineral exploration, hydrogeology conditions, structural, geomorphological, lithological features, depth, and thickness of the aquifer system and other geological areas [32]. Furthermore, in the area of groundwater studies, researchers have used thematic layers such as geomorphology, geology, drainage patterns, lineaments, vegetation, intensities of rainfall, and slopes [33]. A geophysical resistivity survey was performed by [34] and borehole lithology results were compared for aquifer characterization with groundwater potential mapping which was created by RS and GIS. For a hydrogeological study, [35] integrated electrical resistivity survey data with RS outputs in a GIS environment. Moreover, [36] suggested that the geophysical resistivity data integrated with high-resolution satellite data collected from RS and GIS techniques provide more accurate information on geological and hydrological characteristics and also give possible groundwater potential zones in the hard rock formations.

5.4 Seepage meter

Seepage meter is one of the most common instruments for directly measuring SW-GW seepage flux. Initially, it was developed to measure water loss from a canal in 1940 by [37], and also it is used for other purposes such as measuring seepage flux in small lakes, estuaries, rivers, and several other environments [38]. The basic concept of the seepage meter is the difference between initial (V_o) and final (V_f) volume of water through a surface area (A) in time (t) and is given as

$$Seepageflux\,(Q) = \frac{\left(V_f - V_o\right)}{tA} \tag{11}$$

This method was described as a plastic bag type seepage meter, which is based on isolating principle and covers a portion of SW-GW interactions with a bottomless cylinder which important in determining the directions of water exchange at the local scale [38]. The streambed features (riffle-pool sequences) can be recognized by seepage meters at the local scale because this method is useful to investigate the water flux estimates of lower streambed sediments [39]. Seepage meters can also be used to determine the volumetric change in flow, discharge, or recharge zones along with the streambed sediments in the hyporheic zone at a local scale. The seepage meter is favorable for those streams which have low current velocities which represent a local scale stream [17]. For instance, four seepage meters have been used, along with riverbed sediments of Biebrza River, Poland to quantify the hyporheic exchange flux at a local scale [40].

5.5 Heat tracer method

Subsurface temperature variation is associated with the movement of water. This variation affects the chemistry of water which can be traced by the heat tracer method. According to Anderson [41], the heat tracer method is used to determine hyporheic exchange, GW flow patterns, and rate of discharge and recharge at the local regional scale. SW temperature fluctuates throughout the season and also daily while GW temperature remains constant throughout the year. This method has been used by Schmidt, Raich, and Schirmer [42] for SW-GW interactions at the local scale and suggested that streambed temperatures can be quickly, reliably, and cheaply assessed the SW-GW interactions at several locations. The successful combination of their conceptual methods described by Constantz [43] with these technical improvements to assess SW-GW interactions, GW discharge or recharge, SW movement through the streambed, and GW flow systems. In the past, heat tracers' methods have been used to evaluate losing and gaining stream. For instance, the temperature was investigated by Cox, Su, and Constantz [44] and also determined the special conductance, and chloride between the aquifers system using the heat tracer method in the Russian River, Mendocino, California. Their results indicated that the special conductance and chloride data were not correlated with RW data. It means GW was not significantly varied by the exchange of SW and GW system and temperature variations in GW were negligible.

5.6 Darcy's law method

Darcy's law [45] measures the hydraulic gradients, aquifer hydraulic conductivity, cross-sectional area of the aquifer perpendicular to the flow, and to evaluate the rate of GW flow. Darcy's law expressed as,

$$\text{Darcy's}\, law\,(q) = -K\frac{(dh)}{dl} \tag{12}$$

where q is a specific discharge $(\mathrm{L/T})$, K is hydraulic conductivity, l is the distance (L), and h is the hydraulic head. The hydraulic gradient is measured by piezometers and mini-piezometer at both local and regional scales. Piezometer indicates that the hydraulic head difference at great depth or vertical GW flow while mini-piezometer at shallow depths indicate GW downwelling or upwelling processes. The hydraulic conductivity is based on streambed sediments. It can be utilized as the estimation of streambed sediments by the slug test. A slug test is based upon the immediate increase or fall of the water level in the bore and the conformity of the water level to the original position when the water returns. The velocity and direction of GW flow can be determined by the mini-piezometer method. Furthermore, the estimation of flow between the SW-GW aquifer systems through semi-impervious stratum in one dimension which is used based on Darcy's law.

$$\text{Darcy's}\, law\,(q) = K\Delta h \tag{13}$$

where q flows between SW-GW, Δh is the river head and aquifer head, and K is hydraulic conductivity of the semi-impervious streambed stratum. The investigation of the Platte River by Chen et al. [45] in eight tributaries of eastern and south-central Nebraska, the USA, with the help of this method. The river joins the Missouri River in the eastern part of Nebraska and its interactions with high plains aquifer systems.

5.7 Environmental tracer method

Environmental tracer method is used to analyze the SW-GW interactions on both local and regional scales which are based on isotope data and geochemical data such as major ions or heavy metals. Stable hydrogen and oxygen isotopes are useful for assessing the flow of precipitation, source of water, age of water, and hydrological processes. In addition, this method has been used to determine the GW influx to a tropical river with major ions to supplement GW flux results [46]. Furthermore, it has been used to determine the gaining, losing disconnected and connected stream reach condition based on the geochemical parameters and stable isotopes [47–53]. Moreover, it is important to understand the recharge or discharge zone along with the GW flow system.

6. Conclusions

The number of groundwater studies continues to increase globally. This chapter discusses the SW-GW interactions and groundwater characteristic with appropriate assessment methodologies encompassing subsurface investigation (resistivity), hydraulics aspects of groundwater (pumping test), and mapping (RS and GIS).

Furthermore, understanding the SW-GW interactions through available methods (seepage meter, heat tracer, and environmental tracer) is useful in watershed management, that is, risk management and assessment of the aquifer system. Moreover, environmental tracer method is also a useful for the evaluation of the hydrological process, source of water, age of water, and gaining and losing disconnected and connected stream conditions.

Conflict of interest

The authors declare no conflict of interest.

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References

[1] Akhtar N, Rai SP. Heavy metals concentrations in drinking water and their effect on public health around moth block of Jhansi District, Uttar Pradesh, India. Indian Journal of Environmental Protection. 2019;**39**(9):945-953

[2] Fleckenstein JH, Krause S, Hannah DM, Boano F. Groundwater-surface water interactions: New methods and models to improve understanding of processes and dynamics. Advances in Water Resources. 2010;**33**(11):1291-1295

[3] Conant B, Robinson CE, Hinton MJ, Russell HAJ. A framework for conceptualizing groundwater-surface water interactions and identifying potential impacts on water quality, water quantity, and ecosystems. Journal of Hydrology. 2019;**574**:609-627

[4] White DS. Perspectives on defining and delineating hyporheic zones. Journal of the North American Benthological Society. 2006;**12**(1):61-69

[5] Brunner P, Simmons CT, Cook PG. Spatial and temporal aspects of the transition from connection to disconnection between rivers, lakes and groundwater. Journal of Hydrology. 2009;**376**(2):159-169

[6] Barthel R, Banzhaf S. Groundwater and surface water interaction at the regional-scale – A review with focus on regional integrated models. Water Resources Management. 2016;**30**(1):1-32

[7] Dahl M, Nilsson B, Langhoff JH, Refsgaard JC. Review of classification systems and new multi-scale typology of groundwater-surface water interaction. Journal of Hydrology. 2007;**344**(1-2):1-16

[8] Todd DK, Mays LW. Grounwater Hydrology. 3rd ed. USA: Hamilton Printing; 2005

[9] Western AW, Zhou SL, Grayson RB, McMahon TA, Blöschl G, Wilson DJ. Spatial correlation of soil moisture in small catchments and its relationship to dominant spatial hydrological processes. Journal of Hydrology. 2004;**286**(1-4):113-134

[10] Western AW, Blöschl G. On the spatial scaling of soil moisture. Journal of Hydrology. 1999;**217**(3-4):203-224

[11] Mahmood R. Scale issues in soil moisture modelling: Problems and prospects. Progress in Physical Geography. 1996;**20**(3):273-291

[12] Hupet F, Lambot S, Javaux M, Vanclooster M. On the identification of macroscopic root water uptake parameters from soil water content observations. Water Resources Research. 2002;**38**(12):36-1-36-14

[13] Driscoll FG. Groundwater and wells. 2nd ed. St. Paul, Minnesota: Johnson Filtration Systems; 1986

[14] Freeze RA, Cherry JA. Groundwater. Bergen Country, NJ, USA: Prentice Hall Inc; 1979

[15] Kalhor K, Ghasemizadeh R, Rajic L, Alshawabkeh A. Assessment of groundwater quality and remediation in karst aquifers: A review. HHS Public Access. 2019;**8**:104-121

[16] Johnson AI. Specific yield compilation of specific yields for various materials—Hydrologic properties of earth materials. Geological Survey Water-Supply Paper. 1967;**1662-D**:80

[17] Tóth J. A theoretical analysis of groundwater flow in small drainage basins. Journal of Geophysical Research. 1963;**68**(16):4795-4812

[18] Sophocleous M. Interactions between groundwater and surface

water: The state of the science. Hydrogeology Journal. 2002;**10**(1):52-67

[19] Harish HK, Arina K. Ground water and surface water interaction. In: GIS and Geostatistical Techniques for Groundwater Science, Section C., G. Q. A. U. G. and G. A. 14.2. Elsevier Inc; 2019. pp. 197-207

[20] Peterson DM, Wilson JL. Variably Saturated Flow Between Streams and Aquifers. Socorro: New Mexico Water Resources Research Institute; 1988

[21] Brunner P, Cook PG, Simmons CT. Connected and disconnected streams. Water Resources Research. 2010;**66**:37-39

[22] Larsen F et al. Controlling geological and hydrogeological processes in an arsenic contaminated aquifer on the Red River flood plain, Vietnam. Applied Geochemistry. 2008;**23**(11):3099-3115

[23] Almadani S, Ibrahim E, Al-Amri A, Fnais M, Abdelrahman K. Delineation of a fractured granite aquifer in the Alwadeen area, Southwest Saudi Arabia using a geoelectrical resistivity survey. Arabian Journal of Geosciences. 2019;**12**(15)

[24] Gowda JP, Kumar SCP. Segmentation of white blood cell using k-means and Gram-Schmidt Orthogonalization. Indian Journal of Science and Technology. 2017;**10**(6):1-5

[25] Keller GV. Dipole method for deep resistivity studles. Geophysics. 1966;**31**(6):1088-1104

[26] Arshad M, Shakoor A, Ahmad I, Ahmad M. Hydraulic transmissivity determination for the groundwater exploration using vertical electric sounding method in comparison to the traditional methods. Pakistan Journal of Agricultural Sciences. 2013;**50**(3):487-492

[27] Calvache ML, Sánchez-Úbeda JP, Duque C, López-Chicano M, De La Torre B. Evaluation of analytical methods to study aquifer properties with pumping tests in coastal aquifers with numerical modelling (Motril-salobreña aquifer). Water Resources Management. 2015;**30**(2):559-575

[28] Thiem G. Hydrologische Methoden. Leipzig, Germany: JM Gebhardt; 1906a

[29] Theis CV. The relation between the lowering of the piezometric surface and the rate and duration of discharge of a well using ground-water storage. Eos Transactions American Geophysical Union. 1935;**16**(2):519-524

[30] Cooper HHJ, Jacob CE. A generalized graphical method for evaluating formation constants and summarizing well-field history. Eos Transactions American Geophysical Union. 1946;**27**(4):526-534

[31] Caputo MC, Carlo DL, Masciale R, Masciopinto C. Long-term pumping test and ert to visualize hydrogeologic barriers in heterogeneous and Karstic coastal aquifers. Journal of Geology and Geophysics. 2017;**6**(5):1-5

[32] Alfy ME, Lashin A, Faraj T, Alataway A, Tarawneh Q , Al-Bassam A. Quantitative hydro-geophysical analysis of a complex structural karst aquifer in Eastern Saudi Arabia. Scientific Reports. 2019;**9**(1):1-18

[33] Senthilkumar M, Gnanasundar D, Arumugam R. Identifying groundwater recharge zones using remote sensing & GIS techniques in Amaravathi aquifer system, Tamil Nadu, South India. Sustainable Environment Research. 2019;**29**(1):1-9

[34] Elubid BA et al. Geospatial distributions of groundwater quality in Gedaref state using geographic information system (GIS) and

drinking water quality index (DWQI). International Journal of Environmental Research and Public Health. 2019;**16**(5):1-20

[35] Anbazhagan S, Jothibasu A. Geoinformatics in groundwater potential mapping and sustainable development: a case study from southern India. Hydrological Sciences Journal. 2016;**61**(6):1109-1123

[36] Oseji JO, Atakpo EA, Okolie EC. Geoelectric investigation of the aquifer characteristics andgroundwater potential in Kwale, Delta state, Nigeria. Journal of Applied Sciences and Environmental Management. 2005;**9**(1):157-160

[37] Kumar MG, Bali R, Agarwal AK. Integration of remote sensing and electrical sounding data for hydrogeological exploration—A case study of Bakhar watershed, India. Hydrological Sciences Journal. 2009;**54**(5):949-960

[38] Rashid M, Lone MA, Ahmed S. Integrating geospatial and ground geophysical information as guidelines for groundwater potential zones in hard rock terrains of south India. Environmental Monitoring and Assessment. 2012;**184**(8):4829-4839

[39] Israelson OW, Reeve RC. Bulletin No. 313—Canal Lining Experiments in the Delta Area, Utah, UAES Bulletins; 1944

[40] Brodie RS, Baskaran S, Ransley T, Spring J. Seepage meter progressing a simple method of directly measuring water flow between surface water and groundwater systems. Australian Journal of Earth Sciences. 2014;**42**(14):37-41

[41] Lee DR. A device for measuring seepage flux in lakes and estuaries. Limnology and Oceanography. 1977;**22**(1):140-147

[42] Kalbus E, Reinstorf F, Schirmer M. Measuring methods for groundwater-surface water interactions: A review. Hydrology and Earth System Sciences. 2006;**10**(6):873-887

[43] Anibas C, Buis K, Verhoeven R, Meire P, Batelaan O. A simple thermal mapping method for seasonal spatial patterns of groundwater-surface water interaction. Journal of Hydrology. 2011;**397**(1-2):93-104

[44] Anderson MP. Heat as a ground water tracer. Ground Water. 2005;**43**(6):951-968

[45] Schmidt C, Raich MB, Schirmer M. Characterization of spatial heterogeneity of groundwater-stream water interactions using multiple depth streambed temperature measurements at the reach scale. Hydrology and Earth System Sciences. 2006;**10**(3):1419-1446

[46] Constantz J. Heat as a tracer to determine streambed water exchanges. Water Resources Research. 2008;**46**(4):1-20

[47] Cox MH, Su GW, Constantz J. Heat, chloride, and specific conductance as ground water tracers near streams. Ground Water. 2007;**45**(2):187-195

[48] Darcy H. Les fontaines publiques de la ville de Dijon. Paris: Victor Dalmont; 1856

[49] Chen X, Dong W, Ou G, Wang Z, Liu C. Gaining and losing stream reaches have opposite hydraulic conductivity distribution patterns. Hydrology and Earth System Sciences. 2013;**17**(7):2569-2579

[50] Cook PG. A Guide To Regional Flow in Fractured Rock Aquifers. CSIRO Land and Water, Glen Osmond: Henley Beach, South Australia: Seaview Press; 2003

[51] Engelhardt I et al. Comparison of tracer methods to quantify

hydrodynamic exchange within the hyporheic zone. Journal of Hydrology. 2011;**400**(1-2):255-266

[52] Xu W, Su X, Dai Z, Yang F, Zhu P, Huang Y. Investigations multitraceurs des interactions nappes-rivières: cas d'étude du bassin de la rivière Nalenggele, nord-ouest de la Chine. Hydrogeology Journal. 2017;**25**(7):2015-2029

[53] Aguilar BJ, Banks EW, Batelaan O, Kipfer R, Brennwald MS, Cook PG. Groundwater residence time and aquifer recharge in multilayered, semi-confined and faulted aquifer systems using environmental tracers. Journal of Hydrology. 2017;**546**:150-165