Groundwater-Surface Water Exchange

William W. Woessner

THE GROUNDWATER PROJECT

GROUNDWATER-SURFACE WATER EXCHANGE

The Groundwater Project

William W. Woessner

Emeritus Regents' Professor of Hydrogeology
University of Montana
Missoula, Montana USA

GROUNDWATER-SURFACE WATER EXCHANGE

The Groundwater Project
Guelph, Ontario, Canada

Published by the Groundwater Project, Guelph, Ontario, Canada, 2020.

Woessner, William W.

Groundwater-Surface Water Exchange / William W. Woessner - Guelph, Ontario, Canada, 2020.

158 pages

ISBN: 978-1-7770541-5-1

THE GROUNDWATER PROJECT

Cover Photo: William W. Woessner

Dedication

I dedicate this book to all who want to learn about and freely spread their knowledge of groundwater-surface water exchange.

The Groundwater Project Foreword

The United Nations Water Members and Partners establish their annual theme a few years in advance. The theme for World Water Day of March 22, 2022, is “Groundwater: making the invisible visible.” This is most appropriate for the debut of the first Groundwater Project (GW-Project) books in 2020, which have the goal of making groundwater visible.

The GW-Project, a non-profit organization registered in Canada in 2019, is committed to contribute to advancement in education and brings a new approach to the creation and dissemination of knowledge for understanding and problem solving. The GW-Project operates the website https://gw-project.org/↗ as a global platform for the democratization of groundwater knowledge and is founded on the principle that:

“Knowledge should be free and the best knowledge should be free knowledge.” Anonymous

The mission of the GW-Project is to provide accessible, engaging, high-quality, educational materials, free-of-charge online in many languages, to all who want to learn about groundwater and understand how groundwater relates to and sustains ecological systems and humanity. This is a new type of global educational endeavor in that it is based on volunteerism of professionals from different disciplines and includes academics, consultants and retirees. The GW-Project involves many hundreds of volunteers associated with more than 200 hundred organizations from over 14 countries and six continents, with growing participation.

The GW-Project is an on-going endeavor and will continue with hundreds of books being published online over the coming years, first in English and then in other languages, for downloading wherever the Internet is available. The GW-Project publications also include supporting materials such as videos, lectures, laboratory demonstrations, and learning tools in addition to providing, or linking to, public domain software for various groundwater applications supporting the educational process.

The GW-Project is a living entity, so subsequent editions of the books will be published from time to time. Users are invited to propose revisions.

We thank you for being part of the GW-Project Community. We hope to hear from you about your experience with using the books and related material. We welcome ideas and volunteers!

The GW-Project Steering Committee

August 2020

Foreword

This book introduces how groundwater and surface water features such as rivers, streams, lakes and wetlands are linked and function as a continuous hydrologic system. Groundwater systems underpin these terrestrial surface waters. Surface water and groundwater are continuously linked in the hydrological cycle but often are assessed separately even when it is recognized that water exchanges occur between them. Conceptually, surface waters and the associated shallow groundwater systems are best viewed as a single interacting system, a continuum that is one water resource. Investigations reveal that water enters and leaves surface water features and the groundwater zone at multiple rates, locations, scales and time frames. This exchange physically moves both water and dissolved constituents between groundwater and surface waters and supports associated groundwater dependent ecological communities. Those who understand the conceptual and field linkages will be able to determine how natural and impacted streams, lakes and wetlands function, and which preservation or restoration actions can resolve issues and meet goals.

Some examples include: what information is needed to plan and execute remediation efforts if stream, lake and wetland features become impacted from physical and chemical alterations? In river systems, what level of river water exchange with the bed, banks and floodplain is required to support a natural geomorphic set of conditions, and appropriate aquatic and terrestrial ecological systems? If a groundwater contaminant is migrating towards a group of lakes, which ones are likely to be impacted and at what locations? When lakes act both as groundwater discharge sites and sources of groundwater recharge, how can specific aquatic ecosystems be maintained? How can vegetated areas associated with spring wetlands be restored? When water is pumped from groundwater near a lake, will be the amount of lake water available for irrigation diversions be impacted? Though these questions are not answered specifically in this book, the book builds the necessary foundation upon which to understand the issues and formulate resolutions. This book presents the conceptual models, descriptions of field-based methodologies, and modeling tools needed to understand surface-groundwater exchanges at multiple scales and in varied hydrogeological conditions.

This book has been prepared by a senior groundwater scientist who has pioneered research concerning groundwater-surface water interactions and taught a course on the subject for 30 years as well as a range of introductory and advanced hydrogeology classes at the university level. He is a specialist in field investigations, groundwater problem analysis, modeling building and computer simulation.

John Cherry, The Groundwater Project Leader

Guelph, Ontario, Canada, September 2020

Acknowledgements

I thank the following individuals for their thorough and useful reviews of and contributions to this book:

- Marc Hinton, Research Scientist-hydrogeologist with Natural Resources Canada;
- Mary P. Anderson, Emeritus Professor of Hydrogeology, University of Wisconsin-Madison; and,
- Eileen Poeter, Emeritus Professor of Geologic Engineering and Hydrology, Colorado School of Mines.

I thank Professors Ric Hauer and Willis Weight for inviting me to contribute material related to groundwater-surface water exchange to their textbooks, parts of which were rewritten and expanded upon for this book. I am grateful for Amanda Sills' oversight of this book and to Elhana Dyck for copyediting, both of the Groundwater Project, Guelph, Ontario, Canada, and my wife Jean for a final read through. I appreciate Eileen Poeter's layout editing and production of this book.

William W. Woessner

Table of Contents

1 Introduction and Importance

Surface water and groundwater resources have traditionally been treated as separate or weakly linked systems in some university courses, legal considerations, international cross boundary negotiations, and local and federal regulatory rules. In truth, groundwater and surface-water resources are a fully connected resource responding to changes in hydrologic conditions. Exchange of surface water and groundwater occurs at multiple scales, rates and time frames. Scientific literature clearly supports this model. However, in some jurisdictions, legal interpretations and government regulations isolate surface water and groundwater resources. In these situations, legal concepts need revision to bring them into line with current scientific knowledge.

The United States Geological Survey refers to surface water and groundwater as a single resource (Winter et al., 1998) and it is this framework that guides the material presented in this book. Even though this view is sound, Conant and others (2019) suggest that a gap exists between conceptualizing interactions and holistically integrating the role of groundwater-surface water exchange within physical, geochemical, biological and ecological frameworks. Clearly, interactions are complex and extend beyond solely identifying locations of exchange.

The groundwater system is an important component of the surface-water system and, conversely, surface-water features are linked, at multiple scales, to local and regional groundwater (e.g., Toth, 1963; Winter et al., 1998). Physically, the movement of water from one system to another is largely controlled by differences between surface-water stages (i.e., elevation of the water surface) and groundwater levels, as well as hydrologic conditions at the interfaces (Conant et al., 2019). These interconnections, referred to simply as exchange, result in the creation and sustainability of rivers, lakes, wetlands and coastal systems, support or limit the nature of aquatic and adjacent terrestrial ecosystems (e.g., Meyer, 1997), impact surface water and groundwater quality, and affect the fate of contaminants as they migrate within the hydrologic system (Conant et al., 2019). Exchange occurs under both natural and hydrologically modified conditions such as channelized stream reaches and at reservoirs (Winter et al., 1998). Exchange varies in time as hydrologic conditions naturally change and/or are purposely manipulated.

Investigations of natural or impacted groundwater-surface water exchanges and forecasting the consequences of proposed modifications and remediation efforts require the development of conceptual models describing exchange processes. In addition, methods to locate and quantify exchanges under natural, disturbed and remediated settings are requisite. For example, the design of a lake ecological study or plans to manage the lake stage requires information on how groundwater systems interface with the lakeshore and bottom. A stream restoration project with a goal of enhancing exchanges needs to be built on an appropriate conceptual model. The effort to protect a wetland calls for identification of the presence or absence of groundwater and surface water exchange. Resolution and quantification of natural or modified exchange processes necessitate well supported conceptual models, and the application of appropriate qualitative and quantitative methods and tools.

1.1 Principles and Concepts

This section describes the basic physical principles driving groundwater-surface water exchange. Groundwater is defined as the water that occurs in the zone of saturation. Most groundwater exchange is between near surface unconfined (i.e., water table) systems and surface-water features. Surface water is water that occurs on the land surface as rivers, streams, lakes, and wetlands. This section addresses the spatial and temporal variations in the transfer of water between groundwater systems and surface-water features. It also presents conceptual models of groundwater exchange with streams, lakes, and wetlands. The ocean and associated components (e.g. estuaries) are also considered surface water. Exchange with coastlines is discussed briefly in this first section. Exchange also influences the water quality of surface water and groundwater and how each system interfaces with the adjacent aquatic and terrestrial ecosystems. To expand the physically based conceptual models presented in this book, the reader is directed to work by Conant and others (2019) that presents multiple conceptual flow charts used to build a framework of physical, biological, ecological and geochemical processes that both influence and are a result of the exchange processes.

Physically, groundwater moves from recharge areas to discharges areas, often originating from or discharging to surface-water features. Recharge and discharge areas vary greatly and are dictated by hydrologic conditions and the landscape framework (including hydrogeologic properties). In addition, surface-water systems exchange water with groundwater systems. Such relationships should be reflected when conceptualizing multi-scale groundwater and surface-water budgets (Figure 1).

The exchange process is driven by: 1) the relative elevation of surface-water features and associated groundwater head distribution in the adjacent and underlying groundwater system; 2) the hydraulic properties and composition of the sides, banks and bottoms of surface-water features; and, 3) the underlying geologic framework composition and structure. Challenges in documenting exchange sites and rates include the significant differences in residence times and flow paths of connected groundwater and surface-water systems.

Generic cross sections and map views are typically used to illustrate the steady-state and three-dimensional exchange processes. Sequences of cross sections are used to provide illustrations of changes in exchange under transient conditions.

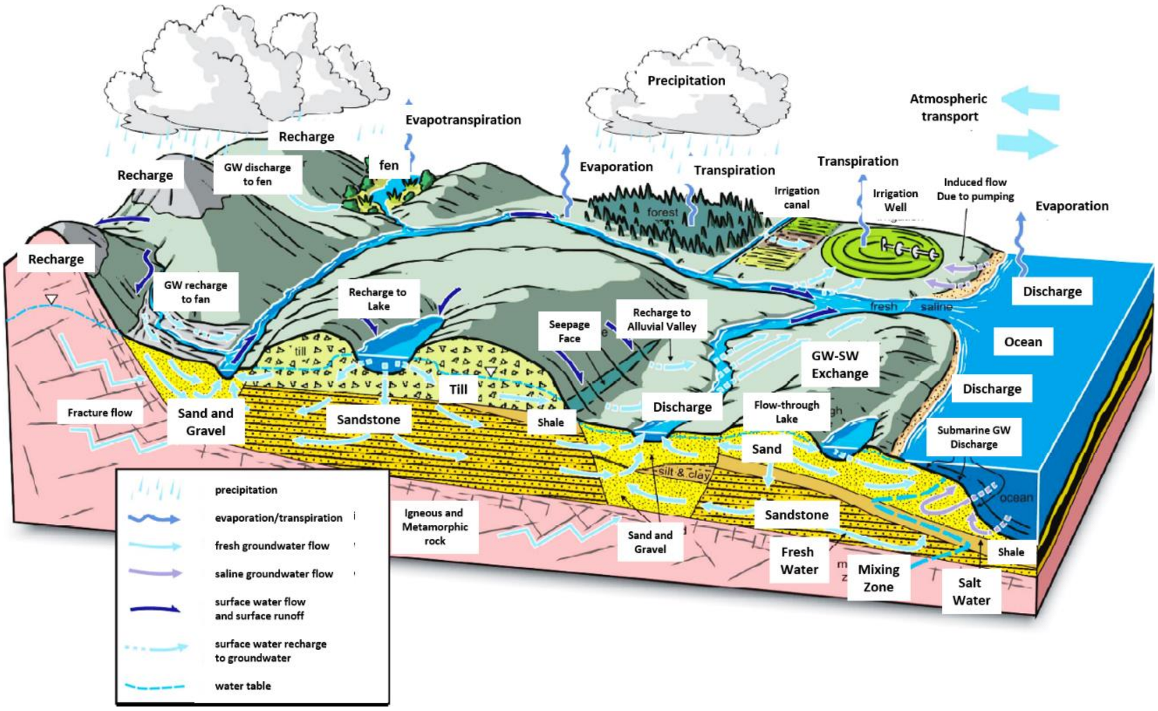


Figure 1 - Schematic of the hydrologic cycle driven by precipitation recharge**.** Key water budget components include water lost by evaporation and transpiration, and the exchange of water in the surface water and groundwater system. Groundwater recharge and discharge (light blue arrows) creates exchanges with surface-water features such as rivers, streams, lakes and irrigation ditches. Groundwater discharges to and is recharged by surface-water features. Groundwater exchange also occurs at the coastline with discharge to the ocean (Hinton, 2014).

Conceptual models of rivers, lakes, and wetlands have been developed by a number of authors (e.g. Freeze and Cherry, 1979; Winter et al., 1998; Woessner, 1998; Fetter, 2001; Woessner, 2000; Anderson et al., 2015; Weight, 2019). Depending on their discipline, researchers use different terms to describe the same conditions (Table 1). Models for rivers are presented in Figure 2.

Table 1 - Groundwater-Surface Water Exchange Terminology (Woessner, 2020).

Discipline	Groundwater flow to Surface Water	Surface Water Flow to Groundwater	Groundwater Flow to and from Surface Water	No Groundwater Exchange
Groundwater Scientists	Effluent Conditions	Influent Conditions	Flow-Through Conditions	Zero-exchange or Parallel Flow
Surface water Scientists	Gaining Conditions	Losing Conditions	Flow-Through Conditions	NA
Aquatic Ecologists	Upwelling Conditions	Downwelling Conditions	Flow-Through Conditions	NA

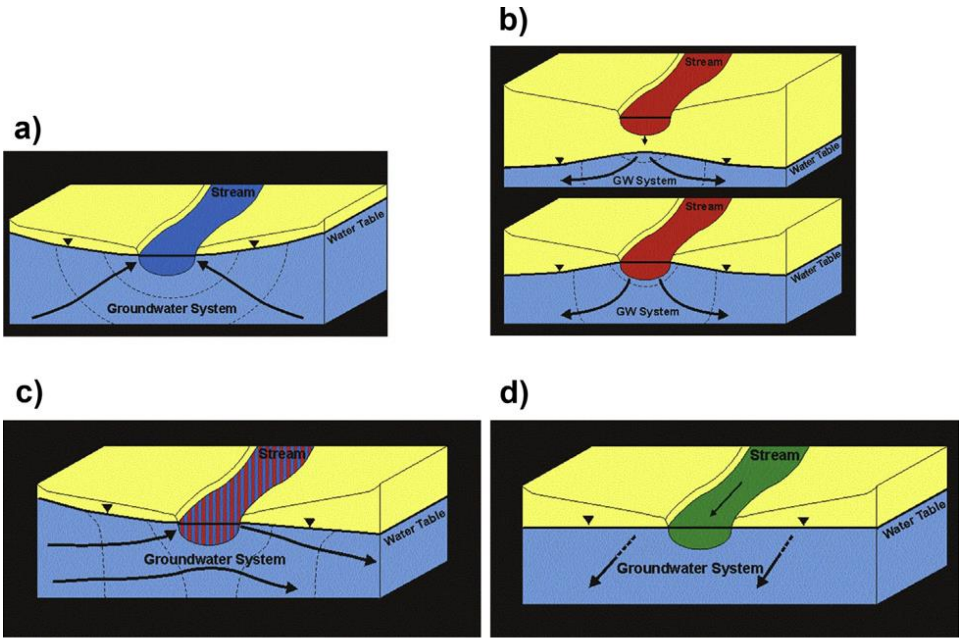


Figure 2 - Conceptual models of groundwater stream exchange. a) Effluent conditions (gaining stream); b) Influent conditions (losing stream); c) Flow-through conditions (flow-through stream); d) Parallel flow conditions, zero-exchange stream (Anderson et al., 2015).

Hydrogeologists refer to settings where groundwater is discharging to surface water as effluent conditions; when groundwater is being recharged by surface water as influent conditions, and in situations where both conditions are present in a single feature as flow-through (e.g. lake flow-through conditions). This terminology is focused on the groundwater system (e.g., Todd and Mays, 2004). An effluent stream, lake, or wetland setting is receiving flow from the groundwater (groundwater is discharging, exiting the groundwater system). Scientists focused on characterizing surface-water features often describe the exchange process as it affects the surface-water feature. For example, they define an effluent reach of a stream as a gaining stream, the stream is receiving groundwater flow and the stream discharge downstream is increasing (e.g. Winter et al., 1998). A losing stream reach is one where groundwater is being recharged as the stream water leaks out of its bed and banks (influent conditions) and downstream discharge is decreasing. Flow-through conditions are generally used by both groups to describe conditions when water is entering and exiting a surface-water body at multiple locations. For streams embedded in a groundwater system in which no exchange is occurring, the term zero-exchange condition or parallel flow can be used (e.g., Figure 2). A third group of scientists have developed a descriptive set of terms also based on stream-groundwater interactions. Stream ecologists often refer to effluent streams as a site of upwelling (groundwater discharge) and influent streams as downwelling (surface-water recharge) (e.g., Hauer and Lambert, 2017). They rarely describe flow-through or zero-exchange settings. Thus, a portion of a surface-water body or coastline receiving groundwater discharge can be described as under the influence of effluent, gaining or upwelling conditions, and when surface water infiltrates its banks, shoreline and bed, conditions are described as influent, losing or downwelling.

The physical exchange process can be generalized and conceptualized using four schematic representations: 1) effluent or gaining; 2) influent or losing (two settings); 3) flow-through; and, 4) parallel flow or zero exchange. Generic conceptual models are presented in a cross section with a rectangular surface-water feature, which can be visualized as representing a stream, lake or wetland at various scales (Figure 3). Coastline exchange is discussed briefly at the end of this section. Map views of groundwater conditions associated with rivers, lakes, and wetlands are presented later in this book when each surface-water feature is specifically addressed. The general conceptual models illustrate that the groundwater water framework (geology) and interface conditions (surface-water body beds and banks) could have heterogenous and anisotropic hydraulic conductivity values (e.g., Figure 3). Of course, the world is three dimensional as is the geologic framework controlling groundwater flow rates and directions, so heterogeneity is present, and flow occurs in the third dimension that is not shown in these two-dimensional illustrations. In transient settings, each hydrogeologic material also requires an appropriate value of storativity to account for changes in groundwater storage.

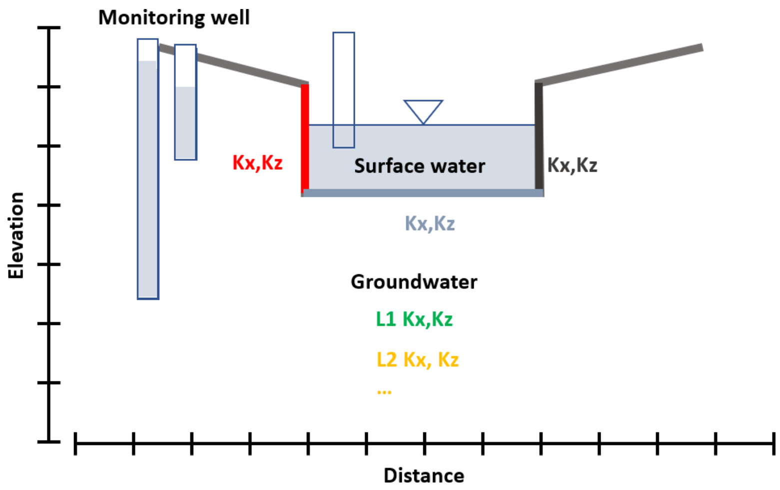


Figure 3 - Generalized conceptual cross section used to illustrate the exchange process for streams, lakes, and wetlands at various scales. Kx is the horizontal hydraulic conductivity and Kz is the vertical hydraulic conductivity. Hydraulic conductivities (Kx, Kz) of the surface-water boundaries (bed and banks) and underlying sediments (e.g., layer L1, L2 Kx, Kz …) may vary spatially, as indicated by the red and black banks and the blue bottom of the surface-water feature, and the green and orange values of underlying sediments. The scale can be varied to represent small or large areas thus, no units are indicated, and no vertical exaggeration is incorporated in the illustrations (Woessner, 2020).

Further simplifying the generalized model shown in Figure 3, conceptual cross sections are presented in subsequent sections for each of the exchange processes. In the conceptual cross sections, horizontal and vertical scales are generalized (no units are shown) in order to represent multiple scales (mm, cm, m, km, and so on). Groundwater heads are represented as shaded water levels in monitoring wells (vertical rectangles) and surface-water stage by the shaded water and water level shown in the vertical rectangle. The general head distribution is indicated by water levels in monitoring wells that are open at the bottom and equipotential lines are labeled with unitless relative values. For illustration purposes, flow lines are constructed assuming no vertical exaggeration in the cross sections, each cross section is constructed parallel to groundwater flow, and a single hydraulic conductivity value is assigned to the earth materials such that conditions are isotropic and homogeneous. Certainly, there are many possible combinations of conditions that influence the final groundwater flow lines and exchange flux rates; these are not accounted for in the schematic representations. Exchange is inherently transient in nature; however, to simplify the discussion, steady-state groundwater conditions are assumed.

1.2 Effluent or Surface Water Gaining Conditions

Under hydrologic conditions where the water table is higher than the surface-water stage (river, lake, and wetland) and groundwater is discharging into the surface water, effluent conditions occur (Figure 4). The surface-water feature gains from groundwater discharge when effluent conditions are present. Representative monitoring wells open only at the bottom illustrate that wells open at different depths have water levels higher than the surface-water stage. In this setting, groundwater exits to the sides and bottom of the surface-water feature. Flux rates of groundwater are dependent on the magnitude of the hydraulic gradient at the boundary, as well as the hydraulic conductivities of the geologic sediments and the banks and bottom of the surface feature. The surface-water stage reflects the local water table.

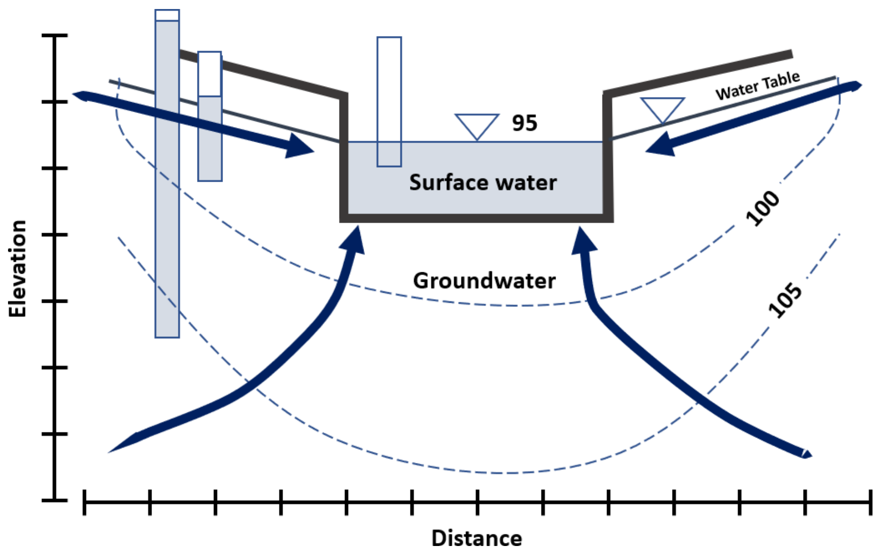


Figure 4 - Conceptual model of effluent conditions under steady state, isotropic and homogeneous conditions. Numbers represent relative values of head. Dashed lines are equipotential lines and arrows represent general groundwater flow directions. Water levels in monitoring wells open only at the bottom show heads are higher than the surface-water stage (Woessner, 2020).

1.3 Influent or Surface Water Losing Conditions

Influent conditions require the surface-water stage to be higher than the underlying and adjacent water table. There are two general scenarios when this occurs: 1) the water table remains hydrologically connected to the stage and slopes away from the feature (Figure 5), or 2) the surface-water feature is separated from the groundwater system by a vadose zone (partially saturated) (Figure 6). In both settings, the groundwater is recharged by surface water. Under these influent conditions, a volume of surface water over a specified time interval is lost to the groundwater (losing conditions). Representative monitoring wells illustrate that water levels in wells finished at multiple depths adjacent to or beneath the influent stream, lake or wetland are lower than the surface-water stage. Flux rates to groundwater are dependent on the magnitude of the local hydraulic gradient (difference between the surface-water stage and groundwater head), as well as the hydraulic conductivities of the geologic sediments and of the banks and bottom of the surface feature. The surface-water stage reflects the elevation of the local water table when the water table is connected to the feature. Representative water levels in monitoring wells are lower than the stage. However, when water percolates through a vadose zone (partially saturated sediments) the surface-water stage is disconnected from the water table (Figure 6). Depending on the leakage rate and the hydrologic properties of the saturated sediments, the water table beneath the source area may become higher than the surrounding area, mounded as shown in Figure 6. When leakage rates are low and/or aquifer hydraulic conductivities large, mounding may not be observable.

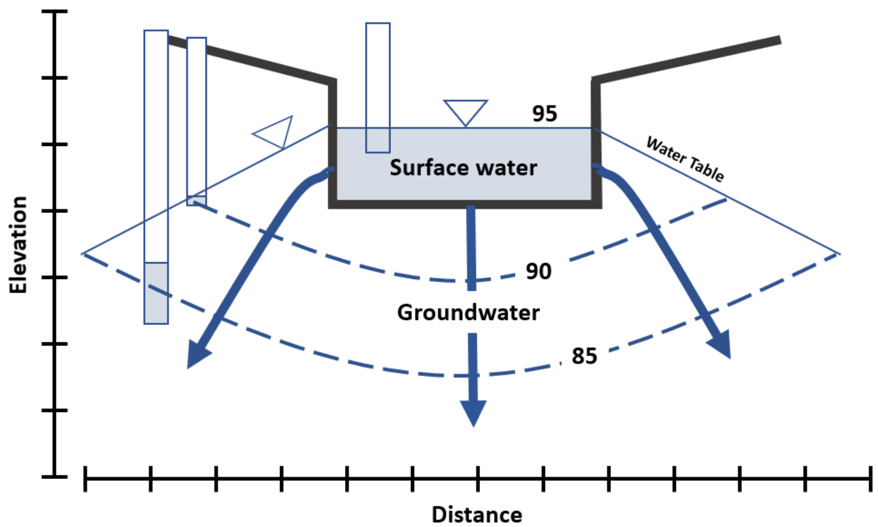


Figure 5 - Conceptual model of influent conditions under steady-state, isotropic and homogeneous conditions. Numbers represent relative values of head. Dashed lines are equipotential lines and arrows represent general groundwater flow directions. Water levels in monitoring wells open only at the bottom are lower than the surface-water stage. In this example, the water table is connected to the surface-water feature (Woessner, 2020).

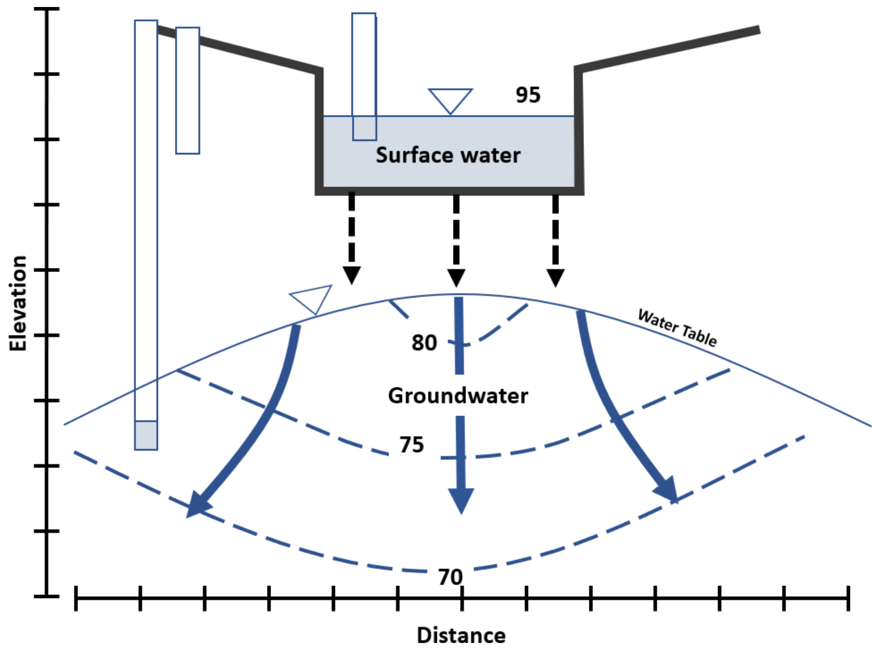


Figure 6 - Conceptual model of influent conditions under steady-state isotropic and homogeneous conditions. Numbers represent relative values of head. Dashed lines are equipotential lines and arrows represent general groundwater flow directions. Black dashed arrows show leakage through the vadose zone. Water levels in monitoring wells open only at the bottom are lower than the surface-water stage. In this example, the water table is below the surface-water feature (Woessner, 2020).

1.4 Flow-through Conditions

A flow-through condition occurs when the water table is higher at one side of the feature than the surface-water stage (river, lake, and wetland) and lower at another location (Figure 7). In this scenario, groundwater is discharging into the surface water and surface water is seeping into the groundwater system. Under flow-through conditions, surface water is added to and lost from the surface feature's volume and/or flow over a specified time interval. Generally, the surface-water stage reflects the water table in most settings. Representative monitoring wells illustrate that water levels in wells at different depths (open only at the bottom) are higher than the surface-water stage in the area where effluent conditions are present and lower in areas that are influent. Flux rates of groundwater are dependent on the magnitude of the hydraulic gradient at the boundaries, and the hydraulic conductivities of the geologic sediments and the banks and bottom of the surface-water feature.

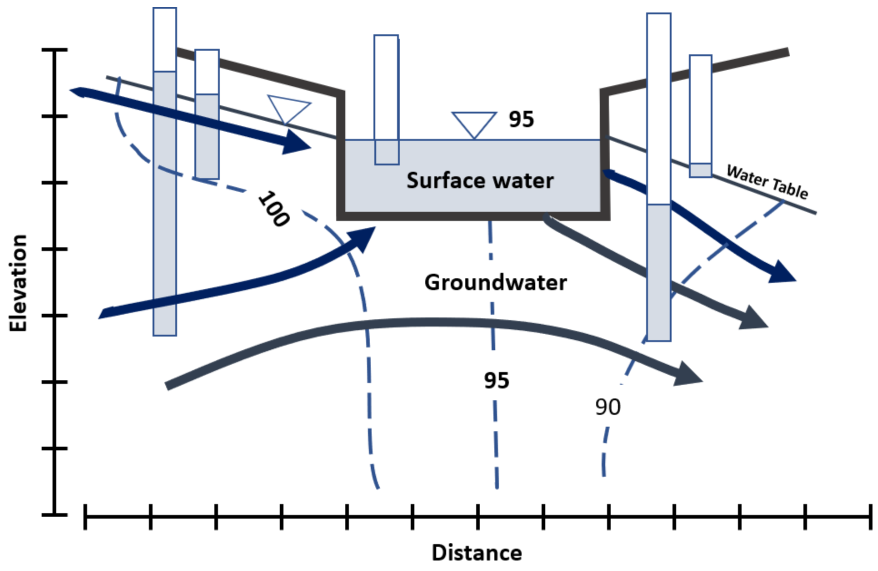


Figure 7 - Conceptual model of flow-through conditions under steady-state, isotropic and homogeneous conditions. Numbers represent relative values of head. Dashed lines are equipotential lines and arrows represent general groundwater flow directions. Monitoring wells are only open at the bottom. Water levels in up-gradient monitoring wells (left) are higher than the surface-water stage. Water levels in down-gradient monitoring wells (right) are lower than the surface-water stage. In this example, the surface-water stage represents the water table (Woessner, 2020).

1.5 Zero-Exchange or Parallel Flow Conditions

A zero-exchange or parallel flow condition occurs where no exchange is observable. This is found when the water table elevation is equal to the surface-water stage (river, lake, and wetland) (Figure 8). As both the stage and water table are at a common elevation, no hydraulic gradient is present, and the groundwater flow is parallel to the surface-water flow. Representative monitoring wells, open only at the bottom, illustrate that the water levels in wells finished at multiple depths are equal to the surface-water stage. No equipotential lines are shown in Figure 8 because wells and the surface-water stage are a common equipotential line. Flow is illustrated as parallel to the surface-water feature (out of the page). In these settings the stage represents the local water table elevation.

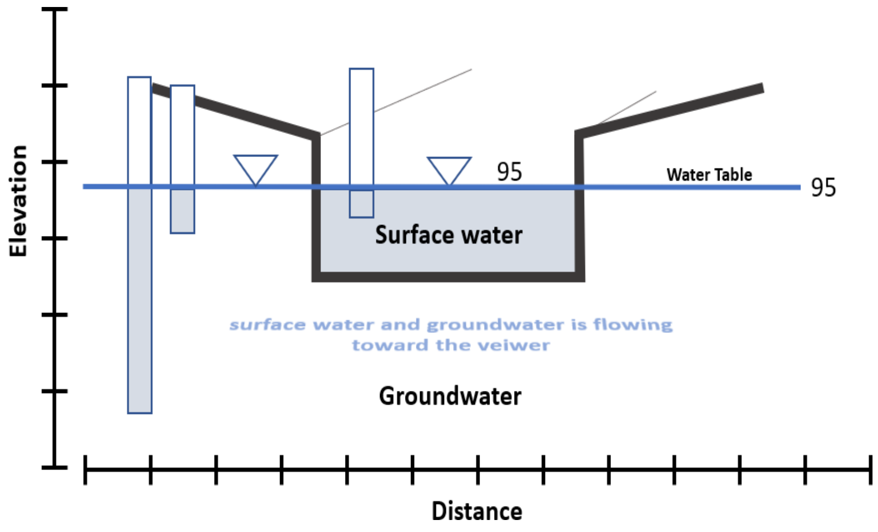


Figure 8 - Conceptual model of parallel flow or zero-exchange conditions under steady-state isotropic and homogeneous conditions. Numbers represent relative values of head. No equipotential lines are shown as the groundwater heads are the same as the surface-water stage. Water levels in monitoring wells open only at the bottom equal the surface-water stage elevation. Flow is at right angles to the cross section (Woessner, 2020).

1.6 Coastline Exchange

The generalized conceptual models of exchange are applicable to coastal settings. However, the presence of the more-dense brackish and seawater adds complexity to the exchange process forming a boundary to freshwater flow (Figure 9). Sea level stages, and groundwater heads and rates of groundwater flux control the locations of the groundwater-seawater interface. For example, groundwater discharges when the water table near the shoreline is higher than the ocean stage, effluent conditions (Figure 9). When ocean levels rise higher than the groundwater levels along the shoreline, sea water infiltrates into the shallow groundwater system, influent conditions. Flow-through conditions are not applicable in coastline settings but zero-exchange occurs temporarily in portions of the flow system when sea level and groundwater heads are equal. Many research efforts focus on the development of water supplies along coastlines and how extraction of freshwater influences the nature of the sea-water interface/transition zone (e.g., Fetter, 2001; Jiao and Post, 2019). This book focuses on groundwater-surface water exchange associated with rivers, lakes and wetlands.

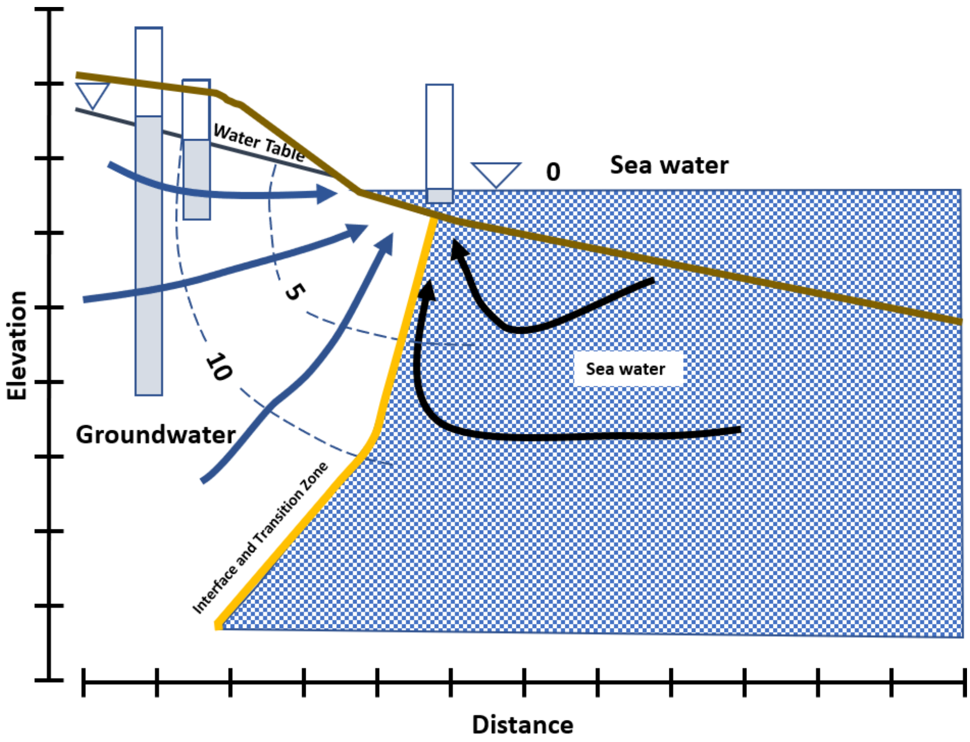


Figure 9 - Conceptual model of groundwater effluent conditions along a coastline under steady-state, isotropic and homogeneous conditions. Sea water is represented by a blue stippled pattern. Blue arrows show the flow of groundwater and black arrows the relative movement of subsurface sea water. The location of the interface and transition zone between fresh groundwater and sea water is indicated by an orange line. Numbers represent relative values of head. Equipotential lines are dashed. Water levels represent values for monitoring wells open only at the bottom (Woessner, 2020).

1.7 Heterogeneity in Exchange

The conceptual models of groundwater-surface water exchange presented in previous sections are uncomplicated as they illustrate exchanges under isotropic and homogeneous hydrogeological conditions. Location and magnitude of exchange to and from surface-water features are dependent on the natural distribution of heads, hydraulic conductivity, anisotropy distributions, and boundary conditions (Figure 10). The presence of anisotropic conditions, within the constraints of the head distribution, will direct flow preferentially to zones of higher hydraulic conductivity. In general, earth materials with low permeabilities will limit the movement of exchange waters in the subsurface. When evaluating exchange conditions at some sites, an increased level of detail is required to capture locations and magnitudes of exchange as the study area becomes smaller.

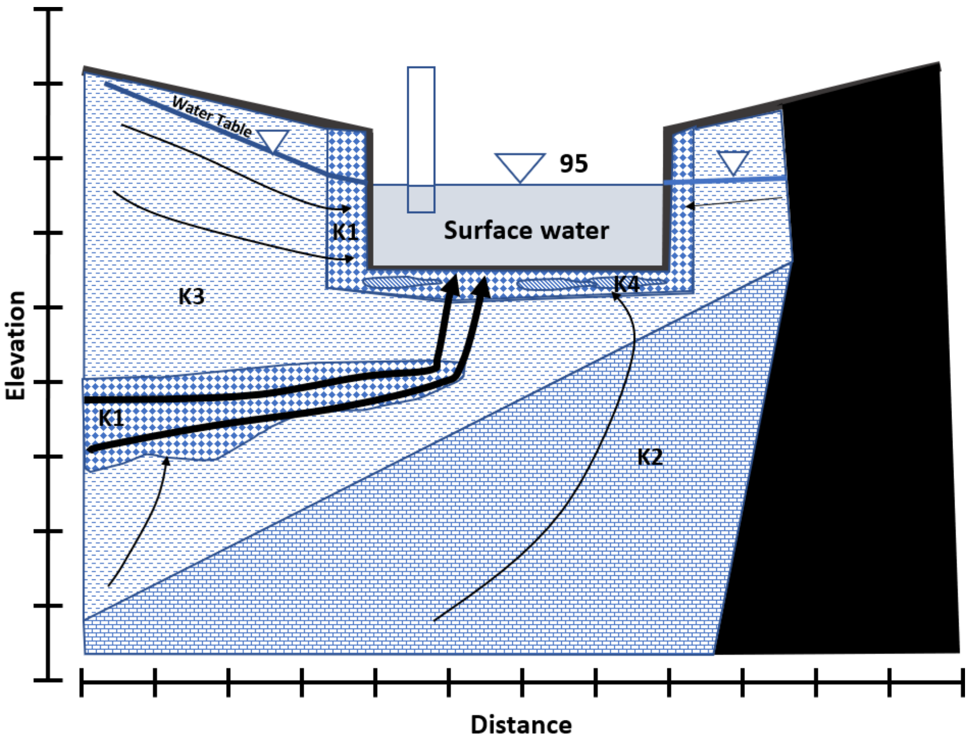


Figure 10 - Schematic of an effluent exchange site showing the impact of a boundary (black shading represents impermeable material), and heterogenous and isotropic conditions. Spatial units represented by patterns have differing hydraulic conductivities, K1>K2>K3>K4. Black arrow thickness represents relative groundwater flux and magnitude. Flow is refracted into and out of zones of differing hydraulic conductivity. In this example concentrated groundwater discharge occurs in only a portion of the bed of the surface-water feature (Woessner, 2020).

1.8 The Scale of Exchange

Cross sections can be used to represent the exchange process at various dimensions depending on the site context. For example, at the landscape scale, hydrogeologists generally visualize surface-water features as either receiving groundwater discharge or acting as sources of groundwater recharge (e.g., Figure 11). Toth (1963) developed a regional conceptual model that shows nested groundwater flow systems, and multiple groundwater recharge and discharge locations (exchange sites). In the context of groundwater-surface water exchange, surface-water features corresponding with discharge areas are effluent and features located in recharge areas are influent. When a topographically varying landscape receives sufficient groundwater recharge so that the water table is higher and mirrors topographic highs, then local, intermediate and/or regional flow systems develop (Figure 11a). In settings where recharge rates and geologic conditions do not create a water table that mirrors the topography, some surface-water features may occur where water collects in topographic lows. These features can act as a recharge sources under these conditions (Figure 11b). Regional groundwater exchange is also influenced by the distribution of the hydrogeologic properties of the underlying geologic framework. Hinton (2014) schematically shows variations in flow system exchange locations as influenced by the hydraulic conductivity distribution (Figure 12).

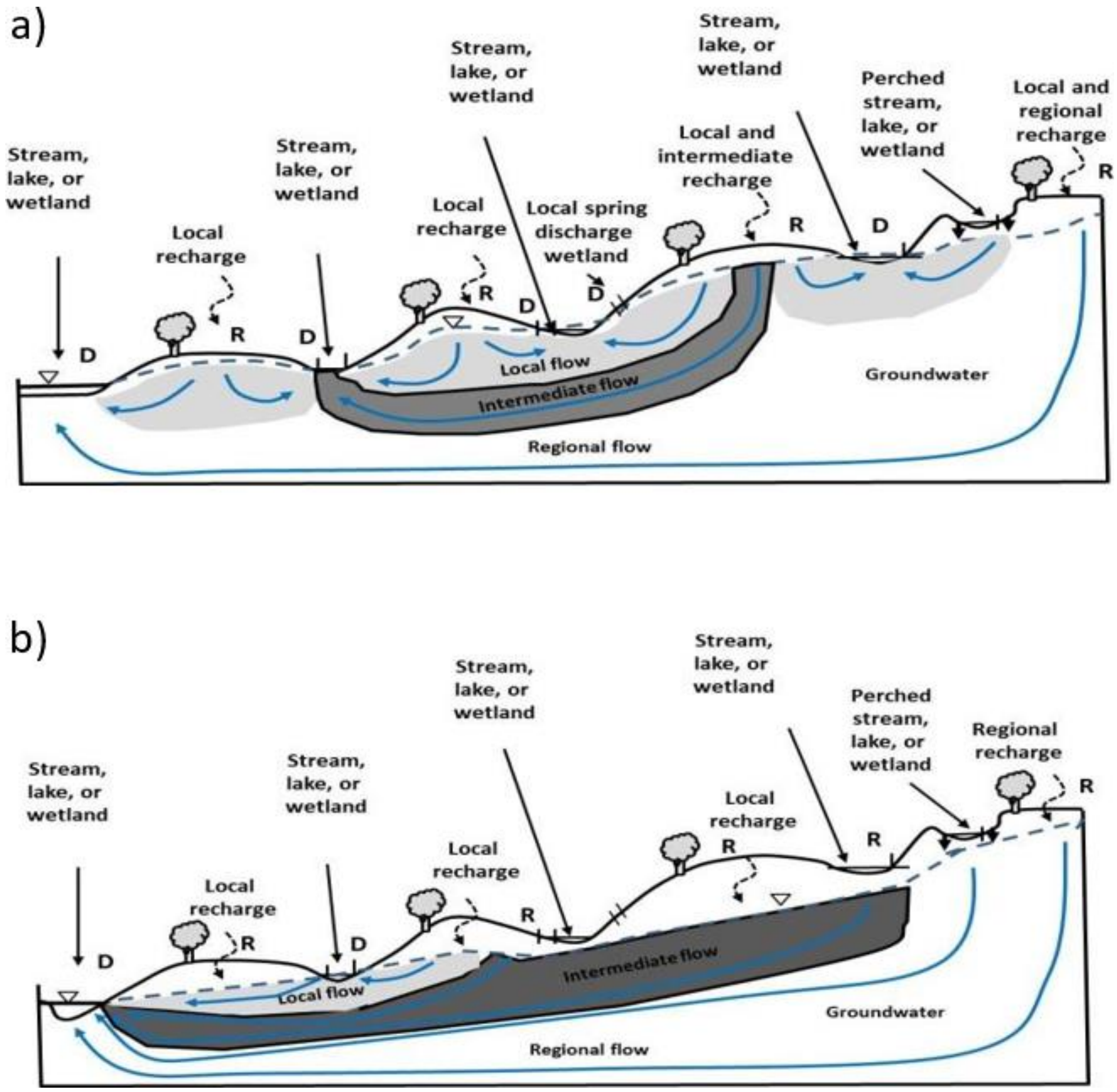


Figure 11 - Schematic cross sections showing the exchange of local, intermediate and regional groundwater flow systems with streams, lakes and wetlands. Groundwater flow lines are blue, shaded areas represent local and intermediate groundwater flow systems. R is placed over a zone receiving recharge and D is over a zone where groundwater discharge is occurring. a) Conditions where recharge is enough to create a water table that mirrors the topography (Carter, 1996). b) Setting in which recharge rates and geologic conditions do not result in a topographically dominated water table configuration. (Woessner, 2020).

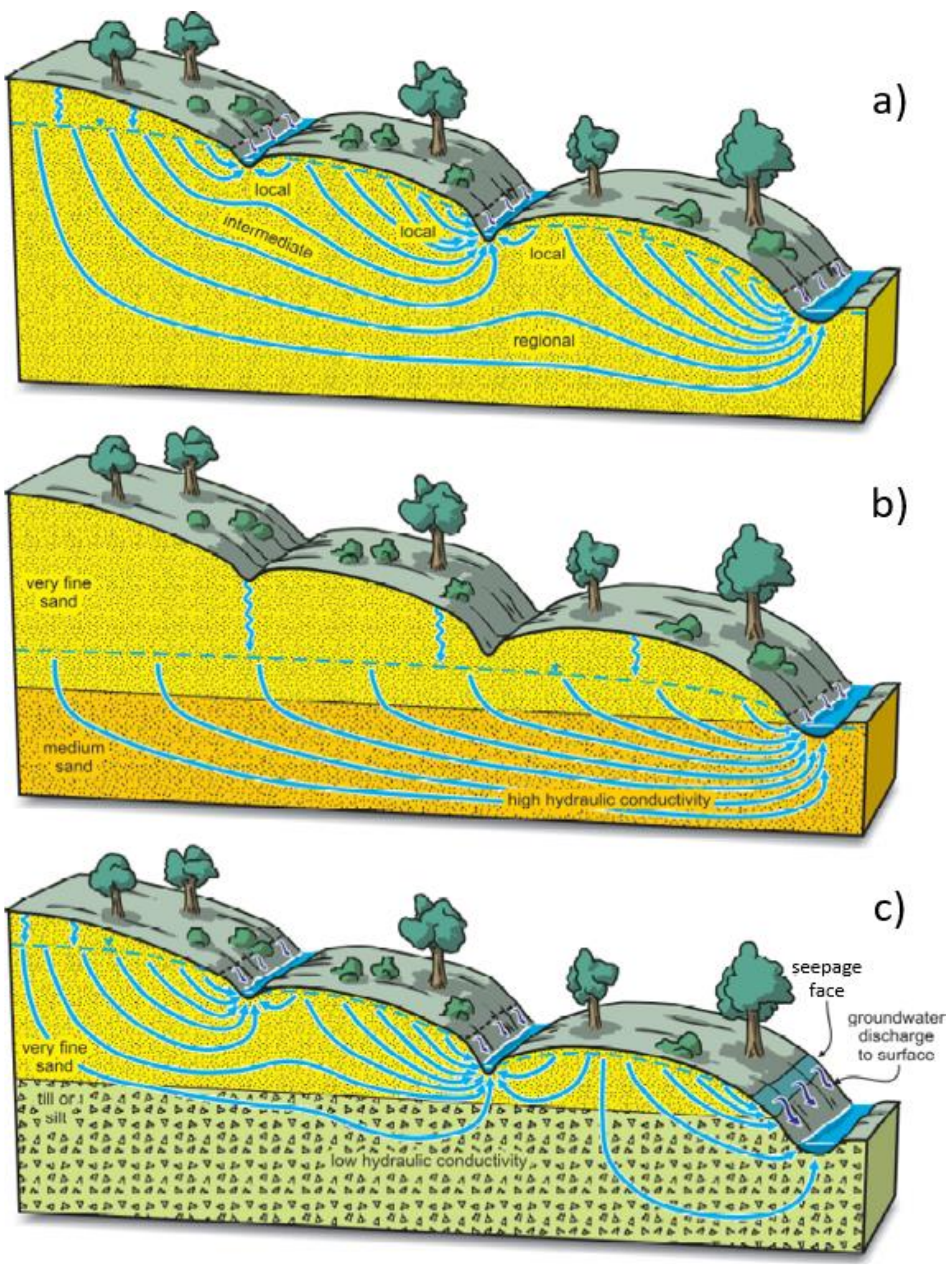


Figure 12 - Examples showing how variations of hydraulic conductivity can impact local, intermediate and regional groundwater flow systems. It is assumed recharge rates are the same in all three settings and the changes in flow systems shown are caused by the hydraulic conductivity of underlying earth materials. a) Flow paths under isotropic and homogenous conditions with the water table mirroring the topography. b) Flow paths when the first layer of earth materials is underlain by a saturated layer with a higher hydraulic conductivity. c) Flow paths when the first layer of earth material is underlain by a saturated layer with a lower hydraulic conductivity (Hinton, 2014).

Exchange occurring at the scale of tens of meters to meters, and the sub-meter scale add complexity as slight variations in bed and bank configuration and elevation, hydraulic properties and hydraulic gradients drive local exchanges (Figure 13). Though often requiring extensive instrumentation, exchange at these scales is not only of interest to hydrogeologists, but also to those linking ecological systems to surface-water features and those focusing on exchange of contaminants (e.g., Hauer and Lambert, 2017; Conant, 2004; Conant et al., 2019).

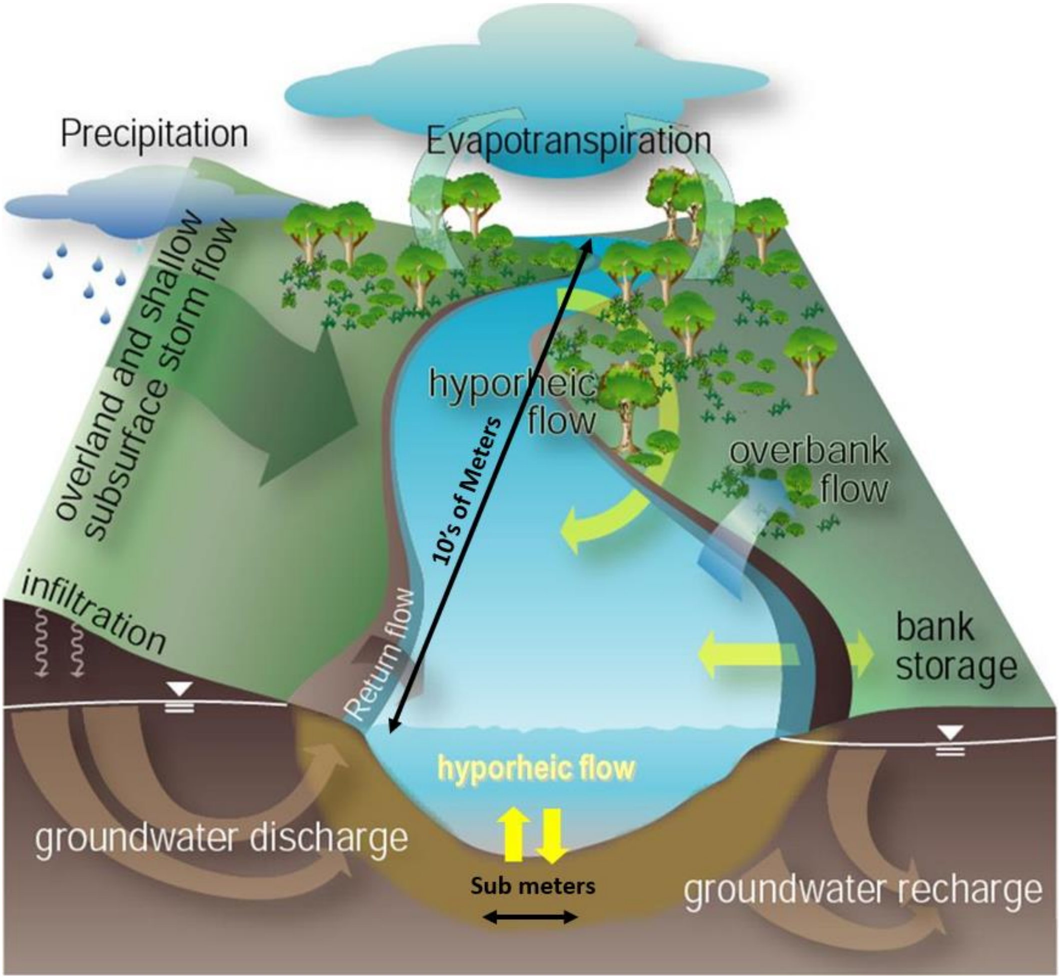


Figure 13 – Schematic of a river systems with groundwater exchange (hyporheic flow) at the 10's of meters to sub-meter scale (modified from USGS, 2015)

Transient Changes in the Exchange Process

In some settings, exchange conditions are influenced by temporal long- or short-term hydrologic conditions such as seasonal variations in water availability, stage responses to individual storm and flood events, and changes in surface-water temperatures. Such conditions result in temporal changes in rates and locations of groundwater exchanges with streams, lakes and wetlands (e.g., LaBaugh and Rosenberry, 2008) (Figure 14). For example, when the water table rises in response to a wet period a surface-water feature may become a gaining feature that was previously influent. However, when groundwater recharge becomes limited during drying periods and/or drought, a gaining feature could become flow-through and/or losing later in the year (Figure 14). When rapid changes in surface-water stage occur in response to a short-term precipitation event or resulting flooding, exchange conditions may change from gaining to losing and then, as surface-water stages decline, gaining conditions are re-established (e.g., bank storage, Freeze and Cherry, 1979, p. 225-226). Influent seepage rates can also be impacted by surface-water temperature changes (e.g., Constantz et al., 1994). This occurs because vertical bed hydraulic conductivities increase and decrease slightly as surface-water temperatures vary (changes in the specific weight and viscosity of water) (e.g., Freeze and Cherry, 1979; Fetter, 2001). Zamora (2007) suggests temperature impacts on seepage rates should be evaluated to determine if seasonal surface-water temperatures impact local seepage rates. Activation of production wells impacting the shallow groundwater system near surface-water features may locally reverse the surface-water exchange (e.g., Barlow and Leake, 2012). Studies of exchange using only a snapshot analysis (single moment or period) are often insufficient to characterize seasonal and longer-term trends in the movement of water between surface-water features and groundwater. Thus, in most settings, researchers assessing exchange processes should design studies that account for changing conditions.

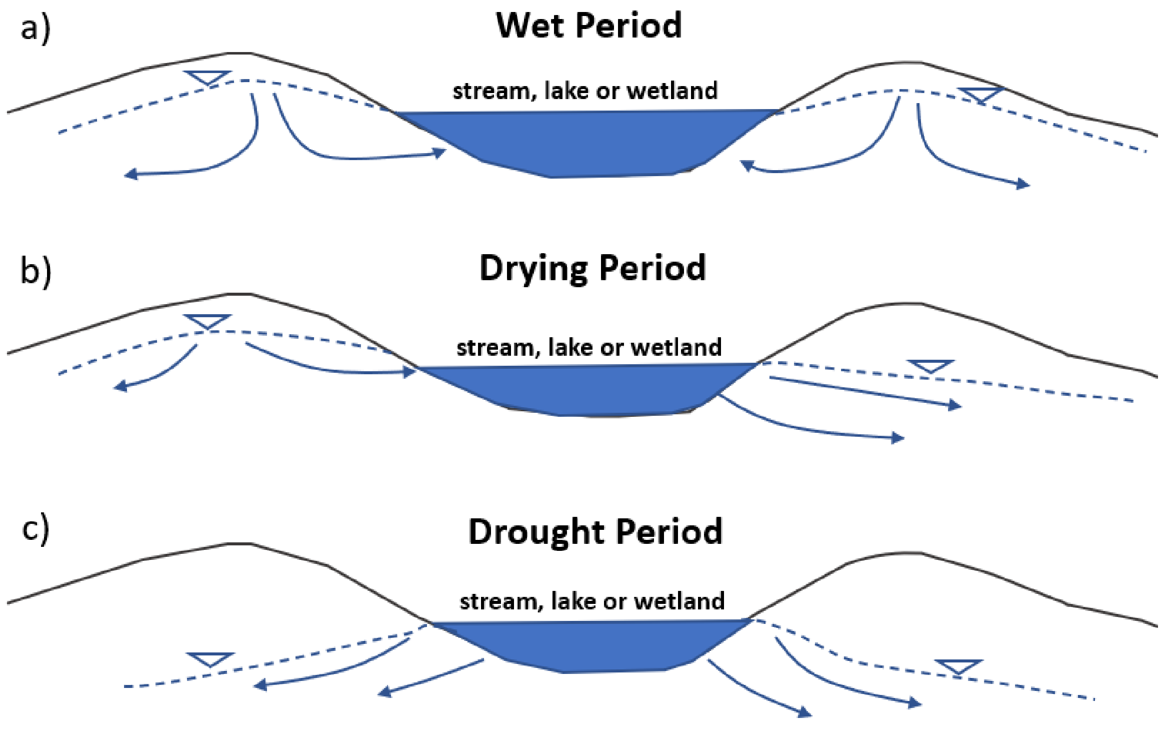


Figure 14 - Seasonal variation (periods of wet and dry) of groundwater exchange with streams, lakes and wetlands in a hypothetical setting. a) The wet period corresponds to a setting where groundwater recharge drives the system and effluent conditions dominate (dashed line is the water table and blue arrows represent groundwater flow). b) Drying conditions represent times with a lower surface-water level and reductions in groundwater inflow resulting in a shift to flow-through conditions. c) Drought conditions may cause influent conditions to dominate (Woessner, 2020).

The conceptual models presented in Figures 4 through 14 represent the baseline conditions underlying similarities and differences of exchange process in rivers, lakes, and wetlands. The next three sections use these conceptual models to describe exchanges with streams, lakes and wetlands.

2 Streams and Groundwater Exchange

Stream-groundwater exchange is described at the watershed/basin (50 to 500 km^2), valley segment (100 to10,000 m^2), reach (10 to 100 m^2) and channel/habitat-unit (1 to10 m^2) scales in this section as delineated in Figure 15 (Bisson et al., 2006).

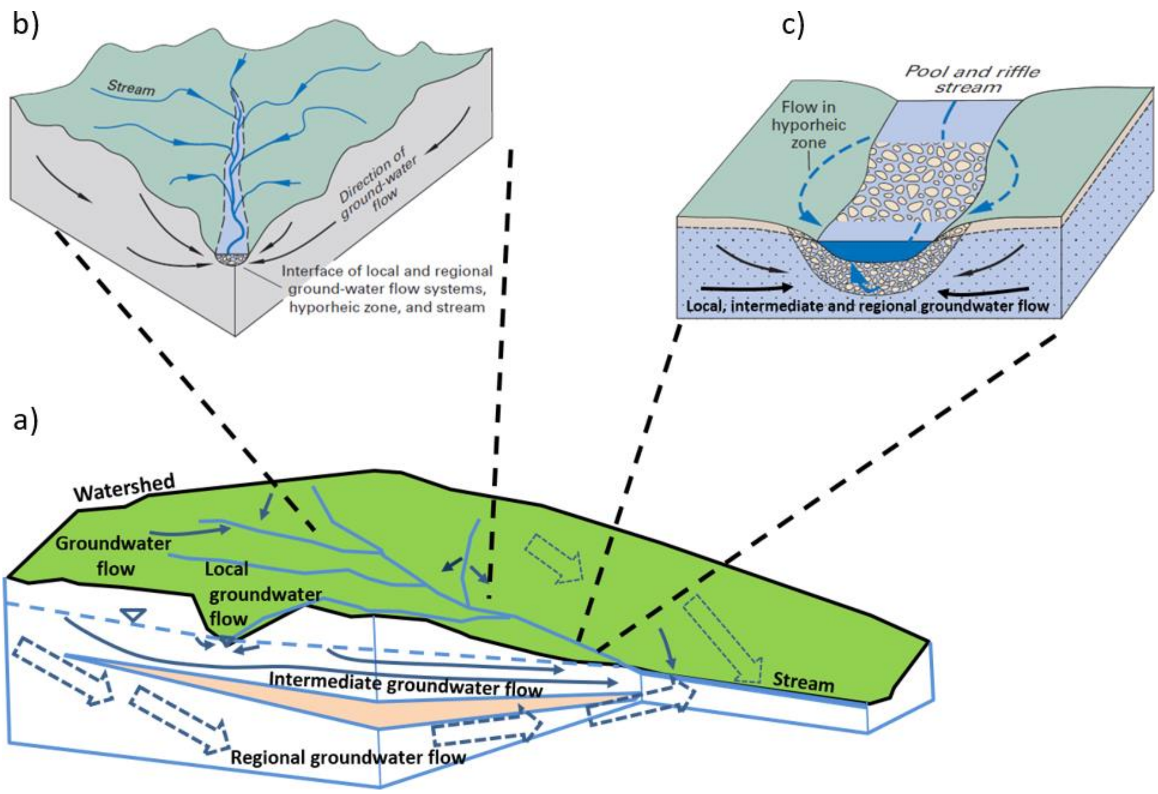


Figure 15 – Stream-groundwater exchange at multiple scales. a) At the watershed/landscape scale, shallow local and intermediate groundwater discharges to, or is recharged by, tributaries and the main channel. Basin scale groundwater flow discharges to the main channel (large dashed arrows). b) Smaller stream valleys and floodplains induce groundwater exchange. c) Channel-scale exchange includes surface water circulating within the beds and banks as well as local, intermediate and/or regional groundwater exchange (black arrows) (modified from Healy et al., 2007).

2.1 Effluent or Gaining Stream

Effluent or gaining streams occur when the adjacent water table and groundwater head beneath the stream are greater than the stream stage (Figure 4, Figure 16). Groundwater gradients are upward (Figure 16a). In map view, groundwater flow converges towards the stream and equipotential lines point upstream as a "V" (Figure 16b). Groundwater entering the stream becomes streamflow and the stream stage represents the water table elevation at the edge of the stream.

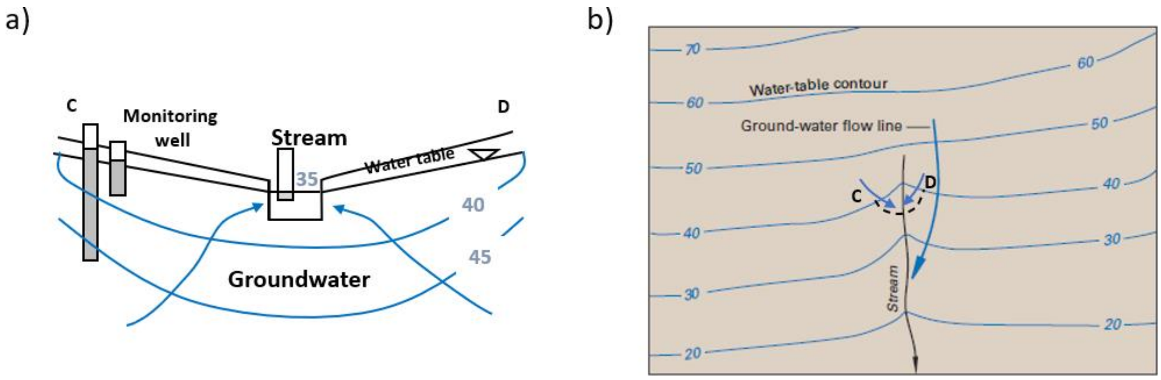


Figure 16 - Cross section and map views of effluent groundwater-stream interactions (gaining stream). Equipotential lines and relative head values are shown as blue water table contours. Groundwater flow is in the direction indicated by blue arrows. Aquifer conditions are assumed to be isotropic and homogeneous. Monitoring wells are open only at the bottom. a) Cross section showing upward groundwater gradient and groundwater discharging to the channel (Woessner, 2018). b) In map view, equipotential lines cross the stream and point in the upstream direction (V points upstream). The approximate location of the cross section in Figure 16a (C-D) is the dashed black line located along flowlines that converge at the stream (after Healy et al., 2007).

2.2 Influent or Losing Stream

Influent or losing streams occur when the adjacent water table and the heads beneath the stream are lower than the stream stage. When the stream and water table are well connected (e.g., Figure 5) water flows directly from the stream channel to the adjacent groundwater (Figure 17a). In map view, when the groundwater and stream are fully connected, the groundwater flow diverges from the stream (Figure 17b). Equipotential lines point as a "V" downstream. Under these conditions, the stream stage represents the water table elevation.

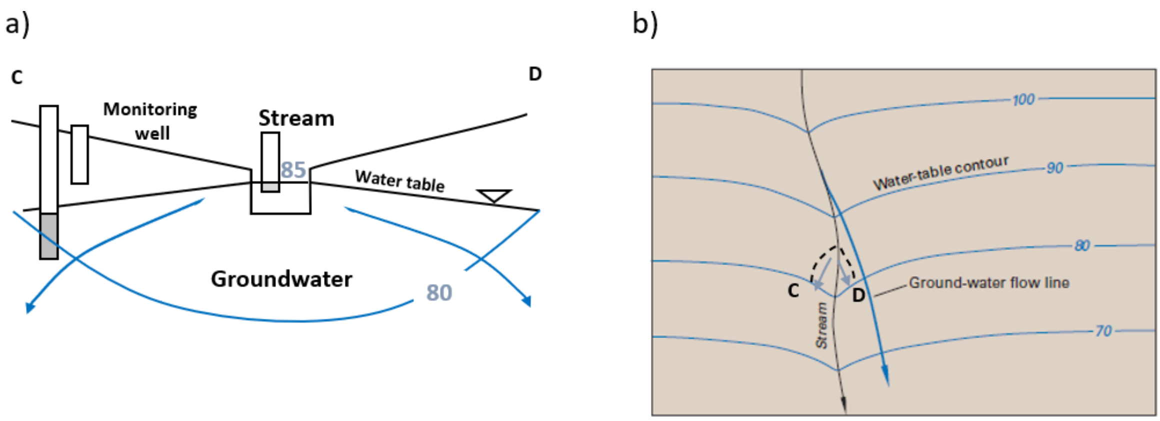


Figure 17 - Cross section and map views of influent groundwater stream interactions (losing stream). Equipotential lines and relative head values are shown as blue water table contours. Groundwater flow is in the direction indicated by blue arrows. Aquifer conditions are assumed to be isotropic and homogeneous. Monitoring wells are open only at the bottom. a) Cross section showing a downward groundwater gradient and groundwater being recharged by the channel (Woessner, 2018). b) In map view, equipotential lines cross the stream and point in the downstream direction (V downstream). The location of the cross section in (a) (C-D) is approximated by the position of the dashed black line (after Healy et al., 2007).

In contrast, an influent stream can also be disconnected from the underlying water table in which case leakage to the groundwater is by percolation (Figures 18 and 19). When the stream is perched above the water table, the stream stage does not represent the local water table elevation. In some settings the leakage from the stream creates a groundwater mound beneath the stream and diverging flow paths occur (Figure 18a and b). When leakage rates are low and aquifer hydraulic conductivity is large the leakage can have little effect on the groundwater flow direction (Figure 19a and b).

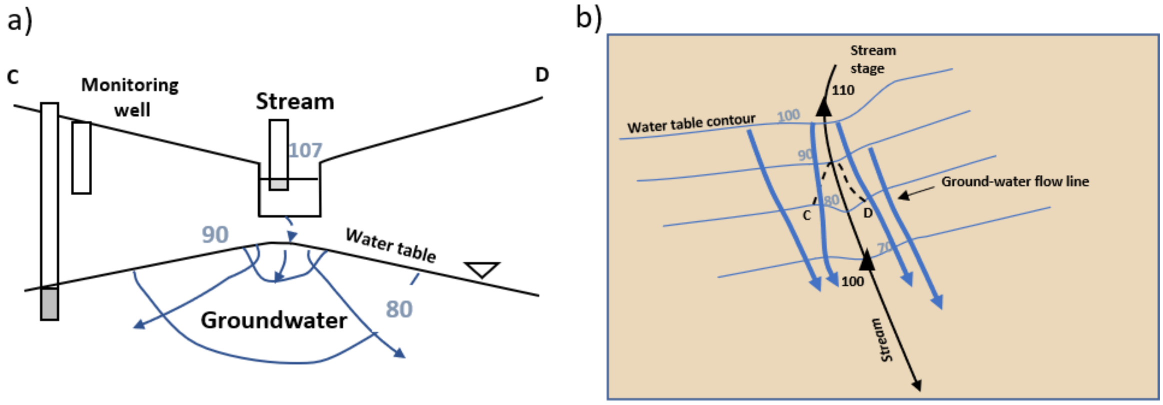


Figure 18 - Cross section and map views of an influent stream (losing) perched above the groundwater flow system. The dashed arrow shows stream water percolating through the vadose zone to the water table. Black triangles are stream stage measurement locations and black numbers are stream stages. Equipotential lines and relative head values are shown as blue water table contours. Groundwater flow is in the direction indicated by blue arrows. Aquifer conditions are assumed to be isotropic and homogeneous. Monitoring wells are open only at the bottom. a) Cross section showing a groundwater mound formed beneath the channel. b) A map view showing equipotential lines that cross the stream, bowing downstream. Groundwater flow lines (blue arrows) are shown in this example as nearly parallel to the channel. The location of the cross section (C-D) is approximated by the position of the dashed black line (Woessner, 2020).

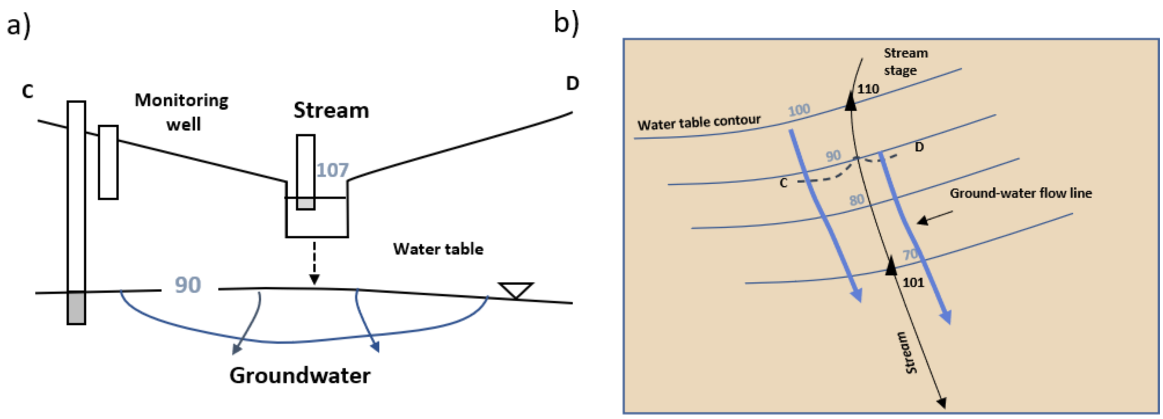


Figure 19 - Losing (influent) stream perched above the groundwater flow system where stream recharge rates and hydrologic conditions cause minimal groundwater mounding beneath the stream. Groundwater flowlines (blue arrows) are shown as parallel to the channel in this example. Monitoring wells are open only at the bottom. a) The dashed arrow represents stream water percolating to the water table. b) Black triangles are stream stage locations and black numbers are stream stages. The water table is below the stream bottom and stream stages do not represent the water table elevation. The location of the cross section (C-D) is approximated by the position of the dashed black line (Woessner, 2020).

2.3 Flow-through Stream

Flow-through streams occur when the water table adjacent to the stream is higher on one side of the channel and lower on the opposite side of the channel (Figure 20).

Groundwater enters the stream through one section of stream bed and channel bank, and stream water leaves the opposite bed and bank recharging the local groundwater system (Figure 20a). Equipotential lines are parallel to the stream channel (Figure 20b). Under these conditions, the stream stage is connected to the groundwater system and represents the water table elevation.

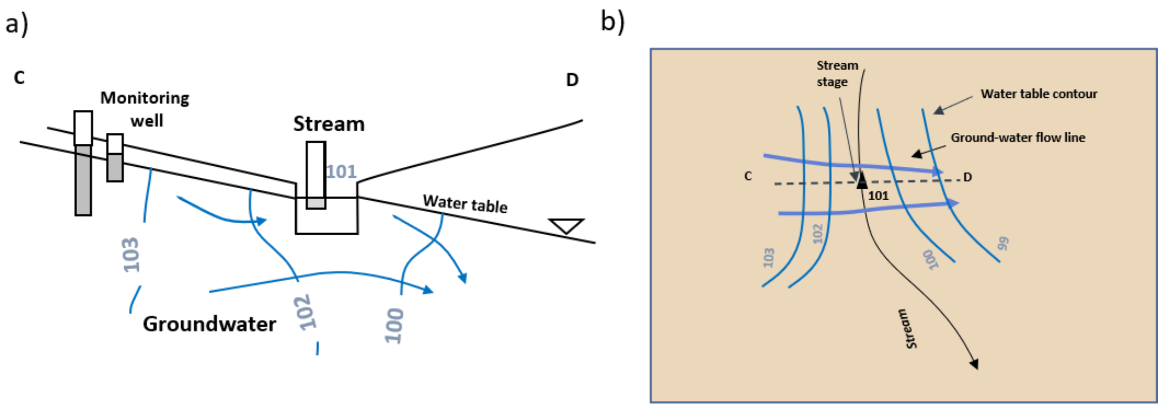


Figure 20 - Cross section and map views of a flow-through stream. The black triangle is a stage location and the black number is the stream stage. Equipotential lines and relative head values are shown as blue water table contours. Groundwater flow is in the direction indicated by blue arrows. Aquifer conditions are assumed to be isotropic and homogeneous. Monitoring wells are open only at the bottom. a) Cross section showing groundwater discharging on the left bank and recharging the adjacent groundwater system at the right bank. b) The map view shows equipotential lines are parallel to the stream channel. Groundwater flow lines are perpendicular to the stream channel. The location of the cross section is approximated by the position of the dashed black line (C-D) (Woessner, 2020).

2.4 Zero-Exchange Stream

When the local water table elevation mirrors the stream stage, groundwater flows parallel to the channel and zero-exchange conditions occur (Figures 21a). As no gradient is present between the stream and the groundwater system, water table contours cross the stream at right angles and flow is parallel to the channel (Figure 21b). In this setting, the stream stage represents the local water table elevation. This setting occurs when portions of the stream transition between effluent and influent conditions. Its extent and duration are not commonly discussed in the literature.

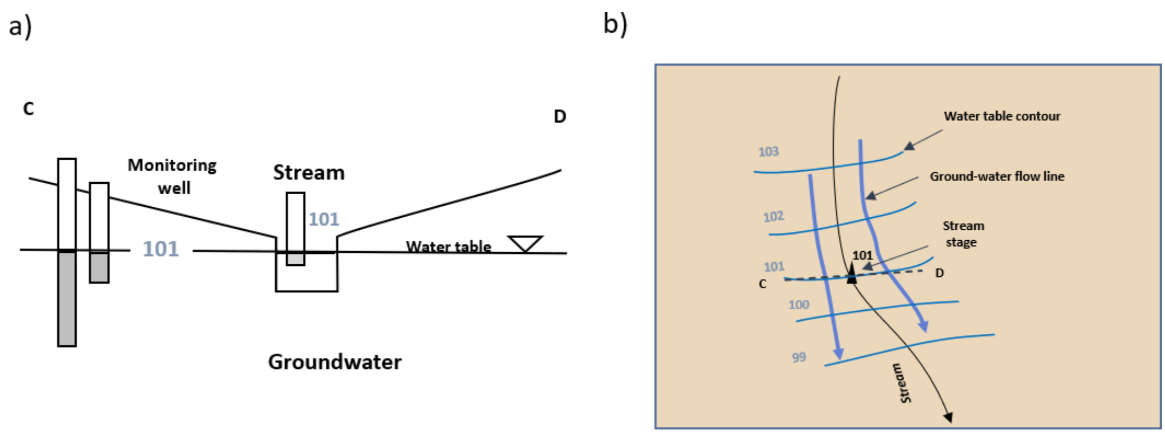


Figure 21 Cross section and map views of a zero-exchange (parallel flow) conditions. The black triangle is the location of the stream stage and the black number is the stage measurement. Equipotential lines and relative head values are shown as blue water table contours. Groundwater flow is in the direction indicated by blue arrows. Aquifer conditions are assumed to be isotropic and homogeneous. Monitoring wells are open only at the bottom. a) Cross section showing zero-exchange. The stream stage and groundwater head are equal at this location and groundwater flow is parallel to the stream. b) Map view showing equipotential lines cross the channel at right angles. Groundwater flow lines are parallel to the stream channel. In this example, the stream stage reflects the local water table. The location of the cross section (C-D) is approximated by the position of the dashed black line (Woessner, 2020).

2.5 Perennial, Intermittent and Ephemeral Streams

In general, streamflow conditions can be described in terms of the presence/duration of flow, as continuous and discontinuous, and in most cases, are directly related to the nature of the groundwater exchange process. Streams referred to as perennial have flows year-round and are most often supported by base flow (effluent conditions) when runoff is insufficient to maintain discharge (Figure 22a). Intermittent streams flow only when sufficient groundwater discharge (effluent) and/or precipitation support flows. However, there are periods when the water table drops below the effluent portions of the channel and streamflow recharges the underlying groundwater. If channel leakage is high, all or portions of the channel become dry for a period of time. Ephemeral streams only flow in response to runoff as the water table generally remains below the channel bottom (Figure 22c). Most of the time influent conditions occur during channel flow. These streams remain dry when no runoff occurs.

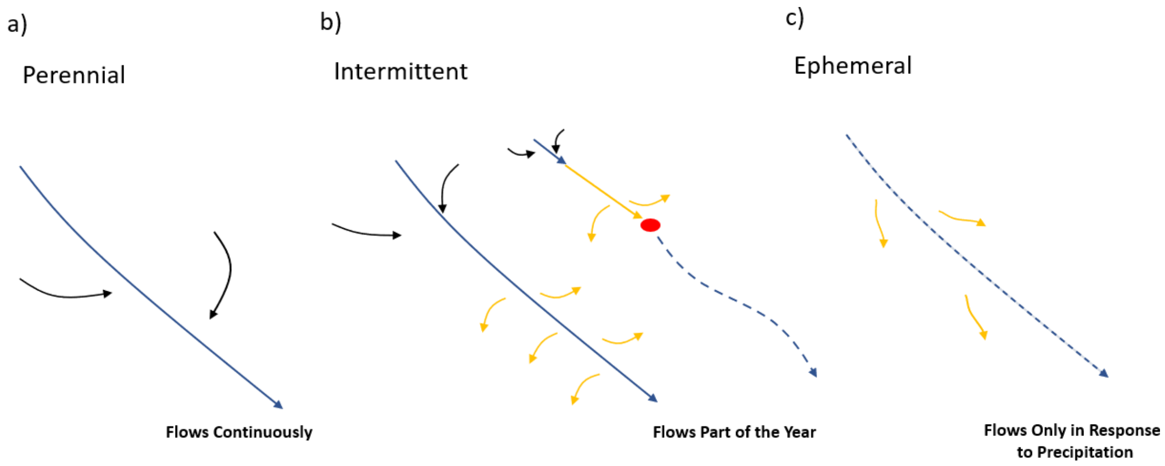


Figure 22 - Map view of stream channels (long blue solid and dashed arrows; surface flow is from the upper left to the lower right) illustrating groundwater exchange in three settings. Small arrows indicate general exchange directions of groundwater (black arrows) and surface water (orange arrows). a) Perennial effluent stream where groundwater discharges to the stream and streamflow is maintained year-round. b) Intermittent streamflow is driven by precipitation and groundwater discharge. Sections of the stream may be gaining or losing during periods of full channel flow (left diagram). During a portion of the year groundwater inflows decrease and stream leakage (orange channel arrow) increases (right diagram of b) such that all or portions of the channel will become dry (red dot and dashed blue line). c) Ephemeral streams contain no streamflow until precipitation causes runoff. Streamflow seeps into the channel (influent). Channels become dry when runoff ceases and/or seepage rates exceed streamflow (Woessner, 2020).

2.6 Exchange at the Watershed/Basin Scale

Hydrogeologists generally characterize rivers, streams, creeks, brooks, canals and other linear flowing surface-water features as either receiving groundwater discharge or acting as a source of groundwater recharge (Figures 16 and 17). On a regional scale, this categorization may be appropriate, yet too simplified when assessing conditions at sites with smaller areas (e.g., Figure 15).

Toth (1963) addressed how regional/watershed scale groundwater systems covering large areas behave when recharge and geological conditions allow water tables to build up and reflect the topographic highs and lows of the landscape. In these settings if general groundwater discharge areas are assumed to represent exchange with a surface-water feature (river, lake or wetland), landscape-scale exchange locations and processes can be shown as illustrated in Figure 11 and 12. Toth examined how flow path length, water table topography, water table slope, anisotropy, and aquifer thickness impact groundwater exchange with discharge locations (viewed here as surface-water features). His work showed regional gaining streams can receive groundwater flow from a large regional flow system, an intermediate system, and/or a local flow system (Figure 11). In Toth's (1963) landscape context, influent surface-water conditions occur if features located at topographic highs and groundwater divides recharged the groundwater. However, as stated previously, when combinations of recharge and geologic conditions do not create water table configurations that reflect surface topography, regional groundwater conditions may dominate exchange, and influent conditions can occur at topographic lows as well as features located in areas of groundwater divides (Figure 11 and Figure 12).

At the basin scale, when streams are dominated by effluent conditions, the flow records at stream gauging stations can be used to quantify a portion of the annual contributions of groundwater to the stream. Stream base flow represents a net addition of groundwater to the stream within the basin area above the gauging station (Figure 23). When streamflow is measured and other components of the basin water budget computed, groundwater contributions to streamflow are quantified (as discussed in Section 5.2).

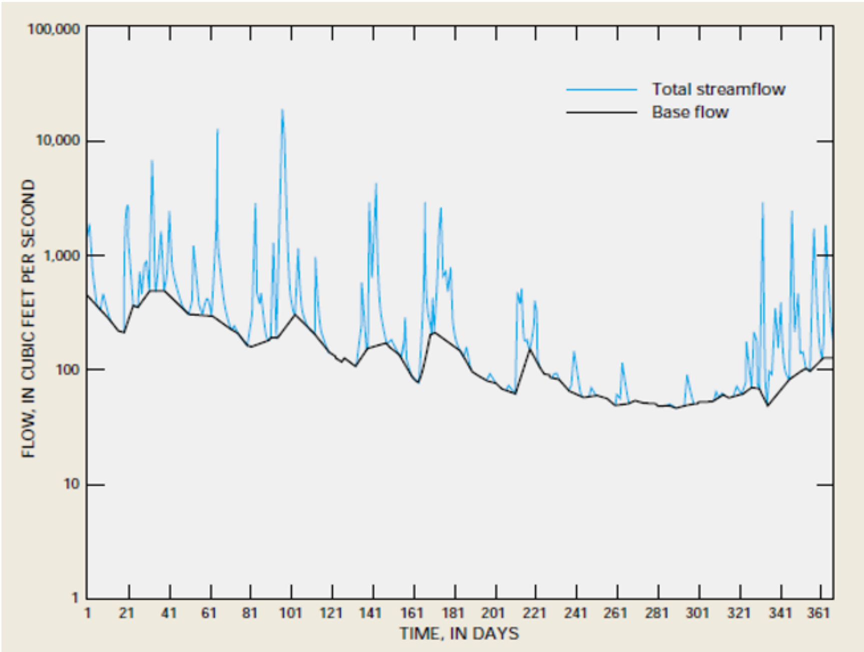


Figure 23 - A stream hydrograph of the Homochitto River, Mississippi, USA, showing total discharge over 361 days. The graph area below the back line represents base flow, that portion of the total discharge contributed by groundwater discharge in the basin area above the location of the gauging station (Winter et al., 1998).

Certainly, regional exchange locations and rates vary over time. Stream hydrographs document this variation. In addition, the locations and rates of exchange with headwater tributaries and arid-land streams vary depending on their interaction with groundwater and runoff.

2.7 Exchange at the Valley Segment/River Corridor Scale

The valley-segment scale includes exchange within the entire valley corridor over a few to 10's of kilometers of stream channels and associated riparian zones, active and inactive floodplains, and older river terraces. Woessner (2000) described this multiple featured area as the fluvial plain (Figure 24). Except for bedrock lined channel settings, generally, the fluvial-plain sediments include deposits of anisotropic and heterogeneous materials.

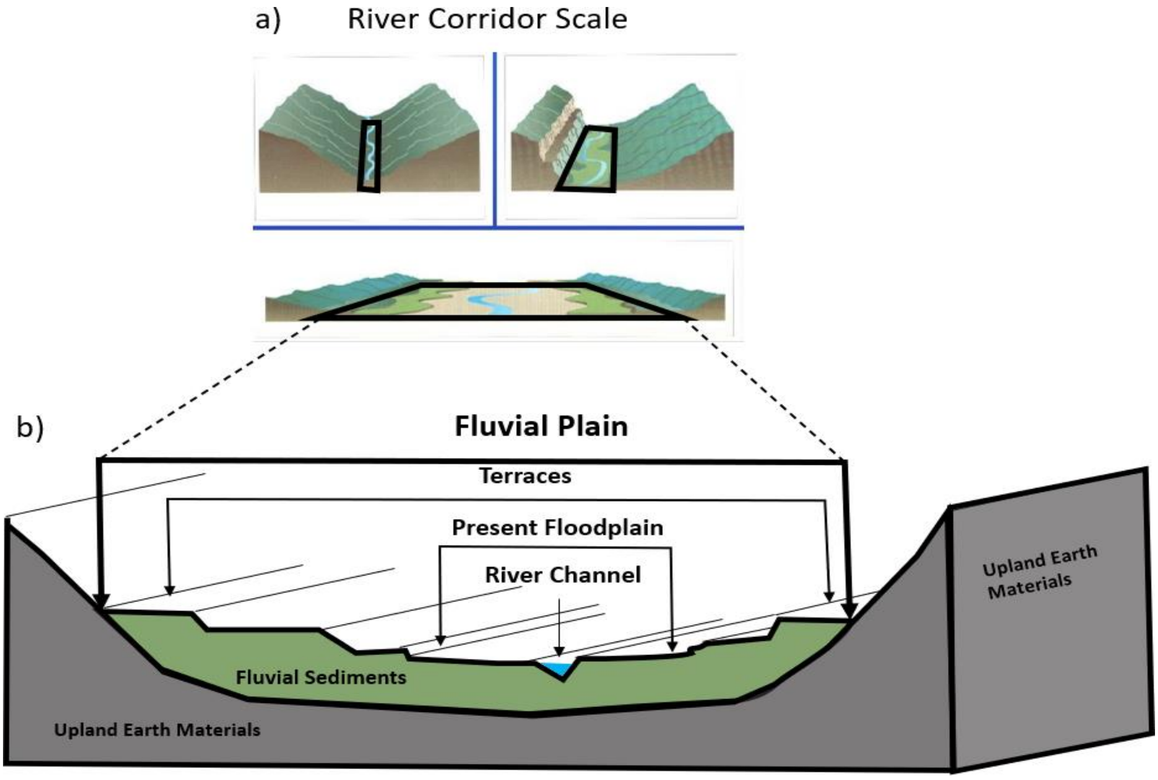


Figure 24 - Valley segment/river corridor and the fluvial plain. a) The river corridor includes the channel and surrounding lowlands (black rectangle) and represents kilometers of stream channel length. b) The landscape making up the river corridor is referred to as the fluvial plain. It is usually composed of unconsolidated fluvial sediments and includes the river channel, present floodplain, and older river terraces. It is bounded by the adjacent uplands that can be composed of unconsolidated or consolidated earth materials (gray) (modified from USEPA, 2019).

Exchange at this scale is most commonly quantified as water-balance-computed changes in streamflow between two stream gauging stations or sites. In the simplest form, an upgradient and downgradient stream gauging site are selected, the discharge is determined at both sites and a water budget is computed. The results determine if the river corridor section is gaining, losing or showing no change in flow (Figure 25). Multiple types of stream channel exchanges may be occurring over the selected river segments; however, the streamflow analyses will yield the net change for a given segment without identifying if exchange processes in some sections of the segment differ from the net exchange. Water budgets are more complex when additional sources or losses of water occur within the study section (Figure 25). It is necessary to quantify flow measurement errors to determine if measured flow differences are significant (e.g., Healy et al., 2007) as discussed in Section 5 of this book.

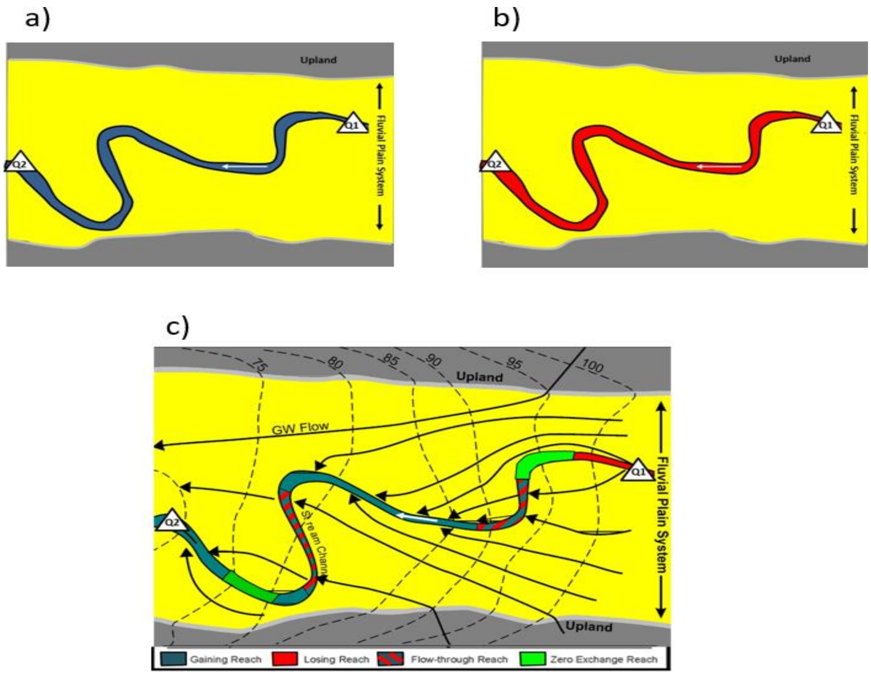


Figure 25 - Map views of a valley segment and two stream gauging locations, Q1 upstream and Q2 downstream (triangles). a) A gaining stream corridor where the measured upstream flow rate, Q1, is less than the downstream flow rate, Q2 (indicated by the gray-blue channel). b) A losing stream corridor where the measured upstream flow rate, Q1, is greater than the downstream flow rate, Q2 (indicated by the red stream channel). c) An example of a gaining stream corridor in which Q1<Q2; however, exchange within the segment is complex (Woessner, 2020).

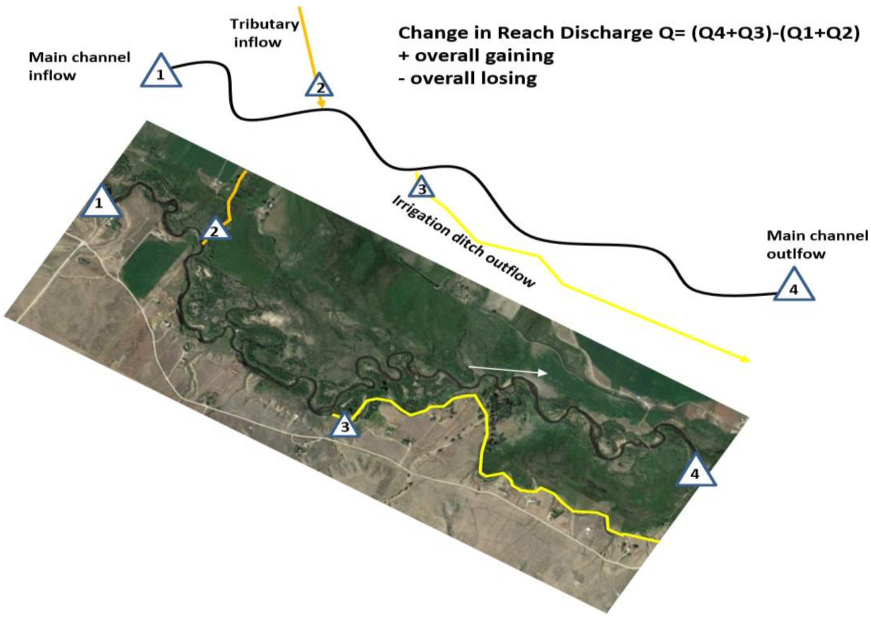


Figure 26 - Example of a 3 km stream corridor that has an inflow of tributary water at gauge 2, and a loss of water by an irrigation canal diversion at gauge 3. The dominant exchange process is computed by the equation shown in the figure. This assumes that no other significant inflows and outflows occur over the time period when flow is measured (e.g., evaporation, transpiration). The photo image is from Google Earth in 2015 (Woessner, 2020).

Characterizing exchange at the valley-segment/river-corridor scale lumps conditions into the two streamflow discharge measurements. Exchange may vary as streamflow changes over time. If river corridors are dominated by losing streamflow, stream water recharge will enter the surrounding fluvial and upland geologic materials. Some of this water may be recirculated in the fluvial plain and discharge back to the stream channel in a downstream section of the stream. In other settings, water may enter a larger regional groundwater system and not return to the river corridor.

2.8 Exchange at the Reach/Floodplain Scale

Groundwater and river exchanges viewed at the reach scale, 10's of meters to kilometers, can contain a single exchange process or multiple exchange conditions. Exchange at the reach scale is influenced by the fluvial processes generating the reach/floodplain sediments. Woessner (2000) noted that in most settings dominated by unconsolidated sediments where groundwater is flowing towards the stream reach, the exchange is influenced by the presence of a contrast in hydraulic conductivity of fluvial the sediments that generally directs groundwater flow down floodplain (Figure 27).

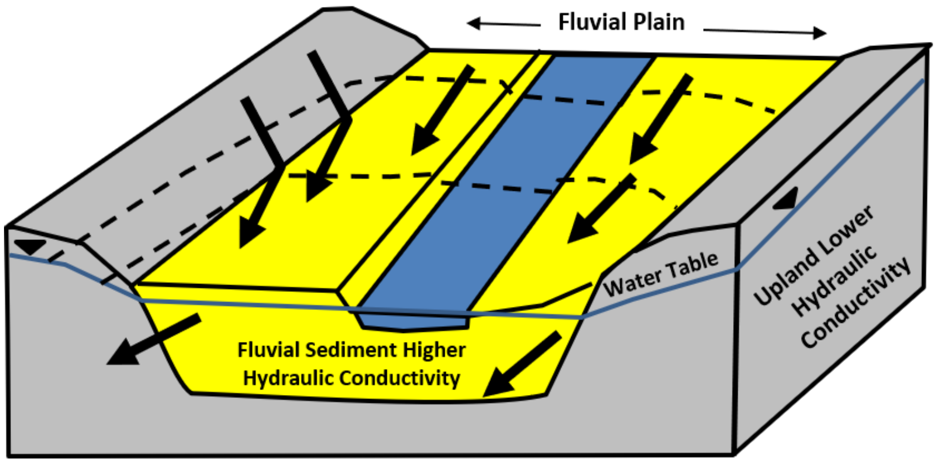


Figure 27 - Groundwater flow in the vicinity of a river reach. The river is flowing from the back of the block diagram to the front. Dashed lines are water table contours (heads decreasing down valley) determined from a network of monitoring wells and stream stage measurement locations. Black arrows are groundwater flow directions. Flow conditions are represented as isotropic and homogeneous. Groundwater flows from the uplands to the fluvial plain. The sloping fluvial plain and higher hydraulic conductivity fluvial sediments direct flow in the down-plain direction (modified from Woessner, 2000).

Quantifying exchange in a stream reach is most often based on how streamflow changes between two stream gauging locations used to define the reach (as was described in the valley segment exchange section above). In addition, the details of exchange along the reach is obtained by mapping the relationship of the stream stage to the water table in the vicinity of the reach (Figure 28).

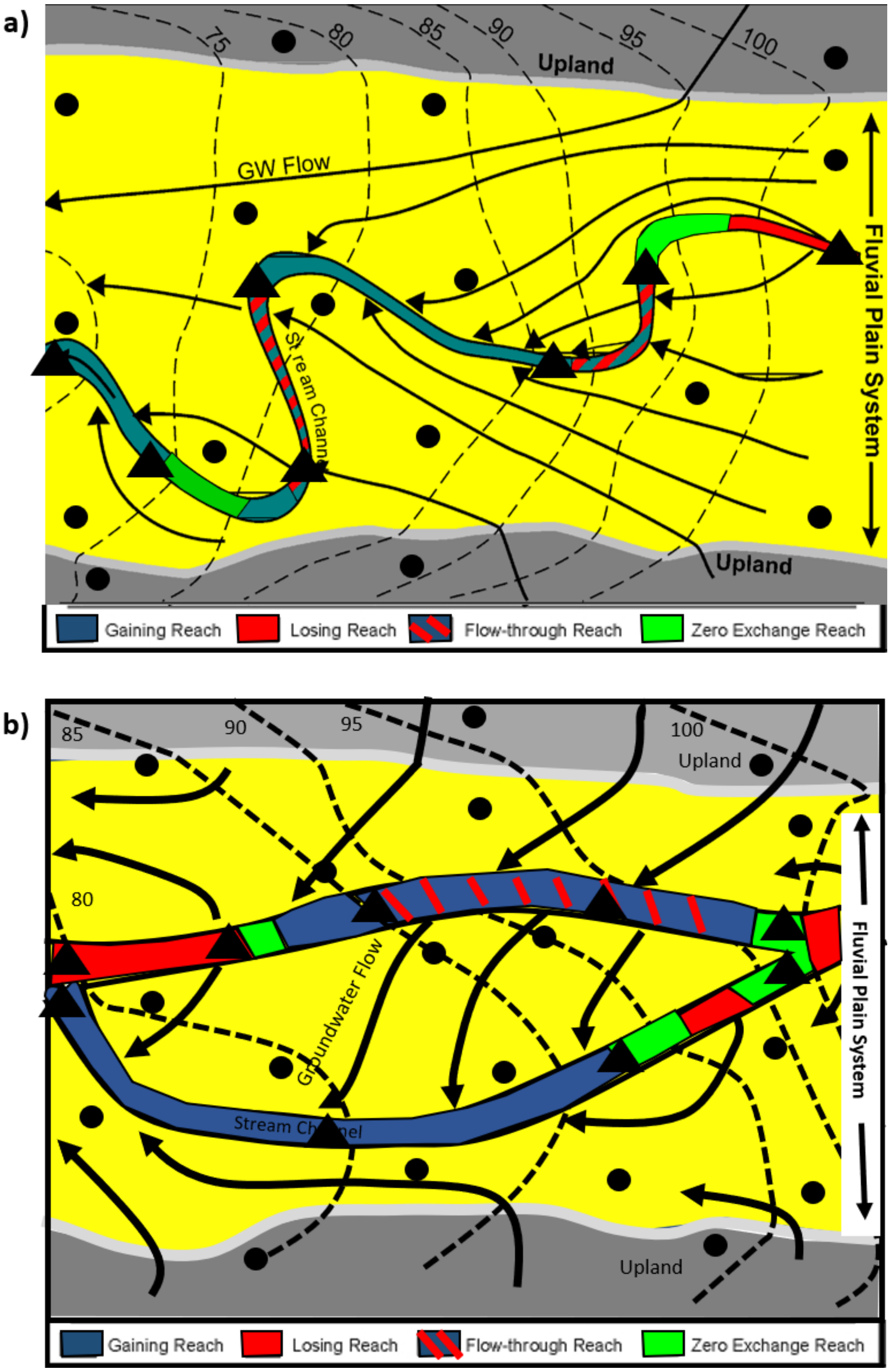


Figure 28 - Groundwater flow and exchange in portions of reaches defined as gaining, losing, flow-through and zero-exchange. Dashed lines are water table contours determined from a network of monitoring wells (black dots) and stream stage measurement locations (black triangles). Black arrows are groundwater flow directions. Flow conditions are represented as isotropic and homogeneous. Groundwater flow is from the uplands to the fluvial plain in these examples. a) A schematic of a meandering section of stream with reaches that show streamflow leaving and entering the channel. b) A schematic of a braided stream section showing complex exchange patterns (Woessner, 2018).

Water levels in groundwater monitoring networks can be paired with stream stage measurements to derive two-dimensional and three-dimensional groundwater exchange patterns and locations. Reach-scale groundwater exchange is often simulated in groundwater models when rates, locations and timing of exchange is required to meet modeling goals. A variety of approaches and tools are available to simulate exchange (e.g., Anderson et al., 2015; Cardenas, 2015).

As presented earlier, a stream reach that is dominated by groundwater discharge (effluent or gaining stream) has increasing discharge between the up-stream and down-stream observation points. Under these conditions, the stream may exhibit a change in water chemistry that represents the addition of a different water type to the stream water. In losing reaches of streams the stream chemistry generally remains fairly constant because it is not mixing with groundwater. Instead, the groundwater chemistry associated with the stream will reflect its mixing with the stream water flowing into the groundwater system. Flow-through reaches may show chemical changes in both the surface water and the downgradient groundwater.

When the surface water and groundwater exchange, the surface water and groundwater may exhibit changes in ionic and isotopic compositions, and/or temperature (e.g., Healy et al., 2007; Boana et al., 2014). If there are significant contrasts in either the surface water or groundwater chemical compositions, chemical-mixing models may yield additional information on rates and locations of exchange at the reach scale, and/or reflect the geochemical signature of discharging groundwater. Section 5 describes appropriate methodologies for using such information to evaluate exchange locations and rates.

In some losing and flow-through stream reaches, stream water that recharges groundwater in the floodplain follows a flow path to an effluent stream section where the water that originated from the stream re-enters the stream. This process is observed over short flow paths associated with specific channel features as well as at the larger floodplain scale (Winter et al., 1998; Woessner, 2000; Diehl, 2004) (Figure 29). Water that flows from the stream through the groundwater system and returns to the stream is referred to as hyporheic water and the area where it occurs as a hyporheic zone (e.g., Woessner, 2017). At the reach scale this exchange process is three dimensional and influences the geochemical and ecological conditions in portions of the floodplain (e.g., Poole et al., 2008; Buss et al., 2009; Boana et al., 2014).

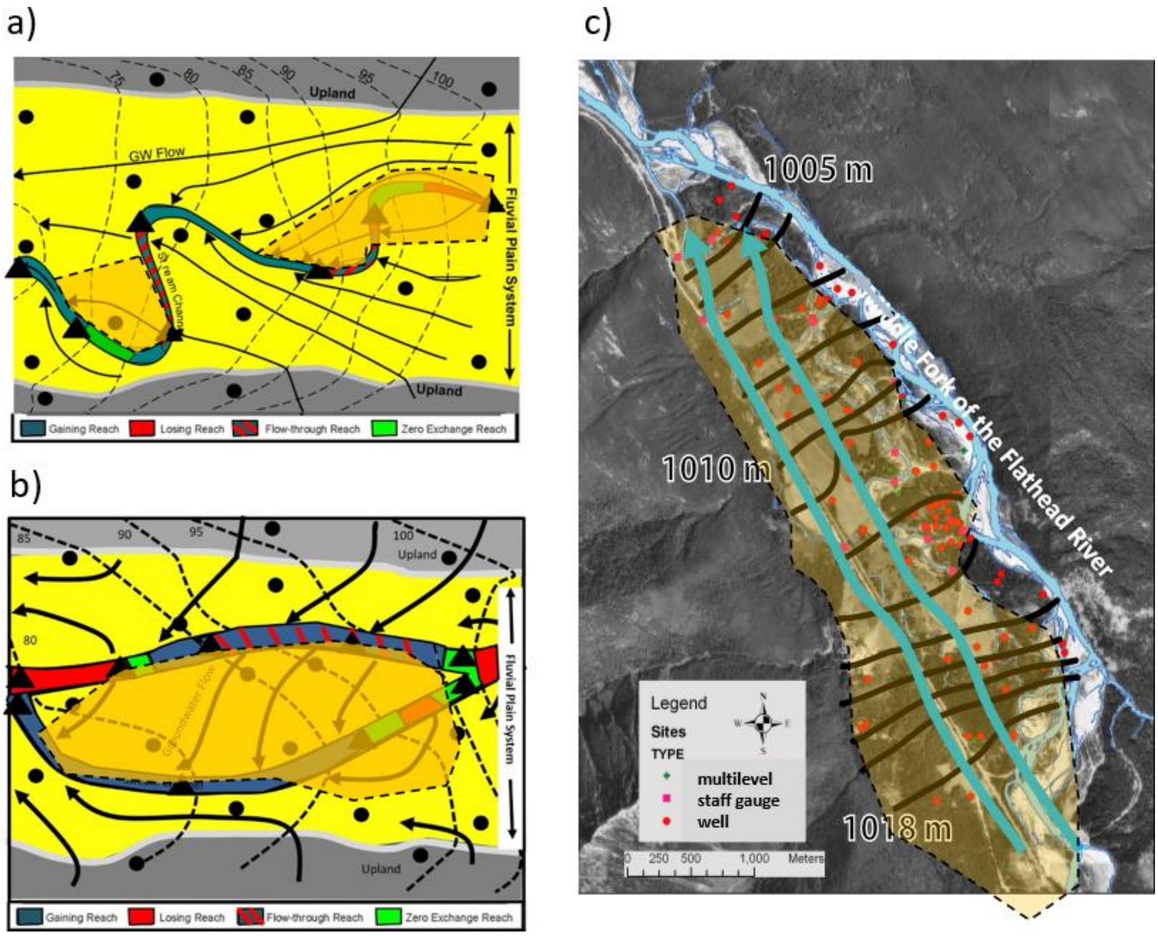


Figure 29 - Examples of reach scale hyporheic flow zones (orange area with the dashed back line boundary). a) Schematic of a meandering reach and the locations of hyporheic flows (Figure 28a). b) Schematic of a braided reach and the locations of hyporheic flows (Figure 28b). c) Water table map (black equipotential lines) of the shallow groundwater system in the eight-kilometer-long sand, gravel and cobble floodplain of the Middle Fork of the Flathead River, Montana, USA. Long light blue arrows represent hyporheic flow paths from the upstream (southeast) portion of the channel. The river is losing flow to the adjacent floodplain. Locations of multilevel piezometers, staff gauges and wells are shown (modified from Diehl, 2004).

Again, it is important to note that, in some settings, even though the overall exchange process may be described as effluent or influent, multiple groundwater-surface water exchanges may be taking place between the defined reach boundaries (Figure 28). If the exchange investigations focus on shorter lengths of channel, different exchange processes may dominate.

Reach-scale exchanges are dynamic and can vary temporally and spatially. Variations in stream stage and groundwater flow systems in response to precipitation events, snowmelt, and droughts, as well as surface water and groundwater management can impact the type, location and flow rates of reach-scale exchange process.

2.9 Exchange at the Channel Scale

The conceptual models presented in Figures 15 through 21 suggest that groundwater can enter or leave a stream channel though its bed and banks at multiple locations along a channel (Figure 28). At the channel scale, in addition to the groundwater exchanges previously described as effluent, influent, flow-through and zero-exchange, river water also circulates into the bed and bank sediments and associated floodplain (Figure 29). When head differences between the stream stage and head in the bed, banks and floodplain contrast, stream water can flow into shallow groundwater and circulate back into the stream as described above. Stream water that leaves the stream channel bed and banks and then returns to the channel at a downstream location is defined as hyporheic water (Figure 30). Stohedahl et al. (2013) view this hyporheic exchange as driven by changes in channel geomorphology (e.g., meander- and bar-driven); variations in the channel profile (i.e., pool- and riffle-driven), and by the heterogeneous bed surface (bottom-driven). Channel segments can be dominated by hyporheic exchange, groundwater exchange or both depending on the surface water and groundwater hydraulics operating in the segment. Exchange investigations at the channel scale are mostly focused on hyporheic exchange and areas of groundwater discharge (Figure 30).

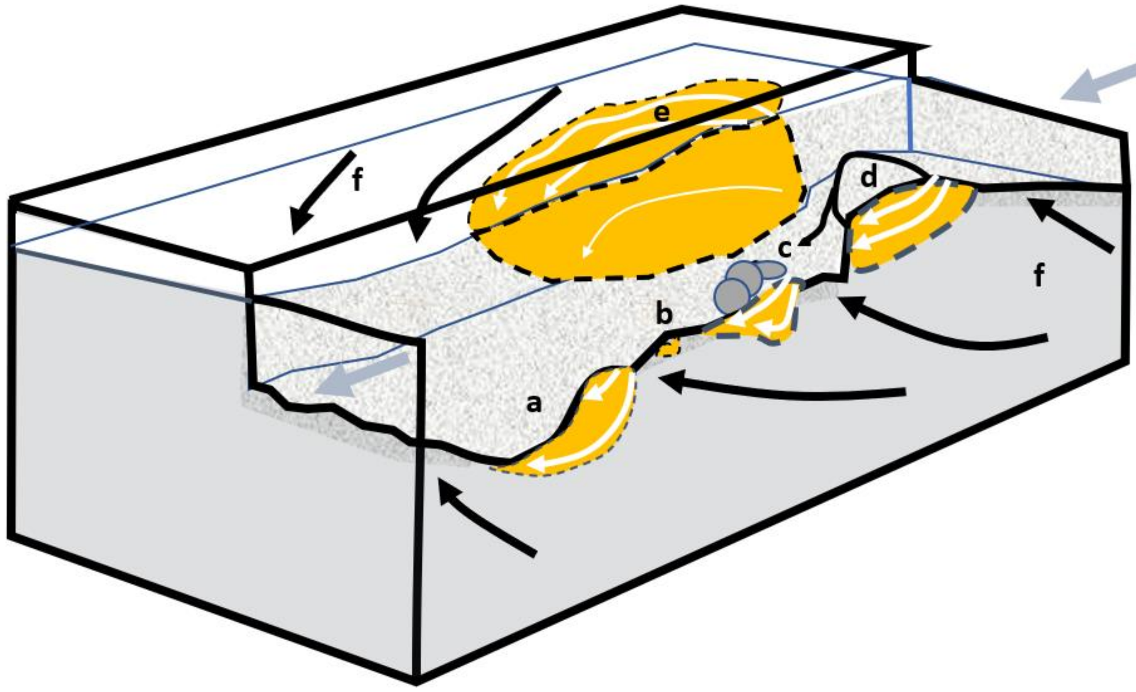


Figure 30 - Stream channel (stippled pattern) with surface flow from right to left (light blue-gray arrows). The hyporheic zone locations are shown in orange. Stream water circulating into the bed is represented by white arrows and outlined with dashed lines. Stream water leaves the stream and mixes with groundwater then re-enters the stream: a. Pool and riffle sequence; b. Bedform; c. Obstruction; d. Mid-channel bar; e. stream water flowing into the adjacent floodplain and returning to the stream; f. Local, intermediate and regional groundwater discharge is focused in portions of the stream bottom (modified from Woessner, 2000).

Researchers also recognize that the extent and magnitude of hyporheic exchange can be physically difficult to document. Measuring and mapping exchanges in stream channels increases in complexity as study sites become smaller (e.g., Woessner, 2000). Often studies rely on in-channel instruments, floodplain monitoring well networks, chemical analyses of stream water and hyporheic water, along with numerical modeling of groundwater and surface-water dynamics. Instrumentation typically includes mini-piezometers, seepage meters, and temperature monitoring and modeling (e.g., LaBaugh and Rosenberry, 2008; Woessner, 2017; Weight and Woessner, 2019). Differences between surface water and regional groundwater, including temperature, chemistry, natural and environmental isotopes, and radon 222, are often used to identify the hyporheic waters illustrated in Figure 30 (e.g., Healy et al., 2007; Boana et al., 2014) and discussed in Section 5 of this book. In some cases, transition zones occur where discharging regional groundwater appears to fully or partially mix with infiltrating stream water. When groundwater discharge to the channel occurs at a sufficiently high rate, all water in the channel bed and bank sediments may be dominated by the groundwater chemistry thus signaling the absence of hyporheic flow (e.g., Cardenas and Wilson, 2006). However, in local losing-channel settings (e.g., riffles), bed and bank water and adjacent floodplain water will be dominated by river water characteristics (Figure 30). Simulations of hyporheic flow have been used to assess the likely extent of hyporheic zones and subsurface flow complexity (e.g., Woessner, 2000; Cardenas and Wilson, 2006; Tonina and Buffington, 2007; Boano et al., 2014).

2.10 Hyporheic Exchange: Links to Physical and Ecological Systems

Hyporheic exchange occurs as stream water circulates into and out of the stream channel, bed, and banks, to mix with the adjacent groundwater system. Excellent overviews and specifics regarding the physical and ecological role of hyporheic systems in stream and groundwater settings are provided in many works (e.g., Winter et al., 1998; Buss et al., 2009; Boano et al., 2014; Cardenas, 2015; Ward, 2016; Woessner, 2017; Hauer and Lamberti, 2017; and Conant et al., 2019).

The hyporheic zone encompasses that portion of the groundwater system where a mixture of surface water and groundwater occur as shown in Figure 29 and Figure 30 (Woessner, 2017). Conant and others (2019) refer to the hyporheic zone as a transition zone between surface water and groundwater systems in which various biogeochemical processes occur. In addition to circulating and mixing waters, hyporheic systems also create habitat and refuge for macroinvertebrates, microbes, and fish. These aquatic ecotones are both influenced by water chemistry and stream biota (Figure 31). The circulating waters process carbon, nutrients, and solutes, while fueling ecosystem metabolism (Woessner, 2017). Hyporheic exchange is viewed by aquatic ecologists as an ecotone between groundwater and river ecosystems characterized by hydrologic, zoologic, chemical and metabolic features (e.g., Burke and Gonser, 1997; Ward, 2015; Hauer and Lamberti, 2017).

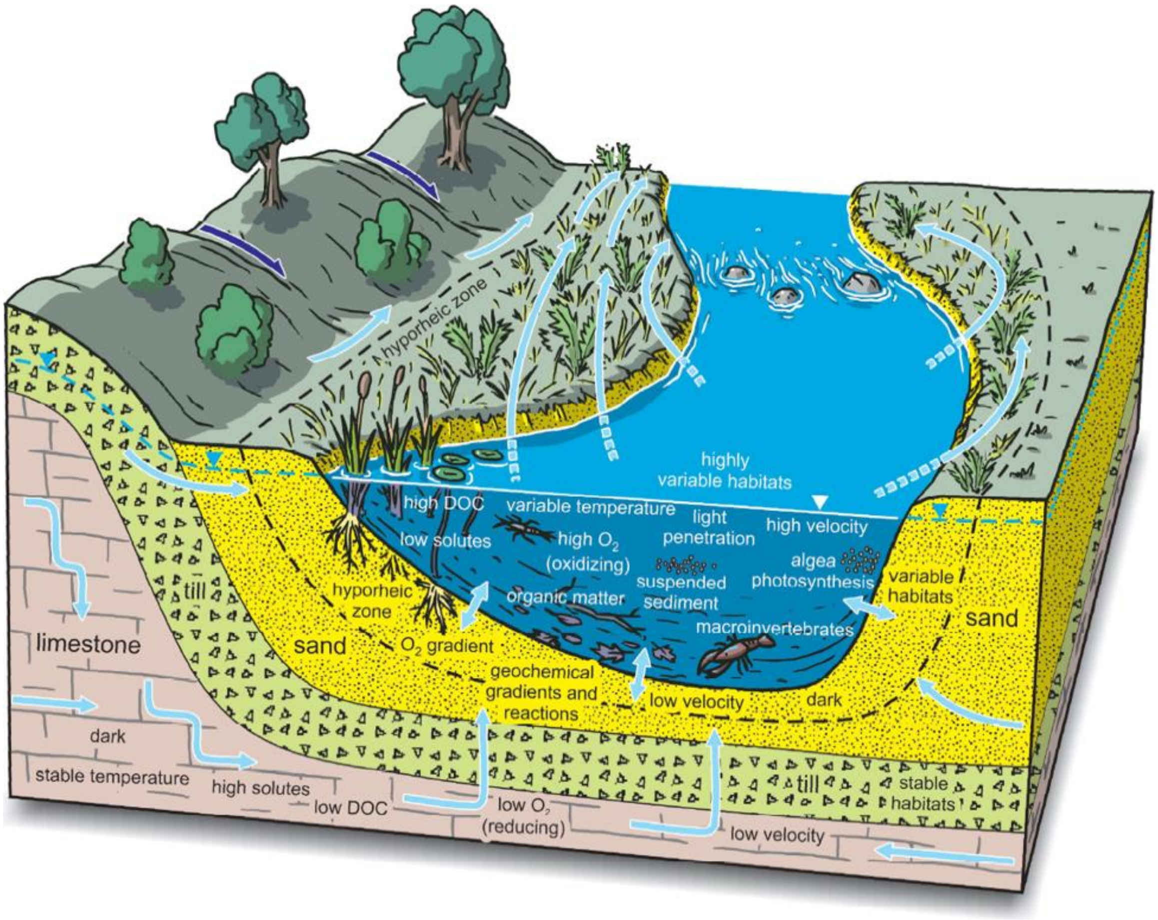


Figure 31 - Schematic of the physical, biogeochemical and ecological components of a hyporheic zone, stream and groundwater system. The outer boundary of the hyporheic zone is represented by the dashed black line. Light blue arrows represent groundwater flow to the hyporheic zone. Double arrows show exchange of river water with the sediments (hyporheic flow). Arrows with dashed ends show hyporheic flow from the channel into the bank and floodplain, water that returns to the stream at some point downstream (Hinton, 2014).

The hyporheic exchange process is enhanced in the presence of channel complexity (e.g., braided and meandering channels), bars, variations in channel topography, the distribution and magnitude of permeable bed sediments, in-channel and bank vegetation, and variations in flow regimes (e.g., Harvey and Bencala, 1993; Carling et al., 1999; Woessner, 2000; Malcolm et al., 2005; Storey et al., 2003; Buffington et al., 2004; Anderson et al., 2005; Gooseff et al., 2006; Worman et al., 2007; Cardenas and Wilson 2007abc; Greig et al., 2007; Tonina and Buffington, 2007; Cardenas, 2008ab; Arrigoni et al., 2008; Cardenas, 2009; Bean et al., 2013; Boana et al., 2014). In contrast, stream modifications that reduce channel complexity (e.g., channelizing, and damming) often result in degrading hyporheic exchange areas, locations and rates.

The hyporheic zone can form a temporary or permanent refuge and habitat for aquatic organisms including fish and invertebrates (e.g., Boulton, 2007; Datry and Larned, 2008; Stubbington et al., 2009); Buss et al., 2009; Ward, 2016) including zoobentos in various stages of life histories (e.g., Hauer and Lamberti, 2017; Lamberti and Hauer, 2017). It also cycles solutes, including nitrates and phosphorus, and organic matter between the river and hyporheic zone (e.g., Fisher et al., 1998; Boulton, 2007). Hyporheic zones can act to modify surface water chemistry and focus biological production at locations where hyporheic water is discharging back to the stream known as hot spots (e.g., Valett et al.,1990, 1994; Coleman and Dahm, 1990; Pepin and Hauer, 2002; Boulton, 2007). In some settings, contaminated groundwater discharging to a stream or contaminated surface water circulating in the hyporheic zone may be altered by processes operating in the hyporheic exchange system (e.g., Conant, 2004; Conant et al. 2019).

Hyporheic zones are studied by developing conceptual models based on the exchange literature, modeling and field data. Initial conceptual models are tested and revised after undertaking a site characterization program based on specifics of the physical setting and results of biogeochemical sampling and analyses (e.g., LaBaugh and Rosenberry, 2008; Boano et al, 2014; Buss et al., 2009; Cardenas, 2015; Woessner, 2017; Weight and Woessner, 2019).

3 Lakes and Groundwater Exchange

Groundwater exchange with lakes can be conceptualized using the cross sections presented in Figures 4 through 8. The surface water represented in these generic conceptual models can be viewed as representing ponds and lakes at various scales. Following Winter and others (1998), lake-groundwater exchange is shown using five, cross-sectional conceptual models (Figure 32a-e). A mixed-exchange condition has been added to the lake conceptual model and the zero-exchange scenario has been omitted as it is not likely to occur or be identified in most settings. Early conceptual models of lakes assumed they were isolated from the groundwater systems by low conductivity bottom sediments. As groundwater-lake research expanded, the systems were recognized as interconnected (e.g., Born et al., 1974).

Lakes also tend to trap sediments from runoff, streamflow and or shoreline erosion. These deposits typically cover all or portions of the lake bottom and are usually organic-rich, finer-grained and of lower hydraulic conductivity than the materials in which the lake formed. The presence of low hydraulic conductivity sediments can limit the rates and locations of exchange. For the conceptual models of lake and groundwater exchange presented here, it is assumed that exchange is a function of the location of the lake in the adjacent groundwater system and that bottom sediment characteristics do not control overall groundwater exchange conditions. A discussion of more complex exchange settings follows the initial development of exchange models.

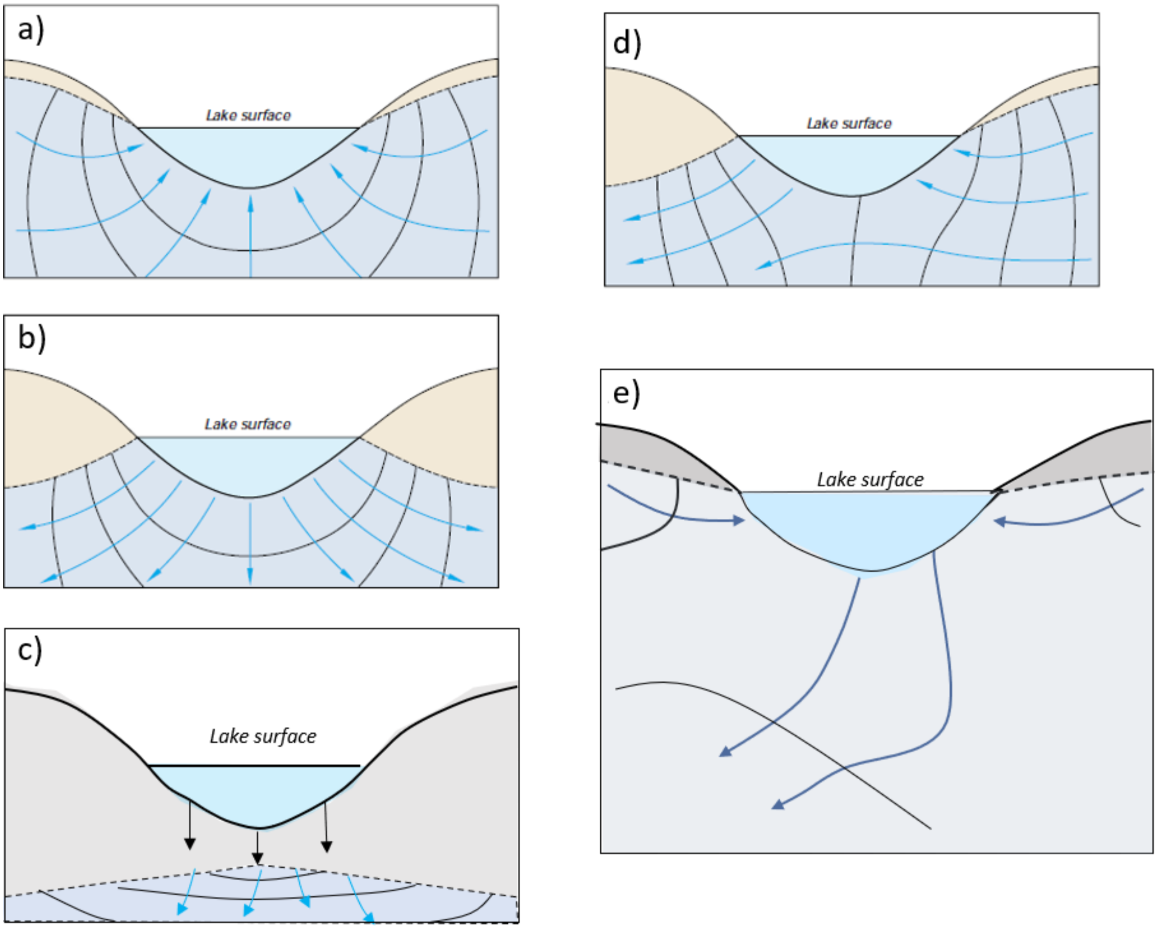


Figure 32 - Conceptual cross sections of lake-groundwater exchange. Blue arrows represent groundwater flow. Black lines are equipotential lines. The water table is a black dashed line. Conditions are isotropic and homogeneous. a) Effluent or gaining lake. b) Influent or losing lake. c) Influent or losing lake perched above the water table (black arrows represent leakage). d) Flow-through lake. e) Mixed exchange lake (Winter et al., 1998 and Woessner, 2020).

3.1 Effluent or Gaining Lake

A lake located in a groundwater flow system in which all groundwater flow is into the lake is an effluent or gaining lake (Figures 32a and 33). The lake surface is an expression of the water table. In this setting, flow discharging to the lake causes the lake level to rise unless it is balanced by loss of water from the lake by way of direct evaporation, evapotranspiration or surface-water outflow. Water levels in the lake adjust in response to changes in the lake water budget.

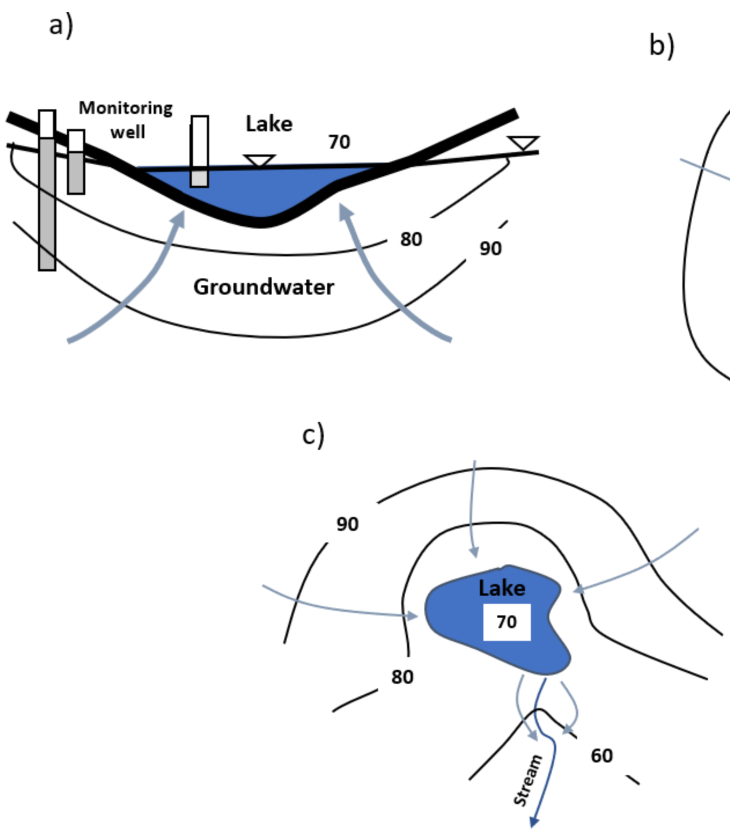


Figure 33 - Cross section and map views of effluent (gaining) lake exchange. Equipotential lines and relative head values are shown in black. Groundwater flow is in the direction indicated by blue arrows. Aquifer conditions are assumed to be isotropic and homogeneous. Monitoring wells are open at the bottom. a) Cross sectional representation showing an upward groundwater gradient and groundwater discharging to the lake. Lake stage is shown as a water level on the vertical rectangle. b) A map view showing equipotential lines and groundwater flow converging at the lake. c) A map showing an effluent lake that has a stream discharge. Some groundwater may flow from the lake to the stream under these conditions (Woessner, 2020).

3.2 Influent or Losing Lake

Influent or losing lakes leak lake water into the underlying and adjacent groundwater system (Figure 32b and c, Figure 34 and Figure 35). Hydraulic gradients between the lake surface and groundwater are downward. When earth materials are fully saturated, the lake is directly connected to the groundwater system as the lake loses water. Under these conditions the lake surface reflects the elevation of the water table. In contrast, when the zone of saturation is disconnected from the lake, water percolates to the underlying groundwater system (Figure 35). Influent conditions may be affected by lower permeability lake sediments restricting the rate of leakage to the groundwater system. The lake surface elevation under such conditions does not represent the water table. If leakage rates are large and hydraulic properties of the sediment low, mounding of groundwater below a perched lake can occur.

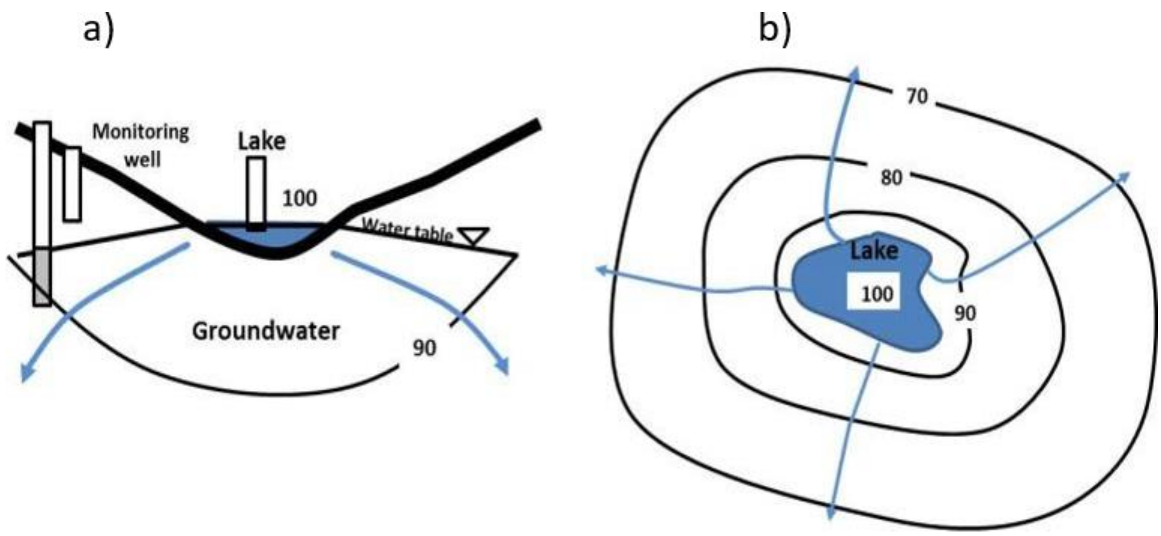


Figure 34 - Cross section and map view of influent (losing) lake exchange. Equipotential lines and relative head values are shown in black. Groundwater flow is in the direction indicated by blue arrows. Aquifer conditions are assumed to be isotropic and homogeneous. Monitoring wells are open at the bottom. a) Cross sectional representation showing a downward groundwater gradient and groundwater being recharged by the lake. Lake stage is shown as a water level on the vertical rectangle. b) A map view showing equipotential lines and groundwater flow diverging from the lake (Woessner, 2020).

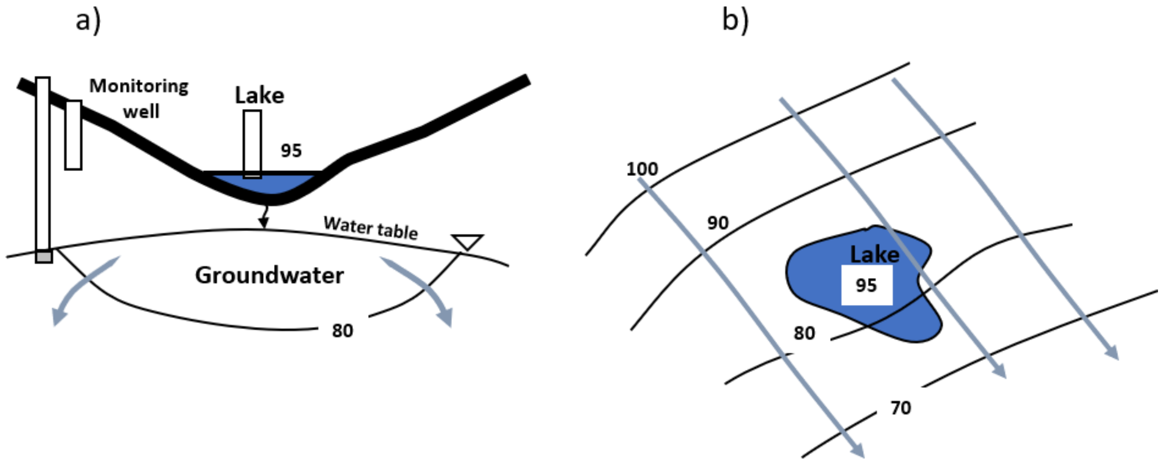


Figure 35 - Cross section and map view of influent (losing) lake exchange with the lake perched above the water table. Lake water percolating through the vadose zone is shown by a small black arrow beneath the lake. Equipotential lines and relative head values are shown in black. Groundwater flow is in the direction indicated by blue arrows. Aquifer conditions are assumed to be isotropic and homogeneous. Monitoring wells are open at the bottom. a) Cross sectional representation showing a downward groundwater gradient and the water table disconnected from the lake. The groundwater is being recharged by the lake. Lake stage is shown as a water level on the vertical rectangle. b) A map view showing equipotential lines and groundwater moving underneath the lake (Woessner, 2020).

3.3 Flow-through Lake

Flow-through lakes occur when the water table is higher on one side of the lake than the other, creating a gradient for groundwater to enter and leave the lake (Figure 32d and Figure 36). In some settings these lakes have no surface-water outlet or inlet. The lake surface represents the elevation of the local water table.

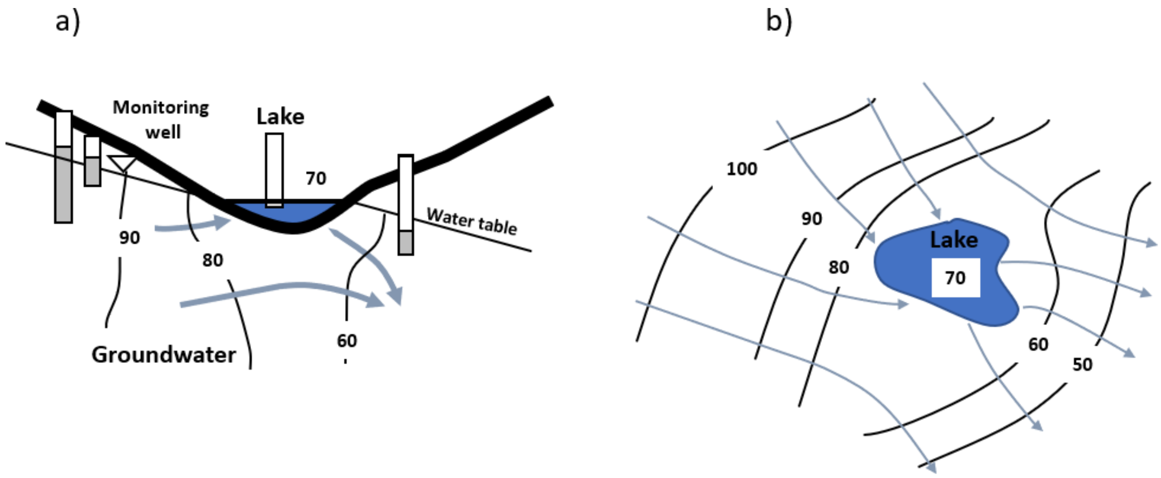


Figure 36 - Cross section and map view of flow-through lake exchange. Equipotential lines and relative head values are shown in black. Groundwater flow is in the direction indicated by blue arrows. Aquifer conditions are assumed to be isotropic and homogeneous. Monitoring wells are open at the bottom. a) Cross sectional representation showing an upward groundwater gradient at the up-gradient side (left) and a downward gradient as lake water flows into the groundwater system (right). Lake stage is shown as a water level on the vertical rectangle. b) A map view showing equipotential lines and groundwater flow converging at the lake at the up-gradient side and diverging from the lake on the downgradient side (Woessner, 2020).

3.4 Mixed Exchange Lakes

A mixed exchange lake suggests that the lake system is dominated by groundwater flowing into the lake; however, lake water flows through the bottom into the underlying groundwater system. Mixed exchange lakes usually occur where variations in lake bottom sediment properties and the presence of lower head values in earth material result in the loss of water from the lake. The mixed term is used here to suggest that exchange directions at the lake perimeter and lake bottom can be different (Figure 32e and Figure 37). This condition is presented here to alert investigators to consider the possibility of complex exchanges in some settings.

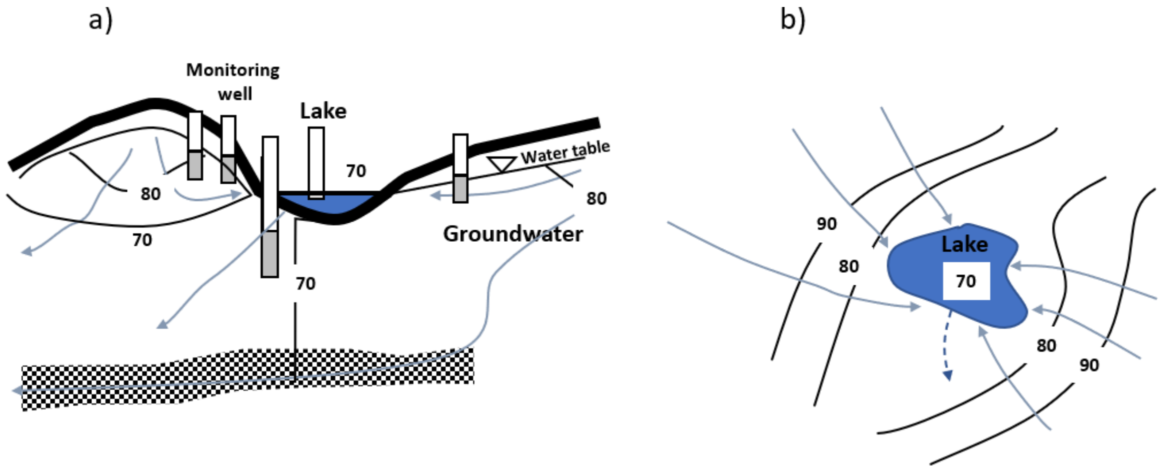
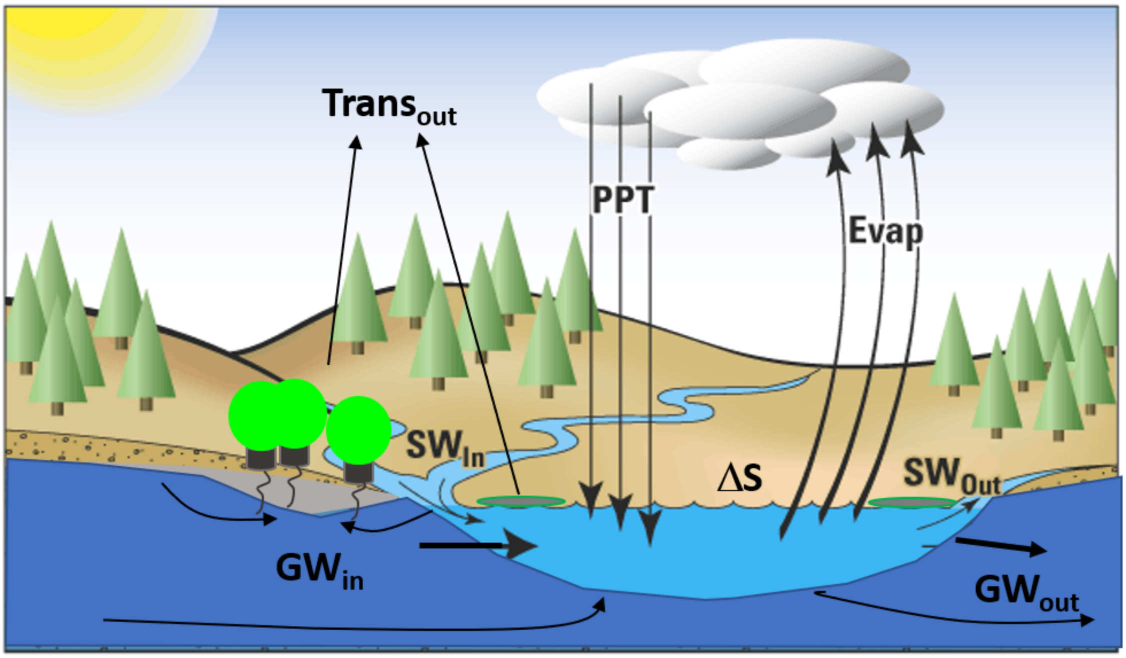


Figure 37 - Cross section and map view of a mixed lake exchange. Equipotential lines and relative head values are shown in black. Groundwater flow is in the direction indicated by blue arrows. Aquifer conditions are assumed to be isotropic and homogeneous. The stippled pattern represents a zone of higher hydraulic conductivity. Monitoring wells are open at the bottom. a) Cross sectional representation showing an upward groundwater gradient near the shore (effluent conditions), and a downward gradient beneath the lake. This causes leakage from the lake bottom in this setting. Lake stage is shown as a water level on the vertical rectangle. b) A map view showing equipotential lines and groundwater flow for a mixed lake exchange. The dashed arrow represents the loss of water from the lake bottom to the underlying groundwater flow system (Woessner, 2020).

3.5 Lakes in Landscapes

In many settings, lakes occur within complex hydrogeologic settings, including those containing intermediate and regional flow systems and heterogeneity (e.g., Figure 1). Evaluating groundwater exchange with lakes is most commonly accomplished by generating a lake water budget and measuring and computing groundwater inflow and outflow to the lake as illustrated in Figure 38 (Winter, 1981). Box 1↴ provides details of Winter's conceptualizations of lake-groundwater exchange. Water budgets can be used to examine exchange for a portion of the lake shore as well as the entire lake when components can be isolated. Section 5 of this book discusses methods for using water budgets to assess groundwater exchange with lakes.



Not to Sale

EXPLANATION

Evap	**Evaporation**
Gw_{in}	**Groundwater inflow**
Gw^{out}	**Groundwater outflow**
PPT	**Precipitation**
Sw_{in}	**Surface-water inflow**
Sw_{out}	**Surface-water outflow**
$Trans_{out}$	**Transpiration outflow**
ΔS	**Net change in Storage (lake, soil, and groundwater)**

Figure 38 – Flow-through lake water budget components. This setting includes streams entering and leaving the lake. Phreatophytes (bright green plants) cause water flow out of the lake through the groundwater system and aquatic vegetation (dark green) transpiration causes outflow of lake water (modified from Robertson et al., 2003).

Groundwater-lake exchange studies generally include instrumentation of lakeshores with monitoring well networks, and near-shore areas (littoral zones) and lake bottoms with instruments that physically measure seepage direction and magnitude known as seepage meters (Figure 39).

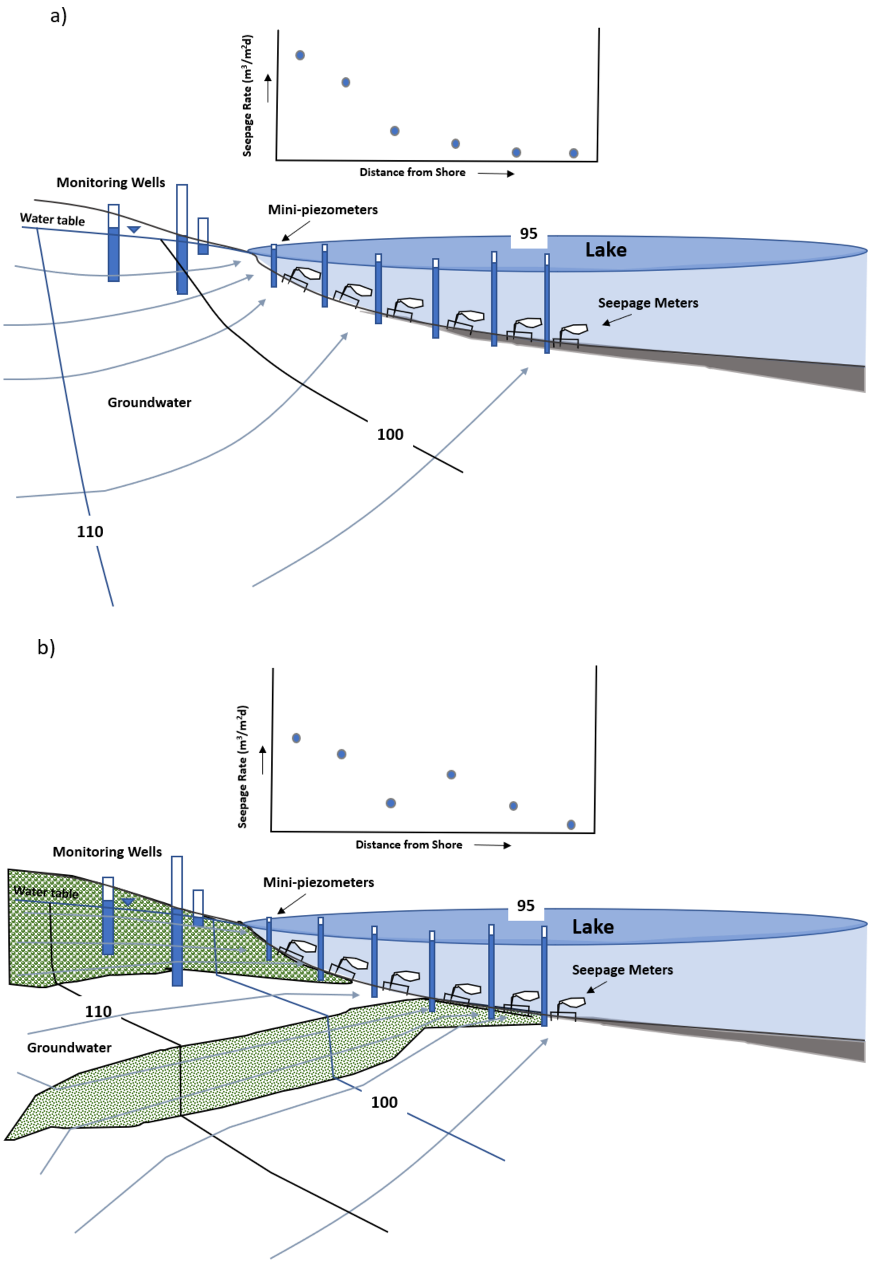


Figure 39 - Schematic cross section of a lake shoreline instrumented to characterize exchange. Monitoring wells are open only at the bottom. Seepage meters (discussed in Section 5) are inserted into the lake-bottom sediments to directly measure flux (m^3/m^2d). Groundwater is discharging to the lakebed (effluent conditions). Mini-piezometers are driven into the lake sediments adjacent to each of the seepage meters with tops extending above lake surface. In this example, all mini-piezometer measured gradients are upward as conditions are effluent. Hydrogeological conditions are simplified as isotropic and homogeneous in (a), and isotropic and heterogeneous in (b). Groundwater flow lines are shown in blue and relative equipotential lines in back. a) An effluent lake where groundwater discharge (flux) decreases logarithmically with distance from shore (blue dots on graph) as first noted by McBride and Pfannkuch (1975). The presence of finer-grained lake bottom sediments (brown) increase as the lake water deepens. b) An effluent lake intersected by two zones of higher hydraulic conductivity materials green stippled patterns). Groundwater flux varies with distance from the shoreline as the higher hydraulic conductivity materials concentrate groundwater flow (blue dots on graph) (Woessner, 2020).

Small diameter monitoring wells (mini-piezometers) placed in the littoral zone and deeper parts of some lakes provide direct comparisons of groundwater heads and lake stages (Figure 39). These site investigations usually include chemical sampling of groundwater and lake water to assess whether variations in chemical characteristics of seepage water and lake water can be used as additional support of exchange. Rosenberry and LaBaugh (2008) provide an excellent set of chapters describing stream and lake characterization methodologies. Exchange in the near shore areas is typically evaluated using monitoring well networks, and in-lake paired installations of seepage meters and mini-piezometers (Figure 39). The application of seepage meters and mini-piezometers is described in the literature (e.g., McBride and Pfannkuch, 1975; Lee and Cherry, 1978; Rosenberry et al., 2008) and discussed in Section 5 of this book.

Lake groundwater exchange locations, rates and timing vary with changes in the local lake water budget. Lakes and reservoirs may be dominated by gaining conditions at one time of year and losing conditions at another (e.g., Figure 10). Anderson and Munter (1981) found transient groundwater conditions associated with a flow-through lake in Wisconsin, USA, resulted in the occurrence of seasonal stagnation points at some locations within the local groundwater flow system. Rosenberry (2000) noted that as groundwater levels adjacent to a lake fall more rapidly than the lake level the near-shore portion of the lake can become perched (Figure 40).

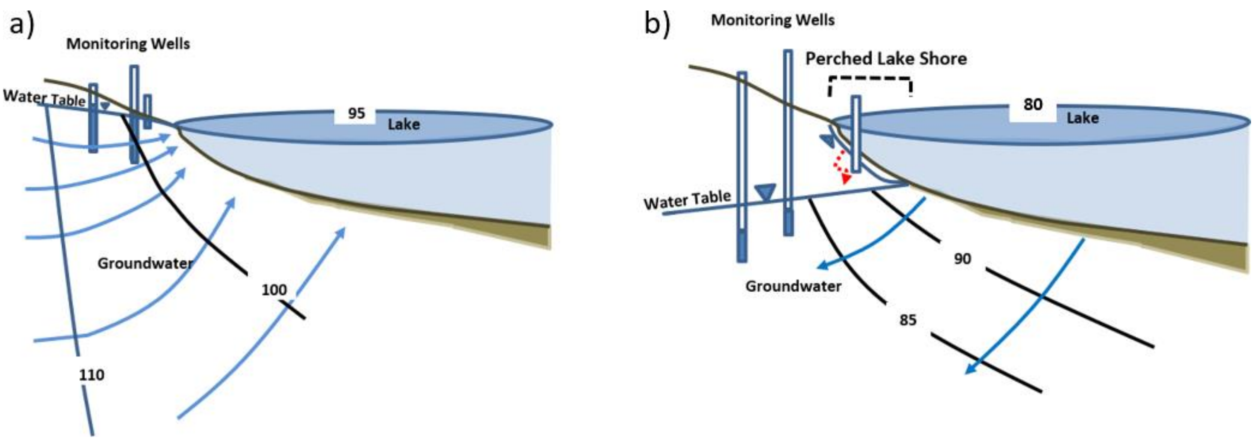


Figure 40 - Schematic of a lake transitioning between an effluent lake and influent lake. Groundwater flow is under isotropic and homogeneous conditions (blue arrows). Equipotential lines are shown in black. Monitoring wells are open only at the bottom. a) Groundwater lake exchange under effluent conditions. b) Groundwater lake exchange under conditions where the reduction in the water table exceeded the lowering of the lake level and influent conditions dominate. A zone of perched lake shoreline groundwater is indicated. It is underlain by an unsaturated (vadose) zone where lake water percolates to the water table (dotted red arrow) (modified from Rosenberry, 2000).

Construction of reservoirs by damming streams also results in exchanges between groundwater and the reservoir. When original stream conditions are generally effluent, groundwater typically feeds the upper end of the reservoir and receives leakage of reservoir water in the downstream area associated with the dam (Figure 41a). When reservoirs are placed in mostly influent stream conditions, reservoir water may remain higher than the surrounding groundwater, or be perched above the water table and create conditions where reservoir water leaks into the adjacent groundwater (Figure 41b). The stream below the dam may become effluent for a portion of the reach if sufficient reservoir leakage occurs.

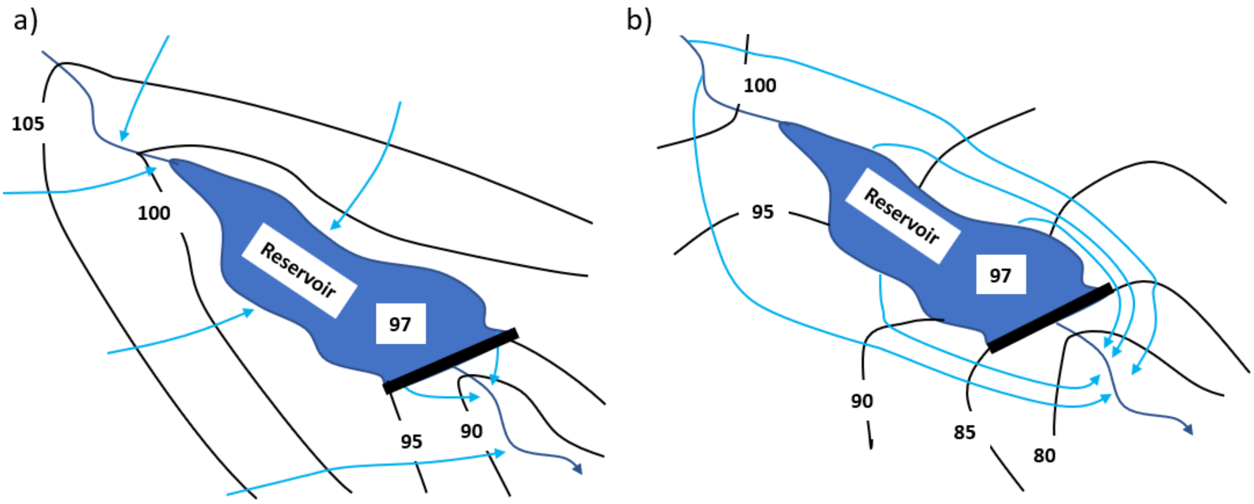


Figure 41 - Conceptual model of groundwater flow (light blue arrows) and heads associated with a reservoir and dam (black lines represent equipotential lines and numbers are representative heads). The dam is shown as a heavy black line. Hydrogeologic conditions are isotropic and homogeneous. A stream is flowing into and out of the reservoir (dark blue line). a) An effluent reservoir or gaining reservoir. Groundwater is mostly flowing into the reservoir as the water table above the dam is higher than the reservoir pool. Leakage occurs at the dam as some reservoir water seeps from the reservoir. b) Influent reservoir or losing reservoir. The water table is lower than the reservoir and water is leaving the reservoir. The stream below the dam is illustrated as gaining groundwater flow. These conditions could also be illustrated with the stream below the dam being influent (Woessner, 2020).

Dams create bodies of surface water that exchange water in a similar manner as lakes and often will include effluent, influent and flow-through conditions at various locations and times. Reservoir stages are often managed to control flooding, provide for a municipal water supply, generate power, and/or divert irrigation water to croplands. As such, stages can be dynamic, changing with seasonal conditions and water use demands. Reservoirs, as with most lakes, slow stream inflows and promote deposition of finer sediment, materials that often blanket the bottom resulting in a reduction of groundwater-surface water exchange over time. In some tributary settings small dams are being installed to promote temporary storage of surface-water runoff. In these settings the associated water table tends to increase in elevation as water builds up in the small reservoirs. These reservoirs and associated wetlands promote delayed water transfer by new hyporheic flow of stream water and stored groundwater to the downstream channel. These modifications may slow streamflow declines during dry periods of the year.

4 Wetlands and Groundwater Exchange

Groundwater exchange with wetlands includes sites where water is continually or intermittently present and a local groundwater system is effluent, influent, flow-through or a combination of exchange conditions (mixed). Wetlands occur in depressions, low areas of the landscape, and, in contrast to lakes, on slopes. Wetlands can be dominated by groundwater, precipitation/runoff, or a combination of conditions that vary over time. Commonly, a wetland is viewed as wet area created by either precipitation or by a water table that is at or near land surface some or all of the year. Wetlands have specific vegetation types (hydrophytes) and underlying soils that reflect the constant or periodic saturation of the soils (hydric) (Mitsch and Gosselink, 2000; Tiner, 1996).

The literature defines wetlands using a wide range of terms based on location, topography, biological composition, and the influence of runoff or groundwater (e.g., bog, fen, marsh, mire, moor, carr, oxbow, peatland, pothole swamp, slough, vernal pool, wet meadow) as discussed by Mitsch and Gosselink (2000). Wetland definitions vary depending on whether a physical science or ecological science approach is applied. A broadly used definition was developed by the U.S. Fish and Wildlife Service (Cowardin et al., 1979).

“Wetlands are lands transitional between terrestrial and aquatic systems where the water table is usually at or near the surface or the land is covered by shallow water. Wetlands must have one or more of the following three attributes: 1) at least periodically, the land supports predominantly hydrophytes; 2) the substrate is predominately undrained hydric soil; and 3) the substrate is nonsoil and is saturated with water or covered by shallow water at some time during the growing season each year.”

Cowardin and others (1979) suggest wetlands can occur with some or all the listed criteria. Legal and regulatory definitions most often require all three attributes be present for a feature to be classified as a wetland. Wetlands are valued for the role they play in temporally storing runoff, impacting water quality, and supporting diverse ecological systems (Mitsch and Gosselink, 2000; Fretwell et al., 1996).

Wetlands often contain fine-grained sediments and decomposed organic sediments both at the perimeter and bottom. When such conditions are present, exchange rates with groundwater systems can be constrained. In contrast, lakes, with fine-grained bottom sediments, are often shallow at the shoreline where wave action limits fine sediment deposition resulting in less restricted groundwater-surface water exchange. Groundwater exchange with a wetland can be conceptualized using the cross sections presented in Figures 4 through 8 and map views representing lakes (Figures 33 through 36). Wetlands also can occur without a connection to the local groundwater system and be dominated by surface runoff.

4.1 Effluent or Gaining Wetland

A wetland located in a groundwater flow system in which all groundwater flow is into the wetland is an effluent or gaining wetland (Figure 42). The wetland surface is an expression of the water table. In this setting, groundwater discharging to the wetland causes the water level to rise unless water is lost from the wetland as surface runoff and/or direct evaporation and transpiration. Water levels in the wetland change as the water budget varies (including changes in rates of evaporation and transpiration) and/or the discharge of water by streamflow (Figure 42c).

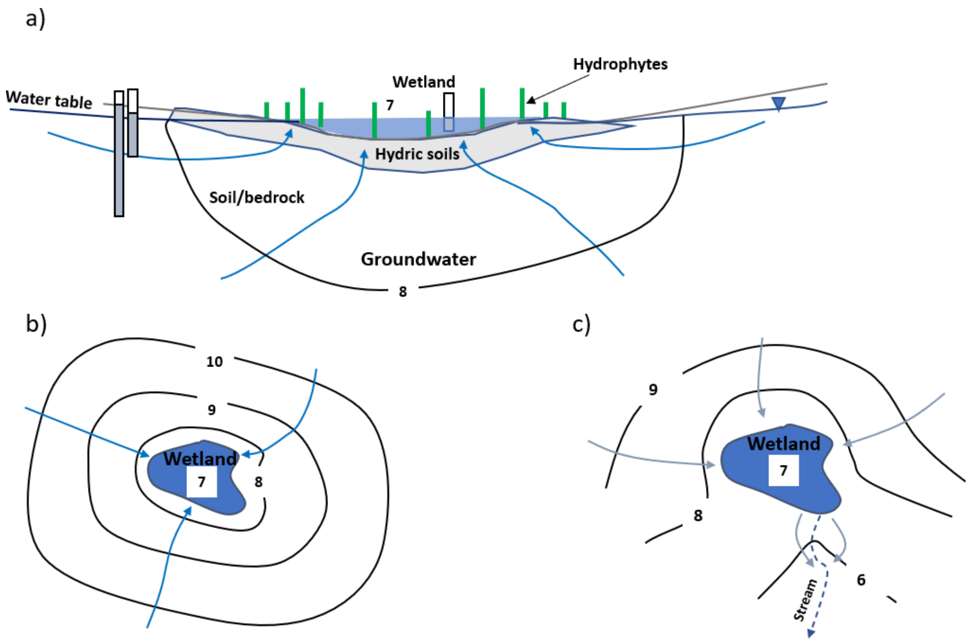


Figure 42 - Schematic cross section and map view showing a wetland dominated by groundwater discharge conditions, producing an effluent or gaining wetland. Equipotential lines and relative head values are in black. Groundwater flow lines are in blue. Isotropic and homogeneous conditions are assumed. Monitoring wells are open at the bottom. a) Cross section of a wetland showing groundwater flow to the wetland. Groundwater gradients are upward and the wetland stage represents the water table. b) Map view showing converging groundwater flow. c) Map view of an effluent wetland with a stream outlet (dashed line). The stream could be perennial or only flow occasionally. Some groundwater flows from the wetland when the stream is active (Woessner, 2020).

Springs and Wetlands

Wetlands often occur associated with springs. Box 2 briefly describes how groundwater discharge creates springs in a variety of hydrogeologic settings.

Wetlands are associated with spring and seepage-face discharge formed on slopes where groundwater discharge is sufficient to establish wetland vegetation and appropriate soil conditions as illustrated in Figure 43 (Mitsh and Gosselink, 2000). In locations where the topography intersects the water table, wetlands can be associated with a depression spring (Figure 43a). In some settings, groundwater discharge from bedding planes, fractures and joints in sedimentary and crystalline bedrock can provide sufficient flow to support wetlands at discharge points (Figure 43b and c).

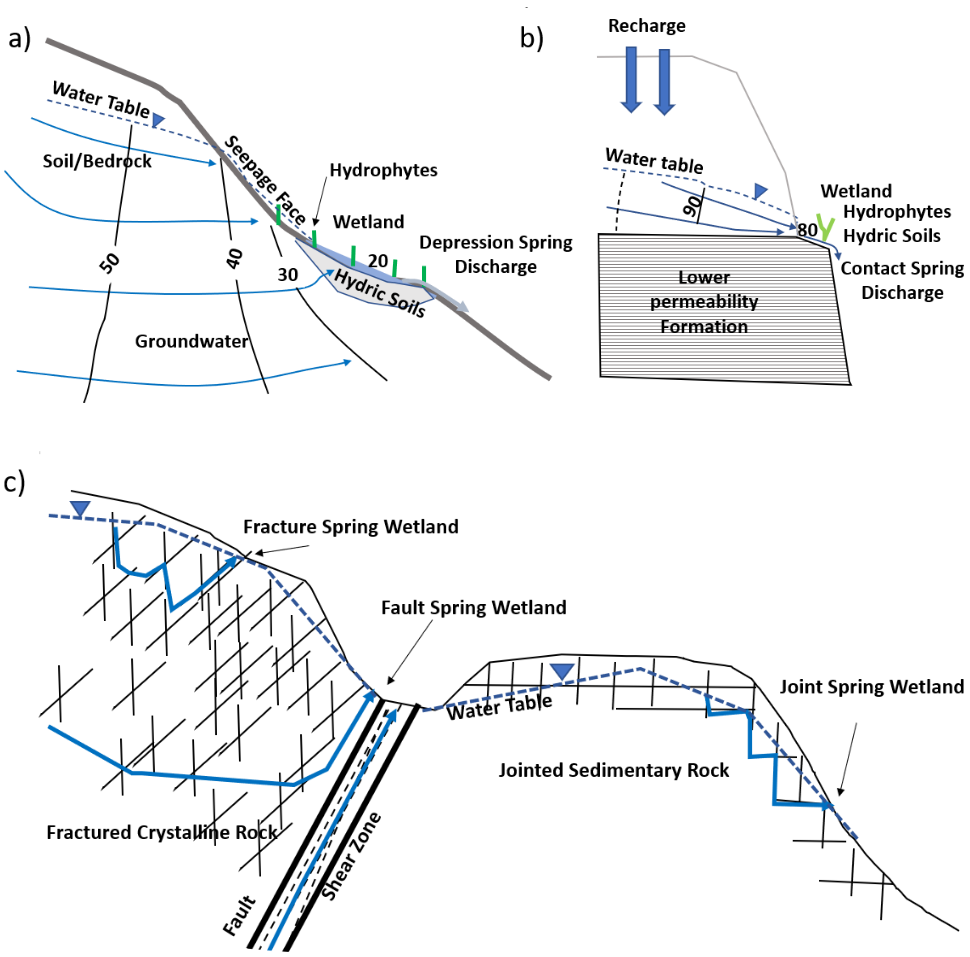


Figure 43 - Schematic cross section showing wetland formation from spring discharge at slopes and depressions. Equipotential lines and relative head values are in black. Groundwater flow lines are in blue. Isotropic and homogeneous conditions are assumed in (a) and (b). Fracture and joint controlled flow are assumed in (c). a) Wetland formation when the water table intersects the slope forming a seepage face and intersects a depression in the slope. b) Wetland formed at the discharge point of a contact spring. Groundwater seeps out at the contact between higher permeability material and lower permeability earth materials. The spring may also form from a perched groundwater system. c) Schematic cross section of other geologic settings where spring flow may create wetlands if appropriate vegetation and soils are established. Not all spring discharge settings are conducive to wetland formation (Woessner, 2020).

4.2 Influent or Losing Wetland

Influent or losing wetlands leak water downward into the underlying and adjacent groundwater system (Figure 44). In this setting the wetland surface represents the water table. An influent wetland also occurs when the zone of saturation is disconnected from the wetland and the wetland percolates water through the unsaturated zone (vadose zone) to the underlying groundwater system (Figure 45). The water surface elevation of the wetland does not represent the water table under perched conditions.

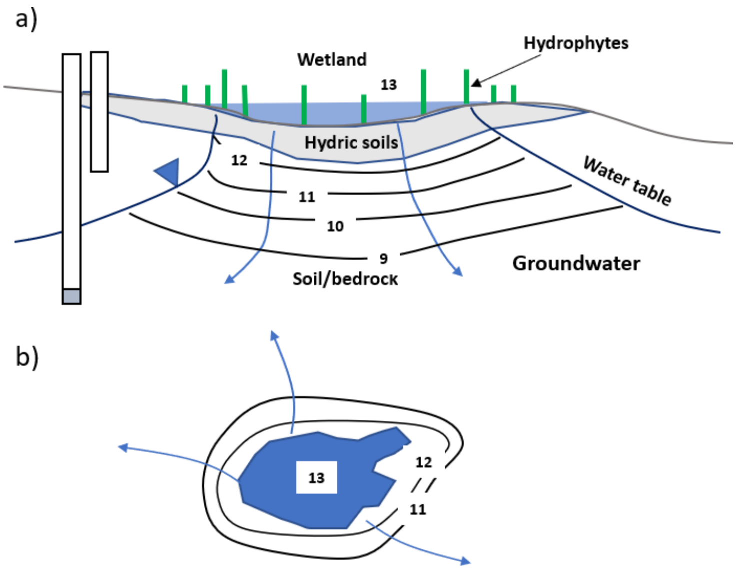


Figure 44 - Schematic cross section and map view of an influent wetland. Equipotential lines and relative head values are in black. Groundwater flow lines are in blue. Isotropic and homogeneous conditions are assumed. Monitoring wells are open at the bottom. The wetland stage represents the water table. a) Cross section of an influent or losing wetland that is connected to the water table. b) Map view of an influent wetland (Woessner, 2020).

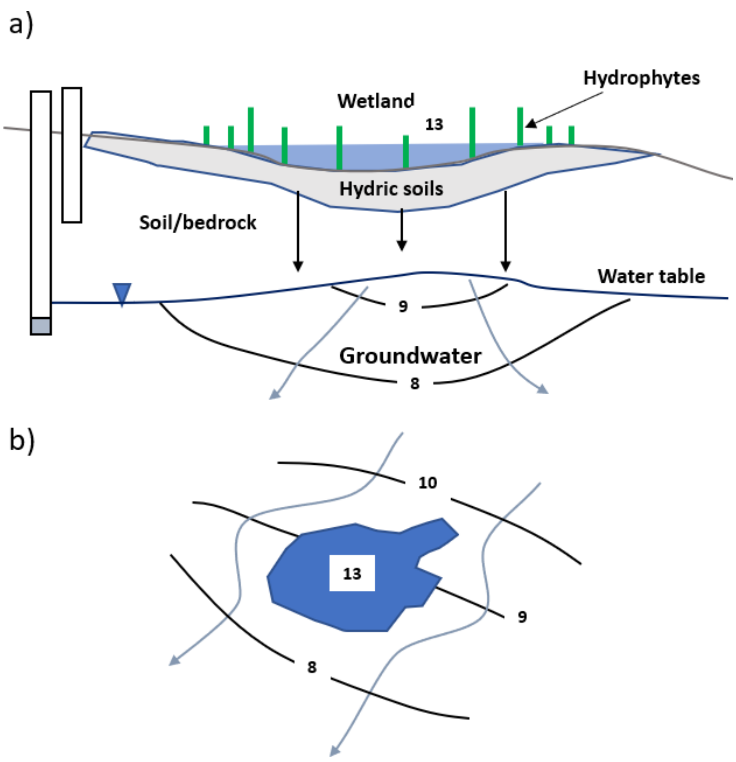


Figure 45 - Schematic cross section and map view of an influent wetland perched above the water table. Equipotential lines and relative head values are in black. Groundwater flow lines are in blue. Isotropic and homogeneous conditions are assumed. Monitoring wells are open at the bottom. The wetland stage does not represent the water table. a) Cross section of an influent or losing wetland that is perched above the water table. b) Map view of an influent wetland perched above the water table (Woessner, 2020).

4.3 Flow-through Wetland

Flow-through wetlands occur when the water table is higher on one side of the wetland than the other side creating a gradient for water to enter and leave the wetland as illustrated by Figure 20 and Figure 36 by substituting the label wetland for stream and lake, respectively, and by Figure 46. Often these wetlands have no surface-water inlet or outlet. The wetland surface represents the elevation of the local water table.

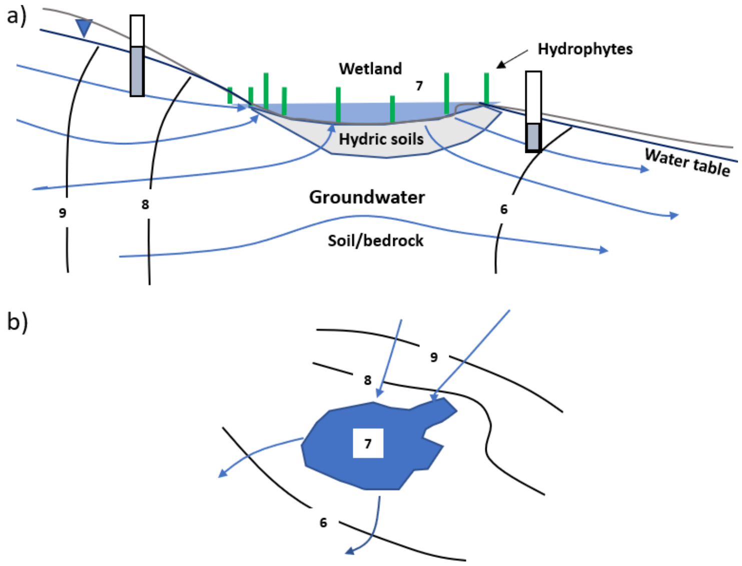


Figure 46 - Schematic showing a wetland dominated by flow-through groundwater conditions. Equipotential lines and relative head values are in black. Groundwater flow lines are in blue. Isotropic and homogeneous conditions are assumed. Monitoring wells are open at the bottom. The wetland stage represents the water table elevation. a) A cross sectional view. b) A map view (Woessner, 2020).

A flow-through wetland may also form a depression spring where the water table intersects the land surface. In a spring setting, water exits the wetland as surface flow, down-slope groundwater discharge and/or is lost by evapotranspiration.

4.4 Mixed Exchange Wetland

As presented in the discussion of streams and lakes, wetland types are commonly represented in the map views and cross sections as shown in Figures 16 through 22; and in Figures 32 through 36 by substituting the word wetland for the labeled streams and lakes, respectively. However, wetland exchange can be more complex in these mixed exchange settings (Figure 47). Field investigation of wetlands may require detailed analyses of groundwater exchange. Section 5 presents methods for such analysis.

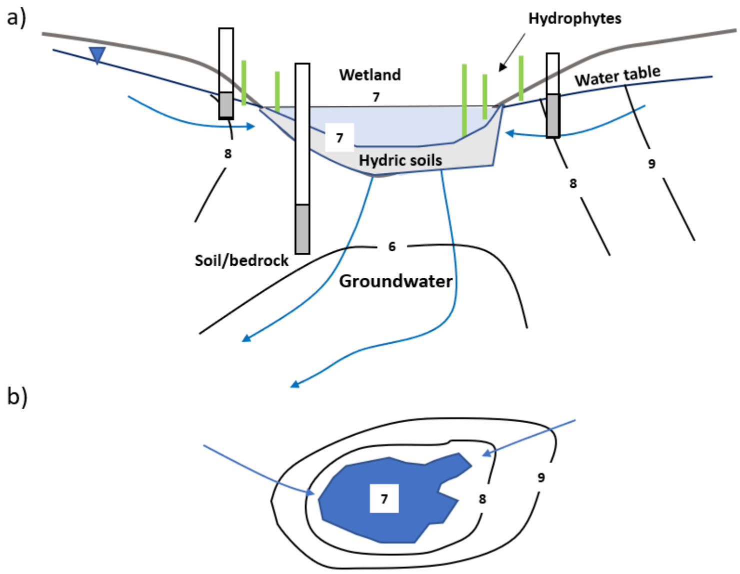


Figure 47 - Schematic cross section and map view showing a mixed exchange wetland dominated by both effluent and influent groundwater conditions. In this setting some surface-water outflow may occur. Equipotential lines and relative head values are in black. Groundwater flow lines are in blue. Isotropic and homogeneous conditions are assumed. Monitoring wells are open at the bottom. a) A cross sectional view of a mixed exchange wetland. b) A map view of a mixed exchange wetland (Woessner, 2020).

4.5 Disconnected Wetland

Most lakes and streams exchange water with the groundwater system. However, wetlands underlain by low permeability hydric soils can be disconnected from groundwater in settings dominated by surface runoff, precipitation, and high evaporation and evapotranspiration (Figure 48).

Disconnected wetlands occur when precipitation and surface runoff fill topographic depressions with water. These water-filled depressions promote wetland vegetation and soil development. Percolation to an underlying groundwater system is inhibited by the low permeability soils and high rates of direct evaporation and evapotranspiration (Figure 48).

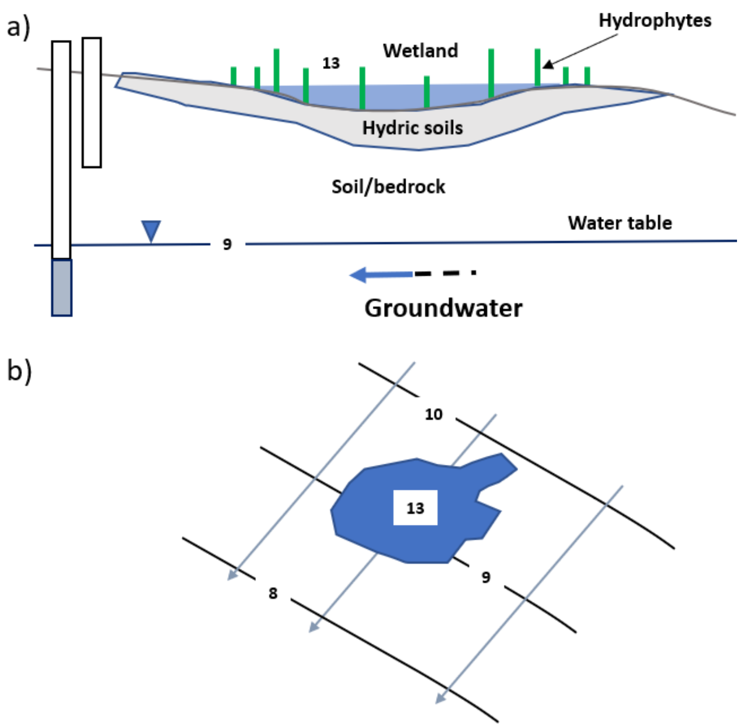


Figure 48 - Schematic cross section and map view of a disconnected wetland. Hydric soil saturation and the wetland water level varies seasonally. Equipotential lines and relative head values are in black. Groundwater flow lines are in blue. Isotropic and homogeneous conditions are assumed. A disconnected wetland occurs when direct evaporation and evapotranspiration exceeds downward movement of wetland water. Monitoring wells are open at the bottom. The water table remains below the wetland soils. In this setting no exchange occurs. a) Cross sectional view of a disconnected wetland. Blue dashed arrow represents groundwater flow beneath the wetland. b) Map view of a disconnected wetland (Woessner, 2020).

4.6 Conceptual Models of Wetlands

Wetlands in some locations are dominated by one of the exchange conditions described above. However, variations in hydrologic conditions over seasons or multiple years may cause transitioning through one or more exchange mechanisms (e.g., effluent to influent, effluent to flow-through to effluent, and so on). Water budgets are extremely useful tools for determining the degree of groundwater exchange.

Generating a wetland water budget requires quantifying the same components found in budgets for rivers and lakes (Figure 49). Instrumentation is required to measure each component as described in Healy and others (2007). Often groundwater characterization includes creating a network of monitoring wells surrounding the wetland, installing mini-piezometers and/or seepage meters in the wetland, sampling for surface water and groundwater quality, and, in some studies, placing water temperature monitors in the wetland water and underlying groundwater (see Rosenberry and LaBaugh, 2008) as discussed in Section 5 of this book. The presence, absence and duration of water in a wetland is dependent on the changes in the magnitudes of inflows and outflows (Figure 49).

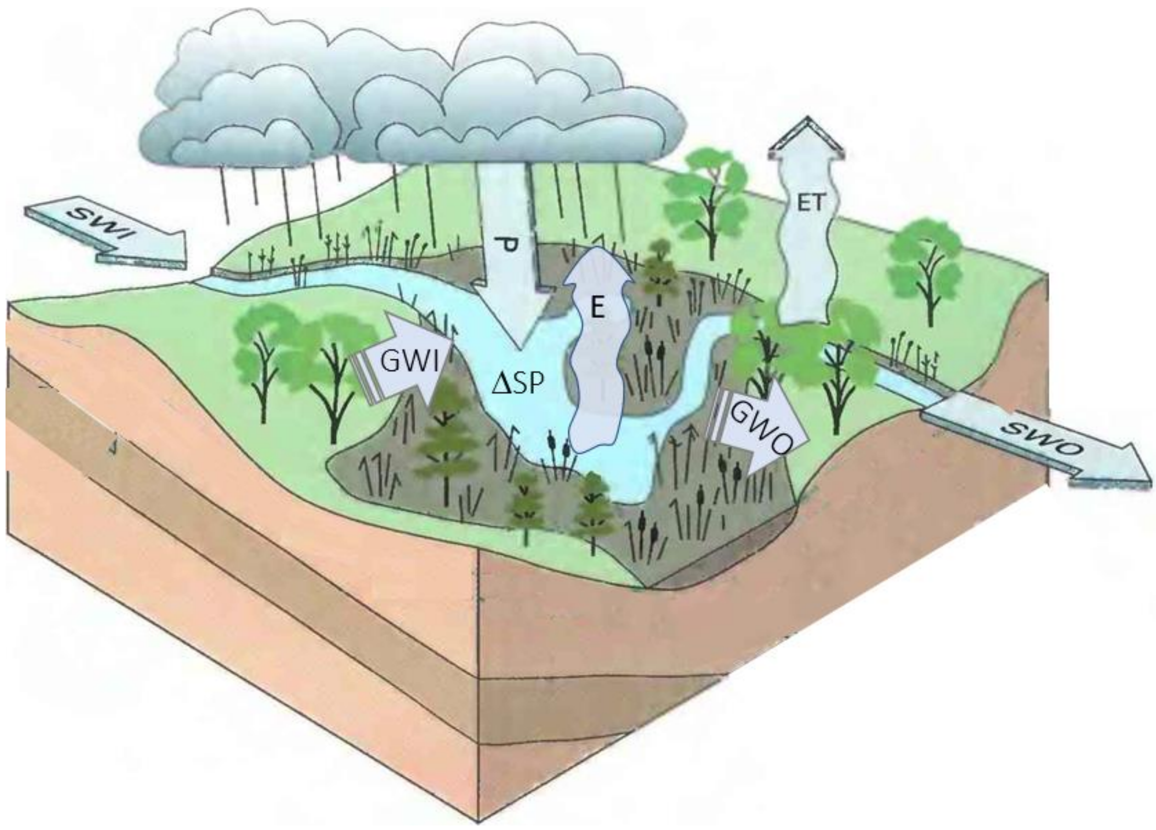


Figure 49 - Wetland water budget components ($P+SWI+GWI=E+ET+SWO+GWO-\Delta SP$). Inputs are the volumes of precipitation (P), surface-water inflow (SWI), groundwater inflow (GWI). Outputs are direct evaporation from the water surface of the wetland pool (E), evapotranspiration (ET), surface-water outflow (SWO), and flow out to the groundwater system (GWO). The change in volume of the wetland pool (ΔSP) is positive if the volume increases and negative if the volume decreases, so the negative sign of the equation produces the appropriate balance (modified from Tiner, 1996).

The development of wetland vegetation and hydric soils is dependent on the length of time water is present in a wetland. The seasonal changes in the wetland surface and subsurface water levels define a wetland hydroperiod (Figure 50). Mitsch and Gosselink (2000) present numerous examples of wetland hydroperiods. Analyses of hydroperiods are used to compare wetland stability from year to year. With additional water budget information, including the degree and timing of wetland exchange conditions, a wetland hydroperiod can be used to examine changes in exchange conditions over time (Figure 50).

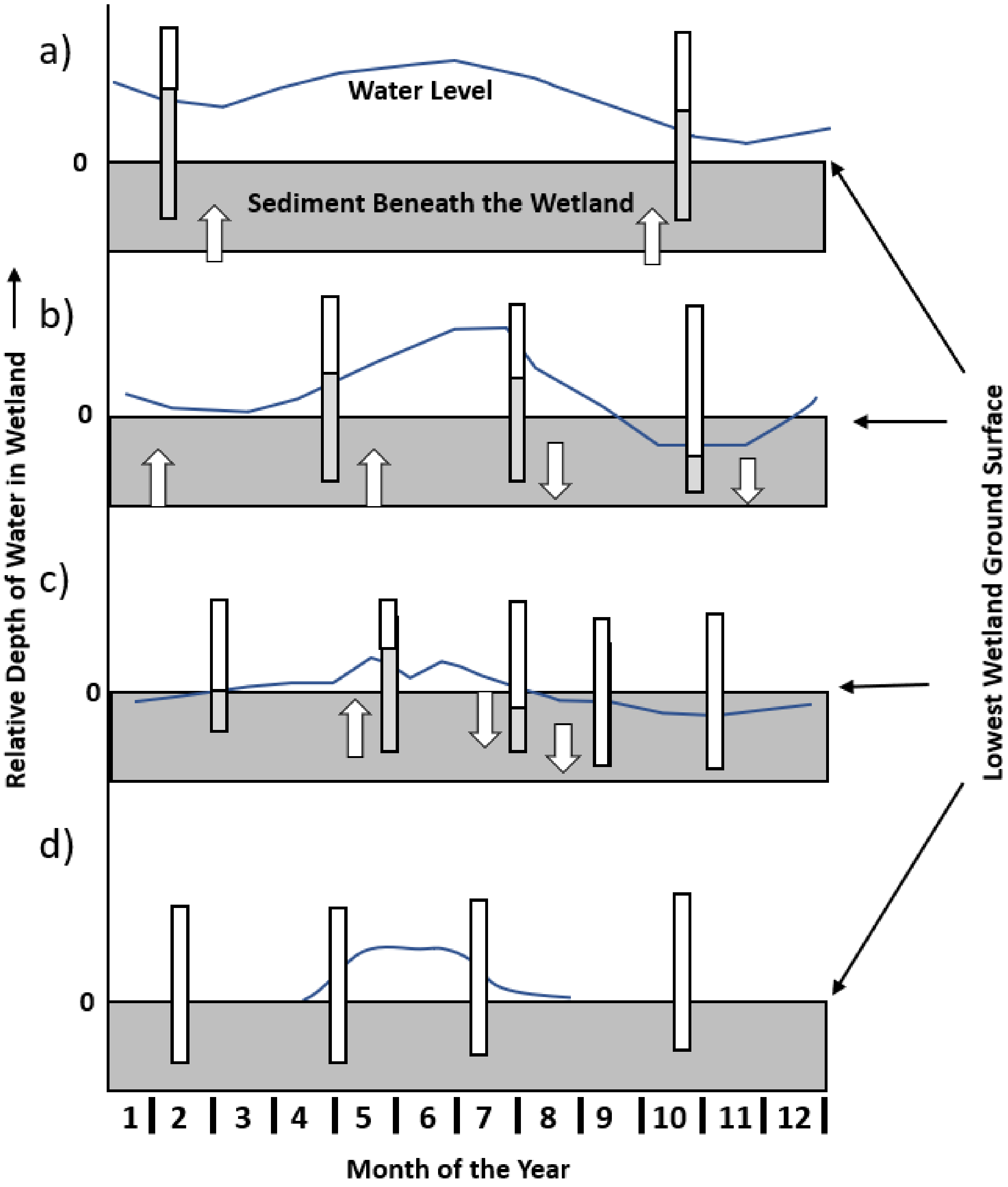


Figure 50 - Examples of hydroperiods, wetland water level changes over time (blue lines), for four hypothetical temperate wetland settings. Relative water depths are plotted against months1 through 12 for one year in a semi-arid climate of the northern hemisphere. The wetland base (lowest ground surface elevation) is shown as the black horizontal line (0) at the top of the rectangle that represents the subsurface (gray). Monitoring wells/mini-piezometers, which are open only at the bottom and located in the wetland, show groundwater levels in light gray. Groundwater exchange direction is indicated by white arrows. a) A wetland hydroperiod dominated by effluent exchange and the influence of spring rains. Wetland base water levels are sustained above the ground surface for the entire year. b) A wetland hydroperiod where groundwater exchange (effluent) and rainfall raise the water level in the spring. In the summer months the wetland becomes influent and by month 9 (September) the wetland has dried, and the water level is below the wetland base. c) A wetland hydroperiod during a drought as compared to conditions in (b). Limited exchange is present in the winter and the wetland receives groundwater discharge (effluent) and some precipitation in the spring combined with some rainfall until summer. Then the wetland becomes influent and by September remains influent or becomes disconnected from the underlying groundwater system. d) A wetland hydroperiod for a disconnected wetland. Water in the wetland is from precipitation and runoff. Evaporation/evapotranspiration dominate the water balance and no groundwater exchange occurs (after Mitsch and Gosselink, 2000).

4.7 Wetlands in Landscapes

Wetlands can be generally classified based on geologic/ hydrologic conditions and vegetation. Classes include marine, estuarine, riverine, lacustrine and palustrine. Terrestrial systems include riverine (associated with rivers), lacustrine (associated with topographic depressions, that are like a lake), and palustrine (includes most inland wetlands and a subset of estuarine and intertidal systems) as shown in Figure 51.

Wetlands that are well connected to groundwater commonly occur at groundwater discharge sites associated with local, intermediate and regional groundwater systems (e.g., Figure 10). In addition, they can also occur on slopes associated with spring discharge (Figure 43), in topographic lows in recharge areas, and low-relief depression dominated landscapes.

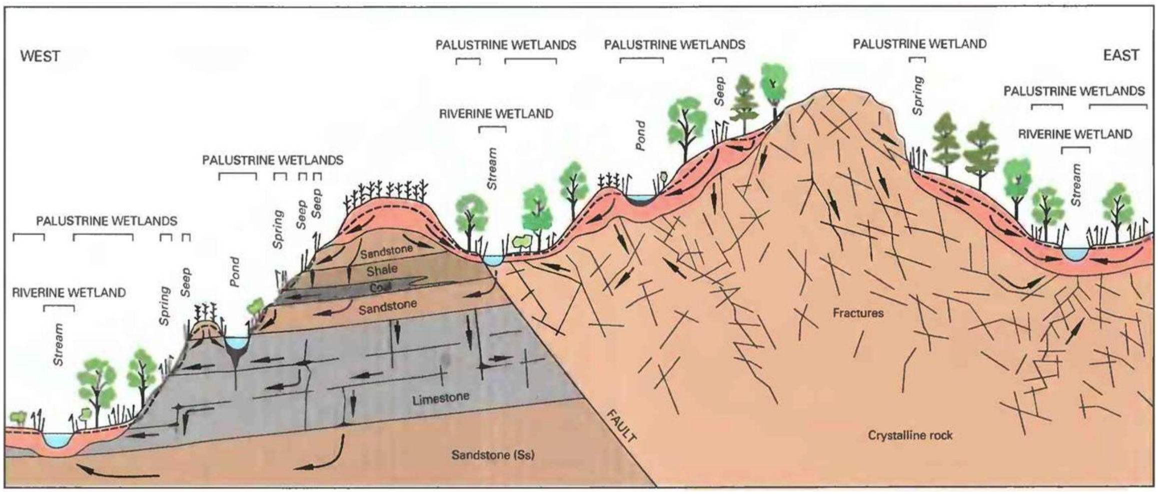


Figure 51 - Schematic cross section of a region underlain by complex sedimentary and crystalline rock deposits showing terrestrial wetlands. A palustrine wetland is used to represent terrestrial wetlands other than riverine wetlands as shown in this diagram. The water table is indicated by a dashed black line and general groundwater flow is indicated by black arrows. Recharge focused at topographic highs drives groundwater exchange in this landscape. Wetlands form where groundwater discharges as springs and seepage faces created by groundwater flow from bedding contacts, fractures and joint systems, at discharge zones originating from shallow local flow systems and at locations where more regional groundwater flow discharges (Fretwell et al., 1996).

Wetland systems illustrated in Figure 51 show wetlands in sites with large topographic relief. However, many wetlands occur in areas with low topographic relief and cover large areas. Often wetlands form in topographic settings where the water table intersects the land surface for all or part of the year. For example, in North America large expanses of wetlands occur in low relief regions including the boreal reaches of Canada, and the interior and coastal plains of Canada, the United States and Mexico (Figure 52). Mitsch and Gosselink (2000) and Fretwell and others (1996) provide additional examples of wetland landscape settings and groundwater flow systems.

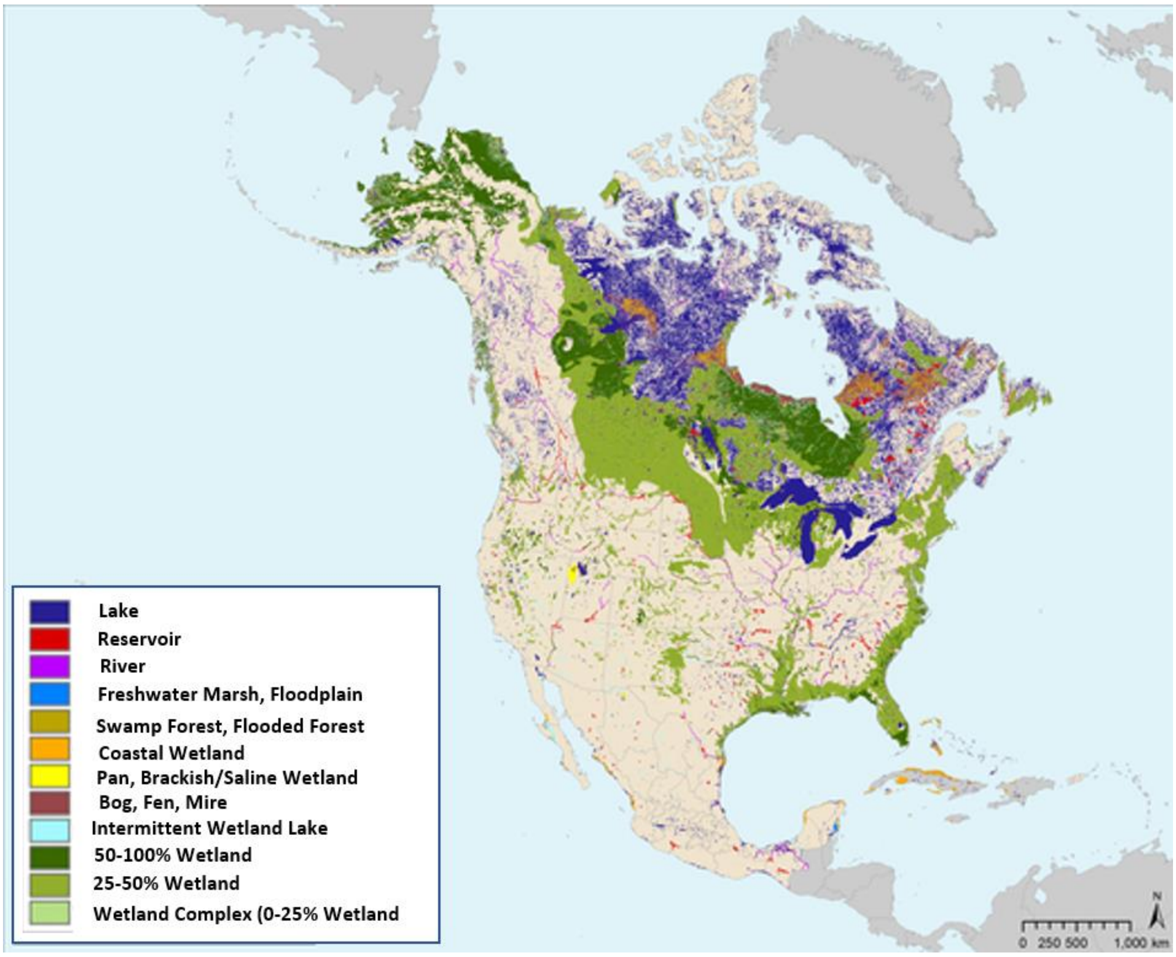


Figure 52 - A map showing the extent of wetland types in North America. Many wetlands are located in lowlands or areas of low relief associated with interior and coastal landscapes. Wetland types including lacustrine (lake/reservoir, intermittent wetland lake), riverine (river), palustrine (freshwater marsh, floodplain, swamp forest, flooded forest, bog, fen, mire,), marine (coastal wetland, pan, brackish/saline wetland) are represented as well as generalized mapping where a portion of the land area is covered by wetlands (e.g., 25-50% wetland). Lehner and Doll (2004) distributed by Commission for Environmental Cooperation, Montreal, Quebec↗.

5 Methods to Investigate Groundwater-Surface Water Exchange

This section presents the general methods used to qualitatively and quantitatively describe groundwater-surface water exchange with streams, lakes and wetlands. Included are examples of applying methods and interpreting the exchange processes in natural and modified groundwater and surface-water systems. Characterizing exchanges requires a number of approaches including calculating exchanges using water budgets, applying hydrogeological principles and modeling, physical instrumentation, remote sensing, and geochemical analyses (Figure 53). In some cases, these procedures are intertwined as analyses overlap. Some methods provide a broader view of exchange (e.g., water budgets) and other methods characterize conditions at a specific location (e.g., seepage meters). Rosenberry and LaBaugh (2008) provide an excellent public-domain book describing methodologies applied to characterizing exchange sites, magnitudes and durations. They address the use of streamflow records; wells, piezometers and seepage meters; methods to characterize karst hydrology; and, the application of temperature to assess exchanges. Once data are obtained, field and remote sensing data sets require multiple levels of analysis and interpretation. A variety of modeling tools are also used to build conceptual models, test uncertainty, and quantify exchange conditions and timing (e.g. Anderson et al., 2015).

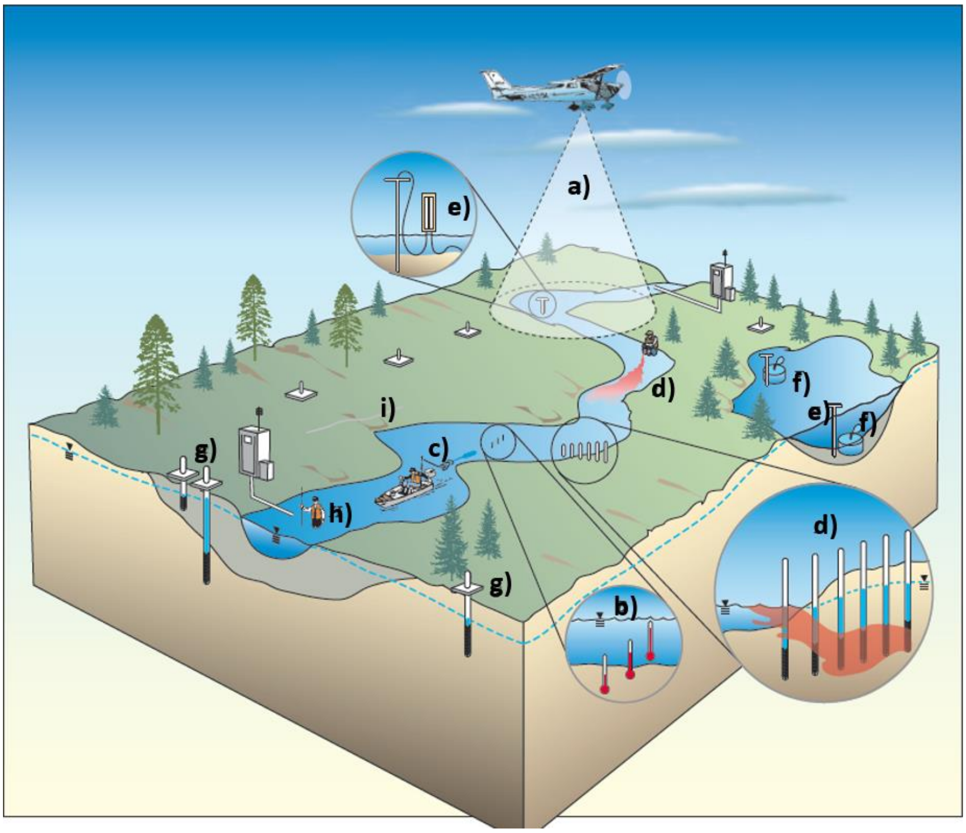


Figure 53 - Schematic showing field-based techniques used to assess exchange of surface water and groundwater. Letters on the figure represent methods as follows: a) airborne, drone or land-based temperature sensing imagery; b) characterizing temperatures by vertical thermal profiling; c) temperature and specific conductance probe mapping of pore water in surface-water-bottom sediments; d) tracer and geochemical studies; e) mini-piezometers; f) seepage meters; g) monitoring well networks; h) streamflow measurements; and i) fiber-optic, bottom-temperature mapping (modified from LaBaugh and Rosenberry, 2008). Methodologies used to characterize exchange at the landscape and feature scale involve application of water budgets and landscape scale mapping tools. Some of these methods can also be applied to study sites covering smaller areas. Other methods yield data related to site scale investigation (monitoring well networks) and at point locations within study sites (e.g., seepage meters).

5.1 Water Budgets

Water budgets can be applied at multiple scales to assess groundwater-surface water exchange for streams, lakes and wetlands. They account for all water sources, sinks and changes in storage within a defined boundary and time frame (e.g., Figure 1; Figure 38; Figure 49). Healy and others (2007) prepared an excellent public-domain publication on water budgets for resource and environmental management that addresses formulation of and methods used to quantify components of water budgets at various scales. Earlier work by Winter (1981) addressed the challenges of computing groundwater components of lake budgets. He cautioned against setting groundwater components as the budget unknowns because this approach causes budget uncertainties and errors to be incorporated in the computed (residual) groundwater exchange values. Winter (1981) stressed the importance of independently quantifying components of the water budget.

At an individual feature scale (reach of a stream, lake, and wetland) a water budget boundary and time interval are required when developing a water budget. The budget requires quantification of the effluent and influent exchange of groundwater with the defined budget boundaries (e.g., a watershed; shoreline of a surface-water feature). For example, the groundwater inflow for a lake setting (Figure 38) is defined by Equation 1 when using a lake water budget with the lake shoreline as the boundary. Each component requires quantification and an estimate of error (e.g., Healy et al., 2007).

$$Pin + Swin + GWin = GWout + SWout + Eout + ETout - S \tag{1}$$

where:

Pin = precipitation falling within the lake domain (L^3/T)

$SWin$ = surface-water flow into the lake (L^3/T)

$GWin$ = groundwater discharge to the lake (L^3/T)

$GWout$ = lake water leakage to groundwater (may include the flow of water to phreatophytes located near the shoreline) (L^3/T)

$SWout$ = surface-water flow out of lake (L^3/T)

$Eout$ = direct evaporation from the lake water surface (L^3/T)

$ETout$ = loss of water from plants within the lake by evapotranspiration (L^3/T)

S = volumetric net change in storage of lake water, an increase in volume is positive and a decrease is negative (L^3/T)

Standard hydrology texts (e.g., Hornberger et al., 1998; Dingman, 1994; Watson and Burnett, 1993) and hydrogeology texts (e.g., Freeze and Cherry, 1979; Domenico and Schwartz, 2000; Fetter, 2001; Schwartz and Zhang, 2003; Fritts, 2012; Weight, 2019) discuss techniques to quantify each component of a water budget. Box 3↴ presents the derivation of a water budget for Mirror Lake in New Hampshire (Healy et al., 2007) and addresses estimation of error associated with the budget components. Though water budgets are valuable, they should be used in conjunction with other methodologies when finer resolution of site-specific exchange is needed.

5.2 Stream Hydrograph Separation Methods

Stream discharge records provide information on water yield of the basin above the location of the stream gauging site. Exchange between the stream and groundwater system can be inferred from analyses of both short- and long-term discharge records.

Stream discharge records are generated at a gauging site where stream stage is measured and a relationship between stage and discharge is available. Stream stage is reported relative to a reference datum, and is measured using a ruled staff, transducer, bridge-mounted measuring reel, acoustic sounder, electric water level gauge, and/or with equipment housed at official stream gauging stations (e.g., float, bubbler, transducer) as illustrated in Figure 54. In some settings, established stream gauging stations provide stage, discharge and geochemical data that are accessible in data bases and/or real time (e.g., USGS National Water Information System↗). When stream stage measurements are combined with discharge measurements at a wide range of flows, a stage-discharge relationship can be developed for the station (Figure 54c). A stage-discharge relationship, makes it possible to estimate stream discharge as a function of time when only stage is measured. These records can then be analyzed to determine net basin-scale exchange.

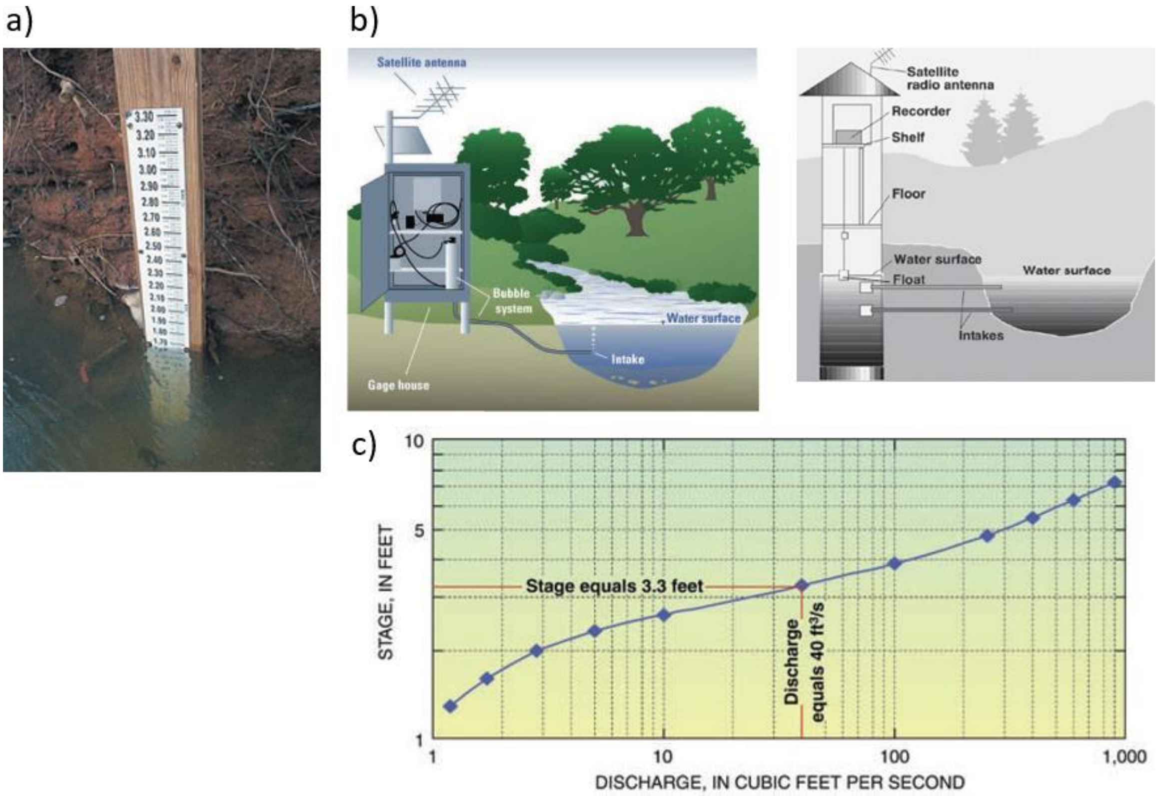


Figure 54 - Stream gauging stations: a) staff Gauge (USGS, 2005); b) stream gauging stations (USGS, 2011 and USGS, 2019); and, c) stage discharge curve (USGS, 2011).

At the basin scale, groundwater exchange can be inferred using hydrograph separation techniques when effluent conditions dominate (e.g., Wisler and Brater, 1959; Meyboom, 1961; Rorabaugh, 1964; Rorabaugh and Simons, 1966; Hannula et al., 2003; Rutledge, 1993; Rutledge, 1998; Combalicer et al., 2008). Hydrograph separation involves interpreting the groundwater contribution to the measured stream discharge, base flow (Figure 23). Though hydrograph separation is sometimes viewed as deriving an estimate of total basin groundwater recharge (groundwater exchange), it actually represents the net discharge of groundwater to the stream over a given time period. In cases where basin groundwater flow bypasses the stream channel (e.g., underflow) or other water budget components are significant (e.g., evapotranspiration, pumping, and diversion of groundwater from the basin), base flow will not reflect total basin groundwater recharge (e.g., Rutledge, 2005; Combalicer et al., 2008). However, base flow separation proves useful when estimating annual, or event, groundwater exchange (effluent conditions) as shown in Figure 55.

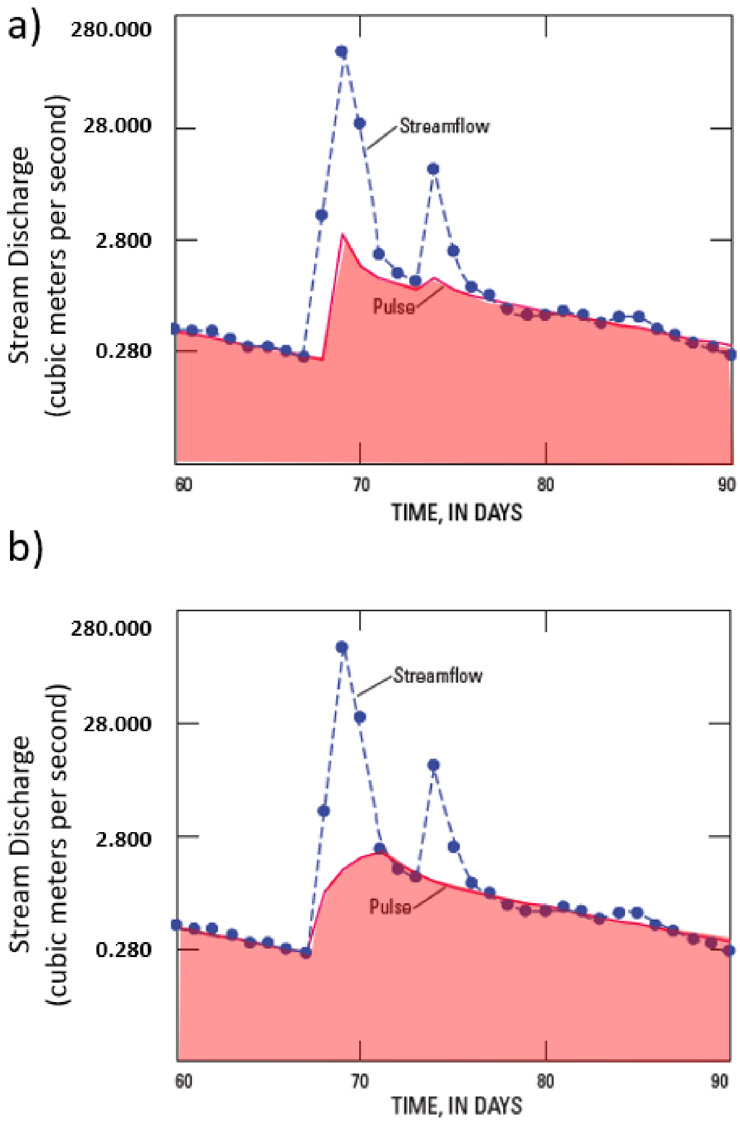


Figure 55 - Application of hydrograph separation techniques to a March 1974 hydrograph of Big Hill Creek, Cherryvale, Kansas, USA. The blue dashed line and circles are the streamflow records and the red solid line the interpreted base flow supplied by groundwater (shaded in red). Modeling was performed using the PULSE model (from Rutledge, 2000). a) Recharge was modeled as 0.65 inches (1.6 cm) and 0.08 inches (0.2 cm) on days 69 and 74, respectively. b) Modeling results when recharge was simulated as constant from day 68 to day 72. The total recharge modeled was 0.73 inches (1.8 cm) in both cases (modified from LaBaugh and Rosenberry, 2008, including conversion to metric units).

Graphical, analytical, and forward and inverse numerical modeling tools have been developed to perform hydrograph separation. Standard hydrology and hydrogeology textbooks describe the methodologies (e.g., Fetter, 2001). As an example, Hannula and others (2003) used a hydrograph separation technique reported by Moore (1992) to estimate groundwater discharge to a creek in the Iron Mountain West Shasta Mining District in California, USA. They also computed average specific yield and average peak transmissivity of the unconsolidated materials adjacent to the creek.

A comparison of the results of six software packages developed to perform hydrograph separation is presented by Combalicer and others (2008). A variety of software including: PART (Rutledge, 2007a); RORA (Rutledge, 2007b); PULSE (Rutledge, 2002); Base Flow Index (BFI) (Wahl and Wahl, 2001); River Analysis Package (RAP) (Marsh et al., 2003); and, WHAT (Lim et al., 2004) are each applied to the same river hydrograph data set. Neff and others (2005) also assess multiple hydrograph separation methods in their analysis of base flow to the Great Lakes, USA and Canada.

5.3 Basin Groundwater Modeling

In addition to hydrograph separation models, basin scale exchange processes can be examined using numerical groundwater models, coupled surface water-groundwater models and lumped streamflow routing models (e.g., Anderson et al., 2015). Modeling is used to conceptualize and assess river exchanges at the basin, valley segment, reach and channel scales (including hyporheic zones); lake and wetland exchange.

Modeling emphasizing the groundwater system (e.g., the MODFLOW↗ family of codes; and FEFLOW↗ (Diersch, 2014) account for groundwater exchanging with surface-water features under effluent, influent, flow-through and disconnected or zero-exchange conditions. Models contain specific coding that directs how, when and where groundwater passes to and from rivers, lakes and wetlands. In addition, model and local water budgets are computed from which exchange can be inferred (e.g., Anderson et al., 2015). Groundwater focused models allow the evaluation of anisotropic and heterogeneous three-dimensional groundwater flow, and exchange with surface-water features. The book by Anderson and others (2015) covers a wide range material including the basic development of conceptual groundwater models and methods to address uncertainty in modeling results. They describe how groundwater models can be used to assess exchange with rivers, lakes and wetlands, and how some codes link standard hydrologic (surface water) models to groundwater models. A more detailed discussion of model application is discussed elsewhere in this book.

5.4 Stream Synoptic Surveys

Synoptic surveys or seepage runs are used to assess exchange conditions in streams. They are conducted by measuring stream discharges at two or more locations and taking the difference between the measurements to determine if discharge is increasing or decreasing, that is whether the stream is gaining or losing. This method is based on physical streamflow gauging and requires consideration of the entire water budget between survey boundaries (Figure 56). Seepage runs are typically conducted over a short time period during which flow conditions are assumed to be in a steady state. Net groundwater inflow is computed from the difference between inputs and outputs as shown in Equation 2.

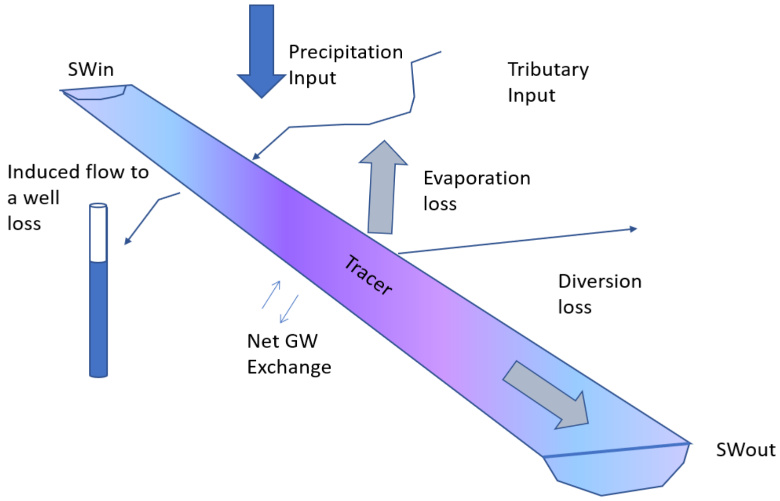


Figure 56 - Schematic of a river reach used for a synoptic survey and a tracer test to characterize net groundwater exchange over the reach. All inputs and outputs must be quantified between the upgradient and downgradient monitoring locations (SWin and SWout) to allow estimation of the net groundwater exchange. Exchanges can also be assessed using the introduction of a known concentration of tracer and monitoring its arrival over time at a downstream station as discussed in another subsection of this book. (Woessner, 2020).

$$(SWout + other\ channel\ losses) - (SWin + Inputs) = Net\ GW\ Exchange \tag{2}$$

where:

$SWout$ = discharge of the reach measured at the downstream location (L^3/T)

$other\ channel\ losses$ = quantified or estimated (e.g. diversions, evaporation, infiltration and capture of river water at a nearby shallow pumping well) (L^3/T)

$SWin$ = discharge of the stream measured at the upstream monitoring point (L^3/T)

$Inputs$ = quantified or estimated gains of surface water between the upstream and downstream monitoring locations (e.g. precipitation and inflow from tributaries) (L^3/T)

$Net\ GW\ Exchange$ = additional gain or loss of stream discharge attributed to groundwater (L^3/T)

Synoptic surveys, though useful, must be carefully planned as measurement errors can be large. Ideally, when discharge at the upstream location is subtracted from discharge at the downstream location and value is positive, the reach is effluent, gaining groundwater; while when negative, the channel is losing water to the underlying groundwater system (some reporting methods may reverse the sign of the net difference). It is important that seepage runs are designed so that measurement errors and uncertainties do not mask the computed net gains or losses.

Donato (1998) conducted two seepage runs (synoptic surveys) on the Lemhi River, Idaho, USA. A section of channel approximately 100 km in length was subdivided into 14 reaches (Figure 57). Current meter gauging methods were used to determine stream discharge (assuming a 5% measurement error). Measurements included diversions and tributaries within reaches and were completed at 117 sites over a five-day period. The first seepage run was conducted in August (under high groundwater elevations during the irrigation season) and the second in October 1997 (after irrigation ceased). Results of seepage gains and losses for each seepage run showed the river was effluent during August at most all reaches (Figure 57). During the October run, at least six reaches changed from gaining to losing. For the 28 computed gains or losses only 5 were deemed compromised by measurement error overlap (5%). In this study, a large number of measurements were required to characterize the system because the river was influenced by intensive irrigation. The study also illustrates the transient nature of exchange.

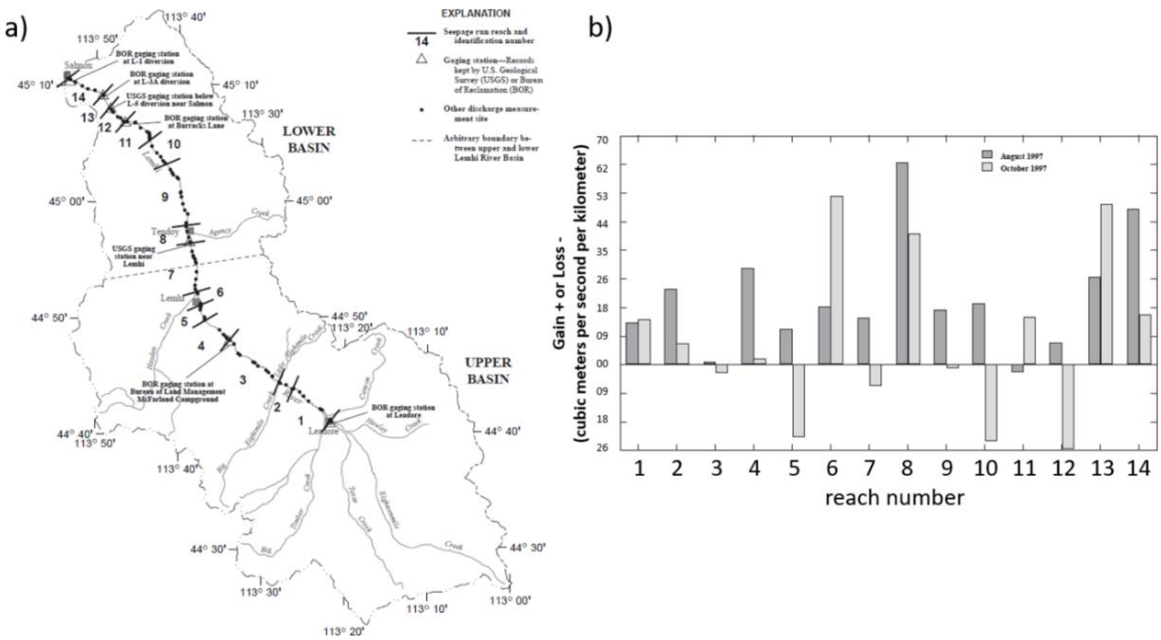


Figure 57 - Seepage runs conducted on the Lehmi River, Idaho, USA. Seepage runs were conducted over 5 days in August (irrigation season) and five days in October, 1997 (after irrigation stopped). The river is flowing from southeast to northwest (the top of the figure is north). To account for irrigation diversions and tributary flows, 117 measurements were taken for each survey. Streamflow measurement error was assumed to be 5%. a) A map view of river gauging locations. The gauging intervals are numbered sequentially from the upstream to the downstream gauging site. b) Computed reach groundwater inflows (plotted as positive numbers) and outflow (plotted as negative numbers). Error bars are not shown (Donato, 1998).

Individual instruments and methods need to be researched to define likely ranges of measurement errors. The reach length selected for a synoptic survey needs to be sufficiently long so that discharge measurement differences are larger than the error envelope for each measurement as discussed in Box 4⤵. Conducting seepage runs in lower order streams (smaller flows) is less complicated than using the techniques to characterize higher order streams (large flows). Generally, larger streamflow is more difficult to measure (thus typically have larger error) and longer reaches are needed to quantify changes. It should also be noted that methods yield net exchange, but multiple exchange processes may occur within a reach as explained in Section 2 of this book.

In regions with seasonal variation, seepage runs are often planned for a time when transpiration by streamside vegetation is small (e.g., after a killing frost) to simplify the reach water budget. However, usually surveys are repeated during other seasons to characterize how seasonal factors influence exchange.

5.5 Surface-water Stage and Groundwater Monitoring Networks

Installing a monitoring-well and surface-water-stage network allows for collection of head and stage data to generate two- and three-dimensional equipotential distributions. Interpretations of these data yield groundwater flow directions, rates, and changes in heads over time. Two-dimensional maps and cross sections, three-dimensional renderings, as well as model simulations and associated graphics provide visualizations of exchange conditions.

Map views and cross sections showing head and stage distributions allow groundwater flow directions and exchange conditions to be inferred (Figure 58). Groundwater elevations are derived from wells penetrating the associated, usually shallow, unconfined groundwater system. Networks of monitoring wells or piezometers are established using either existing wells or specifically-installed, small-diameter (~5 to 10 cm) wells. Well designs usually include a short, slotted section near the bottom of the well, a cap on the bottom of the well, and closed piping extending to the surface (Figure 59) (e.g., Sterrett, 2008). Other well designs are appropriate when installation conditions are difficult or budgets are restrictive (e.g., hand driven sand points, jackhammer-driven tubes, and jetted piezometers; Sterrett, 2008).

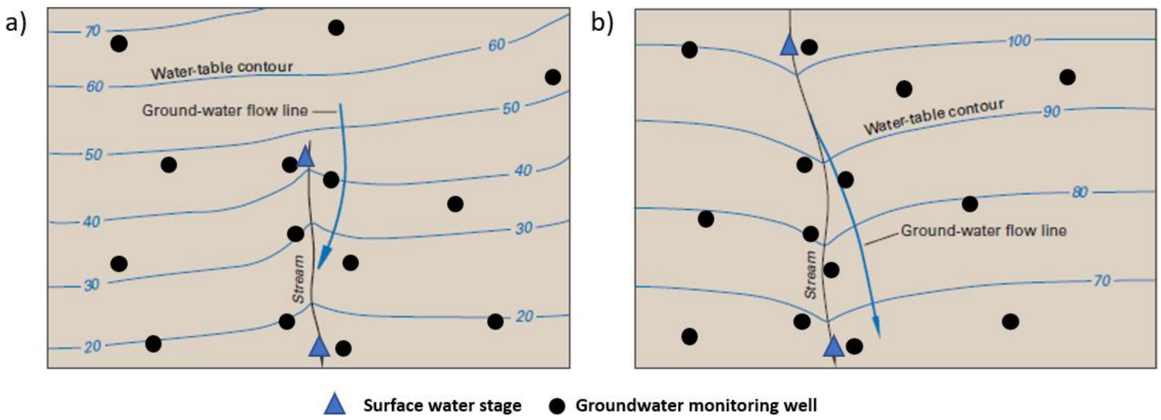


Figure 58 - Schematic of monitoring well networks and stream stage stations used to assess groundwater-stream interaction. Geologic information, head data, and river stage elevations are used to construct an equipotential map (blue lines and relative numbers) and generate groundwater flow lines (assuming steady state, isotropic and homogeneous conditions). a) Gaining stream. b) Losing stream connected to the water table (modified from Healy et al., 2007).

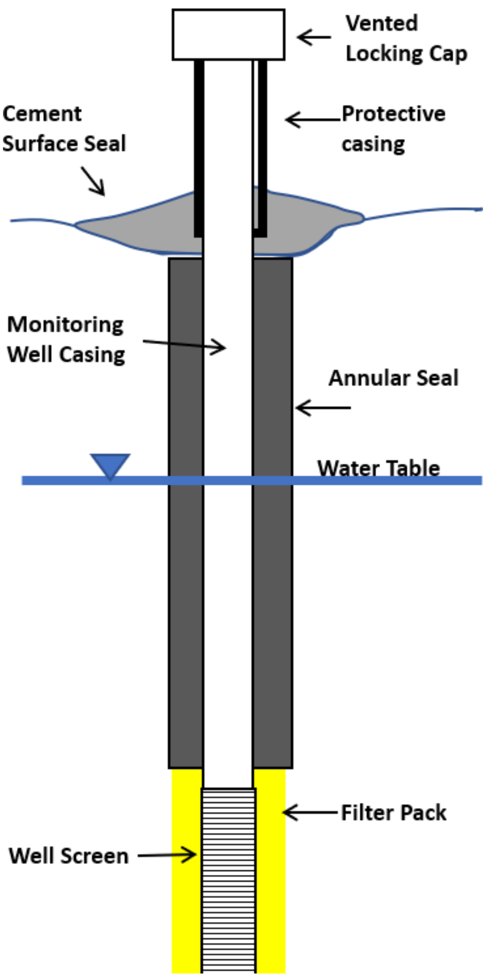


Figure 59 - Typical groundwater monitoring well design. The cement surface seal and low permeability annular seal material limit the leakage of surface-water runoff around the well casing. The filter pack is typically a higher hydraulic conductivity material that facilitates the movement of water to the portion of the well open to the groundwater system (the well screen in this example) (Woessner, 2020).

In most cases, wells are installed in the unconfined groundwater system interacting with the surface-water feature. In addition to wells completed near the water table, a portion of the wells should be paired with one or more deeper wells to measure the direction and magnitude of vertical hydraulic gradients (Figure 60). Gradients are computed as the difference in heads (100-105 in Figure 60) divided by the difference in well depths (30 in Figure 60). The gradient's negative sign shown in Figure 60 is a convention indicating that flow is from areas of high to low hydraulic head. Thus, when Darcy's Law is used to generate groundwater discharge, flux, velocity or estimates of hydraulic conductivity, gradients are computed by subtracting the head at some distance down gradient from the head at the starting point. Woessner and Poeter (2020) discuss the use of Darcy's Law to quantify flow and provide further detail on the convention for calculating gradient in Section 4.3 of the Groundwater Project book↗ titled "Hydrogeologic Properties of Earth Materials and Principles of Groundwater Flow". The vertical movement of groundwater is described as upward (head in deep piezometer is greater than head in shallow piezometer) or downward (head in deep piezometer is less than head in shallow piezometer). Multiple monitoring well depths at a location also allow for detailed water quality sampling and aquifer property determination.

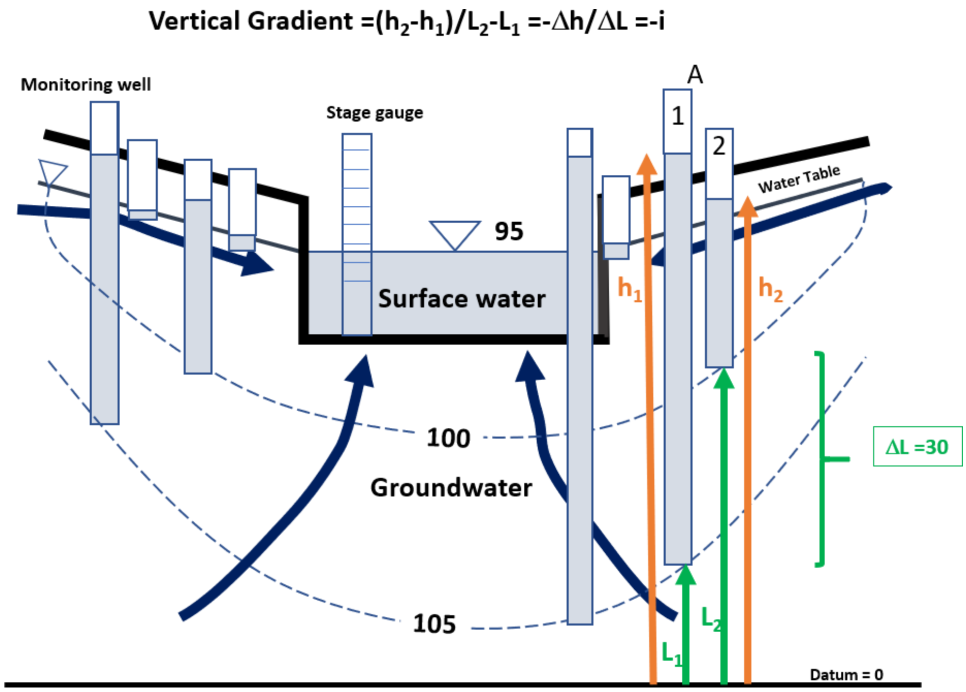


Figure 60 - Pairs of shallow and deep monitoring wells provide information on vertical gradients for an effluent condition. Equipotential lines are dashed, and groundwater flow arrows are shown in dark blue. Flow direction interpretation assumes isotropic and homogeneous conditions. There is no vertical exaggeration in this section. Water levels in open-ended monitoring wells are shown in light blue. The vertical gradient at location A would be computed as the difference in heads between wells 2 and 1 (orange arrows) (in this case 100-105=-5) divided by the distance separating the ends of the piezometers (L_2-L_1, green arrows), ΔL=30, -5/30=-0.17(Woessner, 2020).

Water levels in monitoring wells are determined using steel or electric water-level measuring tapes, transducers, and/or acoustical tools. The tops of the well casings, from which water levels are measured, and adjacent stream stage stations are surveyed to a common datum for the study area (e.g., mean average sea level). The elevation of groundwater levels, streams, lakes and wetlands are typically determined by standard surveying techniques (e.g., level, theodolite, survey grade GPS). Survey errors should be sufficiently small so that differences between stages and groundwater elevations are identifiable at the relevant scale of investigation.

Water level data are contoured, and groundwater flow lines mapped using methods appropriate for the identified anisotropy in aquifer materials (e.g. Woessner and Poeter, 2020). Representations of groundwater flow in cross sections need to account for both cross-sectional vertical exaggeration and anisotropy (e.g., Freeze and Cherry, 1979; Anderson et al., 2015; Woessner and Poeter, 2020). Monitoring well designs should also accommodate likely water quality sampling tools and methods, and be made of materials that do not compromise groundwater geochemistry (e.g., Sterrett, 2008). The extent of the well network is based on the domain boundaries and the importance of the far-field groundwater conditions (requiring wells farther from the site-specific study site) to describe exchange (e.g., Anderson et al., 2015).

To determine vertical gradients between a surface-water body and the groundwater using an inexpensive tool, small diameter tubes (often less than 2.54 cm diameter and a meter or two long) can be installed to various depths directly in the bed of the surface-water feature. They can be installed as a temporary instrument or left in place for future measurements. Most commonly, these mini-piezometers are placed below the unconsolidated bed of the surface-water feature to examine the distribution of vertical gradients (e.g., Lee and Cherry, 1978; Simonds and Sinclair, 2002; Cox et al., 2005; Buss et al., 2009). The Rosenberry and others (2008) publication is an excellent reference describing mini-piezometer design and use.

5.6 Mini-Piezometers

Mini-piezometers (or piezometers) are commonly open at the end or include a perforated interval a few centimeters long with a series of cut slots or drilled holes near the bottom end of a metal or PVC pipe (Woessner, 2017). In areas with fine sediments steel wool is sometimes installed inside the end of the pipe or a sleeve of nylon mesh wrapped around the perforated interval to prevent plugging (Lee and Cherry, 1978; Simonds and Sinclair, 2002). Mini-piezometers can be driven and pounded into floodplain sediments with shallow water tables (e.g., Rivett et al., 2008; Brodie et al., 2007) and, less commonly, installed in drilled holes in bedrock stream bottoms (Kennedy, 2017).

The mini-piezometer is typically hand driven into the bed to a specified depth (e.g., 10 to 100+ cm) (Figure 61). In some cases, a conductor metal casing with an expendable bottom plug or center rod is driven to depth and then a slightly smaller diameter piezometer is inserted, and the conductor casing removed. The unconsolidated sediment is assumed/encouraged to settle around the piezometer, sealing it into the sediments (preventing short circuiting of surface water). In some cases, a flexible clear tube long enough to extend above the surface of the surface-water feature can be inserted in the conductor casing and used as the mini piezometer (Lee and Cherry, 1978). Installation of mini-piezometers in consolidated sediments (bedrock) requires mechanical drilling of boreholes and piezometer installation that seals the perforated interval from the surface-water body (Figure 62).

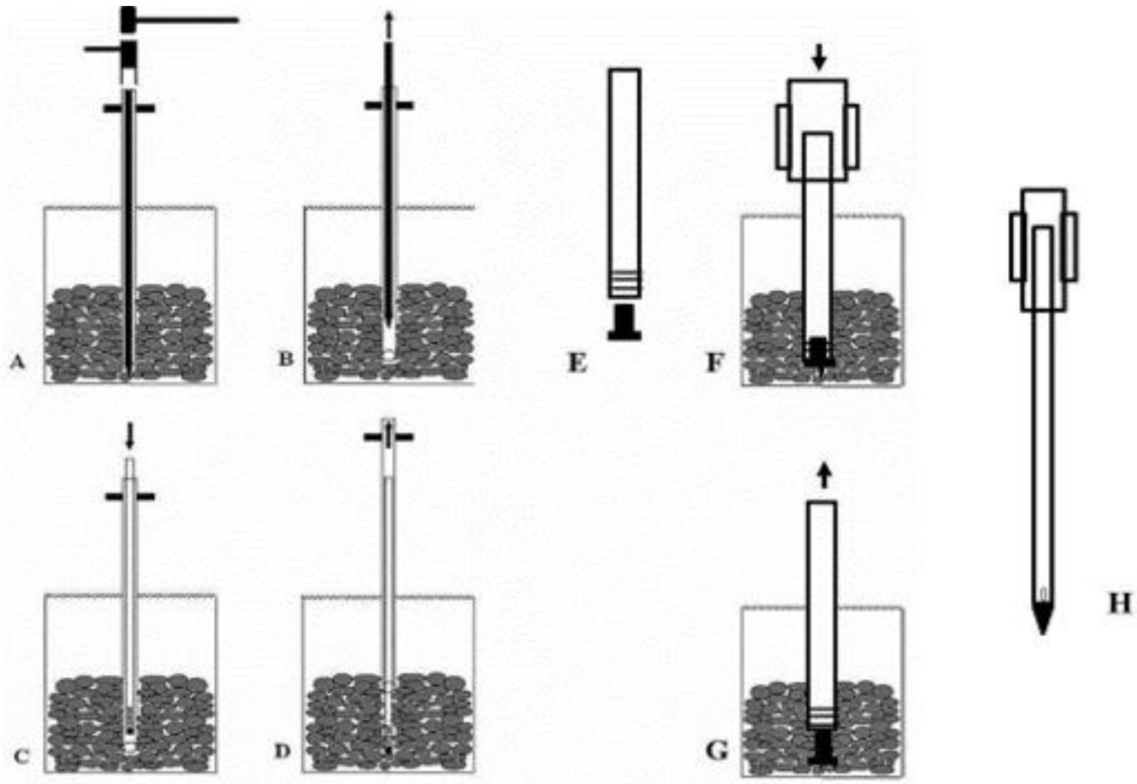


Figure 61 – Mini-piezometer installation in unconsolidated sediments. a) through d) installation using a driving system with an inner solid rod and outer hollow conductor casing. The dual system is driven to the desired depth and the center rod is removed. A smaller diameter perforated mini-piezometer is inserted and the outer casing is removed (from Baxter et al., 2003). e) through g) installation using the driven casing as a mini-piezometer. A loose-fitting disposable tip (e.g., a bolt) is placed in the open end and the pipe is driven to depth using either a post driver or a sledgehammer. Once the desired depth is achieved, the piezometer is pulled back a few centimeters to allow the disposable tip to be retained in the sediment. The piezometer is then open to the sediments. Once a mini-piezometer is installed, the bottom sediment should be tamped around the piezometer to prevent water short circuiting the instrument. h) A fixed, pointed-tip, perforated, ridged-steel pipe (drive point piezometer) driven into the bed (modified from Woessner, 2017).

Kennedy (2017) used small coring tools to install mini-piezometers and seepage meters in a dolostone stream bottom. A Shaw Backpack Drill (Shaw Tool Ltd., Yamhill, OR, USA) was used to install riverbed piezometers (Figure 62a). Lucius (2016) also described mini-piezometer bedrock installation methods to measure heads and nitrogen rich groundwater discharge to a lake (Figure 62b).

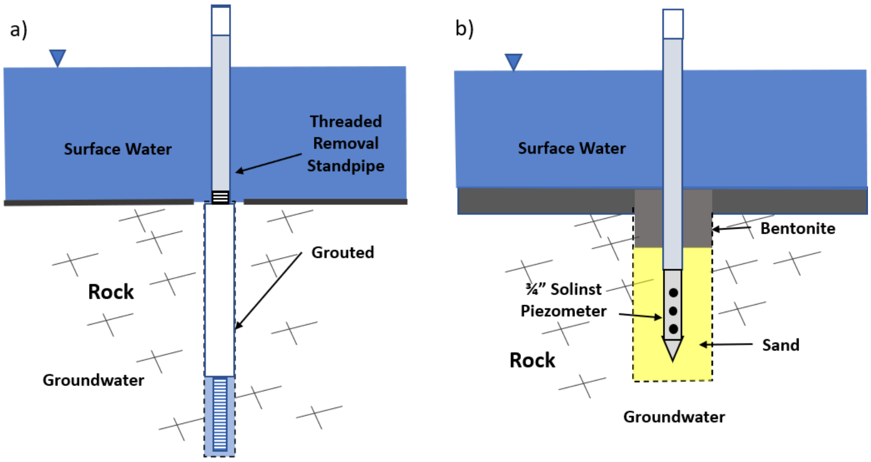


Figure 62 - Bedrock mini-piezometer installation. a) Installation of a river piezometer in a 5 cm diameter core hole in a dolostone riverbed. The steel piezometer was grouted using a grout-in-place method described by Pierce and others (2018). Head and temperature of the groundwater were measured both by a transducer suspended in the open hole and manually in the removable standpipe (modified from Kennedy, 2017). b) Mini-piezometer installed in weathered limestone. A 15.2 cm diameter PVC pipe was driven into the bottom sediments and weathered bedrock, and cleaned out with an auger and pump. The 1.9 cm diameter Solinst piezometer was installed on a PVC standpipe. As the large diameter PVC pipe is removed, sand was installed around the piezometer tip and bentonite was installed in the upper portion of the hole (modified from Lucius, 2016).

To determine the magnitude of the vertical groundwater gradient at the mini-piezometer location, the difference between the groundwater elevation (head) as measured in the mini-piezometer and surface-water stage (Δh) is divided by the depth to which the instrument penetrates the sediment (L) (Figure 63a and b). When a slotted or perforated interval is used, the midpoint of the perforations is used as the penetration depth. The convention used in the exchange literature is that gradients referred to as vertical hydraulic gradients, VHGs, are computed relative to the groundwater level in the mini-piezometer. A positive gradient (groundwater level is higher than the surface-water stage) indicates effluent conditions, the upward movement of groundwater, thus groundwater discharges to the surface-water body (e.g., Simonds and Sinclair, 2002; Woessner, 2017). A negative gradient indicates surface water is recharging the underlying groundwater, downward movement of groundwater (influent conditions) as shown in Figure 63b. The VHG may also be computed from measuring heads from the top of the mini-piezometer casing using the casing as the local datum (Figure 63c and d). VHGs are commonly cited in groundwater-surface water exchange literature and mapped spatially and temporally to indicate where groundwater is entering and leaving a surface water body bed. The concept of upward flow of groundwater as a positive gradient is a convention of the groundwater exchange literature. However, exchange directions and the VHG should be clearly defined when reporting values. It should be noted that in some works, only the vertical difference between the groundwater head and surface-water elevation is reported and used to infer the direction of water movement. This information is qualitative, though useful. However, it is important to note that this difference is not the VHG as the water level difference is not divided by the depth of mini-piezometer penetration (L) into the bed. Gradients are needed to compute groundwater discharge, flux and velocities.

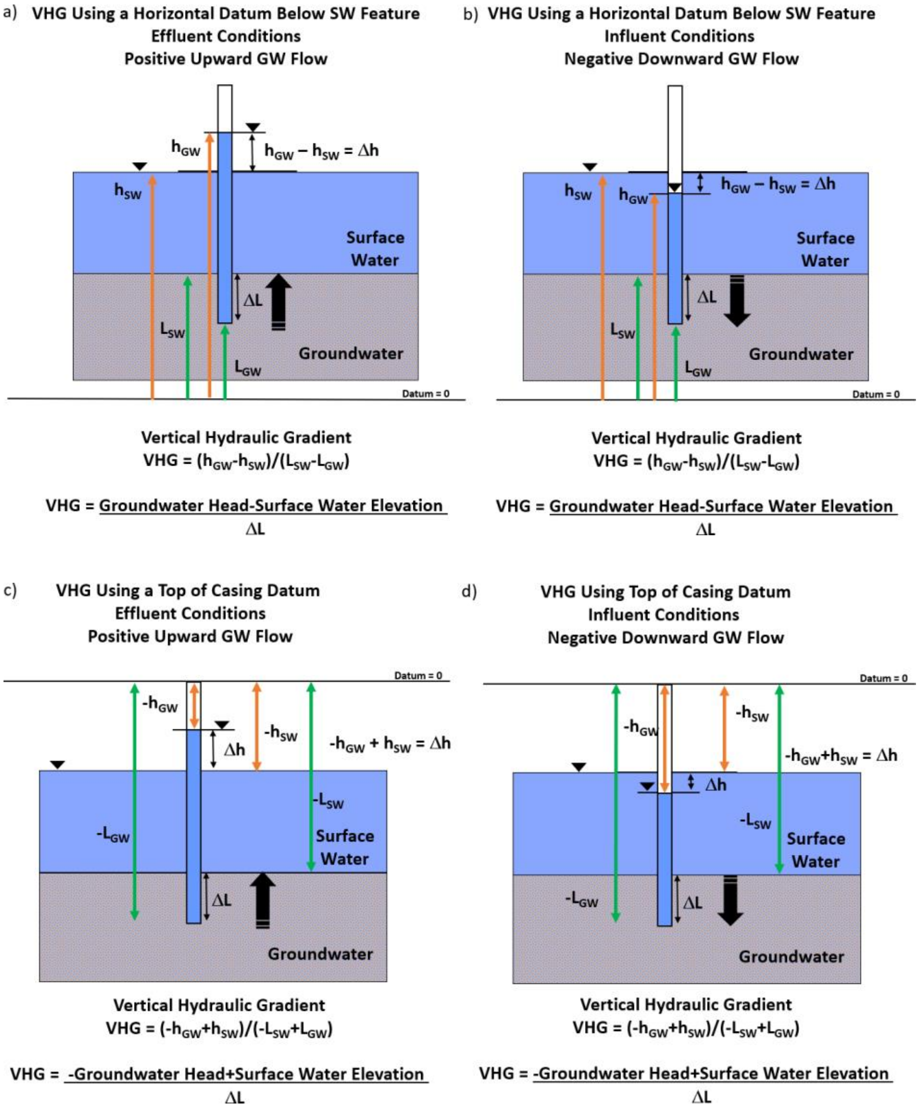
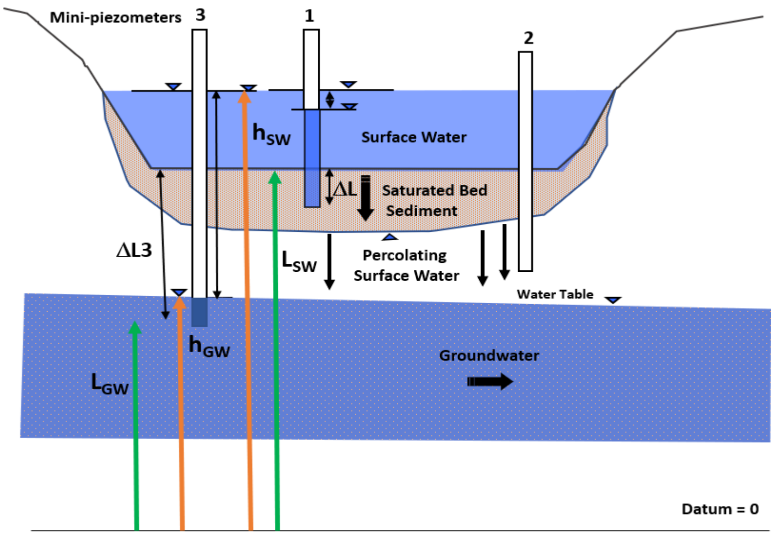


Figure 63 - Vertical hydraulic gradient (VHG) is used to indicate the direction of movement of groundwater at a surface water site. It is computed relative to the position of the groundwater head. When the groundwater head is higher than the surface water head (level) groundwater flow is upward into the bed. This is referred to as a positive VHG. When the surface water head is higher than the groundwater head influent conditions are present. This is referred to a negative gradient, downward flow into the underlying ground water system. Orange arrows represent heads (water levels) measured in the mini-piezometer and the surface water feature. Green arrows are distances (L) to the bottom of the piezometer or bed of the surface water feature. The difference between the groundwater head measured in a mini-piezometer installed to a depth ΔL in the bottom sediments and the surface water stage is Δh. The large black arrow represents the direction of groundwater flow. The mini-piezometer is open only at the bottom. a) When the groundwater head is greater than the surface-water elevation, the VHG is positive and groundwater flow is into the surface-water feature (effluent). b) When the groundwater head is lower than the surface-water stage the VHG is negative and conditions are influent. c) Using a datum of the top of the mini-piezometer, when the groundwater head is above the surface water level flow is effluent. d) When the top of the mini-piezometer is used as a reference, flow is downward when the groundwater head is below the surface water head (Woessner, 2020).

When mini-piezometers are driven into the bed of an influent feature and the water table is connected to the surface-water feature, VHG represents saturated flow conditions (Figure 63b and d). However, if the surface-water feature is disconnected from the water table (water table is below the bottom of the saturated sediments, e.g., losing river, lake or wetland, or disconnected wetland) then mini-piezometer data will reflect vertical hydraulic gradients that do not represent saturated flow conditions (Figure 64 mini-piezometer 3). If such conditions are present, theoretically, a piezometer could be installed in the percolation zone (vadose zone) and no groundwater level would be present (mini-piezometer 2 in Figure 64). VHG values computed for conditions shown in mini-piezometer 3 may yield large negative gradients, in some cases values larger than -1, when head change and penetration ratios are large. In general, when a large negative VHG is computed, data should be reviewed to determine if the influent feature is disconnected from the water table.



Vertical Hydraulic Gradient at Mini-piezometer 3

$$VHG = (h_{GW}-h_{SW})/(L_{SW}-L_{GW})$$

$$VHG = \frac{\text{Groundwater Head-Surface Water Elevation}}{\Delta L}$$

Figure 64 – Mini-piezometers installed in an influent surface-water feature with saturated bed sediment that is disconnected from the water table. Water moves from the surface-water feature to the saturated bed sediment (large vertical back arrow). Perched groundwater then enters the vadose zone and percolates to the underlying water table (small black arrows). The head in mini-piezometer 1 reflects saturated flow conditions between the surface water and its saturated bed. A downward VHG would be computed. Mini-piezometer 2 is located in the vadose zone and no water enters the bottom because the sediments are not fully saturated, so the piezometer is dry. The absence of water in the piezometer does not allow for a calculation of a VHG value. Mini-piezometer 3 is completed in the underlying groundwater system. Orange arrows illustrate heads and green arrows distances (Figure 63 provides additional details). The computed VHG at Mini-piezometer 3 is not representative of continuous saturated flow. In this illustration the head difference is large and the depth of penetration (ΔL at mini-piezometer 3) is also not representative of saturated conditions. (Woessner, 2020).

Water levels used to compute VHGs can be measured using several tools. Most often, the top of the mini-piezometer is used as a local reference as shown in Figure 63c and d. Measurements recorded are the depth to water inside the mini-piezometer, the distance from the top of the mini-piezometer to the surface of the surface-water feature and the depth of penetration into the bed (Figure 65). The type of tool used to measure the water level is a function of the inside diameter of the mini-piezometer. Thus, when designing piezometers, the measurement tool should be considered. Tools include a small diameter electric tape, steel tape, and/or chalked rod and measuring tape (Figure 65) (Woessner, 2017; Baxter et al., 2003). Transducers installed both in the mini-piezometer and surface-water body can be used to record head difference over time (e.g., Freeman et al., 2004). In some cases, where head differences are small or difficult to measure, separate clear flexible tubing can be inserted in the water body and attached to the mini-piezometer, linked, and a vacuum applied to this loop. This manometer board setup draws water levels above the surface and to a board with a vertical scale. Water levels are compared, and the differences noted (Figure 65) (e.g., Lee and Cherry, 1978; Simonds and Sinclair, 2002; Cox et al., 2005).

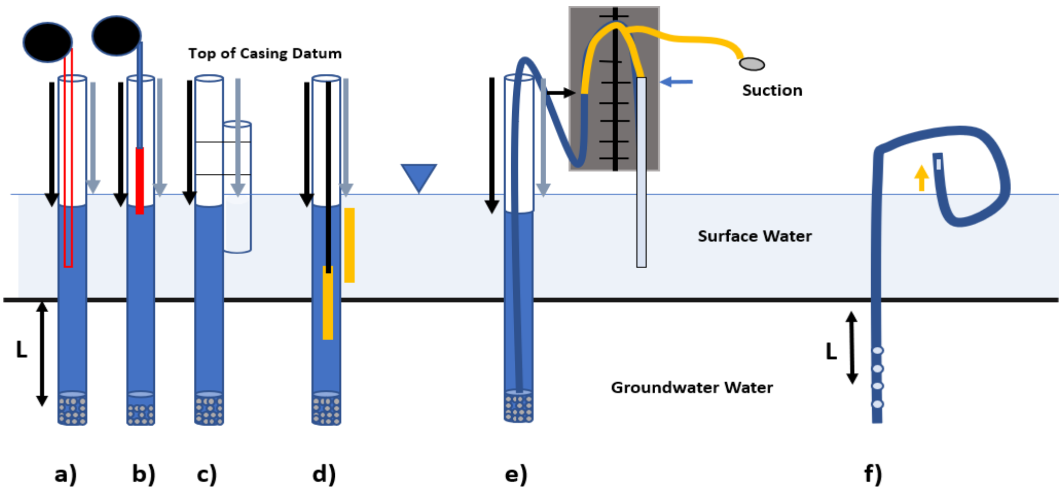


Figure 65 - Measuring water levels in mini-piezometers. L is the penetration depth of the mini-piezometer measured from the surface-water feature bottom to the midpoint of the perforated interval. The datum is the top of the mini-piezometer casing. The black arrow represents the depth of water in the mini-piezometer and the light blue arrow the distance to the surface-water surface measured along the outside of the casing. Examples a) through e) are representative of influent conditions and f) is an effluent condition. a) A steel tape is lowered below the water level in the piezometer and the depth below the top of the casing is calculated as the hold measurement (value at the top of the casing) minus the portion of the tape that is wet. The tape is then stretched along the outside of the casing to measure the relative surface water level. b) Measurement using an electric tape. The probe is lowered into the well until the water level is indicated. The hold point (top of casing) is then read as the depth of water. The probe is then lowered along the outside of the casing to obtain the stream elevation. c) A mini-piezometer setup with a second hollow tube extending into the stream. This design is used to make a measurement of the water level of a turbulent stream surface. d) Installation of a pressure transducer (orange rectangle) in the mini-piezometer and a second one attached to the outside of the casing to record surface-water levels (Woessner, 2017). e) The use of a manometer board. An open-ended clear flexible tube is submerged both in the well and in the stream. It is fitted with a tee and suction is applied to raise water levels to the manometer board. The resulting difference in water levels is measured with the mounted ruler (Winter et al., 1988; Cox et al., 2005). f) Effluent conditions are instrumented with a small diameter mini-piezometer using a hollow root feeder tube (section with small circles) that is inserted into the sediment and an attached to a flexible clear tube. The water level in the clear tube is observed as the positive difference in water levels (small orange arrow). VHGs are calculated as shown in Figure 63c and d. (modified from Rosenberry et al., 2008).

Mini-piezometer gradient data can be used with measurements of sediment and bedrock hydraulic properties to compute local flux rates from Darcy's law, and with seepage meter data to compute local bed-sediment hydraulic conductivities. When mini-piezometers are designed with a perforated interval that allows water to freely enter and leave the piezometer, falling head or constant head slug tests can be conducted to estimate the horizontal hydraulic conductivity of the sediments penetrated by the perforated interval (e.g., Hvorslev, 1951; Bouwer and Rice, 1976, 1989; Van der Kamp, 1976; Butler, 1997; Butler et al., 2003; Butler and Healey, 1998). Sampling of bed and floodplain sediments and coring of bedrock can be used to estimate and measure hydraulic conductivities using lab and field methods (e.g., Freeze and Cherry, 1979; Fetter, 2001; Cedergren, 1997; Woessner and Poeter, 2020). As the exchange of water at the bed is assumed to be vertical, vertical hydraulic conductivity values are needed. Often vertical hydraulic conductivity is estimated from horizontal hydraulic conductivity by assuming an anisotropy ratio, that is a ratio of horizontal hydraulic conductivity (K_h) and vertical hydraulic conductivity (K_v) (e.g., Fetter, 2001; Anderson et al., 2015). The challenge with this approach is selecting an appropriate ratio; the range is typically between 1 and 1000. Ideally, independent methods to measure or estimate the vertical hydraulic conductivity directly in some of the locations where horizontal values are estimated is recommended. This can be accomplished in some settings by using lab permeameter measurements on undisturbed vertical cores of the site sediments and bedrock. If conditions permit, pushing or pounding of an open-ended pipe into the bottom sediments (e.g., on the order of 20 to 50 cm) can be used to conduct an in situ falling head permeameter test (e.g., Kennedy et al., 2010). Pairing mini-piezometer measurements with seepage meter flux values is a common method used to compute vertical hydraulic conductivities when both instruments are installed at a single site.

When both mini-piezometer VHG data and estimates of vertical hydraulic conductivities are obtained at a site, flux rates can be computed using Darcy's law (assuming vertical flow and steady state conditions) as shown in Equation 3. When gradients derived from VHG determinations are used in Darcy's Law related equations the convention that groundwater flow is always from high to low heads applies and the value of the measured gradient at a site is always entered as a negative value so that the computed term is positive. The sign of the VHG can then be used to describe if the discharge, flux or velocity is related to upward or downward movement of site groundwater. For example, field gradient data from measurements of groundwater levels in mini-piezometers and surface-water stages can be used to calculate the quantity of groundwater flux through the bed as shown in Equation 3.

$$Flux = \frac{Q}{A} = -K_v i \tag{3}$$

where:

Q = discharge (L^3/T)

$$A = \text{cross-sectional area } (L^2)$$
$$Q/A = \text{flux } (L^3/(L^2T))$$
$$K_v = \text{vertical hydraulic conductivity of the sediments } (L/T)$$
$$i = \text{gradient always entered as a negative value of the measured VHG}(L/L)$$

5.7 Seepage Meters

Seepage meters measure the flux of water between groundwater and a surface-water feature. Conceptually, a seepage meter is a container open at one end. It is often a 55-gallon (208-liter) drum with the open end pressed into the bed sediments so that it isolates water exchange. On its closed top a hole fitted with flexible tubing and a valve is present to which a thin-walled clear bag is attached (Figure 66). Exchange rates are computed by monitoring the change in bag volume over time and then dividing the value by the area of the bed the container covers. Rosenberry and others (2008) provide a detailed description of the design and use of seepage meters. Seepage meters have been used successfully in streams, lakes, wetlands and oceans. An excellent reference on seepage meter use and operation is the work of Rosenberry and others (2020) that addresses the history and evolution of seepage meters.

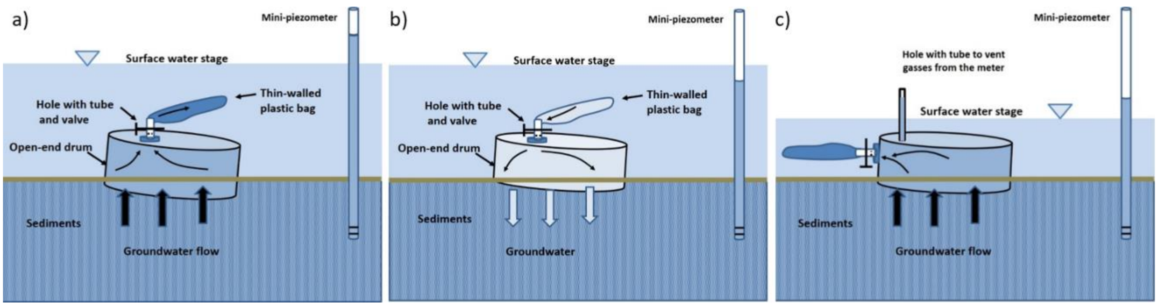


Figure 66 - The seepage meter design consists of an open-end container (like a 55-gallon/208-liter drum that is outfitted with a valve and thin walled bag (e.g., 0.05 mm wall 1-gallon/3.78-liter food storage bag). The bag is used to collect seepage or release water if flow is into the sediments. The meter is installed tightly into the bottom sediments so that leakage does not occur around the edges. In these examples, a mini-piezometer is installed nearby each meter to verify exchange directions and site-specific vertical hydraulic gradients (VHGs). It is assumed mini-piezometers penetrate sediments with the same hydraulic properties as those encapsulated by the seepage meter. a) Installation in an area with effluent groundwater. b) Installation in an area where influent conditions are present. The bag must be prefilled with a known volume of water and once the meter is installed the change in volume over time is computed. c) Seepage meter installed in an effluent shallow water setting. The valve and bag are fitted to the side. A hollow tube is installed through the top of the meter to release gasses that may interfere with seepage rate determination (Woessner, 2018).

Seepage meters are often constructed from containers 10 to 100 cm in diameter sampling about 0.25 m^2 or less of the bottom sediment (e.g., Rosenberry, 2005). They can be sized to meet anticipated site conditions. Natural sediment heterogeneities may result in a wide variation of seepage rates at a study site. In most cases, larger diameter meters are desirable because they integrate seepage over more of the bottom sediments being evaluated (e.g., Isiorho and Meyer, 1999). The standard seepage meter is operated by installing the drum, attaching a partially pre-filled, thin-walled bag sealed to a tube with a valve that is attached to the container (e.g. Israelsen and Reeve, 1944; Lee, 1977; Lee, 1972; Lee and Cherry,1978) (Figure 66 and Figure 67). The valve is opened and the change in water volume in the bag over time is measured. Based on the area of the meter, the flux (discharge per area) is computed. Seepage meters have been used to measure exchange in lakes (e.g., Lee and Cherry, 1978; John and Lock, 1977; Connor and Belanger, 1981; Erickson, 1981; Woessner and Sullivan, 1984; Isiorho and Matisoff, 1990; Shaw and Prepas, 1990b; Lesack, 1995; Rosenberry, 2000; Sebestyen and Schneider, 2001), stream channels (Lee and Hynes, 1977; Connor and Belanger, 1981; McBride, 1987; Libelo and MacIntyre, 1994; Blanchfield and Ridgeway, 1996; Jackman et al., 1997; Cey et al,1998; Fryar et al., 2000; Dumouchelle, 2001; Landon et al., 2001; Murdoch and Kelly, 2003); wetlands (e.g., Choi and Harvey, 2000) and ocean shorelines (e.g., Shinn et al. 2002; Cable et al. 1997; Chanton et al, 2003; Taniguchi, 2002). Zamora (2008) provides a good summary of the seepage meter literature as does Rosenberry and others (2020). In lake and wetland settings, meter shape should not be a limiting factor. However, in flowing water conditions, such as streams and at tidally influenced coastlines, meters and collection bags are subjected to water flowing over the bag, a condition that has been shown to influence seepage results. Box 5 provides a more detailed discussion of seepage meter operation.

In some settings where seepage rates are typically low, interconnected ganged meter setups are used. This is accomplished by linking multiple meters to a single collection bag (Rosenberry, 2005) (Figure 67 C).

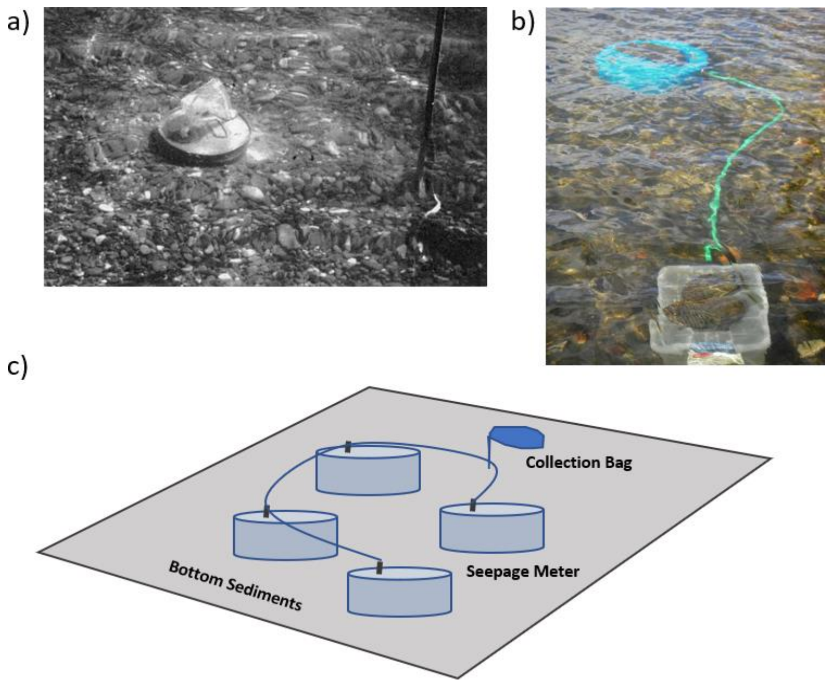


Figure 67 - Seepage meter installations. a) A meter made from a 55-gallon/208-liter drum deployed in Flathead Lake, Montana, U.S.A. (Woessner, 2020). b) A seepage meter in Lake Lacawac, Pennsylvania, USA, with an extended tubing and a plastic box protecting the collection bag (Heaney et al., 2007). c) Ganged seepage meters with seepage collection directed to a single bag (modified from Rosenberry, 2005).

Seepage meters provide a direct measurement of flux for the area captured by the meter. The meter flux is assumed positive when the volume of the collection bag increases and negative when the volume decreases. The flux is computed as shown in Equation 4.

$$\frac{Q}{A} = q = \frac{vol}{t\,A} \tag{4}$$

where:

Q = discharge (L^3/T)

A = area of the interface covered by the meter (L^2)

q = flux ($L^3/(L^2T)$)

vol = net volume of water accumulated or lost from the collection bag (L^3)

t = length of time the meter was operating once the valve was open until the measurement was completed (T)

Seepage meter data can also be used to compute vertical hydraulic conductivity of the sediments in the vicinity of the meter when paired with a mini-piezometer. It should be noted that the mini-piezometer VHG can be locally influenced by the hydraulic properties of the sediment within which it is completed. Assuming steady state, and isotopic and homogeneous conditions, the vertical hydraulic conductivity can be calculated as shown in Equation 5.

$$K_v = -\frac{q}{i} \tag{5}$$

where:

K_v = vertical hydraulic conductivity (L/T)

q = seepage meter flux ($L^3/(L^2T)$)

i = VHG measured by mini-piezometer and entered as a negative value (L/L)

Groundwater velocity (v) at the meter site can also be computed when a measurement or estimate of the sediment effective porosity (n) is available as expressed in Equation 6.

$$v = \frac{q}{n} = \frac{-K_v i}{n} \tag{6}$$

where:

v = vertical groundwater velocity (L/T)

n = effective porosity (L^3/L^3)

i = VHG measured by mini-piezometer and entered as a negative value (L/L)

Seepage meters are most often deployed in shallow water settings and deployed manually by wading or diving. Deeper water installation requires the use of diving equipment to set up a meter and collect data. Cherkauer and McBride (1988) employed a deep-water meter that was lowered to the bottom from a boat. An electronic valve started and stopped meter flow, which entered a tube that carried water to a collection bag located at a float near the water surface. Boyle (1994) also described the design of a deep-water meter deployed in a boreal forest lake.

Another challenge of using the seepage meter is that each time a measurement is taken the collection bag needs to be physically removed and re-installed under water. As a result, researchers developed other flow measuring methods to eliminate the need for the collection bag. Early approaches included a remotely deployed meter that used thermal perturbation methods to measure flows (Taniguchi and Fukuo, 1993). Paulsen and others (2001) developed a submarine seepage meter with an ultrasonic flow meter that measured both forward and backward flow. Data were logged with a tethered data logger. The United States Navy developed a deep water seepage meter, Ultra Seep↗, that is lowered to the bottom using a boat and cable (Figure 68a). It has a self-contained recording ultrasonic flow meter. The meter also collects geochemical data and is equipped to obtain water samples. The United States Geological Survey developed a seepage meter with an electromagnetic flow meter↗ (EFN) that requires external power (Rosenberry and Morin, 2004) as shown in Figure 68b. The EMF meter records rates from 30 ml/min to 30 L/min and works well in both shallow- and deep-water environments.

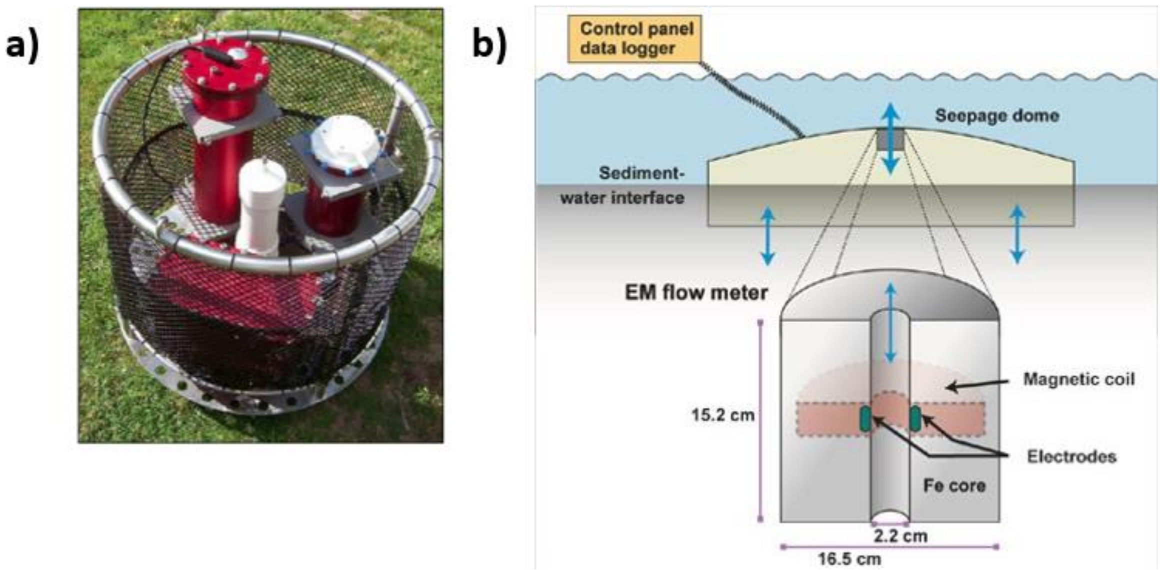


Figure 68 - Seepage meters with flow meters. a) A deep-water seepage meter, Ultra Seep, developed by the United States Navy. It has a self-contained recording ultrasonic flow meter. The Ultra Seep also collects geochemical data and is equipped to obtain water samples. b) Low profile seepage meter equipped with an electromagnet flow meter (EMF) developed by the USGS (Rosenberry and Morin, 2004). The flow meter is powered through a tethered control panel located on shore.

Seepage meters are rarely used in surface-water features with bedrock bottoms, as meters cannot be adequately seated. However, bedrock seepage meters (BSM) have recently been installed in the sedimentary bedrock stream bottom of the Eramosa River in Guelph, Ontario, Canada (Kennedy, 2017). In this study, meters were 5 cm in diameter placed in a hole that was 0.1 m deep bored in the bedrock using a portable concrete drill. The open borehole was outfitted with a low-profile expandable plug (J plug packer) containing a tube and valve as shown in Figure 69 (Kennedy, 2017). Seepage rates at this site were reported to be controlled by the presence or absence of fractures in or adjacent to the bore (e.g., Kennedy, 2017). Preliminary results are promising.

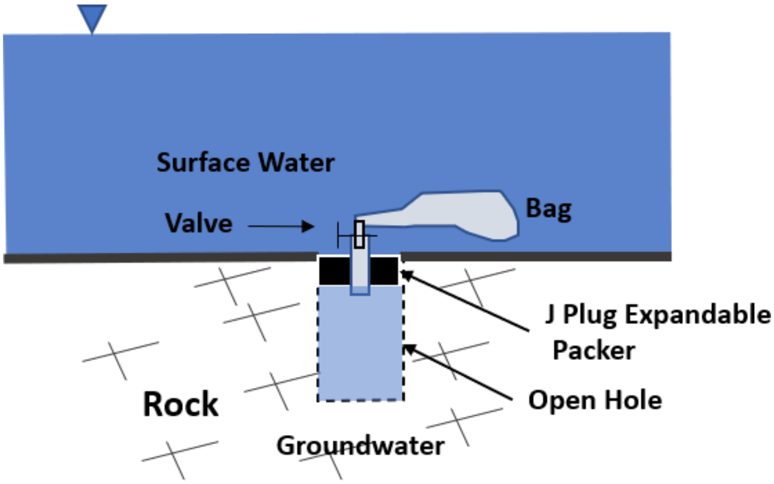


Figure 69 - Bedrock seepage meter (BSM) design. A borehole is drilled into the bedrock, left unlined (open hole), and fitted with an expandable packer (J plug). The packer can be expanded to seal the hole. It was modified to include a stainless-steel tube passing through the plug that is topped with a valve. A tube and thin-walled bag are attached to the valve to collect or supply water. In this application, a transducer was placed in the bedrock chamber to determine the head and temperature inside the meter (modified from Kennedy, 2017).

Exchange studies using seepage meters often deploy a number of meters with the anticipation that seepage patterns will be complex. Studies of seepage in lakes have often found higher seepage rates associated with the littoral zone and near shore areas; rates decreased with distance from the lake shoreline as noted in Figure 39 (e.g., McBride and Pfannkuch, 1975). However, sometimes zones of higher seepage rates are found farther from a shoreline. Toran and others (2014) collected two-dimensional electrical resistivity data and seepage meter data along a lake transect and evaluated the correlation of resistivity data and seepage rates. They concluded that detailed seepage patterns were poorly resolved with resistivity data. Seepage rates are also expected to vary with time. Seepage is dynamic, and measurements over time are recommended.

The seepage meter can also be used to collect a sample of the seepage water when groundwater is discharging to the surface-water feature. This is accomplished by allowing sufficient time for the surface water trapped during meter installation to flush from the meter. Then, a clean deflated bag is installed to collect a water sample once the valve is open.

5.8 Water Temperature

When surface water and groundwater temperatures contrast, exchange locations and, under some conditions, exchange rates can be inferred from temperature measurements (Bartolino and Niswonger, 1999; Anderson, 2005; Rosenberry and La Baugh, 2008; Webb et al., 2008; Buss et al., 2009; Boana et al., 2014). A comprehensive review of using heat as a tracer in groundwater settings is presented by Anderson (2005). Boana and others (2014) summarize the use of heat to characterize hyporheic exchange, and heat-flux/groundwater exchange modeling is described by Glose and others (2017). Examples of field applications that use VHG, seepage meters and temperature differences to characterize exchange are presented in Box 6↴.

Conceptually, when seasonal or daily surface-water temperatures vary, key components of the heat budget change, one of which can be related to the exchange between groundwater and surface water as illustrated in Figure 70 (e.g., Webb et al., 2008). When parameters of the heat budget for a surface-water feature can be measured, the flux of groundwater inflow can be determined either by solving for it as an unknown in the heat budget equation or by simulating the budget and fitting the groundwater exchange term to the balance, for example as was done by Glose (2017) using HFLUX↗ software. Schmidt and others (2007) describe the use of overall stream temperature changes to estimate groundwater discharge along a stream reach.

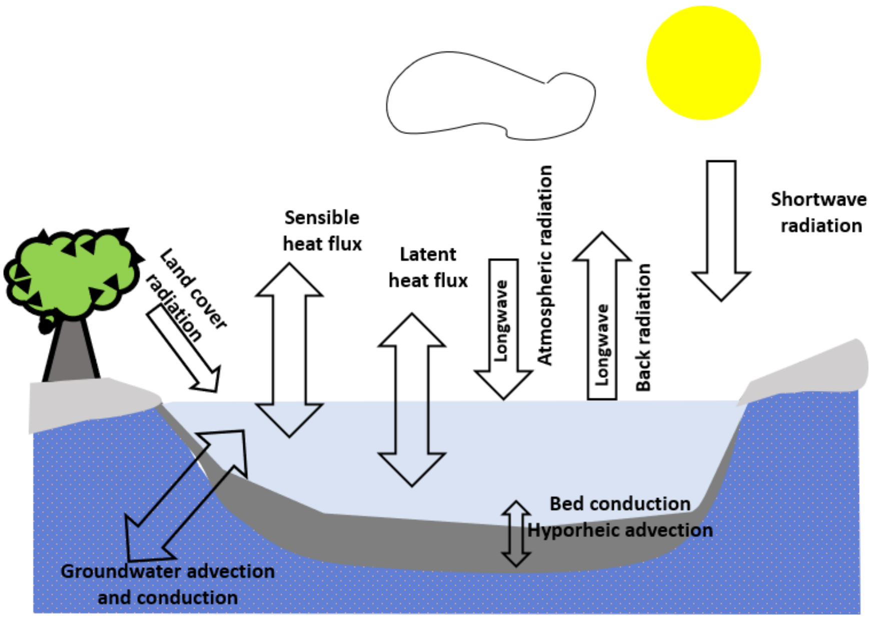


Figure 70 - Components of a heat budget for a surface-water feature that includes interactions with the bed, hyporheic water and groundwater systems (Woessner, 2018).

A basic example of using a heat budget to estimate exchange can be illustrated by examining a gaining and losing stream reach where flows and temperatures of each component are measured is shown in Figure 71. The streamflow and temperature are recorded at an upstream location and the temperature at a downstream location. These data are coupled with a measurement of the temperature of the shallow groundwater system adjacent to the stream. A mixing model is formulated with the quantity of groundwater discharging to the stream as the unknown, GW_{in}. Other factors affecting stream temperature, such as direct surface-water heating or shading are assumed to be small and thus not included in the balance. If this is not the case, then these influences will need to be measured and included in the balance. Influent stream reaches would not be affected by the site groundwater temperature and thus over a short reach no groundwater influenced temperature changes would occur.

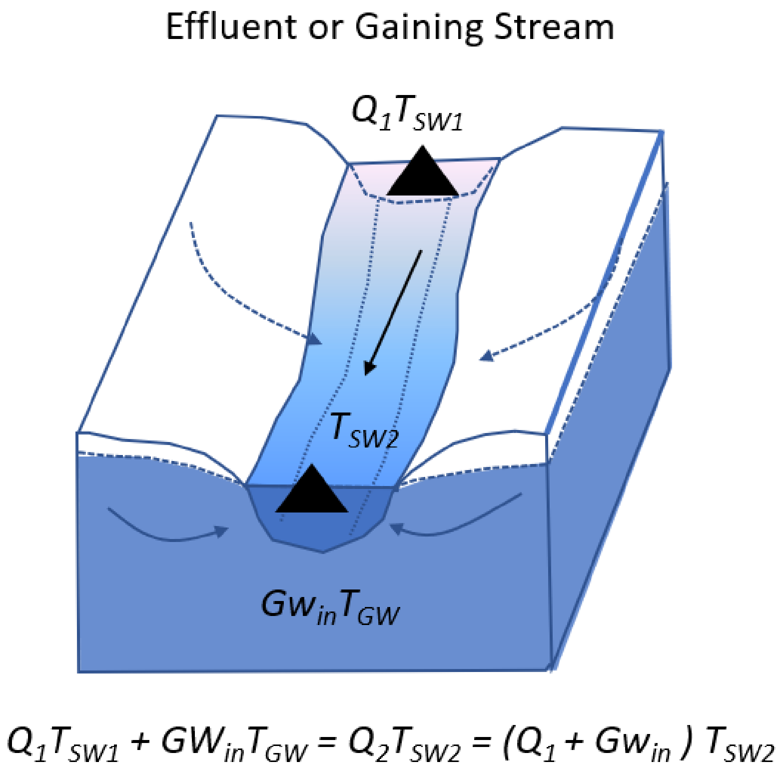


Figure 71 - Using a change in main-stream channel temperature to estimate exchange in an effluent stream reach. In this simple representation, pink is a higher temperature and dark blue is a lower temperature. Stream discharge, Q_1 is measured for the up-stream reach as is the corresponding surface-water temperature, T_{SW1}. Groundwater temperature, T_{GW}, is also measured as defined by shallow wells or samples from mini-piezometers extending into the shallow groundwater system (beyond the hyporheic system if it is present). At some distance downstream where there is a measurable difference in stream water temperatures, a second stream temperature measurement is made, T_{SW2}. The balance is solved for the unknown contributing groundwater discharge, GW_{in}. In this example, the change in stream temperature is a function only of the addition of a groundwater flux. If other factors have a significant influence on the reach heat budget (e.g., atmospheric radiation) the budget equation would need to account for the additional factors as indicated in Figure 70 (Woessner, 2020).

Observations of local changes in surface water, groundwater, and bottom, bank and shoreline temperatures can be used qualitatively to identify locations of diffuse and concentrated exchanges (surface water to groundwater or groundwater to surface water). Large groundwater temperature contrasts and high exchange rates make identification of exchange sites more likely. Often, daily or seasonal shifts (summer or winter) in surface-water temperatures contrast with more constant groundwater temperatures. When arrays of spatially distributed temperature monitors are installed in the bottom of surface-water features, maps of temperature contrasts can be used to infer exchange sites. For example, Loustaunau (2003) created bed-temperature transects by inserting a temperature probe 3 to 6 cm into the bed of Spring Creek near Ronan, Montana, USA (Figure 72). Groundwater was about 5℃ lower in temperature than surface water. Results showed groundwater discharging along the east stream bank and areas of focused groundwater discharge were identified. Conant (2004) mapped the temperatures 0.2 m below the stream bed of the Pine River in Canada with a temperature probe in both summer and winter. He used these data to understand where a plume of contaminated groundwater was discharging to the river (Conant et al., 2019). He also used mini-piezometer measurements adjacent to temperature measurement locations to derive VHGs, performed slug tests to estimate bottom sediment hydraulic conductivities, and computed exchange fluxes (Sections 5.5 and 5.6 describe these methods).

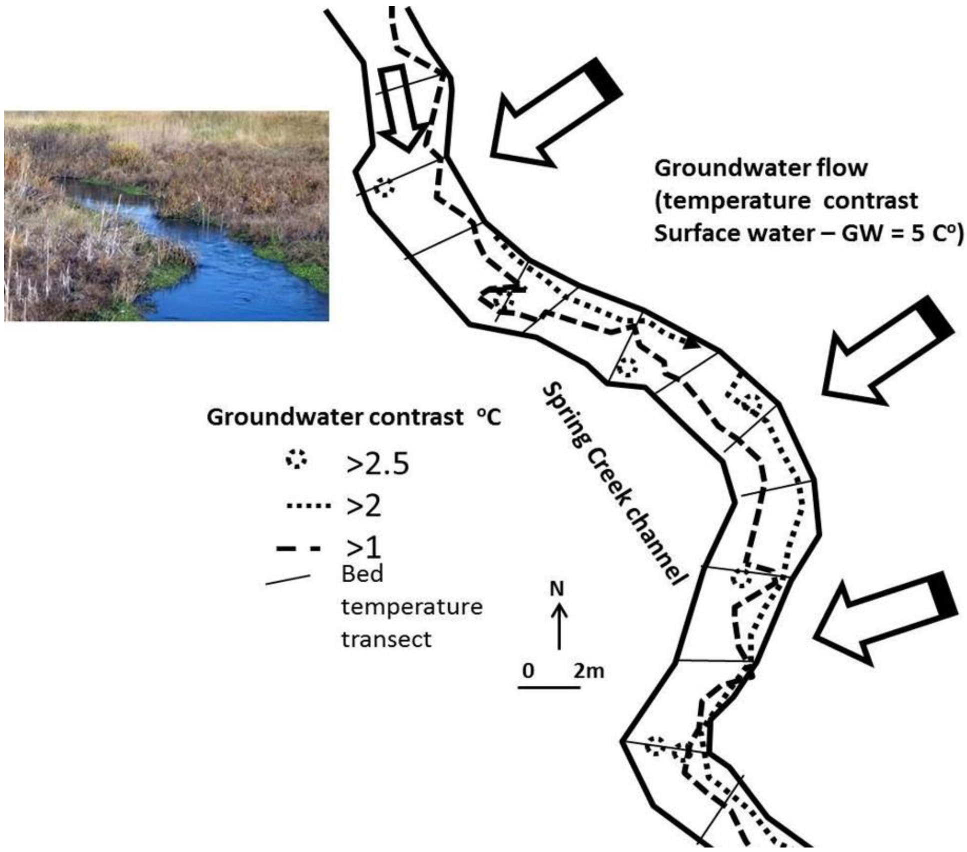


Figure 72 - Temperature mapping of bed waters in Spring Creek near Ronan, Montana, USA. Temperature contrasts show cooler groundwater is entering the right bank and bed of the stream (large arrows and temperature contrast contours). Areas of focused groundwater discharge were also mapped (dashed circles) (modified from Loustaunau, 2003).

Tools to measure water temperatures include a standard laboratory grade thermometer, digital thermocouple temperature monitors with metal probes, recording stand-alone temperature monitors, thermal imaging (remote sensing) cameras, and fiber-optic distributed temperature sensing cables. When groundwater discharge temperatures and rates are sufficient to cause local changes in surface-water temperatures (e.g., cooling or warming along banks or at other locations) methods that sense changes in the surface-water temperature on a local or more regional scale can be applied. These commonly include remote sensing of surface-water temperatures using thermal imaging cameras. This can be accomplished at local site investigations by using a ground-based hand-held thermal camera, one mounted on a drone or using an aircraft mounted FLIR (Forward Looking InfraRed) imaging sensor (Figure 73). Processed images use a color scheme to present temperature contrasts and require field measurements to calibrate the temperature mapping (Cox et al., 2005). The USGS (2020) website provides a good overview of the application of thermal imaging to studying groundwater-surface water exchange. An advantage of using thermal imaging is that it provides a detailed spatial map of temperature distributions; disadvantages are that new images must be acquired to document changes over time.

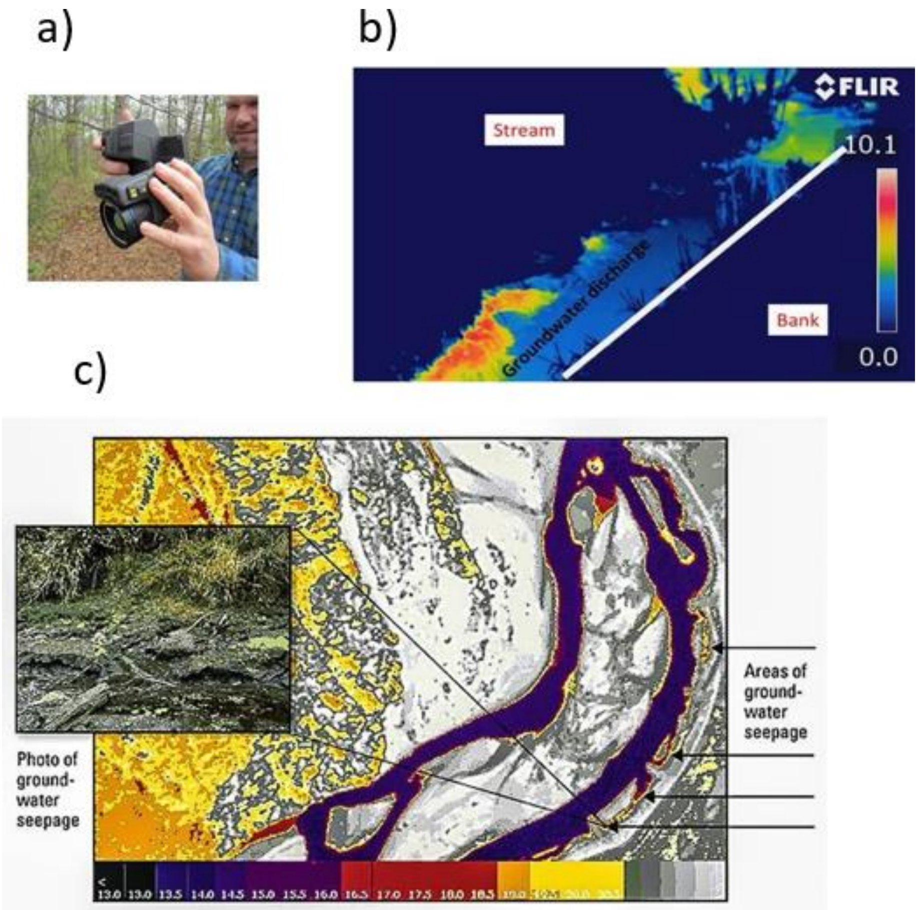


Figure 73 - Thermal remote sensing tools. a) Example of a thermal handheld camera. b) An image taken with a thermal camera of a field site during Autumn showing a 6 m portion of a stream bank (white line) at Tidmarsh Farms, Massachusetts, USA. Lighter colors represent changes in stream surface temperatures caused by warmer groundwater discharge (modified from USGS, 2020). c) FLIR image of a site on the Nooksack River, Everson, Washington, USA, showing warmer surface-water temperatures (brighter colors) along the east shoreline caused by groundwater discharge (Cox et al., 2005).

When bottom temperature patterns are evaluated to assess locations of exchanges, it is advantageous to use temperature sensors embedded in fiber-optic cables that are spaced at regular distances (e.g., 1 m) for example as explained by Selker and others (2006). The cables are then sampled and measurements recorded. A good summary of how fiber-optic distributed temperature sensing (FO-DTS) methods are used to evaluate exchange is found in USGS (2016). Installation requires that the flexible cable be placed on the bottom sediment- or bedrock-water interface. It can be placed linearly (along a river channel, Figure 74) or fitted to a surface area (multiple sections (loops) of the cable covering an area being investigated). Contrasts in temperature at the bed and the surface-water temperature are computed and mapped (Figure 75). Once the cable is installed, multiple measurements can be made as long as the cable remains fixed to the bottom of the surface-water feature. Deployment of the FO-DTS method in the Columbia River at the Hanford Site, Richland, Washington, USA, identified zones of groundwater discharge near the shore during periods of low river stages (Mwakanyamale et al., 2012). However, at high river stages, indications of groundwater discharge were not observable.

Figure 74 - Deploying a fiber-optic distributed-temperature sensor cable in the John Day River, Oregon, USA (OSU, 2020).

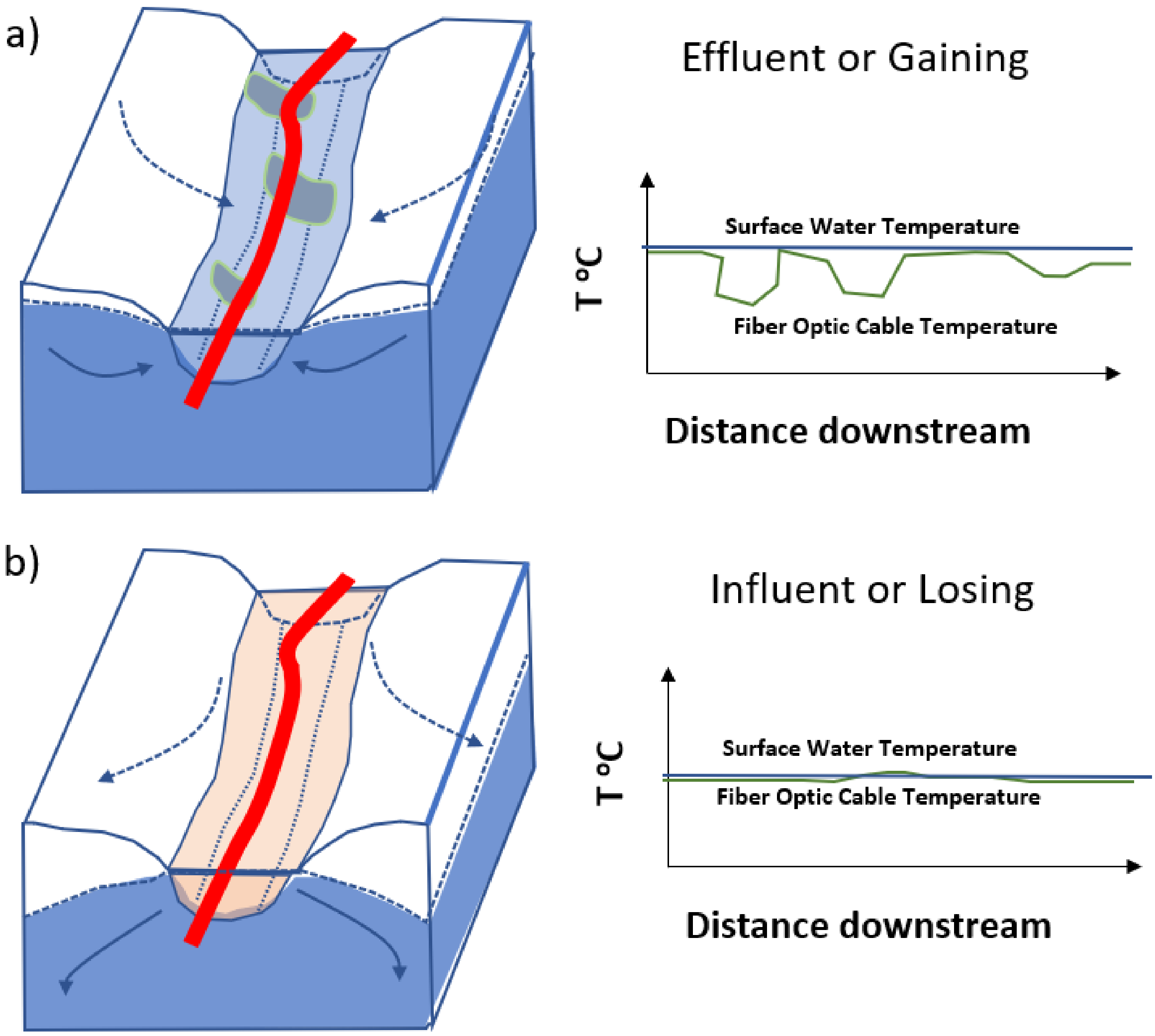


Figure 75 – Fiber-optic distributed-temperature sensor deployed in a stream and hypothetical results. River flow is from the top to the bottom of the page. Darker colors represent cooler temperatures. The cable sensors are spaced evenly. Arrows show the direction of groundwater flow. The dark blue line on each graph is the surface-water temperature. The green line represents the distributed temperature sensor readings. a) Effluent stream with areas of focused groundwater discharge (darker blue patches). Interpreted results show the general location of the exchange. b) Influent stream without areas of groundwater discharge. Because the stream is losing water into the stream bottom, temperatures do not vary and reflect the surface-water temperature (Woessner, 2020)

Point or single-time temperature measurements are used to identify locations and relative magnitudes of exchange. In most exchange settings, the locations and rates of exchange can vary spatially and temporally. To capture the nature of the local exchange process, individual or grouped recording sensors are required. Instrumentation can be tethered by cables to on-shore recording devices or be self-contained having internal batteries, clocks and sensors. Tethered sensors can be used in near-shore applications and where water or wave action does not damage the cable system. Most often, small diameter self-contained sensors are used to record the surface-water temperature and temperatures at multiple depths in the bottom sediments. Examples of self-contained individual sensors used in exchange studies include those shown in Figure 76. Sensors have defined temperature ranges, precision and accuracy limits; these properties must be considered when attempting to observe contrasts in groundwater and surface-water temperatures. Johnson and others (2005) provide information on the costs and operation of several sensors with a focus on their use of iButton sensors (Figure 76).

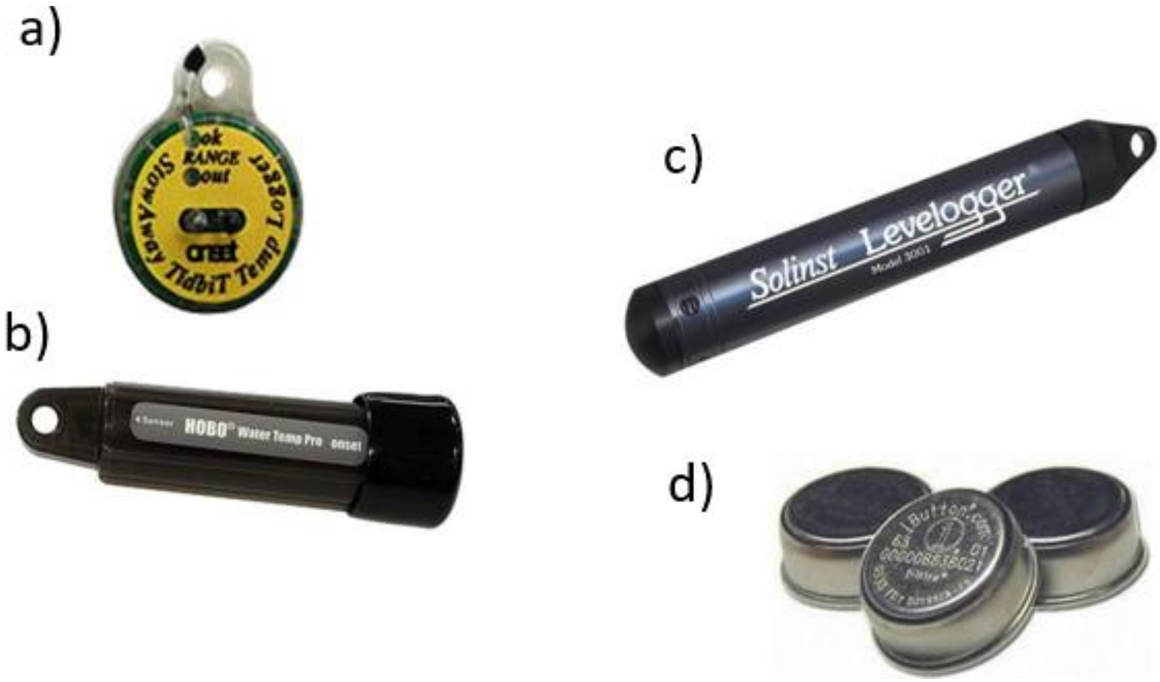


Figure 76 - Examples of commercially available self-contained recording temperature monitoring devices. Each allows recording temperatures on an investigator-specified time interval. a) Onset Tidbit↗. b) Hobo temperature pro↗. c) Solinst transducer Levelogger↗ contains a transducer that records stage and a temperature sensor. d) iButton Maxim-Dallas Semiconductor↗.

Deep water or high stream discharge may prevent placement of instruments by hand, in which case boat deployed tools can be used. One example is the Trident probe↗. The instrument can be pushed into the bottom sediment from a boat, and head, water quality and temperature data derived (Figure 77). It has been used to characterize conditions near freshwater and saltwater shorelines, and in large rivers and lakes. One application includes sampling of bed groundwater associated with salmon spawning areas↗ in the Columbia River near the Hanford Site at Richland, Washington, USA.

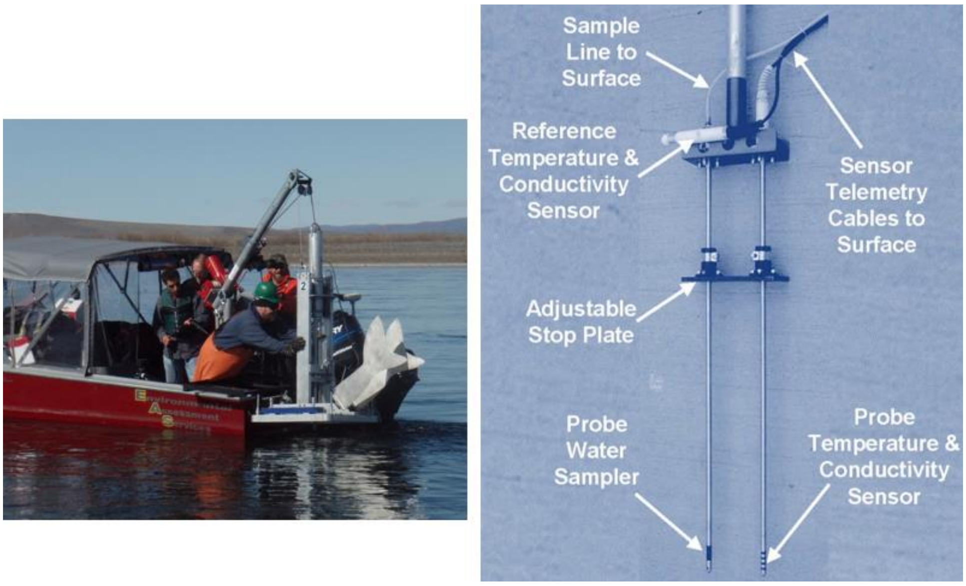


Figure 77 - A groundwater sampling and temperature probe (Trident↗) developed by the United States Navy is pushed or driven into sediment (Steenson and Naugel, 2017).

Temperature can also be used as a non-conservative qualitative or quantitative tracer of surface water or groundwater when temperature information is collected as a function of time (e.g., Stonestrom and Constantz, 2003; Anderson, 2005). This is accomplished by contrasting daily surface-water temperatures with changes in the groundwater (or hyporheic water) temperatures. Constantz (2008) described the qualitative approach to interpreting records of surface water and groundwater temperatures in the bed sediments of a surface-water body (Figure 78). A temperature monitor is placed in the surface-water feature and one or more sensors are installed in the sediments at varying distances below the bed (in the bed sediments or within a mini-piezometer) (Johnson et al., 2005; Constantz, 2008; Woessner, 2017). The paired temperature data are then analyzed to determine exchange locations and direction (e.g., Healy and Ronan, 1996; Hsieh et al., 2000; Stonestrom and Constantz, 2003; Stonestrom and Constantz, 2004; Anigoni et al., 2008; Swanson and Cardanas, 2011; Gordon et al., 2012; Constantz, 2008). Where a stream is gaining groundwater, the daily variations in the surface-water temperature (heating and cooling) are not reflected in the underlying groundwater system (Figure 78a). Instead groundwater temperatures remain relatively constant. Losing stream reaches exchange heat with the bed sediments resulting in bed temperature readings that are dampened, and off set (delayed) from the surface-water temperatures (Figure 78b and Figure 79). When surface water percolates through a vadose zone (losing stream) the groundwater temperature may not fully reflect surface-water temperature as additional heat is lost during percolation (Figure 78c). Ephemeral streams leaking water to the underlying groundwater may show a change in temperature below the bed as sediments become saturated with surface water (Figure 78d).

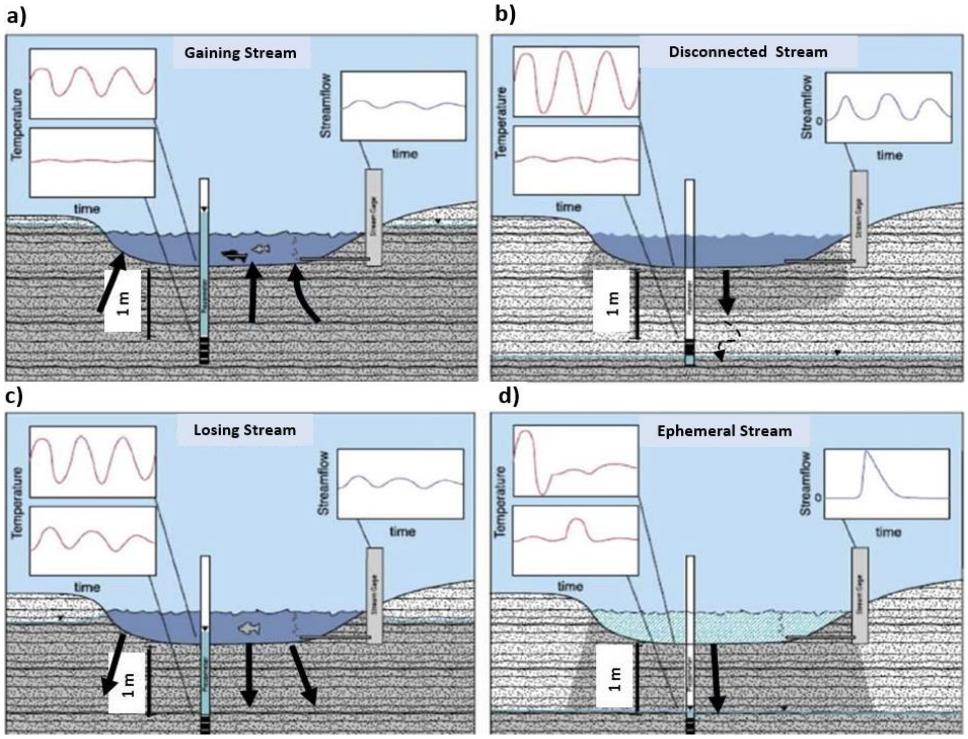


Figure 78 - Daily temperature trends for a surface-water feature and associated underlying groundwater (and hyporheic water). The upper graph on the left in each picture represents the surface-water temperature and the lower left graph the water temperature in the underlying sediments. The graph on the upper right is streamflow. Black arrows show the direction of groundwater flow and the dark gray shading represents saturated sediments. a) Gaining stream. b) Disconnected stream (losing). c) Losing stream. d) Ephemeral stream (modified from Constantz, 2008).

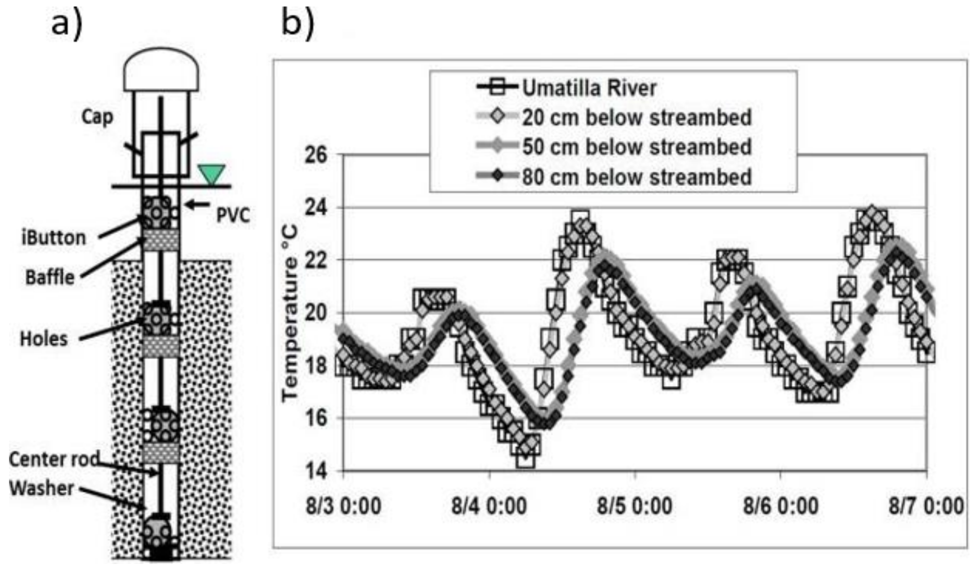


Figure 79 - Use of self-contained iButton temperature instruments to record temperatures in the Umatilla River, Oregon, USA, and at three depths in the gravel river bottom sediment. a) Instruments were placed in a steel mini-piezometer perforated at three locations (small circles) and separated with closed foam packers. iButtons were attached to a center rod for placement and removal. b) Results plotted as temperature vs. time. (8/3 0:00 denotes August 3 at midnight). Temperature peaks at the 50 and 80 cm depths are lagged behind the changes in surface-water temperatures suggesting downward seepage of surface water at this site (modified from Johnson et al., 2005).

Field temperature data can also be used to calculate exchange rates (flux) by applying heat transport theory and modeling. Lien and Ford (2014) provide a good summary of heat flow modeling methods.

Heat is transported through porous media by flowing groundwater, dispersion and conduction (due to Brownian motion) (Stonestrom and Constantz, 2003; Anderson, 2005). During transport, heat also interacts with the solid porous media (heating and cooling). Most commonly, heat transport is modeled as isotropic, homogeneous, steady-state conditions and, in the case of groundwater-surface water exchange, modeled as one-dimensional transport (vertical) (e.g., Constantz and Thomas, 1997; Constantz, 1998; Constantz et al., 2002ab; Goto et al., 2005; Hatch et al., 2006; Keery et al., 2007; Swanson and Cardenas, 2011; Gordon et al., 2012; Bhaskar et al., 2012; Constantz, 2008; Lien and Ford, 2014; Boano et al., 2104). Modeling information is provided in Box 7↴.

5.9 Stream Tracer Methods

Tracer techniques can be used to identify the pathways, timing and quantities of exchange between streams and groundwater and/or hyporheic zones. Tracers can also be used to determine discharge in small streams (e.g., Kilpatrick and Cobb, 1985). Commonly, tracers are introduced into streams to characterize hyporheic exchange, i.e., the transfer of surface water in and out of the immediately adjacent bed, bank and floodplain (Figure 29 and 30).

Tracer studies introduce a known mass of tracer, for example, a salt such as sodium chloride (NaCl), an ion like bromide (Br) and/or an organic dye such as Rhodimine-WT at an upstream location and then, over time, monitor the arrival of the tracer at a downstream location. The method is most manageable when stream discharges are small (<0.5 m^3/s). The larger the discharge the more tracer is needed to generate a measurable change in the tracer concentration in the stream. Results are typically shown as plots of concentration (or conductivity) versus time at various locations along the stream reach. The plots are referred to as breakthrough curves (Figure 80).

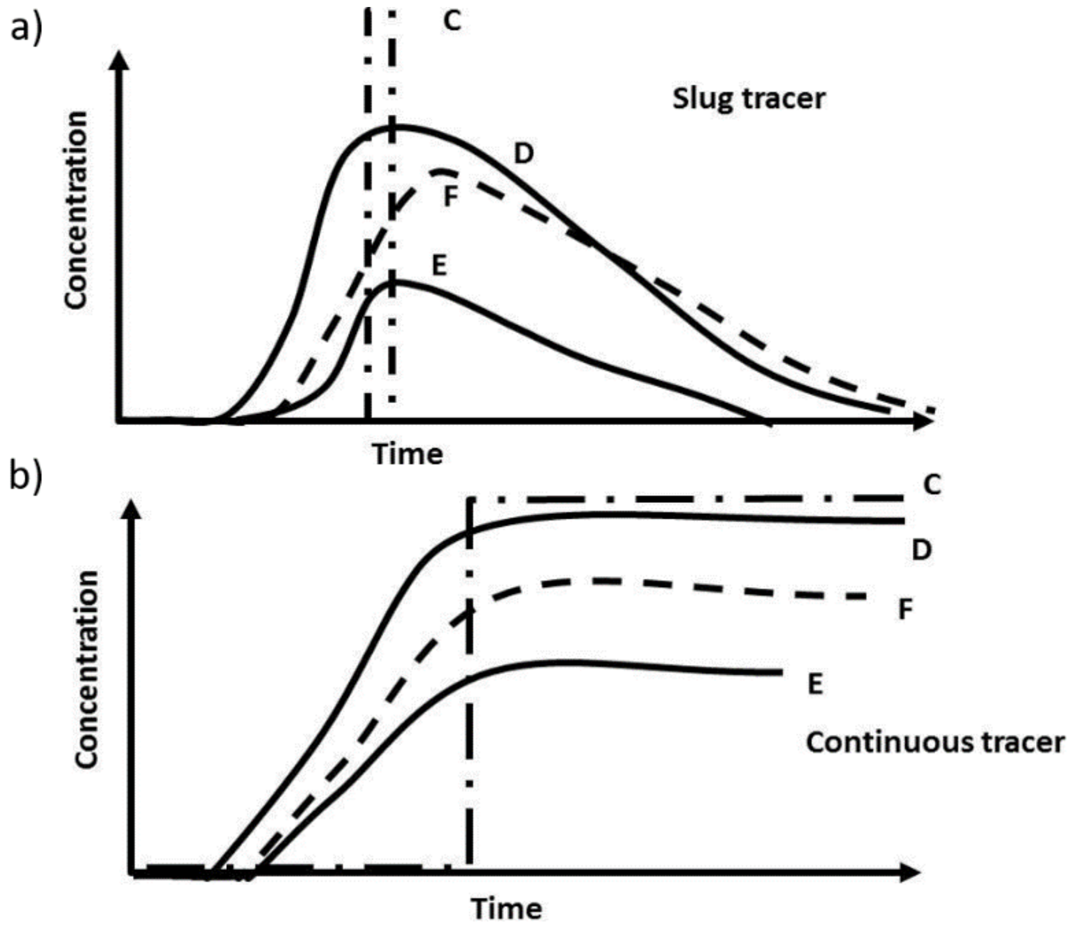


Figure 80 - Examples of tracer breakthrough curves observed at a downstream monitoring location. a) Results of introducing a tracer as an instantaneous source (slug input). b) Observations resulting from a continuous tracer injection. On both diagrams, the C curves (dot and dashed line) represent the observed response if no natural spreading and diffusion occurred between the upstream and downstream monitoring point (plug flow). The D curves represent observations where no channel or hyporheic storage is occurring and concentrations are only impacted by natural in-channel spreading and diffusion. The F curves show a redistribution of mass and the delay of the arrival of the peak concentration. This curve represents the influence of in-channel and hyporheic exchange delaying the tracer mass from reaching the downstream observation point. The E curves show a loss of tracer mass (concentration) suggesting temporary storage of tracer in the hyporheic zone (bed, banks, floodplain). The E curves could also represent the response of the tracer to dilution by inflowing groundwater (Woessner, 2018).

Tracer breakthrough curves collected at the downstream monitoring point are assessed using basic transport and storage theory expressed through analytical and numerical models as described in Box 8↴. Observed curves are compared to theoretical transport conditions (Figure 80). The tracer transport can be impacted by in-channel delays, the input of groundwater, and temporary storage in the channel or hyporheic system. Observed tracer concentrations also reflect the natural spreading and diffusion that occurs in the stream. The data are typically evaluated using one-dimensional analytical models to determine the degree of exchange between the stream and/or the hyporheic zone and groundwater. The field breakthrough curves are matched with parameters that fit results to expected conditions without loss of mass or delay of breakthrough. Cardenas (2015) describes several methods used for data analyses. The OTIS↗ (One-dimensional Transport with Inflow and Storage) software code is often used to assess breakthrough data sets and generate exchange values and components (Runkel, 1998). Data analyses are complicated by the possibility that more than one factor may influence the observed breakthrough curve.

In some studies, monitoring wells located in the floodplain and hyporheic zones and mini-piezometers in the stream channel are sampled during a channel tracer experiment to examine the locations, pathways and rates of circulation of stream water tracer into the hyporheic zone (e.g., Cardenas, 2015; Boana et al., 2014).

Designing a tracer test includes selecting the tracer and determining: the method of input; needed volume and concentration of tracer; and cost of the tracer analysis, and permits or permissions needed to introduce and monitor the tracer. Tracers and detection methods should be inexpensive; tracers should occur in low concentrations in background surface water and groundwater, be non-reactive (conservative), and non-toxic. When background concentrations are low, relatively inexpensive tracer tests include introducing: NaCl accompanied by monitoring of changes in specific conductance using a conductivity meter; Br measured with a specific ion probe; or organic dye tracers such as Rhodamine-WT monitored with a fluorimeter. It is useful to make an estimate of the amount of tracer that will be needed to create a measurable change in constituent concentration at the downstream site. Estimates of the stream concentration of the tracer can be computed using a mixing model that considers a slug or continuous-source input and the fully-mixed concentration once the dye is distributed in the streamflow. Some empirically derived equations have been suggested for slug input of a tracer (e.g., Kilpatrick, 1970). The best approach is to compute tracer inputs and then run a preliminary tracer test to see if concentrations and volumes need adjustment. Experience has shown that attempts at estimating tracer inputs are often poorly constrained.

Batchelor and Gu (2014) investigated how a conservative bromide tracer and a reactive nitrate tracer behaved in two small creeks and the corresponding hyporheic zones of a stream in a North Carolina, USA. At Winkler Creek (mean annual discharge 0.2 m^3/s), a continuous injection of tracer designed to raise the concentrations at the downstream observation point by 1 to 2 mg/l of bromide and 2 to 3 mg/l of nitrate above background concentrations was initiated. The tracers were injected at 30 L/h, and breakthrough concentrations at a location 50 m downstream were reported (Figure 81). The OTIS code (described in Box 7) was used to examine storage in the stream system. Batchelor and Gu observed only small changes in the breakthrough curve and estimated that 1.05% of the streamflow over a 200 m reach of the creek entered transient storage in the channel/hyporheic zone based on analysis of the breakthrough curves observed at the downstream monitoring points (Figure 81).

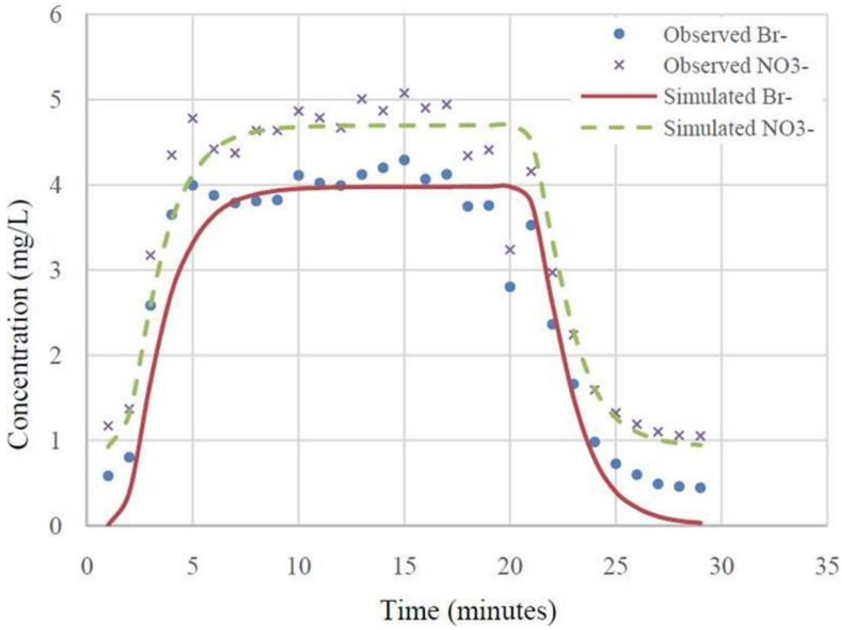


Figure 81 - Breakthrough curves for Br and NO_3 tracers in Winkler Creek, North Carolina, USA, at 50 m downstream from a continuous tracer injection point. Simulated curves represent the results of OTIS modeling (Batchelor and Gu, 2014).

5.10 Brief Summary of Geochemical Methods

Streams, lakes, and wetlands reflect the chemical composition of the sources of water exchanging with the landscapes and hydrologic systems connecting them. For example, the chemical composition of an effluent lake will reflect the precipitation/runoff and evaporative concentration chemistry as well as chemical inputs from stream and groundwater discharges. A discharge wetland may be dominated by the chemistry and evapotranspiration of groundwater input. The geochemical principles used to identify groundwater-surface water exchange such as characterizing source waters, tracing changes in groundwater quality along flow paths, and interpreting flux and velocities, are described in a wide range of hydrogeology, hydrology and geochemical texts (e.g., Freeze and Cherry, 1979; Fetter, 2001; Fritts, 2012; Drever ,1997; Stumm, 1996; Brezonik and Arnold, 2011; Cook and Herczeg, 2000). These texts and other resources also address proper geochemical sampling and analytical methods. Detailed discussion of these concepts and methods are beyond the scope of this book.

Common geochemical techniques applied to deciphering groundwater-surface water exchange include mass balance geochemical mixing modeling, application of stable and radioactive isotope data to differentiate water histories and sources, and use of natural and introduced environmental tracers.

When groundwater and surface water chemistries contrast, components of groundwater-surface water exchange may be detected by evaluating waters for variations in concentrations of ionic constituents, stable and unstable isotopes, organic compounds, dissolved oxygen, pH, temperature, and total dissolved solids (TDS) or specific conductance. Healy and others (2007) list examples of constituents that are useful in water-budget/mass-balance models that often occur in contrasting concentrations in surface water and groundwater (Table 2).

Table 2 - Examples of Tracers Used in Chemical Water Budget Studies (after Healy et al., 2007).

Use	**Natural** occurring in environment	**Historical** added to environment during past human activity	**Applied** introduced for testing	Example Study
Groundwater age time since recharge water isolated from atmosphere	^{35}S, ^{14}C, $^{3}H/^{3}He$, ^{39}Ar, ^{36}Cl, ^{32}Si	^{3}H, ^{36}Cl, ^{85}K, chlorofluorocarbons, herbicides, caffeine, pharmaceuticals		Plummer and others (2001)
Recharge Temperature	N^2/Ar solubility			Plummer (1993)
Tracing groundwater flow paths	^{18}O, ^{2}H, ^{13}C, ^{87}Sr	Chlorofluorocarbons, herbicides, caffeine, pharmaceuticals	Cl, Br, dyes	Renken and others (2005)
Exchange groundwater-surface water	^{18}O, ^{2}H, ^{3}H, ^{14}C, ^{222}Rn		Cl, Br, dyes	Katz and others (1997)
Distance and Travel Time of surface water			Cl, Br, dyes	Kimball and others (2004)

A mass balance or chemical mixing model of exchange between a surface-water system and the associated groundwater system can be used to identify exchange components (Figure 82). For example, a mixing model for a lake under steady-state conditions that is solved for the rate of groundwater inflow would be formulated as shown in Equation 7.

$$GW_{in} = \frac{SW_{out}C_{SWout} - SW_{in}C_{SWin} - PPT_{in}C_{PPTin} + E_{out}C_{Eout} + GW_{out}C_{GWout}}{C_{GWin}} \qquad (7)$$

where:

GW_{in} = groundwater discharge to the lake (L^3/T)
GW_{out} = flow of lake water into the adjacent groundwater (L^3/T)
SW_{in} = flow of surface water into the lake (L^3/T)
SW_{out} = flow of surface water out of the lake (L^3/T)
PPT_{in} = precipitation falling onto the lake (L^3/T)
E_{out} = direct evaporation from the lake (L^3/T)
C_{XXxxx} = Concentrations of a selected constituent in component XXxxx, such as C_{SWout} (M/L^3)

Mixing models can be formulated using a single species or component, or ratios of constituents. Healy and others (2007) and Winter (1981) caution that if some components of mass balances are poorly defined, large errors are likely. An example of a mixing model application is presented in Box 9⤵.

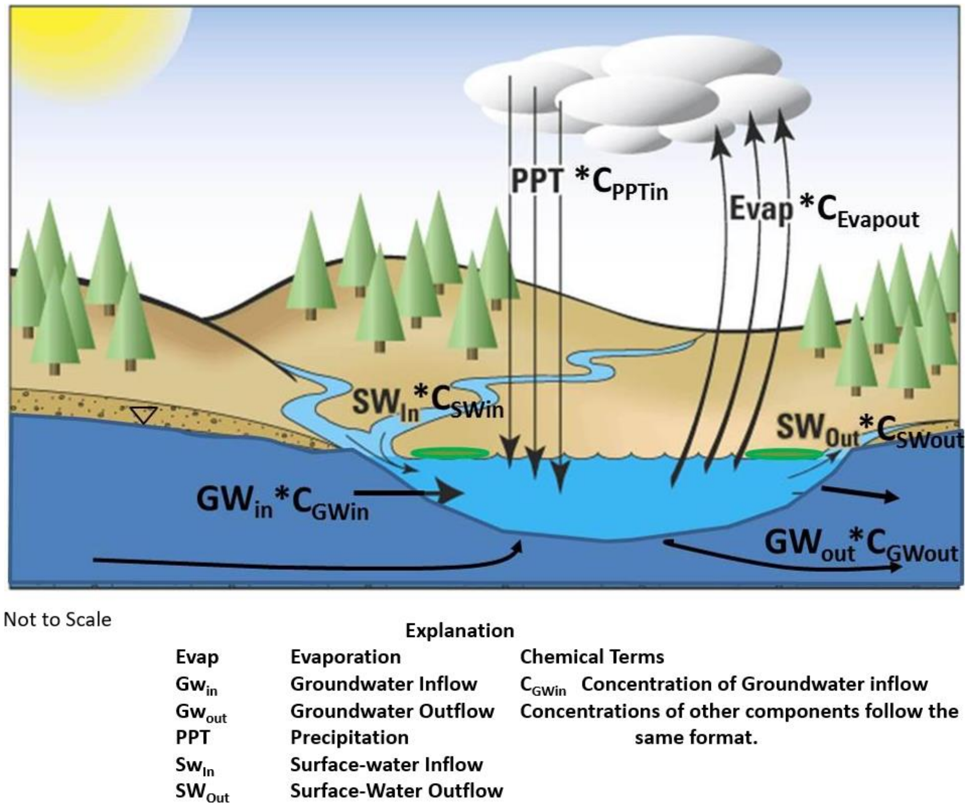


Figure 82 - Water and geochemical balance for a flow-through lake. C_{PPTin} represents the concentration of a geochemical constituent in the precipitation used in the balance. Other C terms represent the same constituent in each part of the balance. The balance is conceptualized under steady-state conditions (modified from Robertson et al., 2003).

In some settings, changes in water chemistry along a groundwater flow path can be used to estimate flow rates. For example, in situations where surface water freely infiltrates the hyporheic or groundwater system the concentration of radon (^{222}Rn) buildup along the flow path is used to establish infiltration rates and sources of water. Most surface-water features have low concentrations of radon as they are open to the atmosphere. Once this water infiltrates, natural radon produced in the sediments is incorporated in the water and concentrations increase until equilibrium is established (e.g., Baskaran et al., 2009; Sacks et al., 1998). Hoehm and Cirpka (2006) describe the use of radon to assess residence times of surface-water exchange in floodplains of the Southern Alps.

Another useful approach is to apply mixing models to constituent concentrations along shorelines and within stream systems to examine sources and contributions of water. Smerdon and others (2012) discuss the use of multiple isotopes to identify water sources and quantify base flow along a 60 km section of the tropical Daly River, Australia. Isotopes of radon (^{222}Rn), sulfur (SF_6), helium (^4He), as well as chlorofluorocarbons (CFCs) were sampled to characterize spring discharges along the channel, the main channel chemistry, and the adjacent groundwater chemistry. Regional groundwater contained concentrations of ^{4}He and very low concentrations of SF_6 and CFCs suggesting long residence times on the order of 10,000 years. Base flow generated by local springs was dominated by SF_6 and CFCs suggesting more localized groundwater exchange. Based on the concentration of constituents in the base flow, they concluded that over 45% of the base flow originated from regional groundwater flow.

Field methods have also been developed to identify exchange locations using geochemistry of pore water in bed sediments (Lee, 1985; Vanek and Lee, 1991; Lee et al., 1993; Harvey et al., 1997; Cey et al.,1998; Kennedy, 2017). The approach uses a conductivity probe to map changes in electrical conductivity (EC) along the bottom sediments of a surface-water feature (Figure 83). This method is used to obtain multiple bed conductivity transects and map contrasts between electrical conductivity of pore water in bed sediments and adjacent groundwater (e.g., Harvey et al., 1997).

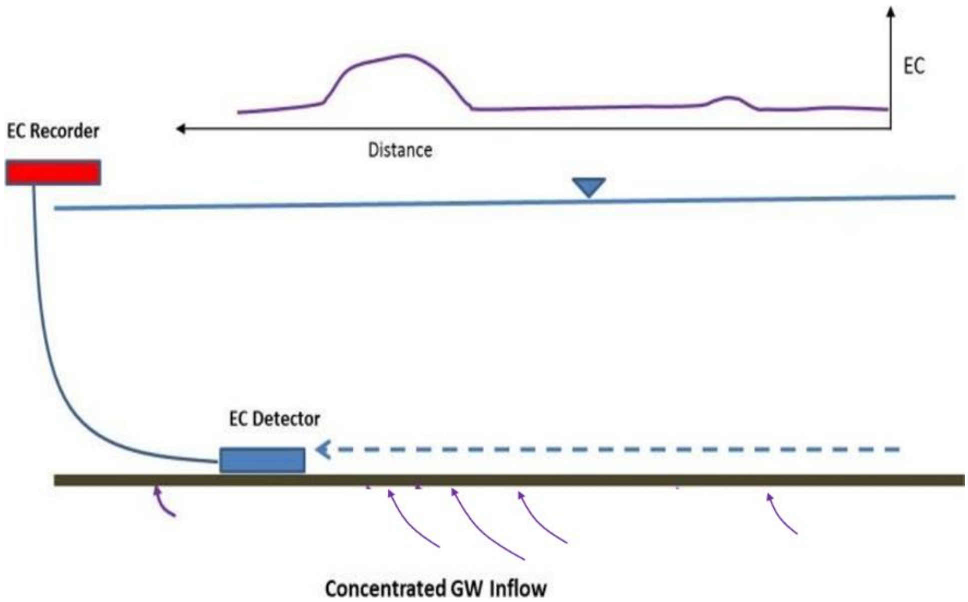


Figure 83 - The weighted electrical conductivity probe (EC recorder) is dragged along the bottom sediments using a boat or raft to carry the recorder. The probe measures the conductivity of the bed and pore water. In this example, groundwater with a higher electrical conductivity (EC) is exiting the bed at concentrated locations (purple arrows). The plot shows the instrument record used to map locations where groundwater discharges (Woessner, 2020).

When exchange rates in fine-grained bed sediments are low, passive diffusive-membrane geochemical samplers can be used to collect pore water sediment samples. These samplers allow pore water to diffuse into one or more collection chambers (Figure 84).

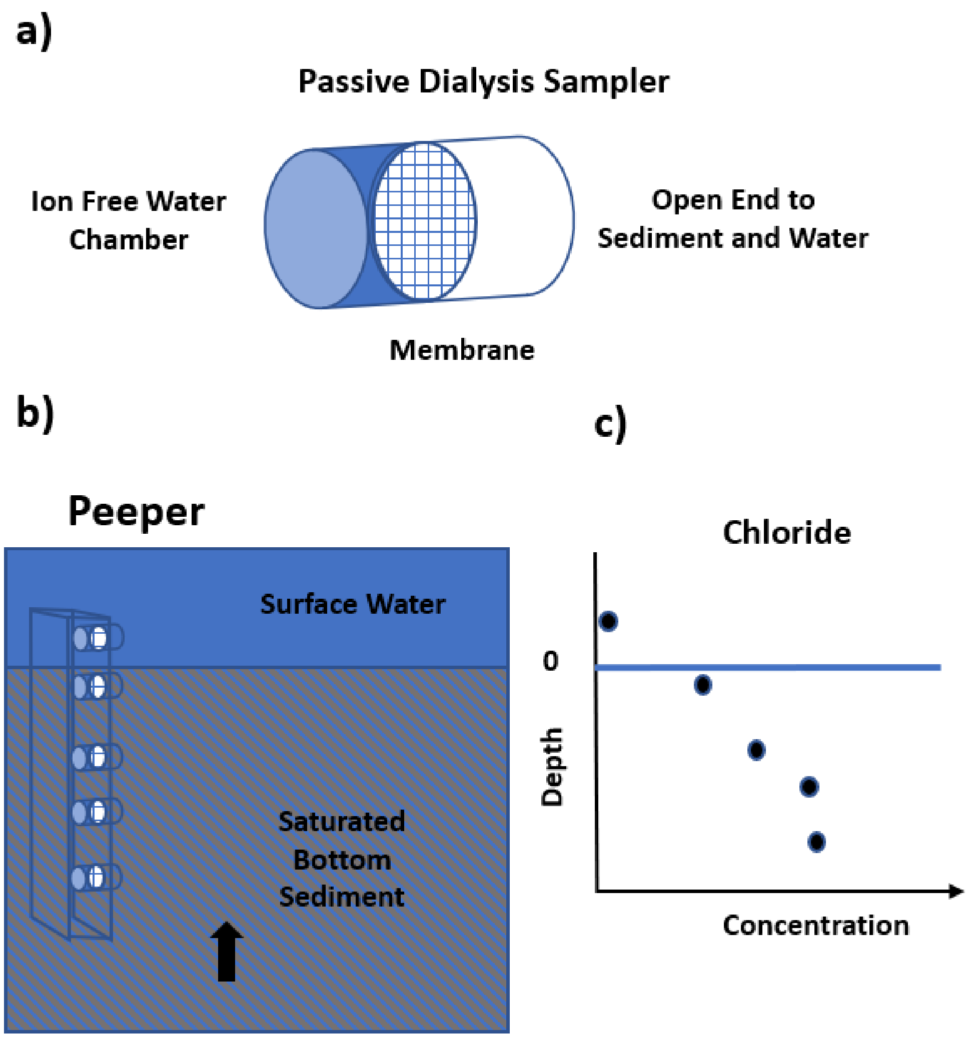


Figure 84 - Passive dialysis sampler. a) Chambers are filled with ultrapure water (blue) and covered with a dialysis membrane. One end of each chamber is open and in contact with the sediment and water. b) The Peeper sampler installed in bottom sediment with one chamber sampling the surface water. The sampler remains in place until pore waters have fully exchanged and the water behind the membrane is in equilibrium with the sediment/water. The black arrow suggests the possible groundwater flow direction. The sampler is extracted, and captured water analyzed. c) An example of a data set collected for chloride concentrations. Depth (0 = bottom sediment interface). Modeling could be applied to determine if diffusive or advective groundwater flow is occurring in the sediments (Woessner, 2020).

Pore-water profiles are often used to examine geochemical processes at the sediment-water interface. However, they can also be used to examine the slow transport of conservative constituents between surface water and groundwater. These data sets are examined to determine locations and exchange rates using transport models (e.g., Freeze and Cherry, 1979; Zheng and Bennett, 2002). An example of a geochemical model used to explore transport though a lakebed is provided by Cornett and others (1989).

Several types of passive samples have been developed for sampling both general ionic chemistry and to target specific inorganic and organic contaminants (e.g., Burgess et al., 2016). The United States Environmental Protection Agency has published an informative manual on the use of passive sediment pore water samplers (Burgess et al., 2016). Samples use low-density polyethylene (LDPE), polyozymethylene (POM), polydimethylsilozane samplers (PDMS) for hydrophobic organic chemicals, and diffusive gradient thin films (DGT) for selective metal evaluations. They provide a table of material used in samplers and suppliers. Care in selecting both the composition and methods of installation are required to obtain representative data.

The chemical composition of pore waters can also be sampled by extracting groundwater from mini-piezometers and pore water from sediment cores. A method to extract and analyze pore water from sediment cores collected in Lake Baldegg, Swizerland, used MicroRhizon samplers and capillary electrophoresis methods (Torres et al., 2013).

6 Concluding Remarks

The exchange of surface water and groundwater is driven by the three-dimensional distribution of hydraulic gradients, and the magnitude of the anisotropy and heterogeneity of the associated earth materials surrounding the surface-water feature. Rivers, lakes and wetlands can gain and lose groundwater, and become disconnected from associated groundwater systems. River systems also exchange water by hyporheic flow where stream water moves into surrounding earth materials and returns to the stream. Exchanges can be identified at the landscape and individual feature scale using a variety of field and laboratory methods. These include water and geochemical mass balances; networks of monitoring wells, mini-piezometers, stage measurement, seepage runs, seepage meters, temperature sensors, and natural, environmental and introduced tracers.

In the broad sense, physical and geochemical groundwater exchanges between rivers, lakes and wetlands were recognized in the early twentieth century. Conceptual models of exchange processes were refined beginning in the last third of the century and intensive investigations of multi-scale exchanges with surface water systems were initiated. More detailed investigations of exchange processes and field, laboratory and modeling methodology development continue in the twenty-first century. The exchange of both natural and contaminated water with connected surface water and groundwater systems has been the focus of a number of investigations. Ideally, specific site investigations and the general literature will provide the qualitative and quantitative data sets needed to manage undisturbed (natural) exchange settings, and to design and remediate sites needing restoration and re-naturalization. However, the literature as of 2020 only partially addresses broad exchange questions such as: How much exchange is needed in a river reach, wetland or along a lake shore, to sustain desired ecological conditions? What are the exchange rates, locations and timing that support the flow, geomorphic character, and water quality of specific surface-water systems? Over-arching exchange concepts still need to be addressed. For example, guidelines are not available for defining the volume of exchange needed to support a trout population in a mountain stream or the type of vegetation required to sustain a land-based constructed wetland, nor are guidelines available to define the fluxes need at a lakeshore to support an endangered species. Development of such guidelines requires extensive study of a wide variety of river, lake and wetland exchange settings under natural and impacted conditions. The ecological literature that focuses on exchange needs to be better integrated with physical and geochemical hydrogeological research. The end goal should be to develop a more complete understanding of how groundwater exchange with rivers, lakes and wetlands supports the surface water system and how surface water exchange supports the associated groundwater systems.

7 Exercises

1) The conceptual models presented in this book keep conditions simple by using a constant value for hydraulic conductivity, cross sections aligned with groundwater flow lines and cross sections plotted without vertical exaggeration. Examine Figure P1 below. The cross section in Figure Exercise 1a, is constructed parallel to flow, as indicated by red line (A-R-B). Under the conditions illustrated in Figure Exercise 1, think about the consequences of using head data to interpret the flow field from a cross section constructed at right angles to the stream as indicated by the line C-R-D. To make the comparison use A-R and C-R. Explain why the cross section along line A-R correctly represents horizontal and vertical groundwater flow, but head and flow data in a cross section along C-R does not appropriately represent flow conditions.

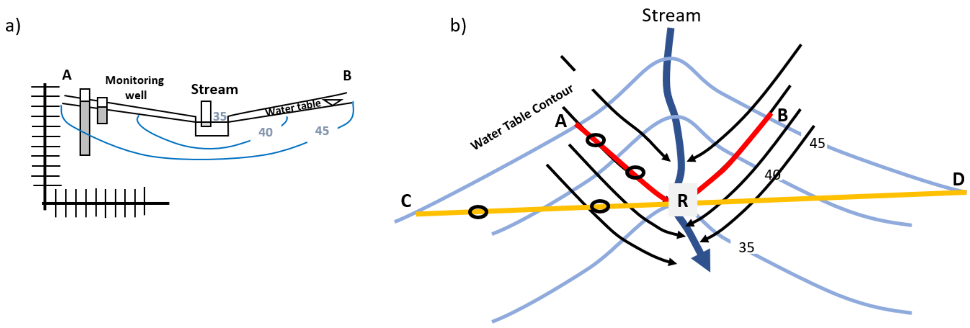


Figure Exercise 1 - Cross section and map view of a stream. Hydrogeologic conditions are isotropic and homogeneous and the groundwater conditions are effluent. The black ovals represent the location of monitoring wells on the cross sections. R is located where the cross sections intersect the stream. a) The cross section has no vertical exaggeration. The section is constructed along groundwater flowlines (red lines A-R and B-R) shown in (b). The left and right boundaries of the cross section coincide with the head contour of 45 (after Healy et al., 2007). b) Map view of flow field for the effluent stream. Water table contours are in blue and groundwater flowlines in black. The red line, A-R-B, shows the location of the cross section in (a). The orange line, C-R-D, is a cross section constructed at right angles to the river (Woessner, 2020).

Click for solution to exercise 1

2) Examine the scaled cross section of Figure Exercise 2 showing two surface-water bodies (solid blue: rivers, lakes or wetlands) in an isotropic and homogeneous setting. Equipotential lines (dashed) were constructed from a monitoring well network containing nests of wells (vertical black lines) open only at the bottom. The water table is indicated by a solid blue line. Equipotential lines represent intervals of 5 (any units can be used).

a) Construct flowlines.

b) Label local, intermediate and regional flow systems if they are present.

c) Identify where stagnation points form and circle the locations on the cross section if they are present.

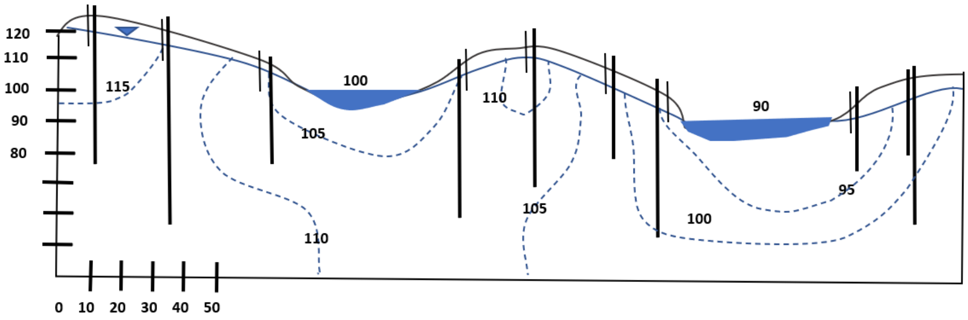


Figure Exercise 2 - Scaled cross section of a hydrologically isotropic and homogeneous system with two water bodies. Elevations are shown on the vertical axis and the stages of the water bodies are relative values. Black vertical lines represent monitoring well locations where head data were collected. Wells are open only at the bottom. Dashed lines are equipotential lines in 5-unit intervals (Woessner, 2020).

Click for solution to exercise 2↴

3) An exchange study of a pond was conducted by installing three seepage meters (A, B, C) and three adjacent mini-piezometers (small black open circles) as shown in Figure Exercise 3. Using the data provided compute the following:

a) The vertical hydraulic conductivity of the bed sediments at location A.

b) The VHG for B.

c) The seepage flux for C.

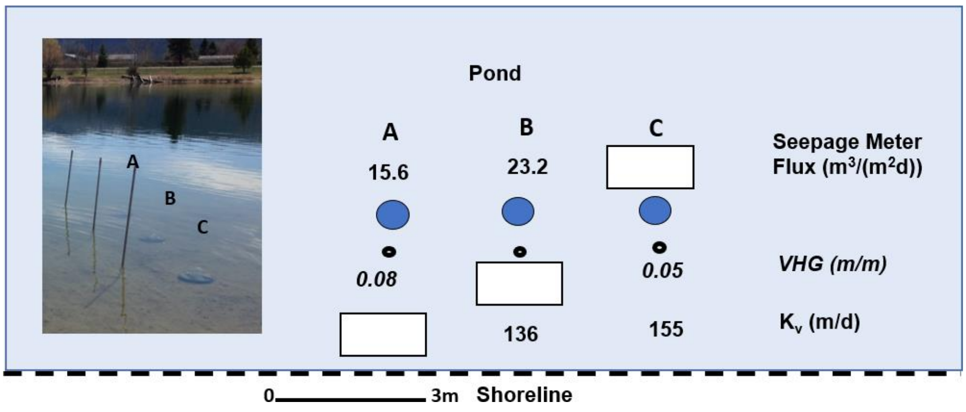


Figure Exercise 3 - An exchange study at a pond constructed in a sand and gravel aquifer. Three seepage meters, A, B and C (blue drums in the picture and blue circles in the map view), were installed about three meters from the shoreline. Adjacent to the meters, mini-piezometers (vertical tubes in the picture and black open circles in the map) were driven about 15 cm into the sand and gravel sediments of the bed. Seepage rates, VHG, and estimates of the bed vertical hydraulic conductivity were obtained for each of the three seepage meter locations (Woessner, 2020).

Click for solution to exercise 3↴

4) A water budget for a shallow subtropical lake found surface-water drainage canal discharges (34%), direct precipitation (24%) and groundwater discharge (14%) dominated inflows. Water left the lake by sheet flow into a swamp (65%), evaporation (34%), and groundwater outflow (1%). Nitrogen loading to the lake was also a concern and the lake was classified as dystropic, (chronic hypoxia and high concentrations of unionized ammonia). Groundwater was computed to deliver 48% of the load. Restoration efforts were initiated in 1996 and included removal of organic sediment, dredging and planting of native vegetation. Additional nitrogen reduction restoration is needed. The lake is in a depression created by limestone dissolution. A seepage study of the lake was conducted using twenty 208-liter seepage meters and mini-piezometers installed next to each meter (Figure Exercise 4). Meters were installed and seepage rates measured every 14 days from October through mid-May. Meters 1-14 were located in the littoral zone and 15-20 were in open water. Meters equilibrated for three months before study operation. Bags were prefilled with 1000 ml of deionized water for each seepage event. Mini-piezometers were placed in holes excavated into the bed sediment and bedrock and backfilled with sand and bentonite. Water quality samples were taken from the mini-piezometers. Mean seepage rates and total nitrogen concentrations were computed for each of the twenty sites (Table Exercise 4).

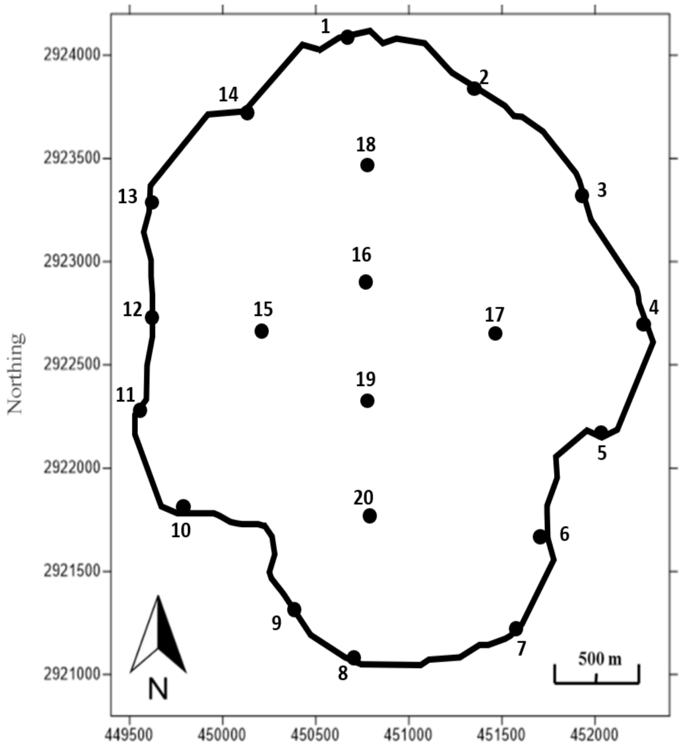


Figure Exercise 4 - Location map of seepage meter and mini-piezometer installations (black dots and numbers) in a lake (modified from Lucius, 2016).

Table Exercise 4 - Mean Seepage Rates and Mean Total Nitrogen Concentrations of Discharging Groundwater in a Shallow Lake. Positive seepage values represent groundwater inflow. Site numbers represent both seepage meters and the adjacent mini-piezometer (Lucius, 2016).

Site	Northing	Easting	Mean Seepage (L/m^2/d)	Mean Total Nitrogen (mg/l)
1	450593.39	2924113.61	1.44	11.84
2	451291.49	2923858.38	0.26	17.7
3	451916.45	2923324.12	1.74	9.06
4	452238.73	2922714.82	1.16	4.38
5	451989.45	2922144.28	0.15	20.32
6	451695.94	2921673.69	0.2	17.1
7	451520.61	2921227.23	0.89	4.55
8	450647.05	2921081.35	7.04	5.45
9	450321.19	2921294.35	0.59	5.64
10	449740.75	2921801.64	0.14	13.02
11	449481.89	2922268.64	1.85	4.16
12	449533.30	2922725.53	-0.08	6.01
13	449535.09	2923278.14	0.25	4.74
14	450033.85	2923738.93	0.83	6.49
15	450155.54	2922664.82	0.89	52.28
16	450716.41	2922891.76	0.67	28.53
17	451418.80	2922658.34	0.87	24.27
18	450717.27	2923464.25	1.02	11.24
19	450717.90	2922308.40	3.11	23.78
20	450725.32	2921747.50	2.11	6.46

a) Plot the mean seepage data on the location map and contour it using an interval of 0.5 L/(m^{2}d). Where is the seepage concentrated? Is water seeping from the lake to the groundwater? Would you classify the lake as effluent, influent, flow-through, or mixed? Support your answer.

b) Plot the mean total nitrogen data on a second copy of the location map. Use a contour interval of 2 mg/l. Where are the concentrations of total nitrogen the highest? Do they correspond with the highest groundwater seepage rates?

c) Compute the total mean total nitrogen loading at each location and contour the data. Use a contour interval of 20 mg/(m^{2}d). Discuss how mean seepage meter and mini-piezometer mean water quality data can be used to target nitrogen-reduction remediation efforts. Think about what other hydrogeologic information is needed to accomplish remediation goals (Think in terms of the "big picture").

Click for solution to exercise 4↴

5) Under relatively simple conditions, the temperature of stream water can be used to estimate the net gain in groundwater for an effluent reach of stream. Figure 71 presents the concept. If the streamflow at the upstream site Q1 is 2.0 m^3/s and the mixed stream water temperature is 12°C, the shallow groundwater system temperature is 8.2°C, and a stream temperature measurement at the downstream site, Q2, is 10.4°C, then:
   a) Assuming that over this 2 km river reach the other components of the heat budget are small, what is the net quantity of groundwater that discharges to this river reach?
   b) How could individual temperature monitors installed in the stream and stream bed be used to verify the stream is gaining in this reach?

   Click for solution to exercise 5↴

6) Heat tracing in the bed of surface-water bodies is an inexpensive and valuable tool for tracing the direction of exchanges, estimating hydraulic properties of bed material, and estimating exchange rates. USGS (2003) Circular 1260 is an excellent resource that explains the relevant principles, methods and modeling approaches. Review the document: https://pubs.usgs.gov/circ/2003/circ1260/pdf/Circ1260.pdf↗

   a) Of the seven case studies presented, choose one and summarize the goals, methods and results. Include two key figures supporting your summary.

   b) The publication also provides information on modeling of heat flow in Appendix B: “Modeling heat as a tracer to estimate streambed seepage and hydraulic conductivity” by Richard G. Niswonger and David E. Prudic. It is useful to study the appendix prior to designing field instrumentation so that field efforts will generate the required data for modeling. Read the appendix and list the parameters and boundary conditions needed when simulating heat transport from a river into the riverbed.

   Click for solution to exercise 6↴

8 References

Anderson, M.P., 2005, Heat as a ground water tracer. Ground Water, volume 43, number 6, pages 951–968.

Anderson, M.P., and J.A. Munter, 1981, Seasonal reversals of groundwater flow around lakes and the relevance to stagnation points and lake budgets. Water Resources Research, volume 17, number 4, pages 1139-1150.

Anderson, M.P., W.W. Woessner, and R.J. Hunt., 2015, Applied Groundwater Modeling: Simulation of Flow and Advective Transport. Academic Press-Elsevier, London, 564 pages.

Arrigoni, A.S., G.C. Poole, L.A.K. Mertes, S.J. O'Daniel, W.W. Woessner, and S.A. Thomas, 2008, Buffered, lagged, or cooled? Disentangling hyporheic influences on temperature cycles in stream channels. Water Resources Research, volume 44, W09418, doi:10.1029/2007WR006480↗.

Bartolino, J., and R. Niswonger, 1999, Temperature profiles of the aquifer system underlying the Rio Grande, New Mexico. Proceedings of Third Annual Middle Rio Grande Basin Workshop, February 22–23, 1999. United States Geological Survey Open-File Report 99-203, pages 66–67.

Baskaran, S., T. Ransley, R.S. Brodie, and P. Baker, 2009, Investigating groundwater-river interactions using environmental tracers. Australian Journal of Earth Sciences, volume 56, number 1, pages 13-19, doi:10.1080/08120090802541887↗.

Batchelor, C., and C. Gu, 2014, Hyporheic exchange and nutrient uptake in a forested and urban stream in the Southern Appalachian's. Environmental and Natural Resources Research, volume 4, number 3, pages 56-66.

Baxter, C., F.R. Hauer, and W.W. Woessner, 2003, Measuring groundwater-stream water exchange: New techniques for installing minipiezometers and estimating hydraulic conductivity. Transactions of the American Fisheries Society, volume 132, pages 493-502.

Bean, J.R., A.C. Wilcox, W.W. Woessner, and C.C. Muhlfeld, 2013, Multiscale hydrogeomorphic influences on bull trout (Salvelinus confluentus) spawning habitat. Canadian Journal of Fisheries and Aquatic Sciences, volume 72, pages 514-526, http://dx.doi.org/10.1139/cjfas-2013-0534↗.

Belanger, T.V., and M.T. Montgomery, 1992, Seepage meter errors. Limnology and Oceanography, volume 37, number 8, pages 1787–1795.

Bencala, K.E., and R.A. Walters, 1983, Simulation of solute transport in a mountain pool and riffle stream. Water Resources Research, volume 19, number 3, pages 718–724.

Bencala, K. E., M.N. Gooseff, and B.A. Kimball, 2011, Rethinking hyporheic flow and transient storage to advance understanding of stream catchment connections. Water Resources Research, volume 47, WH00H03, doi:10.1029/2010WR010066↗.

Bhaskar, A.S., J.W. Harvey, and E.J. Henry, 2012, Resolving hyporheic and groundwater components of streambed water flux using heat as a tracer. Water Resource Research, volume 48, W08524, doi:10.1029/2011WR011784↗.

Bisson, P.A., J.M. Buffington, and D.R. Montgomery, 2006, Valley Segments, Stream Reaches and Channel Units, *in* Methods. *in* Stream Ecology, second edition, F. R. Hauer and G. A. Lamberti, editors. Elsevier, pages 23-49.

Blanchfield, P.J., and M.S. Ridgeway, 1996, Use of seepage meters to measure ground water flow at brook trout redds. Transactions of the American Fishery Society, volume 125, pages 813–818.

Boano, F., A.I. Packman, A. Cortis, R. Revelli, and L. Ridolfi, 2007, A continuous time random walk approach to the stream transport of solutes. Water Resources Research, volume 33, W10425, doi:10.1029/2007WR006062↗.

Boano, F., J.W. Harvey, A. Marion, A.I. Packman, R. Revelli, L. Ridolfi, and A. Worman, 2014, Hyporheic flow and transport processes. Mechanisms, models, and biogeochemical implications, Reviews of Geophysics, volume 52, pages 603–679, doi:10.1002/2012RG000417↗.

Born, S.M., S.A. Smith, and D.A. Stephenson, 1974, The Hydrologic Regine of Glacial Terrain Lakes, with Management and Planning Applications. Upper Great Lakes Regional Commission, United States of America, 73 pages.

Boulton, A.J., 2007, Hyporheic rehabilitation in rivers: restoring vertical connectivity. Freshwater Biology, volume 52, pages 632-650.

Boyle, D.R., 1994, Design of a seepage meter for measuring ground water fluxes in the nonlittoral zone of lakes—evaluation in a boreal forest lake. Limnology and Oceanography, volume 39, pages 670–681.

Brezonik, P.L., and W.A. Arnold, 2011, Water Chemistry: An Introduction to the Chemistry of Natural and Engineered Aquatic Systems. Oxford University Press, Oxford, United Kingdom, 782 pages.

Brodie, R., B. Sundaram, R. Tottenham. S. Hostetler, and T. Ransley, 2007, An overview of tools for assessing groundwater-surface water connectivity. Bureau of Rural Sciences, Canberra, Australia.

Bryan, K., 1919, Classification of springs. Journal of Geology, volume 27, pages 522-561, https://digitalcommons.unl.edu/usgsstaffpub/493/↗.

Buffington, J M., D.R. Montgomery, and H.M. Greenberg, 2004, Basin-scale availability of salmonid spawning gravel as influenced by channel type and hydraulic roughness in mountain catchments. Canadian Journal of Fisheries and Aquatic Sciences, volume 61, pages 2085– 2096, doi:10.1139/F04-141↗.

Burgess, R.M., S.B. Kane Driscoll, G. Allen Burton, Philip M. Gschwend, Upal Ghish, Danny Reible, Sungwoo Ahn, and Tim Thompson, 2016, Laboratory, Field, and Analytical Procedures for Using Passive Sampling in the Evaluating of Contaminated Sediments: User's Manual, United States Environmental Protection Agency, EPA/600/XX-15/071, https://cfpub.epa.gov/si/si_public_file_download.cfm?p_download_id=528886&Lab=NHEERL↗.

Buss, S., Z. Cal, B. Cardenas, J. Fieckenstein , D. Hannah, K. Heppell, P. Hulme, T. Ibrahim, D. Kaeser, S. Krause, D. Lawier, D. Lerner, J. Mant, I. Malcolm, G. Old, G. Parkin, R. Pickup, G. Pinay, J. Porter, G. Rhodes, A. Richie, J. Riley, A. Robertson, D. Sear, B. Shields, J. Smith, J. Tellam, and P. Wood, 2009, The Hyporheic Handbook: A handbook on the groundwater-surface water interface and hyporheic zone for environment managers. Integrated catchment science programme, Science report: SC050070, Environment Agency, Bristol, United Kingdom, 264 pages.

Butler, J. J. Jr., 1997, The Design, Performance and Analysis of Slug Test. Lewis Publishers, Chemical Rubber Company (CRC) Press, Boca Raton, Florida, United States of America 253 pages.

Butler, J. J., and J.M. Healey, 1998, Relationship between pumping-test and slug-test parameters: scale effect or artifact. Ground Water, volume 36, pages 305-313.

Butler, J.J., E.J. Garnett, and J.M. Healey, 2003, Analysis of slug tests in formations of high hydraulic conductivity. Ground Water, volume 41, pages 620-630.

Cable, J.E., W.C. Burnett, J.P. Chanton, D.R. Corbett, and P.H. Cable, 1997, Field evaluation of seepage meters in the coastal marine environment. Estuarine and Coastal Shelf Science, volume 45, pages 367–375.

Cardenas, M.B., 2008a, The effect of river bend morphology on flow and timescales of surface water-groundwater exchange across pointbars. Journal of Hydrology, volume 362, number 1-2, pages 134–141.

Cardenas, M.B., 2008b, Surface water-groundwater interface geomorphology leads to scaling of residence times. Geophysical Research Letters, volume 35, L08402, doi:10.1029/2008GL033753↗.

Cardenas, M.B., 2009, A model for lateral hyporheic flow based on valley slope and channel sinuosity. Water Resources Research, volume 45, W01501, doi:10.1029/2008WR007442↗.

Cardenas, M.B., 2015, Hyporheic zone hydrologic science: A historical account of its emergence and a prospectus. Water Resources Research, volume 51, pages 3601-3616, doi:10.1002/2015WR017028↗.

Cardenas, M.B., and J.L. Wilson, 2006, The influence of ambient groundwater discharge on exchange zones induced by current-bedform interactions. Journal of Hydrology, volume 331, pages 103-109.

Cardenas, M B., and J.L. Wilson, 2007a, Dunes, turbulent eddies, and interfacial exchange with permeable sediments. Water Resources Research, volume 43, W08412, doi:10.1029/2006WR005787↗.

Cardenas, M.B., and J.L. Wilson, 2007b, Exchange across a sediment-water interface with ambient groundwater discharge. Journal of Hydrology, volume 346, number 3–4, pages 69–80, doi:10.1016/j.jhydrol.2007.08.019↗.

Cardenas, M B., and J.L. Wilson, 2007c, Effects of current-bed form induced fluid flow on the thermal regime of sediments. Water Resources Research, volume 43, W08431, doi:10.1029/2006WR005343↗.

Carling, P.A., P. Taylor, B.G. Hankin, and I.A. Benson, 1999, Fluid exchange and oxygen flux through salmonid redds. Research and Development Technical Report, W225, UK Environment Agency, Bristol, United Kingdom, 82 pages.

Carter, V., 1996, Technical aspects of wetlands: wetland hydrology, water quality, and associated functions. National Water Summary on Wetland Resources, United States Geological Survey Water Supply Paper 2425, pages 35-48.

Cedergren, H.R., 1997, Seepage, Drainage and Flow Nets, third edition. Wiley Professional Paperback Series, 496 pages.

Cey, E.E., D.L. Rudolph, G.W. Parkin, and R. Aravena, 1998, Quantifying ground water discharge to a small perennial stream in southern Ontario, Canada. Journal of Hydrology, volume 210, pages 21–37.

Chanton, J.P., W.C. Burnett, H. Dulaiova, D.R. Corbett, and M. Taniguchi, 2003, Seepage rate variability in Florida Bay Drive by Atlantic tidal height. Biogeochemistry, volume 66, pages 187–202.

Cherkauer, D.S., and J.M. McBride, 1988, A remotely operated seepage meter for use in large lakes and rivers. Groundwater, volume 26, pages 165-171.

Choi, J., J.W. Harvey, and M.H. Conklin, 2000, Characterizing multiple timescales of stream and storage zone interaction that affect solute fate and transport in streams. Water Resources Research, volume 36, pages 1511-1518, doi:10.1029/2000WR900051↗.

Choi, J., and J.W. Harvey, 2000, Quantifying time-varying ground-water discharge and recharge in wetlands of the northern Florida Everglades. Wetlands, volume 20, pages 500–511.

Coleman, R.L., C.N. Dahm, 1990, Stream geomorphology: effects on periphyton standing crop and primary production. Journal of the North American Benthological Society, volume 9, pages 293-302.

Combalicer, E.A., S.H. Lee, S. Ahn, D.Y. Kim, and S. Im, 2008, Comparing groundwater recharge and base flow in the Bukmoongol small-forested watershed, Korea. Journal of Earth System Science, volume 117, number 5, pages 553-366.

Conant, B. Jr., 2004, Delineating and quantifying ground water discharge zones using streambed temperature. Ground Water, volume 42, number 2, pages 243–257.

Conant, B.C. Jr., C.E. Tobinson, M.J. Hinton, and H.A.J. Russell, 2019, A framework for conceptualizing groundwater-surface water interactions and identifying potential impacts on water quality, water quantity, and ecosystems. Journal of Hydrology, volume 574, pages 609-627.

Connor, J.N., and T.V. Belanger, 1981, Ground water seepage in Lake Washington and the Upper St. Johns Basin, Florida. Water Resources Bulletin, volume 17, pages 799–805.

Constantz, J., C.L. Thomas, and G. Zellweger, 1994, Influence of diurnal variations in stream temperature on streamflow loss and groundwater recharge. Water Resources Research, volume 30, pages 3253–3264.

Constantz, J., and C.L. Thomas, 1997, Streambed temperatures profiles as indicators of percolation characteristics beneath arroyos in the Middle Rio Grande Basin, USA. Hydrologic Processes, volume 11, number 12, pages 1621–1634.

Constantz, J., 1998, Interaction between stream temperature, streamflow, and groundwater exchanges in alpine streams. Water Resources Research, volume 34, number 7, pages 1609–1616.

Constantz, J., J. Jasperse, D. Seymour, and G. Su, 2002a, Use of temperature to estimate streambed conductance, Russian River, California, Ground Water/Surface Water Interactions. American Water Resources Association Specialty Conference, pages 595-600.

Constantz, J., A.E. Stewart, R.G. Niswonger, and L. Sarma, 2002b, Analysis of temperature profiles for investigating stream losses beneath ephemeral channels. Water Resources Research, volume 38, number 12, pages 52-1 to 52-13.

Constantz, J., 2008, Heat as a tracer to determine streambed water exchanges. Water Resources Research, volume 44, W00D10, doi:10.1029/2008WR006996↗.

Cook, P.G., and A.L. Herczeg, 2000, Environmental Tracers in Subsurface Hydrology. Springer, 529 pages.

Cornett, R.J., B.A. Risto, and D.R. Lee, 1989, Measuring groundwater transport through lake sediments by advection and diffusion. Water Resources Research, volume 25, number 8, pages 1815-1823.

Cowardin, L.M., V. Carter, F.C. Golet, and E.T. LaRoe, 1979, Classification of wetlands and deepwater habitats of the United States. FWS/OBS-79/31, United States Fish and Wildlife Service, Washington, D.C., 103 pages.

Cox, S.E., F.W. Simonds, L. Doremus, R.L. Huffman, and R.M. Defawe, 2005, Groundwater/surface water interactions and quality of discharging groundwater in streams of the Lower Nooksack River basin, Whatcom County, Washington. United States Geological Survey Scientific Investigations Report 2005-5255, 46 pages.

Cuthbert, M.O., and R. Mackay, 2013, Impacts of non-uniform flow on estimates of vertical streambed flux. Water Resources Research, volume 49, number 1, pages 19–28, doi:10.1029/2011WR011587↗.

Datry, T., and S.T. Larned, 2008, River flow controls ecological processes and invertebrate assemblages in subsurface flowpaths of an ephemeral river reach. Canadian Journal of Fisheries and Aquatic Sciences, volume 65, pages 1532-1544.

Diehl, J.C., 2004, Hydrogeological characteristics and groundwater/river exchange in a gravel dominated floodplain Middle Fork of the Flathead River northwestern

Montana. Master of Science Thesis, Department of Geosciences, University of Montana, 151 pages.

Diersch, Hans-Joerg G., 2014, FEFLOW: Finite Element Modeling of Flow, Mass and Heat Transport in Porous and Fractured Medi., Springer, 996 pages.

Dingman, S.L., 1994, Physical Hydrology: Macmillan Publishing Company, 575 pages.

Domenico, P.A., and F.W. Schwartz, 2000, Physical and Chemical Hydrogeology. Wiley, New York.

Donato, M.M., 1998, Surface-Water/ground-Water Relations in the Lemhi River Basin, East-Central Idaho. United States Geological Survey Water-Resources Investigations Report 98-4185, 20 pages.

Drever, J.I., 1997, The geochemistry of natural waters: Surface and groundwater environments, third edition. Pearson, 436 pages.

Dumouchelle, D.H, 2001, Evaluation of ground-water/surface-water relations, Chapman Creek, West-Central Ohio, by means of multiple methods. United States Geological Survey Water-Resources Investigation Report 01-4202, 13 pages.

Duque, C., C.J. Russoniello, and D.O. Rosenberry, 2020, History and evolution of seepage meters for quantifying flow between groundwater and surface water: Part 2-marine settings and submarine groundwater discharge. Earth-Science Reviews, 204, 103168, Elsevier, 12 pages, doi:10.1016/j.earscirev.2020.103168↗.

Erickson, D.R., 1981, A study of littoral ground water seepage at Williams Lake, Minnesota, using seepage meters and wells. Master of Science Thesis, University of Minnesota, USA, 153 pages.

Fetter, C.W., 2001, Applied Hydrogeology, fourth edition. Prentice Hall, New Jersey, USA, 598 pages.

Fisher, S.G., N.B. Grimm, E. Martí, R.M. Holmes, B.J. Jeremy Jr., 1998, Material spiraling in stream corridors. a telescoping ecosystem model, Ecosystems, volume 1, pages 19-34.

Freeman, L.A., M.C. Carpenter, D.O. Rosenberry, J.P. Rousseau, R. Unger, and J.S. McLean, 2004, Use of submersible pressure transducers in water-resources investigations. Chapter A3, United States Geological Survey Techniques of Water-Resources Investigations, Book 8.

Freeze, R.A., and J.A. Cherry, 1979, Groundwater. Prentice-Hall, Englewood Cliffs, New Jersey, 604 pages.

Fretwell, J.E., Williams, J.S., and Redman, P.J., 1996, National water summary on wetland resources. United States Geological Survey Water Supply Paper 2425, doi:10.3133/wsp2425↗.

Fritts, C.R., 2012, Groundwater Science, second edition. Academic Press, 692 pages.

Fryar, A.E., E.J. Wallin, and D.L. Brown, 2000, Spatial and temporal variability in seepage between a contaminated aquifer and tributaries to the Ohio River. Ground Water Monitoring Remediation, volume 20, pages 129–146.

Glose, A.M., L.K. Lautz, and E.A. Baker, 2017, Stream heat budget modeling with HFLUX: model development, evaluation, and application across and contrasting sites and seasons. Environmental Modelling and Software, volume 92, pages 217-228, doi:10.1016/j.envsoft.2017.02.021↗.

Google Earth, 2015, 46 17'49.62 N 112 43'36.41 W Clark Fork River, Montana, United State of America, accessed 6/25/15.

Gooseff, M.N., D.M. McKnight, R.L. Runkel, and B.H. Vaughn, 2003a, Determining long time-scale hyporheic flow paths in Antarctic streams. Hydrological Processes, volume 17, pages 1691-1710.

Gooseff, M.N., S.M. Wondzell, R. Haggerty, and J. Anderson, 2003b, Comparing transient storage modeling and residence time distribution (RTD) analysis in geomorphologically varied reaches in the Lookout Creek basin, Oregon USA. Advances in Water Resources, volume 26, pages 925-937.

Gooseff, M.N., J.K. Anderson, S.M. Wondzell, J. LaNier, and R. Haggerty, 2006, A modeling study of hyporheic exchange pattern and the sequence, size and spacing of stream bedforms in mountain stream networks, Oregon, USA. Hydrological Processes, volume 20, number 11, pages 2443-2457.

Gordon, R.P., L.K. Lautz, M.A. Briggs, and J.M. McKenzie, 2012, Automated calculation of vertical pore-water flux from field temperature time series using the VFLUX Method and computer program. Journal of Hydrology, volume 420, pages 142–158.

Goto, S., M. Yamano, and M. Kinoshita, 2005, Thermal response of sediment with vertical fluid flow to periodic temperature variation at the surface. Journal of Geophysical Research, volume 110, B01106, doi:10.1029/2004JB003419↗.

Greig, S.M., D.A. Sear, and P.A. Carling, 2007, Review of factors influencing the availability of dissolved oxygen to incubating salmon embryos. Hydrological Processes, volume 21, pages 323-334.

Haggerty, R., S.A. McKenna, and L.C. Meigs, 2000, On the late-time behavior of tracer test breakthrough curves. Water Resources Research, volume 36, number 12, pages 3467–3479, doi:10.1029/2000WR900214↗.

Haggerty, R., and P.C. Reeves, 2002, STAMMT-L 1.0, Formulation and User's guide, Technical Report ERMS #520308, Sandia National Laboratory, Albuquerque, New Mexico, United States of America.

Haggerty, R., E. Martì, A. Argerich, D. von Schiller, and N.B. Grimm, 2009, Resazurin as a "smart" tracer for quantifying metabolically active transient storage in stream ecosystems, Journal of Geophysical Research, volume 114, G03014, doi:10.1029/2008JG000942↗.

Hannula, S.R., K.J. Esposito, J.A. Chermak, D.D. Runnells, D.C Keith, and L.E. Hall, 2003, Estimating ground water discharge by hydrograph separation. Ground Water, volume 41, number 3, pages 368-375.

Harvey Judson W., and Kenneth E. Bencala, 1993, The effect of streambed topography on surface-subsurface water exchange in mountain catchments. Water Resources Research, volume 29, issue 1, pages 89-98, https://doi.org/10.1029/92WR01960↗.

Harvey, F.E., D.R. Lee, D.L. Rudolph, and S.K. Frape, 1997, Locating groundwater discharge in large lakes using bottom sediment electrical conductivity mapping. Water Resources Research, volume 33, number 11, pages 2609-2615.

Harvey, J.W., and C.C. Fuller, 1998, Effect of enhanced manganese oxidation in the hyporheic zone on basin-scale geochemical mass balance. Water Resources Research, volume 34, pages 623–636, doi:10.1029/97WR03606↗.

Harvey, F.E., and D.R. Lee, 2000, Discussion of "The effects of bag type and meter size on seepage meter measurements". Ground Water, volume 38, pages 326–328.

Harvey, J.W., M.H. Conklin, and R.S. Koelsch, 2003, Predicting changes in hydrologic retention in an evolving semiarid alluvial stream. Advances in Water Resources, volume 26, pages 939–950.

Hatch, C.E., A.T Fisher, J.S. Revenaugh, J. Constantz, and C. Ruehl, 2006, Quantifying surface water–groundwater interactions using time series analysis of streambed thermal records: Method development. Water Resources Research, volume 42, W10410, doi:10.1029/2005WR004787↗.

Hauer, F.R., and G.A. Lamberti, editors, 2017, Methods in Stream Ecology, third edition, volume 1. Academic Press, Elsevier Incorporated, Burlington, MA, 886 pages.

Hays, J.R., 1966, Mass Transport Phenomena in Open Channel Flow. Doctor of Philosophy Dissertation, Department of Chemical Engineering, Vanderbilt University, Nashville, Tennessee.

Healy, R.W., and A.D. Ronan, 1996, Documentation of the computer program VS2DH for simulation of energy transport in variably saturated porous media-modification of the United States Geological Survey's computer program VS2DT. United States Geological Survey Water-Resources Investigation Report 96-4230, 36 pages.

Healy, R.W., T.C. Winter, J.W. LaBaugh, and O.L. Franke, 2007, Water budgets: foundations for effective water-resources and environmental management. United States Geological Survey Circular 1308, 90 pages.

Heaney, M. J., J.E. Nyquist, and L. Toran, 2007, Marine resistivity as a tool for characterizing zones of seepage at Lake Lacawac, PA. Conference Symposium on the Application of Geophysics to Engineering and Environmental Problems, doi:10.4133/1.2924642↗.

Hinton, M.J., 2014, Groundwater-Surface Water Interactions in Canada, *in* Canada's Groundwater Resources, editors, Rivera, A., Fitzhenry and Whiteside, Brighton Massachusetts, pages 151-185.

Hoehm, E., and O.A. Cirpka, 2006, Assessing residence times of hyporheic ground water in two alluvial flood plains of the Southern Alps using water temperature and tracers. Hydrological Earth Systems Science, volume 10, pages 553-563.

Hornberger, G.M., J.P. Raffensperger, and K.N. Eshleman, 1998, Elements of Physical Hydrology. Johns Hopkins University Press, 312 pages.

Hsieh, P.A., W. Wingle, and R.W. Healy, 2000, VS2DI—a graphical software package for simulating fluid flow and solute or energy transport in variably saturated porous media. United States Geological Survey Water-Resources Investigations Report 99-4130, 16 pages.

Hvorslev, M. J., 1951, Time Lag and Soil Permeability in Ground Water Observations. United States Army Corps of Engineers, Bulletin 36, Waterways Experimentation Station, Vicksburg, Mississippi, United States of America.

Isiorho, S.A., and G. Matisoff, 1990, Ground water recharge from Lake Chad. Limnology and Oceanography, volume 35, pages 931–938.

Isiorho, S.A., and J.H. Meyer, 1999, The effects of bag type and meter size on seepage meter measurements. Ground Water, volume 37, number 3, pages 411–413.

Israelsen, O.W., and R.C. Reeve, 1944, Canal lining experiments in the delta area, Utah. Utah Agricultural Experiment Station, Technical Bulletin 313, Logan, Utah.

Jackman, A.P., R.A. Walters, and V.C. Kennedy, 1984, Transport and concentration controls for chloride, strontium, potassium and lead in Uvas Creek, a small Cobble-Bed Stream in Santa Clara County, California, U.S.A, Part 2: Mathematical Modeling. Journal of Hydrology, volume 75, pages 111–141.

Jackman, A.P., F.J. Triska, and J.H. Duff, 1997, Hydrologic examination of ground water discharge into the upper Shingobee River. United States Geological Survey Water Resources Investigations Report 96-4215, pages 137–147.

Jiao, J. and V. Post, 2019, Coastal Hydrogeology. Cambridge University Press, New York, 418 pages.

John, P.H., and M.A. Lock, 1977, The spatial distribution of ground water discharge into the littoral zone of a New Zealand Lake. Journal of Hydrology, volume 33, pages 391–395.

Johnson, A.N., B.R. Boer, W.W. Woessner, J.A. Stanford, G.C. Poole, S.A. Thomas, and S.J. O'Daniel, 2005, Evaluation of an inexpensive small-diameter temperature logger for documenting ground water–river interactions. Groundwater and Remediation, volume 25, Issue 4, pages 101–105.

Katz, B.G., T.B. Coplen, T.D. Bullen, and H.J. Davis, 1997, Use of chemical and isotopic tracers to characterize the interactions between ground water and surface water in mantled karst. Ground Water, volume 35, number 6, pages 1014–1028.

Keery, J., A. Binley, N. Crook, and J.W.N. Smith, 2007, Temporal and spatial variability of groundwater-surface water fluxes: Development and application of an analytical method using temperature time series. Journal of Hydrology, volume 336, number 1–2, pages 1–16.

Kennedy, C.D., L.C. Murdoch, D.P. Genereux, D.R. Corbett, K. Stone, P. Pham, and H. Mitasova, 2010, Comparison of Darcian flux calculations and seepage meter measurements in a sandy streambed in North Carolina, United States. Water Resources Research, volume 46, W09501, doi:10.1029/2009WR008342↗.

Kennedy, C.S.C, 2017, Groundwater-surface water interactions in the discrete fracture networks of bedrock rivers. Doctor of Philosophy Dissertation, Environmental Sciences, University of Guelph, Canada, 155 pages.

Kilpatrick, F.A., 1970, Dosage requirements for slug injections of Rhodamine BA and WT dyes, *in* Geological Survey Research. United States Geological Professional Paper 700-B, pages B250-B253.

Kilpatrick, F.A., and E.D. Cobb, 1985, Measurement of discharge using tracers. Techniques of Water-Resources Investigations of the United States Geological Survey, Book 3, Chapter A16.

Kim, B.K.A., A.P. Jackman, and F.J. Triska, 1992, Modeling biotic uptake by periphyton and transient hyporheic storage of nitrate in a natural stream. Water Resources Research, volume 28, pages 2743–2752, doi:10.1029/92WR01229↗.

Kimball, B. A., R.L. Runkel, T.E. Cleasby, and D.A. Nimick, 2004, Quantification of Metal Loading by Tracer Injection and Synoptic Sampling, 1997–98. United States Geological Survey Professional Paper 1652, pages 191–262.

Kish, G.R., C.E. Stringer, M.T. Stewart, M.C. Rains, and A.E. Torres, 2010, A geochemical mass-balance method for base-flow separation, Upper Hillsborough, River Watershed, West-Central Florida, 2003-2005 and 2009. United States Geological Survey Scientific Investigations Report 2010-5092, 33 pages.

Lamberti, G.A., and F.R. Hauer, editors, 2017, Methods in Stream Ecology, third edition, volume 2. Academic Press, Elsevier Incorporated, Burlington, Massachusetts, 886 pages.

Landon, M. K., D.L. Rus, and F.E. Harvey, 2001, Comparison of instream methods for measuring hydraulic conductivity in sandy streambeds, Ground Water, volume 39, number 6, pages 870–885, doi:10.1111/j.1745-6584.2001.tb02475.x↗.

Langhoff, J.H., K.R. Rasmussen, and S. Christensen, 2005, Quantification and regionalization of groundwater–surface water interaction along an alluvial stream. Journal of Hydrology, volume 320, pages 342-358.

Lee, D.R., 1972, Septic Tank Nutrients in Groundwater Entering Lake Sallie, Minnesota. Master of Science Thesis, University of North Dakota, 96 pages.

Lee, D.R., 1977, A device for measuring seepage flux in lakes and estuaries. Limnology and Oceanography, volume 22, number 1, pages 140–147.

Lee, D.R., and H.B. Hynes, 1977, Identification of ground water discharge zones in a reach of Hillman Creek in Southern Ontario. Water Pollution Resources Canada, volume 13, pages 121–133.

Lee, D.R., and J.A. Cherry, 1978, A field exercise on groundwater flow using seepage meters and mini-piezometers. Journal of Geological Education, volume 27, pages 6–10.

Lee, D.R., 1985, Method for locating sediment anomalies in lakebeds that can be caused by groundwater flow. Journal of Hydrology, volume 79, pages 187-193.

Lee, D.R., R. Dal Bianco, and M. St. Aubin, 1993, Locating groundwater discharge zones using electrical conductivity. Paper presented at Outdoor Action Conference. National Water Well Association, Las Vegas, May 25-27, 1993.

Lehner, B., and P. Döll, 2004, Development and validation of a global database of lakes, reservoirs, and wetlands. Journal of Hydrology, volume 296, pages 1–22.

Lesack, L.F., 1995, Seepage exchange in an Amazon floodplain lake. Limnology and Oceanography, volume 40, pages 598–609.

Liao, Z., and O.A. Cirpka, 2011, Shape-free inference of hyporheic traveltime distributions from synthetic conservative and "smart" tracer tests in streams. Water Resources Research, volume 47, W07510, doi:10.1029/2010WR009927↗.

Liao, S., D. Lemek, K. Osenbruck, and O.A. Cirpka, 2013, Modeling and inverting reactive stream tracers undergoing two-site sorption and decay in the hyporheic zone. Water Resources Research, volume 49, pages 3406-3422, doi:10.10.1002.WRCR.20276↗.

Libelo, E.L., and W.G. MacIntyre, 1994, Effects of surface-water movement on seepage-meter measurements of flow through the sediment-water interface. Applied Hydrogeology, volume 2, number 4, pages 49–54.

Lien, B.K., and Ford, R.G., 2014, Quantifying seepage flux using sediment temperatures. U.S. Environmental Protection Agency, EPA/600/R-15/454, 23 pages, https://nepis.epa.gov/Exe/ZyPDF.cgi/P100MP4G.PDF?Dockey=P100MP4G.PDF↗.

Loustaunau, K. P., 2003, Transport and fate of methyl tertiary butyl ether (MTBE) in a floodplain aquifer and a stream interface, Ronan, Montana. Master of Science Thesis, Department of Geosciences, University of Montana, Missoula, 86 pages.

Lucius, M.A., 2016, Creating a water and nutrient budget for Lake Trafford, FL, USA. Unpublished Master of Science Thesis, Florida Gulf Coast University, College of Arts and Sciences, 129 pages.

Malcolm, I.A., C. Soulsby, A.F. Youngson, and D.M. Hanna, 2005, Catchment-scale controls on groundwater -surface water interactions in the hyporheic zone: Implications for salmon embryo survival. River Research and Applications, volume 21, number 9, pages 977-989.

Marion, A., M. Zaramella, and A. Bottacin-Busolin, 2008, Solute transport in rivers with multiple storage zones: The STIR model. Water Resources Research, volume 44, W10406, doi:10.1029/2008WR007073↗.

McBride, J.M., 1987, Measurement of ground water flow to the Detroit River, Michigan and Ontario: Milwaukee, Wisconsin. Master of Science Thesis, University of Wisconsin–Milwaukee.

McBride, M. S., and H.O. Pfannkuch, 1975, The distribution of seepage within lakebeds. Journal of Research United States Geological Survey, volume 3, number 5, pages 505-512.

Meyer, J.L., 1997, Stream health: incorporating the human dimension to advance stream ecology. Journal of the North American Benthological Society, volume 16, number 2, pages 439-447.

Meyboom, P., 1961, Estimating groundwater recharge from stream hydrographs. Journal of Geophysical Research, volume 66, pages 1203-1214.

Mitsch, W.J., and J.G. Gosselink, 2000, Wetlands, third edition. Wiley, New York, 920 pages.

Moore, G.K., 1992, Hydrograph analysis in a fractured rock terrane. Ground Water, volume 30, number 3, pages 390-395.

Murdoch, L.C., and S.E. Kelly, 2003, Factors affecting the performance of conventional seepage meters. Water Resources Research, volume 39, number 6, 1163, doi:10.1029/2002WR001347↗.

Mwakanyamale, K., L. Slater, F. Day-Lewis, M. Elwaseif, and C. Johnson, 2012, Spatially variable stage-driven groundwater-surface water interactions inferred from time-frequency analysis of distributed temperature sensing data. Geophysical Research Letters, volume 39, L06401, doi:10.10292011GL050824↗.

Neff, B.P, S.M. Day, A.R. Piggott, and L.M. Fuller, 2005, Base Flow in the Great Lakes Basin. United States Geological Survey Scientific Investigations Report, 2005-5217, 23 pages.

O'Connor, B.L., M. Hondzo, and J.W. Harvey, 2010, Predictive modeling of transient storage and nutrient uptake. Journal of Hydraulic Engineering, volume 136, number 12.

OSU, Oregon State University, 2020, website on Fiber optic cable installation in the John Day, Oregon, accessed September 8, 2020. https://ctemps.org/feature-story/fiber-optic-cable-installation-john-day-oregon↗.

Paulsen, R.J., C.F. Smith, D. O'Rourke, and T.F. Wong, 2001, Development and evaluation of an ultrasonic ground water seepage meter. Ground Water, volume 39, pages 904–911.

Pepin, D.M., F.R. Hauer, 2002, Benthic responses to groundwater-surface water exchange in two alluvial rivers. Journal of the North American Benthological Society, volume 21, pages 370-383.

Pierce, A.A., B.L. Parker, R. Ingleton, and J.A. Cherry, 2018, Novel well completions in small diameter coreholes created using portable rock drills. Groundwater Monitoring and Remediation, volume 38, number 1, pages 43-55.

Plummer, L.N., 1993, Environmental tracers for water movement in desert soils of the American southwest. Soil Science Society of American Journal, volume 56, number 1, pages 15-24.

Plummer, L.N., E. Busenberg, J.K. Böhlke, D.L. Nelms, R.L. Michel, and P. Schlosser, 2001, Groundwater residence times in Shenandoah National Park, Blue Ridge Mountains, Virginia, USA—A multi-tracer approach. Chemical Geology, volume 179, pages 93–111.

Poole, G.C., S.J. O'Daniel, K.L. Jones, W.W. Woessner, E.S. Bernhardt, A.M. Helton, J.A., Stanford, B.R. Boer, and T.J. Beechie, 2008, Hydrologic spiralling: the role of multiple interactive flow paths in stream ecosystems. River Research and Applications, volume 24, pages 1018–1031.

Renken, R.A., K.L. Cunningham, M.R. Zygnerski, M.A. Wacker, A.M. Shapiro, R.W. Harvey, D.W. Metge, C.L. Osborn, and J.N. Ryan, 2005, Assessing the vulnerability of a municipal well field to contamination in a karst aquifer, Environmental and Engineering Geoscience, volume 11, number 4, pages 319–331.

Rivett, M.O., Ellis, P.A., Greswell, R.B., Ward, R.S., Roche, R.S., Cleverly, M.G., C. Walker, D. Conran, P.J. Fitzgerald, T. Willcow, and J. Dowle, 2008, Cost-effective mini drivepoint piezometers and multilevel samplers for monitoring the hyporheic zone. Quarterly Journal of Engineering Geology and Hydrogeology, volume 41, pages 49-60.

Roberts, M., and K. Warren, 1999, North Fork Blackfoot River Hydrologic Analysis. Montana Department of Natural Resources and Conservation, Helena, Montana, United States of America, 36 pages.

Robertson, D.M., W.J. Rose, and H.S. Garn, 2003, Water quality and the effects of changes in phosphorous loading, Red Cedar Lakes, Barron and Washburn Counties, Wisconsin. United States Geological Survey Water-Resources Investigations Report 03-4238, 42 pages.

Rosenberry, D.O., 2000, Unsaturated-zone wedge beneath a large, natural lake. Water Resources Research, volume 36, number 12, pages 3401–3409.

Rosenberry, D.O., 2005, Integrating seepage heterogeneity with the use of ganged seepage meters. Limnology Oceanography, Method 3, pages 131-142.

Rosenberry, D.O., C. Duque, and D.R. Lee, History and evolution of seepage meters for quantifying flow between groundwater and surface water: Part 1 – Freshwater settings. Earth-Science Reviews, volume 204, doi:10.1016/j.earscirev.2020.103167↗.

Rosenberry, D.O., and J.W. LaBaugh, 2008, Field techniques for estimating water fluxes between surface water and ground water. United States Geological Survey Techniques and Methods 4–D2, 128 pages.

Rosenberry, D.O., J.W. LaBaugh, and R.J. Hunt, 2008, Use of monitoring wells, portable piezometers, and seepage meters to quantify flow between surface water and ground water, *in* Rosenberry, D.O. and J.W. LaBaugh, editors, Chapter 2 of Field Techniques for Estimating Water Fluxes between Surface Water and Ground Water. United States Geological Survey, Techniques and Methods 4-D2, pages 39-70.

Rosenberry, D.O. and M.A. Menheer, 2006, A system for calibrating seepage meters used to measure flow between ground water and surface water. United States Geological Survey Scientific Investigations Report 2006-5053, 21 pages.

Rosenberry, D.O., and R.H. Morin, 2004, Use of an electromagnetic seepage meter to investigate temporal variability in lake seepage. Groundwater, volume 42, pages 68-77.

Rorabaugh, M.I., 1964, Estimating changes in bank storage and ground-water contribution to streamflow. International Association of Scientific Hydrology, Publication 63, pages 432-441.

Rorabaugh, M.I., and W.D. Simons, 1966, Exploration of methods relating ground water to surface water, Columbia River basin--second phase. United States Geological Survey Open-File Report, 62 pages.

Runkel, R. L., 1998, One-Dimensional Transport with Inflow and Storage (OTIS): A solute transport model for streams and rivers. United States Geological Survey Water Resources, Investigation Report, 98-4018, 73 pages.

Runkel, R.L., 2002, A new metric for determining the importance of transient storage. Journal of the North American Benthological Society, volume 21, pages 529-543.

Runkel, R. L., and S.C. Chapra, 1993, An efficient numerical solution of the transient storage equations for solute transport in small streams. Water Resources Research, volume 29, pages 211–215, doi:10.1029/92WR02217↗.

Runkel, R.L., D.M. McKnight, and E.D. Andrews, 1998, Analysis of transient storage subject to unsteady flow: Diel flow variation in an Antarctic stream. Journal of the North American Benthological Society, volume 17, pages 143-154.

Russonielllo, C.J., and H.A. Hichael, 2015, Investigation of seepage meter measurements in steady flow and wave conditions. Ground Water, volume 53, number 6, pages 959-966.

Rutledge, A.T., 1993. Computer programs for describing the recession of ground-water discharge and for estimating mean ground-water recharge and discharge from streamflow records. United States Geological Survey Water-Resources Investigations Report 93-4121, 45 pages.

Rutledge, A.T., 1998, Computer programs for describing the recession of ground-water discharge and for estimating mean ground-water recharge and discharge from streamflow records—Update. United States Geological Survey Water-Resources Investigations Report 98–4148, 43 pages, doi:10.3133/wri984148↗.

Rutledge, A.T., 2000, Considerations for use of the RORA program to estimate ground-water recharge from streamflow records. United States Geological Survey Open-File Report 2000-156, 44 pages, http://pubs.usgs.gov/of/2000/ofr00-156/↗.

Rutledge, A.T., 2002, User Guide for the PULSE Program: Open-File Report 2002-455, https://pubs.er.usgs.gov/publication/ofr02455↗.

Rutledge, A.T., 2005a, Appropriate use of the Rorabaugh model to estimate groundwater recharge. Ground Water, volume 43, issue 3, pages 292-293. doi 10.1111/j.1745-6584.2005.0022x↗.

Rutledge, A.T., 2005, Program user guide for PART, http://water.usgs.gov/ogw/part/User.ManualPART.pdf↗.

Rutledge, A.T., 2007a, A program user guide for PART, http://water.usgs.gov/ogw/part/UserManualPART.pdf↗.

Rutledge, A.T., 2007b, Update on the use of the RORA program for recharge estimation. Ground Water, volume 45, number 3, pages 374–382.

Sack, L A., A. Swancar, and T.M. Lee, 1998, Estimating Ground-water exchange with lakes using water-budget and chemical mass-balance approaches of Ten Lakes in ridge areas of Pol and Highlands Counties, Florida. United States Geological Survey Water-Resources Investigations Report 98-4133, 52 pages.

Schmidt, C., B. Conant Jr., M. Bayer-Raich, and M. Schirmer, 2007, Evaluation and field-scale application of an analytical method to quantify groundwater discharge using mapped streambed temperatures. Journal of Hydrology, volume 341, number 3-4, pages 292–307.

Schwartz, F.W., and H. Zhang, 2003, Fundamentals of Ground Water. Wiley, 592 pages.

Sebestyen, S.D., and R.L. Schneider, 2001, Dynamic temporal patterns of nearshore seepage flux in a headwater Adirondack Lake. Journal of Hydrology, volume 247, pages 137–150.

Selker, J.S., L. Thévenaz, H. Huwald, A. Mallet, W. Luxemburg, N. van de Giesen, M. Stejskal, J. Zeman, M. Westhoff, and M.B. Parlange, 2006, Distributed fiber-optic temperature sensing for hydrologic systems. Water Resources Research, volume 42, W12202, doi:10.1029/2006WR005326↗.

Shaw, R.D., and E.E. Prepas, 1989, Anomalous, short-term influx of water into seepage meters. Limnology and Oceanography, volume 34, number 7, pages 1343–1351.

Shaw, R.D., and E.E. Prepas, 1990a, Ground water-lake interactions - I. Accuracy of seepage meter estimates of lake seepage. Journal of Hydrology, volume 119, pages 105–120.

Shaw, R.D., and E.E. Prepas, 1990b, Ground water-lake interactions - II. Nearshore seepage patterns and the contribution of ground water to lakes in central Alberta. Journal of Hydrology, volume 112, pages 121–136.

Shinn, E.A., C.D. Reich, and T.D. Hickey, 2002, Seepage meters and Bernoulli's revenge. Estuaries, volume 25, pages 126–132.

Simonds, F.W., and K.A. Sinclair, 2002, Surface-water/ground-water interactions along the lower Dungeness River and vertical hydraulic conductivity of streambed sediments, Clallam County, Washington, September 1999–July 2001. United States Geological Survey Water-Resources Investigations Report 02–4161, 60 pages.

Smerdon, B.D., W.P. Gardner, G.A. Harrington, and S.J. Tickell, 2012, Identifying the contribution of regional groundwater to the baseflow of a tropical river (Daly River), Australia. Journal of Hydrology, 468-465, pages 107-115.

Steenson, R., and A. Naugle, 2017, Evaluating contaminated groundwater discharges to surface water. Groundwater Resources Association Annual Conference, San Francisco Bay Regional Water Control Board.

Sterrett, R. J., editor, 2008, Groundwater and wells: third edition. Johnson Division, St Paul, Minnesota, 812 pages.

Stewart, M., J. Cimino, and M.A. Ross, 2007, Calibration of base flow separation methods with streamflow conductivity. Ground Water, volume 45, number 1, pages 17-27.

Stonedahl, S.H., J.W. Harvey, and A.I. Packman, 2013, Interactions between hyporheic flow produced by stream meanders, bars, and dunes. Water Resources Research, volume 9, pages 5450–5461, doi:10.1002/wrcr.20400↗.

Stonestrom, D.A., and J. Constantz, 2003, Heat as a tracer of water movement near streams, *in* Heat as a Tool for Studying the Movement of Ground Water Near Streams, Stonestrom, D.A., and J. Constantz, editors, United States Geological Survey Circular 1260, Reston, Virginia, pages 1–6.

Stonestrom, D.A., and J. Constantz, 2004, Using temperature to study stream-ground water exchanges. United States Geological Survey Fact Sheet 2004-3010.

Storey, R.G., K.W.G. Howard, and D.D. Williams, 2003, Factors controlling riffle-scale hyporheic exchange flows and their seasonal changes in a graining stream: A three-dimensional groundwater flow model. Water Resources Research, volume 39, number 2, doi:10.1029/2002WR001367↗.

Stubbington, R., P.J. Wood, A.J. Boulton, 2009, Low flow controls on benthic and hyporheic macroinvertebrate assemblages during a supra-seasonal drought. Hydrological Processes, volume 23, pages 2252-2264.

Stumm, W., and J.J. Morgan, 1996, Aquatic chemistry, third edition. John Wiley and Sons, New York, 1022 pages.

Swanson, T.E., and M.B. Cardenas, 2011, Ex-Stream: A MATLAB program for calculating fluid flux through sediment-water interfaces based on steady and transient temperature profiles. Computational Geoscience, volume 37, number 10, pages 1664–1669.

Taniguchi, M., and Y. Fukuo, 1993, Continuous measurements of ground-water seepage using an automatic seepage meter. Ground Water, volume 31, number 4, pages 675–679.

Taniguchi, M., 2002, Tidal effects on submarine groundwater discharge into the ocean. Geophysical Research Letters, volume 29, number 12, pages 2-1 – 2-3, doi:10.1029/2002GL014987↗.

Thackston, E.L, and K.B. Schnelle, 1970, Predicting the effects of dead zones on stream mixing. Journal of the Sanitary Engineering Division, American Society of Civil Engineers, volume 96, pages 319-331.

Tiedeman, C.R., D.J. Goode, and P.A. Hsieh, 1997, Numerical simulation of ground-water flow through glacial deposits and crystalline bedrock in the Mirror Lake area, Grafton County, New Hampshire. United States Geological Survey Professional Paper 1572, 50 pages.

Tiner, R.W., 1996, *in* National Water Summary on Wetland Resources, Washington, D.C. United States Geological Survey Water, Supply Paper 2445, 431 pages.

Todd, D.K, and L.W. Mays, 2004, Groundwater hydrology, third edition. Wiley, 656 pages.

Tonina, D., and J.M. Buffington, 2007, Hyporheic exchange in gravel bed rivers with pool-riffle morphology: Laboratory experiments and three-dimensional modeling. Water Resources Research, volume 43, W01421, doi:10.1029/2005WR004328↗.

Toran, L., J. Nyquiest, D. Rosenberry, M. Gagliano, N. Mitchell, and J. Mikochik, 2014, Geophysical and hydrologic studies of lake seepage variability. Ground Water, volume 53, number 6, pages 841-850.

Torres, N.T., P.C. Hauser, G. Furrer, H. Brandl, and B. Muller, 2013, Sediment porewater extraction and analysis combining filter tube samplers and capillary electrophoresis. Journal of Environmental Monitoring, volume 15, number 4, pages 715-720.

Tóth, J., 1963, A theoretical analysis of groundwater flow in small drainage basins. Journal of Geophysical Research, volume 68, pages 4795–4812, doi:10.1029/JZ068i016p04795↗.

Turnipseed, D.P., and V.B. Sauer, 2010, Discharge measurements at gaging stations. United States Geological Survey Techniques and Methods Book 3, chapter A8, 87 pages. http://pubs.usgs.gov/tm/tm3-a8/↗.

USEPA, United States Environmental Protection Agency, 2019, https://cfpub.epa.gov/watertrain/pdf/modules/new_streamcorridor.pdf↗.

USGS, United States Geologic Survey, 2005, https://pubs.usgs.gov/sir/2005/5065/images/staff_gage.jpg↗.

USGS, United States Geologic Survey, 2011, How Does a U.S. Geological Survey Streamgage Work? https://pubs.usgs.gov/fs/2011/3001/pdf/fs2011-3001.pdf↗.

USGS, United States Geologic Survey, 2015, Uncovering the Mighty Mississippi's Natural Potential for Nitrogen Removal, https://www.usgs.gov/news/uncovering-mighty-mississippi%E2%80%99s-natural-potential-nitrogen-removal↗.

USGS, United States Geological Survey, 2016, website on Fiber-Optic Distributed Temperature Sensing Technology, https://water.usgs.gov/ogw/bgas/fiber-optics/↗.

USGS, United States Geological Survey, 2019, Streamgage Diagram, https://www.usgs.gov/media/images/usgs-streamgage-diagram↗.

USGS 2020, United States Geological Survey, website on Handheld Thermal Imaging Cameras for Groundwater/Surface-Water Interaction Studies, accessed September 8, 2020, https://water.usgs.gov/ogw/bgas/thermal-cam/↗.

Valentine, E., and I. Wood, 1979, Experiments in longitudinal dispersion with dead zones. Journal of the Hydraulics Division, American Society of Civil Engineers, volume 105, pages 999–1016.

Valett, H.M., S.G. Fisher, E.H. Stanley, 1990, Physical and chemical characteristics of the hyporheic zone of a Sonoran Desert stream. Journal of the North American Benthological Society, volume 9, pages 201-215.

Valett, H.M., S.G. Fisher, N.B. Grimm, P. Camill, 1994, Vertical hydrologic exchange and ecological stability of a desert stream ecosystem. Ecology, volume 75, pages 548-560.

Vanek, V., and D.R. Lee, 1991, Mapping submarine groundwater discharge areas-an example from Laholm Bay, southwest Sweden. Limnology Oceanography, volume 36, number 6, pages 1250-1262.

Ward, A.S., 2016, The evolution and state of interdisciplinary hyporheic research. Wiley Interdisciplinary Reviews Water, volume 3, pages 83-103.

Watson, I., and A.D. Burnett, 1993, Hydrology: An environmental approach (theory and applications of ground water and surface water for engineers and geologists). Buchanan Books, Cambridge, 702 pages.

Webb, B.W., D.M. Hannah, R.D. Moore, L.E. Brown, and R. Nobilis, 2008, Recent advances in stream and river temperature research. Hydrological Processes, volume 22, pages 901–918, doi:10.1002/hyp.6994↗.

Weight, W.D., 2019, Practical hydrogeology. Principles and field applications, third edition. McGraw Hill, 800 pages.

Winter, T.C., 1976, Numerical simulation analysis of the interaction of lakes and groundwaters. United States Geological Survey Professional Paper 1001.

Winter, T.C., 1978, Numerical simulation of steady-state three-dimensional groundwater flow near lakes. Water Resources Research, volume 14, pages 245-254.

Winter, T.C., 1981, Uncertainties in estimating the water balance of lakes. Water Resources Bulletin, volume 17, pages 82-115.

Winter, T. C., J.W. Harvey, O.L. Franke, and W.A. Alley, 1998, Ground water and surface water: A single resource. United States Geological Survey Circular, 1139, 79 pages, https://pubs.er.usgs.gov/publication/cir1139↗.

Winter, T.C., J.W. LaBaugh, and D.O. Rosenberry, 1988, The design and use of a hydraulic potentiomanometer for direct measurement of differences in hydraulic head between groundwater and surface water. Limnology and Oceanography, volume 33, number 5, pages 1209–1214.

Wisler, C.O., and E.F. Barter, 1959, Hydrology, second edition. John Wiley and Sons.

Woessner, W.W., and K.E. Sullivan, 1984, Results of seepage meter and mini-piezometer study, Lake Mead, Nevada. Ground Water, volume 22, number 5, pages 561-568.

Woessner, W.W., 1998, Changing views of stream-groundwater interaction, *in* Van Brahana, J., Y. Eckstein, L.W. Ongley, R. Schneider, and J.E. Moore, editors, Proceedings of the Joint Meeting of the XXVIII Congress of the International Association of Hydrogeologists and the Annual Meeting of the American Institute of Hydrology. American Institute of Hydrology, St. Paul, Minnesota, pages 1–6.

Woessner, W.W., 2000, Stream and fluvial plain ground-water interactions: Re-scaling hydrogeologic thought. Ground Water, volume 38, number 3, pages 423–429.

Woessner, W. W., 2017, Hyporheic Zones, *in* Methods in Stream Ecology, volume 1, third edition, F.R. Hauer and G.A. Lamberi, editors, Academic Press-Elsevier, pages 129-157, doi:10.1016/B978-0-12-416558-8.00008-1↗.

Woessner, W.W., and W.D. Weight, 2019, Chapter 10 Groundwater/Surface-Water Interaction, *in* Practical Hydrogeology: Principles and Field applications, third edition. McGraw Hill, pages 431-521.

Woessner W.W., 2018, With permission various original figures and tables created for publications in 2018, william.woessner@umontana.edu↗.

Woessner, W.W., 2019, With permission, various original figures and tables created for publications in 2019, william.woessner@umontana.edu↗.

Woessner, W.W., 2020, With permission, various original figures and tables created for publications in 2020, william.woessner@umontana.edu↗.

Wörman, A., 1998, Analytical solution and timescale for transport of reactive solutes in rivers and streams. Water Resources Research, volume 34, pages 2703–2716, doi:10.1029/98WR01338↗.

Wörman, A., A.I. Packman, H. Johansson, and K. Jonsson, 2002, Effect of flow-induced exchange in hyporheic zones on longitudinal transport of solutes in stream and rivers. Water Resources Research, volume 38, number 1, doi:10.1029/2001WR00769↗.

Worman, A., A.I. Packman, L. Marklund, J.W. Harvey, and S.H. Stone, 2007, Fractal topography and subsurface water flows from fluvial bedforms to the continental shield. Geophysical Research Letters, volume 34, L07402, doi:10.1029/2007GL029426↗.

Zamora, C., 2008, Estimating water fluxes across the sediment–water interface in the Lower Merced River, California. United States Geological Survey Scientific Investigations Report 2007–5216, 47 pages. http://pubs.usgs.gov/sir/2007/5216/↗.

Zheng, C. and G.D. Bennett, 2002, Applied contaminant transport modeling, second edition. John Wiley and Sons, New York, 621 pages.

9 BOXES

Box 1 Simulating Lake Conceptual Models, Winter's Models

The early work of Winter in the 1970's made an important contribution to conceptualizing lakes in the larger groundwater system. Winter (1976) simulated conceptual models of lakes and groundwater exchange using two-dimensional cross sections orientated parallel to groundwater flow and also in three dimensional models (Winter, 1976; Winter, 1978). This work explored how theoretical groundwater flow and exchange would occur using cross sections constrained by vertical side hydraulic divides and a no-flow boundary at the base. The simulated sections could represent anisotropic sediments, be assigned various hydraulic conductivities, and allow for variation in the position of the water table (Figure Box 1-1). Models were simulated under steady-state conditions.

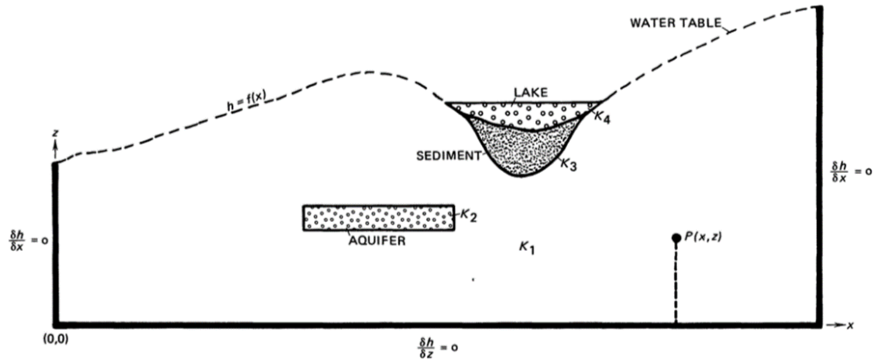


Figure Box 1-1 - Numerical two-dimensional model boundary conditions and parameters used by Winter (1976) to simulate lake exchange and groundwater flow. Both a single-lake and a three-lake model was developed. Hydraulic conductivity values were assigned as K_x and K_z resulting in anisotropic conditions. The water table was assigned as fixed head values, and left, right and bottom boundaries were assigned as zero flux, no-flow. In some models, a rectangle of sediment with differing hydrogeologic properties was included in the simulation, labeled Aquifer in the image (Winter, 1976).

This exercise provided insight into how groundwater and lakes may interact. For example, in the setting shown in Figure Box 1-2, the lake is influenced by the adjacent local flow systems and the regional groundwater flow continues below the lake as hydraulic divides and stagnation points are formed. Winter (1976) describes a stagnation point as "a point on the divide (between flow systems) at which the head is a minimum compared to every other point along the divide". … "It is a point in the flow field at which vectors of flow are equal in opposite directions and therefore cancel … The stagnation point is a point of diversion of ground-water-flow paths." In Figure Box 1-2, the minimum head along the local groundwater divide is 1.8 ft (0.5 m) higher than the lake elevation. When groundwater and lakes are not separated by a hydraulic divide and stagnation point, flow to and leakage from a lake may occur, i.e., mixed exchange (Figure Box 1-3).

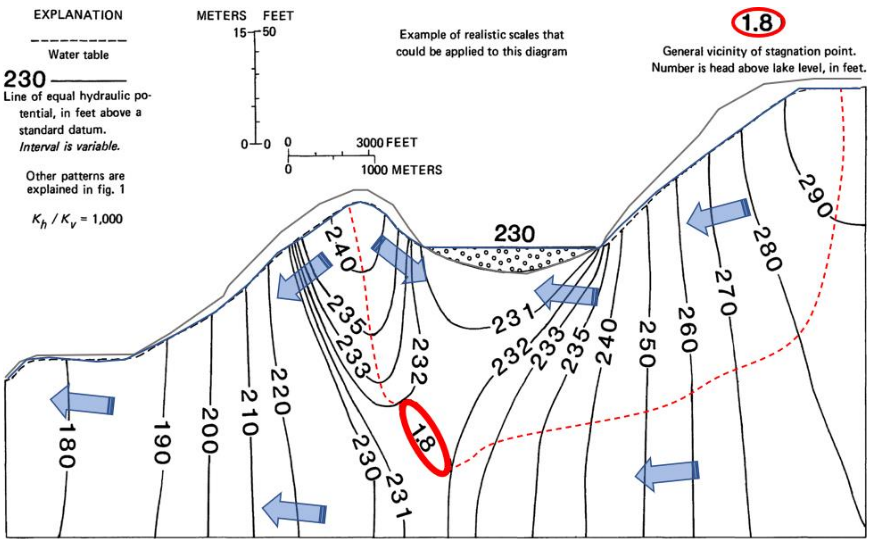


Figure Box 1-2 - Cross section of groundwater flow in a topography with hills (brown line represents land surface/lake bottom; blue line is the water table) and a lake. The lake is effluent or gaining as groundwater locally flows into the lake. Groundwater flow lines are not represented as the system is anisotropic and vertically exaggerated by 80%. Large blue arrows show general groundwater flow. The red dashed line is an approximate representation of the hydraulic divide separating the local and larger more regional groundwater flow system. The heavy red line surrounds an area in which a stagnation point separates local and regional groundwater flow. The head at the stagnation zone is the lowest head on the hydraulic divide but still higher than the lake stage, here 1.8 ft higher (modified from Winter, 1976).

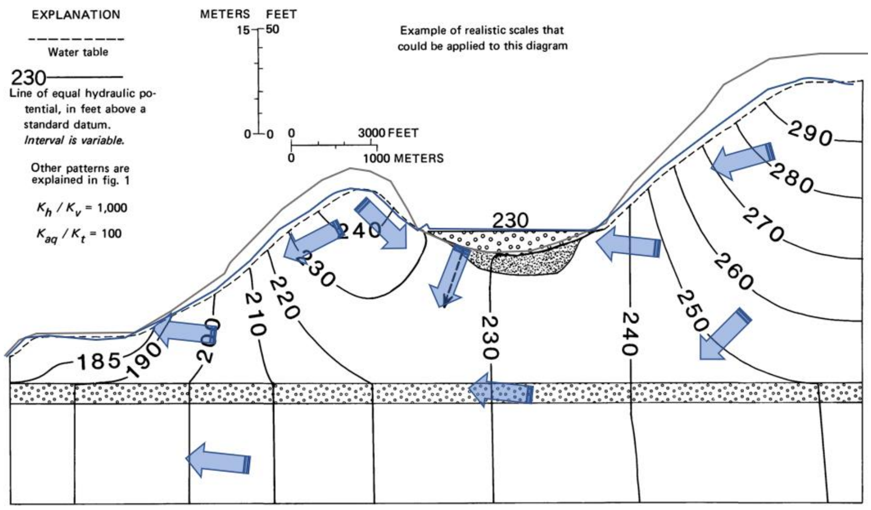


Figure Box 1-3 - Cross section of groundwater flow in a topography with hills (brown line represents land surface/lake bottom; blue line the water table) and a lake. The lake is receiving local groundwater and would be generally considered an effluent or gaining lake. However, the presence of a high hydraulic conductivity horizontal aquifer (stippled horizontal rectangle, K_{aq}) creates a flow system with no stagnation points. This set of conditions results in water being lost from the lake through the bottom of the lake (mixed lake setting). Flow lines are not represented as the system is anisotropic and vertically exaggerated by 80%. Blue arrows illustrate the general groundwater flow direction (modified from Winter, 1976).

Winter (1976) also examined a landscape cross section with three lakes and presented conceptual models showing how these lakes interface with the groundwater flow system (Figures Box 1-4 and 1-5). Note, when multiple flow systems are present, lakes can receive flows from local and intermediate flow systems. A lake located at the regional discharge point on the left side of the land scape (Figure Box 1-4) would receive recharge from local and intermediate flow systems as well as the regional flow system. Further review of this early work and Winter (1978) will provide the reader with insight into the complexity of lake-groundwater exchange.

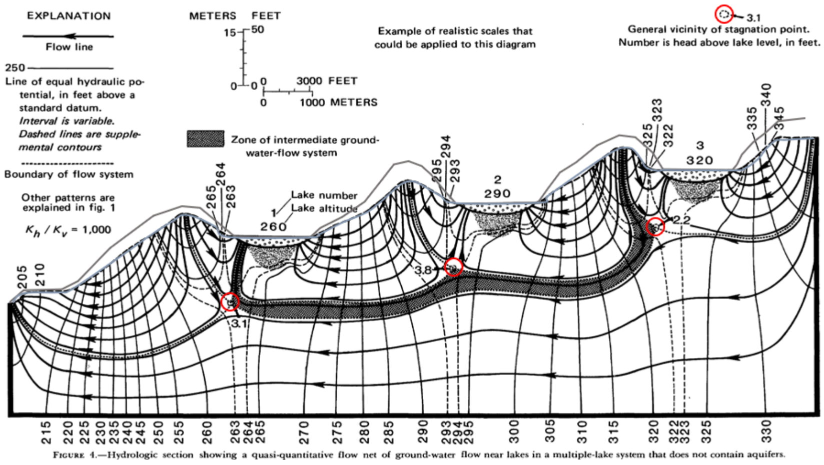


Figure Box 1-4 - Cross section of groundwater flow in a topography with hills (brown line represents land surface; blue line the water table) and lakes. Lakes 1, 2 and 3 are all effluent or gaining lakes. Lakes 2 and 3 are exchanging groundwater with local flow systems and lake 1 receives water from both local and intermediate flow systems. Black flow lines are schematic showing general directions of flow as the system is anisotropic and vertically exaggerated by 80% (modified from Winter, 1976).

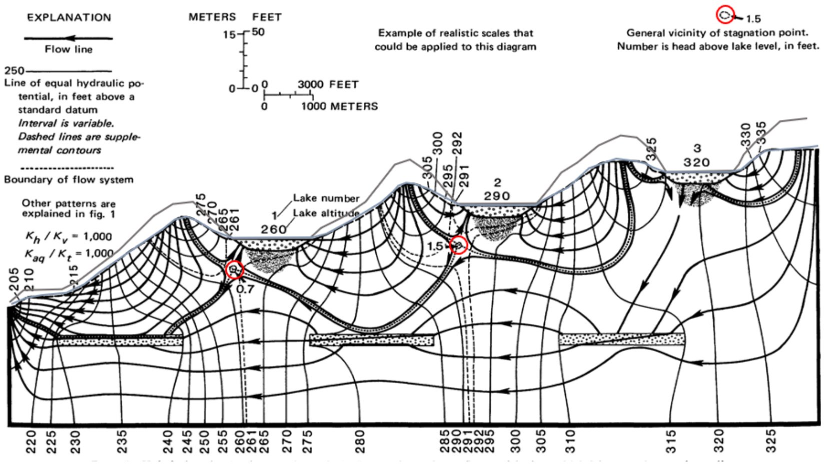


Figure Box 1-5 - Cross section of groundwater flow in a topography with hills (brown line represents land surface; blue line the water table) and lakes. Lakes 1 and 2 are illustrated as effluent or gaining lakes fed by local groundwater flow. Lake 3 receives local discharge and it is also losing water to the regional flow system (mixed system). Flow lines are schematic showing the general direction of flow as the system is anisotropic and vertically exaggerated by 80% (modified from Winter, 1976).

Return to where text links to Box 1

Box 2 Springs

In the most general terms, a spring forms where concentrated groundwater discharges at the land surface. The Groundwater Project book↗, "Groundwater in Our Water Cycle: Getting to Know Earth's Most Important Fresh Water Source' (Poeter et al., 2020) discusses the connection of groundwater with springs (pages 24-27). Springs have been utilized throughout history as water supplies for homes, cities, sources of agricultural water, and have provided the underpinning of unique ecological systems since the dawn of civilization. Springs can be present at the discharge areas of local, intermediate, and regional flow systems (Figure 11). When springs are associated with large scale groundwater systems the discharge, water temperatures, and quality remain fairly constant because the long flow paths and residence time of the groundwater moderates its condition before reaching the spring. In contrast, spring discharge, temperature, and quality are more variable when springs are connected to local flow systems with short flow paths and residence time. Understanding the hydrogeologic conditions that account for springs provides the hydrogeologist with insight needed to answer questions such as: If a large mine is being dewatered will nearby springs be impacted? Will a spring provide water needed to open a water bottling plant? Can impacted springs associated with the formation a wetland be remediated to enhance the discharge to the wetland? The answers to these questions are not directly provided in this material; however, conceptual models of the occurrence of springs in a variety of hydrogeologic settings are described. Springs occur under a number of hydrogeologic settings and general types include: depression springs, contact springs, joint and fracture springs, fault and shear zone springs, and karst springs (Figure Box 2-1). Sources of springs originate from unconfined and confined groundwater systems. Bryan (1919) provides a classification of springs and textbooks such as Fetter (2001) and Todd and Mays (2005) describe how springs fit into the hydrogeological landscape.

Depression springs are found in topographic lows where the water table intersects the land surface (Figure Box 2-1a). When springs occur on slopes, they are associated with a zone of discharging groundwater that forms a seepage face, a wet area above the spring pool where groundwater is seeping from the surface. The source of water at depression springs are usually local flow systems so they typically have variable discharge rates in response to local recharge, however, they may occur in regional topographic lows where large regional groundwater systems, both confined and unconfined, terminate.

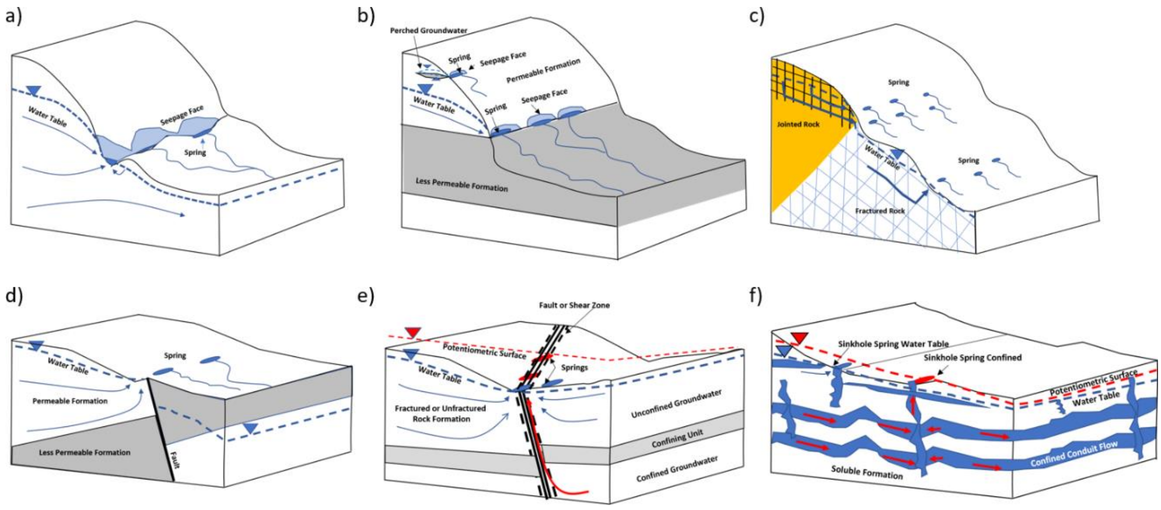


Figure Box 2-1 - Conceptual models of springs. Dashed blue line is the water table. Dashed red line is the potentiometric surface of confined groundwater. Solid blue arrows represent general groundwater flow. Red arrows represent source water from confined systems. Water table springs are blue ovals. Springs originating from confined groundwater are red ovals. a) Depression spring occurs where there is a topographic low in the landscape and the water table intersects the land surface. A seepage face (light blue) occurs above the spring pool as groundwater seeps from the slope. b) Contact springs form when a water table builds up above a lower permeable formation and then intersects the slope. Springs and seepage faces occur at the geologic contact. c) Joint and fracture springs are present when joints and fractures become saturated and the resulting water table intersects the slope. d) Fault springs occur when a fault disrupts a permeable formation and groundwater flow discharges along the fault trace. e) Shear or fault zone springs can form where complex fractured zones intersect the ground surface. Such zones can behave as either barriers to groundwater flow or provide a more permeable zone that concentrates groundwater discharge from shallow or deep groundwater systems. Springs form along the shear zone where the water table intersects the land surface and along the fault trace if deeper confined systems discharge to the sheared zone (which occurs when the potentiometric surface is above land surface). f) Karst springs form in formations that have undergone dissolution and include springs in areas with high water tables and surface solution features, as well as from deeper confined conduit systems when the potentiometric surface is above land surface (Woessner, 2020).

Contact springs occur when water percolating through rocks or sediments encounters a lower permeable material that underlies a more permeable formation (Figure Box 2-1b). If recharge is sufficient, a saturated zone builds up above the contact between the two formations. The water table in the permeable formation can be part of a completely saturated system or part of a perched system. A spring and seepage face occurs on the slope face between where the water table intersects the land surface and the geologic contact. These springs are typically generated by local flow systems.

Springs can also occur in saturated portions of jointed and fractured rocks (Figure Box 2-1c). Spring discharge in some fractured and jointed geological material occurs relatively near the surface as fractures receive local recharge, become saturated, and discharge is focused where features intersect the land surface. However, in some cases large units of jointed and fractured rocks act as equivalent porous media and larger groundwater flow systems develop and form springs in areas where joints or springs intersect the land surface.

Springs are also associated with faults and complex shear zones (Figure Box 2-1 d and e). Water table springs are formed when faulting results in juxtaposition of permeable material with less permeable material so that flow discharges as a spring at the fault interface where the water table is at land surface (Figure Box 2-1d). Faults and shear zones can be either less permeable, more permeable than the surrounding material or provide a combination of permeabilities (Figure Box 2-1e). When fault gouge is created, the fault has low permeability and inhibits groundwater flow. In some cases, the fault and its surrounding area are highly fractured, and the permeability is enhanced. These conditions tend to concentrate groundwater flow and discharge. When these fractured zones extend to depth, confined groundwater systems are often intercepted and when the zone is permeable, it provides an avenue for deeper groundwater to discharge at springs on the land surface. When the potentiometric surface is above the land surface, springs can occur along the fault trace. The water in springs receiving discharge from deep groundwater systems often has a fairly constant discharge and chemical composition, and water is warmer than local groundwater (hot springs occur when water temperatures are above 37°C).

When terrains are composed of soluble formations such as limestones and dolomites, karst springs can form where the water table or potentiometric surface exceeds the land surface elevation, creating depression springs (Figure Box 2-1f). When water levels in solutioned cavities and collapse features such as sinkholes occur in water table depressions, sinkhole springs are formed. Large karst springs can also occur at sinkholes or solution cavities that are connected to networks of saturated conduits in which heads are greater than the land surface elevation.

The water quality of springs varies with the character of the groundwater source. Springs supplied by short groundwater flow paths, like those of local flow systems, are impacted by seasonal changes in recharge, and water typically contains a low quantity of total dissolved solids. Springs receiving intermediate and regional flow have higher total dissolved solids and their quality is more consistent. Spring quality can also be impacted by the composition and solubility of the components of the earth materials that the discharging groundwater passed through. In some settings local flow systems can form springs with high total dissolved solids even though the flow path and water residence times are short.

Return to where text links to Box 2↑

Box 3 Mirror Lake

The USGS initiated studies of Mirror Lake located in a basin composed of glacial deposits and bedrock in New Hampshire, USA, in the late 1970's. They used both physical measurements of water budget components, chemical water budget analyses, and regional groundwater modeling to assess the water balance. Their work also attempted to examine uncertainties in budget parameters. Energy budget methods were used to compute evaporation using data collected on a floating raft and at a land station. Precipitation estimates were derived from two gauges located 400 m from the lake. Stage recorders were used with stream gauging to measure surface-water inflows and outflows. Lake stage measurements were used to compute changes in storage, and groundwater inflow and outflow were computed using Darcy's law with a measured hydraulic conductivity value, hydraulic gradients derived from well networks and lake stage measurements, and estimates of cross-sectional areas based on site geology and well logs.

Annual and monthly chemical and water budgets were computed for the period 1981 to 2000 (Table Box 3-1). The groundwater inflow and outflow components of the annual budget were identified as having the largest uncertainty. The single measured hydraulic conductivity value used in the Darcy's Law calculations generalized the complex nature of the geologic material adjacent to the lake boundaries. Groundwater inflow was estimated at 47,000 m^3/y using the single hydraulic conductivity value to represent the properties of the cross sections. A second method using geochemical modeling and analysis of isotopic ratios of oxygen-18 to oxygen-16 data yielded a groundwater inflow value of 95,000 m^3/y. A third approach used a groundwater basin numerical flow model to generate the groundwater inflow to the lake. Modeling estimated groundwater inflow was 133,000 m^3/y, about 2.8 times greater than the initial computed value (Tiedeman et al., 1997). The researchers recognized the uncertainty in the groundwater inflow term and settled on 113,000 m^3/y as the input estimate.

Applications of multiple methods to estimate groundwater inflow to the lake may shed light on the possible variability of poorly resolved water budget components. The relatively high degree of uncertainty in component measurements supports the need to employ additional dense spatial and temporal instrumentation, and data collection and analyses to reduce uncertainty. Healy and others (2007) estimated uncertainties of 5-10 percent for precipitation measurements, 10 to 15 percent for evaporation values, 5 to 10 percent for streamflow, and 30 to 50 percent for groundwater exchange. They computed an overall budget uncertainty of 13 percent. This study suggests that choosing a representative value for the water budget groundwater inflow and outflow components of a water budget requires an evaluation of uncertainty and professional judgement based on the quality and distribution of parameters used for the computations. Uncertainty analyses suggest additional data collection may be required to obtain a more representative water budget.

Table Box 3-1 - Annual Initial and Final (modeled) Water Budgets for Mirror Lake, New Hampshire, USA. Initial groundwater inflow is estimated using a single hydraulic conductivity value to calculate discharge to the lake. Values are in 100,000 m^3/y. Estimated percent uncertainty in parameters is also shown. The imbalance (residual) is the difference between inflow and outflow and represents the uncertainty (error) in the budget (modified from Healy et al., 2007).

	Initial	**Final**	**Estimated Uncertainty**
Inflows			
Precipitation	182	182	10-15%
Surface-water inflow	417	417	10-15%
Ground-water inflow	47	113	30-50%
Outflows			
Evapotranspiration	77	77	10-15%
Surface-water outflow	257	257	10-15%
Ground-water outflow	281	347	30-50%
Lake volume change	16	16	—
Imbalance	**15**	**15**	**13%**

Return to where text links to Box 3↥

Box 4 Measurement Errors in Synoptic Surveys

Standard stream gauging techniques introduce errors that range from 2% to 20% (Sauer and Meyer, 1992). A good reference for stream gauging methods is the USGS publication by Turnipseed and Sauer (2010). They suggest flow measurement errors usually average between 3 and 6%. When evaluating whether observed streamflow changes are meaningful, gains or losses must be outside of individual measurement error ranges. For example, if streamflow measurements contained an error of +/- 6%, and the upstream discharge measurement was 30 m^3/s (+/- 1.8 m^3/s) and the downstream measurement was 28 m^3/s (+/- 1.7 m^3/s), the net groundwater exchange computed by difference (assuming no other inputs or outputs within the reach) would be -2 m^3/s. If the 6% error is applied to the stream gauging measurements the upstream measurement could be as high as 31.8 m^3/s and the downstream measurement as low as 26.3 m^3/s. That means that for a measurable difference in streamflow to be observed the value obtained would have to exceed the difference of the discharges with the 6% error applied (31.8 m^3/s -26.3 m^3/s = 5.5 m^3/s). Unfortunately, the error ranges of inflow and outflow measurements did not allow differentiation of gain or loss when a difference of 2 m^3/s was observed. Depending on site conditions, increasing the distance separating the measurements may resolve the issue if sufficient additional streamflow gains/losses occur. Roberts and Warren (1999) noted that additional errors are introduced if flow conditions are changing during the period of measurement.

Return to where text links to Box 4↥

Box 5 Seepage Meter Operation

Though seepage meters are useful tools, careful attention must be paid to operational details including meter design, materials, sizing of tubing and bag composition. Applications sometimes require modifications of either the procedure or the meter itself. Rosenberry and others (2008, 2020) present a list of factors affecting meter data sets as shown here in Figure Box 5-1. Their work should be reviewed prior to building or installing meters.

Sources of error when using seepage meters include:

1. Incomplete seal between seepage-meter chamber and sediments, unstable cylinders;
2. Insufficient time between meter installation and first measurement;
3. Improper bag-attachment procedures, bag resistance, and moving water;
4. Leaks;
5. Measurement error;
6. Flexible seepage-meter chamber;
7. Insufficient or excessive bag-attachment time;
8. Accumulation of trapped gas;
9. Incorrect coefficient to relate measured flux to actual flux across the sediment-water interface; and
10. Insufficient characterization of spatial heterogeneity in seepage through sediments.

Figure Box 5-1 - A list of Sources of Error When Using Seepage Meters (Rosenberry et al., 2008)

In addition to Rosenberry and others (2008) summary of meter operation errors, Zamora (2008) describes experiments testing meter operation and describes the challenges of using seepage meters in a river setting. Rosenberry and Menheer (2006) developed a system for calibrating seepage meters. Rosenberry and others (2020) present an extensive table of best seepage meter installation and operations practices that should be reviewed before designing and operating meters in field settings.

Meter design includes the components shown in Figure 66. Metal and plastic materials are commonly used if they can be pushed into the bottom sediments to a sufficient depth so that the sides are buried and sealed without disturbing the sediments. Depending on the type of bottom sediments and difficulty or ease of "seating" the meter, the meter diameter and wall thickness design may need modification. When installing and seating the meter the sides may extend into the bottom a few centimeters to 10+ centimeters. The key is to prevent leakage around the meter walls. In all cases, an unobstructed water-filled space between the meter top and sediment interface is required. Meters are typically installed with a slight tilt so that any gasses collected during meter installation or operation will flow out of the meter. In shallow installations, a vent can be used to prevent gas build-up after installation (Figure 66).

Meter installation may disturb the flux conditions at a site. Once a meter is installed, multiple measurements may be required to assure that equilibrium conditions have been reestablished. Without prior knowledge of flux rates, this is accomplished by running the meter and comparing flux rates until some reproducibility of rates is established. The time interval between installation and equilibrium will depend on the site flux rates. Higher rates require shorter times (hours), while lower rates require longer times (days).

Seepage meters were originally designed for shallow lake and wetland studies, and studies of small streams and irrigation ditches with low flow velocities (Lee, 1977). When meters are used to characterize exchange in river systems with rapid streamflow, additional considerations are needed. Both Zamora (2008) and Rosenberry and others (2020) summarize the relevant literature. When the meter protrudes above the stream bed, it redirects stream water and, in some cases, stream water will flow under the meter sides and enter the meter. This generates false over-estimates of seepage (e.g., Zamora, 2008). Flux rates have also been reported to be impacted by the hydraulics of water flowing over and around the collection bag. Investigators have suggested installing shields to limit hydraulic effects (e.g., Libelo and MacIntyre, 1994; Murdoch and Kelly, 2003; Zamora, 2008; Kennedy et al., 2010) as shown in Figure 66. In some settings, it may be possible to extend the meter tubing and bag set up to a submerged nearby location out of the main current (e.g., near the bank or some other more protected zone). Care must be taken to keep the meter and the submerged bag locations at a stream location that has the same stage as the meter, because locating the bag downstream or upstream of a meter changes the gradient conditions controlling seepage (Rosenberry et al., 2020). Both the bag and tubing should remain submerged and secured to the bottom. When meters are installed and left in streams for future measurements, they may become exposed by scour or covered by sediments (Zamora, 2008). Such conditions may require meter removal and reinstallation. If meter installation is planned in stream systems, the seepage meter literature should be carefully reviewed (e.g., Rosenberry et al., 2008; Zamora, 2008; Rosenberry et al., 2020).

Russoniello and Michael (2015) report meter operation in large lakes and along coastlines can be compromised during periods of wave activity. In flume studies, they found small anomalous readings may occur when ratios of wavelength to water depth is less than five. When meters are deployed in these settings, the literature should be consulted to assure proper methods are used (Duque et al, 2020).

In all meters, the diameter and properties of the ports, tubing, valves and bags must not restrict or alter natural flux rates. Meter port opening, tubing, and valve designs must be of a sufficiently large diameter so that no resistance to flow from the meter to the bag occurs (e.g., Zamora, 2008; Rosenberry et al., 2020).

Over time, several modifications to the original bag design have been suggested. Generally, a light-weight, clear, plastic bag is used. It has been found that bags need to be prefilled with water then deflated to remove air prior to meter operation. Shaw and Prepas (1989) found that when a fully deflated (empty) bag is installed and the valve is opened, it expands slightly, drawing a volume of water into the bag that is not related to the natural seepage (referred to as bag memory). Moreover, deflated bags that are stiff (thick walled) resist filling, causing underestimation of seepage rates. Measurements starting with unfilled bags and connected tubes were found to cause inconsistent measurements (e.g., Shaw and Prepas, 1990a,b; Belanger and Montgomery, 1992; Murdoch and Kelly, 2003; Landon et al., 2001; Zamora, 2008; Rosenberry et al., 2020). Generally, depending on the bag size, bags are prefilled with water, 100 to 1000 ml, and the air squeezed out. Harvey and Lee (2000) reported bags should not be overfilled and kinks should be avoided. Isirho and Meyer (1999) examined how meter size and bag type impacted measurements in the laboratory. They concluded bag type had minimal effects on seepage rates. However, Zamora (2008) conducted a series of laboratory experiments and reported seepage values were more repeatable when connected tubes were full of water before opening valves, bags were prefilled with a volume of water and bags were thin walled (0.04 mm).

Return to where text links to Box 5↥

Box 6 Application of Mini-Piezometers, Seepage Meters, and Temperature Contrasts

The United States Geological Survey assessed the exchange in a 3-mi (4.8 km) section of Chapman Creek that passes by a landfill near Tremount City, Clark County, west-central Ohio, USA (Dumouchelle, 2001) as shown in Figure Box 6-1. The study's purpose was to assess how the stream water and groundwater interacted in the vicinity of the landfill site. The stream reach was instrumented with four piezometers, two seepage meters, five stream/streambed temperature monitors, and a seepage run was conducted by measuring the stream and tributary flows during a period of low flow.

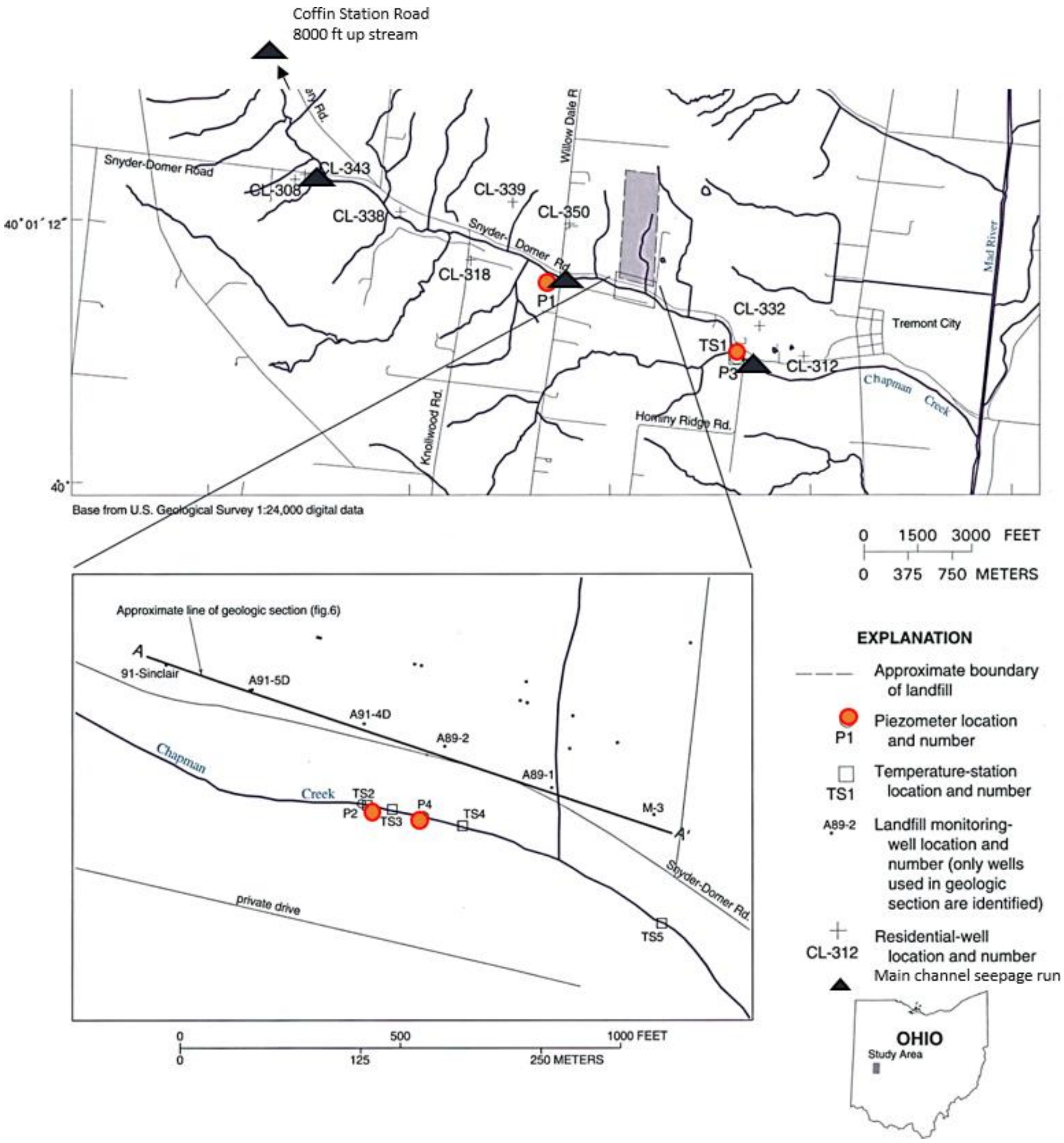


Figure Box 6-1 - Location map for exchange study of Chapman Creek, Ohio. The stream flows from left to right. Seepage meters were installed near P2 and P3. Approximate locations of seepage run gauging sites are shown with triangles (modified from Dumouchelle, 2001).

The 2-foot (0.6 m) deep creek flows through glaciated terrain and the stream bed is composed of fine sand to large cobbles and boulders. The four piezometers were constructed of 1.25-inch (3 cm) diameter pipe with 12 to 18 inches (30 to 46 cm) of wire-wrapped screen, tipped with a drive point (sandpoint). The piezometers were driven into the stream bed so that the top of the screened interval was 2.5 to 3 feet (0.8-0.9 m) below the stream bottom. Water levels were measured (+/- 0.02 feet (0.6 cm)) with an electric tape or chalked steel tape. Seepage meters were constructed as described by Lee (1977) using a 55-galllon/208-liter drum top and a partially pre-filled plastic collection bag. Temporary piezometers were installed adjacent to the seepage meters. Sites with soft clay and little or no gravel were selected for installation. Hourly temperature monitoring of the stream and streambed water was conducted (as discussed in Section 5.8 of this book). Waterproof-cased sensors (+/- 0.7 °C) were installed at sand/cobble sites where a steel fence post was driven into the stream bottom. One sensor was attached to the post just above the stream bottom and the second was buried in the stream bottom sediment about 6 to 8 inches (18-20 cm). A seepage run was conducted in October 2000 at low flow (seepage runs are discussed in Section 5.4 of this book). Discharge measurements were conducted on Chapman Creek and six tributaries using standard methods.

Shallow piezometer installation data (eight measurements) showed heads in piezometers were lower than the stream stage at P1, P2 and P3. The P4 data suggest bed water was moving upward and into the stream for three of the four measurements (Table Box 6-1). Seepage results (two measurements) for the seepage meter located near P2 showed a loss of water from prefilled bags supporting the interpretation of a losing stream section (head difference data from P2). However, seepage meter results near P3 suggested water was flowing into the meter, a contradiction to the head difference data observed at P3 (loss) (Table Box 6-2). Later analysis suggests the meter was located in coarse sediments and may not have been fully seated in the bed. Possibly, surface water flowed under the meter sides and into the seepage meter. Temperature contrast data sets indicated stream bed temperatures were similar to stream water temperatures, and daily highs and lows of stream water were slightly lagged compared to stream bed temperature (Table Box 6-3). These data are characteristic of locations where surface water is infiltrating into the stream bed (losing stream). These data are also generally supported by the piezometer data and seepage meter results at P2. The seepage run study showed that the stream was gaining between the upstream and downstream gauging locations (~3 mi, 4.8 km) as shown in Table Box 6-4.

Table Box 6-1 - Results of piezometer and stream stage head measurements at four piezometer locations shown in Figure Box 6-1. [Piezo, piezometer; P#, piezometer number; ft, feet; bmp, below measuring point; ---, no data; negative differences in water levels indicates flow into the aquifer, positive differences indicated flow into the stream] (Dumouchelle, 2001).

	P1			P2			P3			P4		
Date 2000	Piezo water level (ft, bmp)	Stream level (ft, bmp)	Water level difference	Piezo water level (ft, bmp)	Stream level ft, bmp)	Water level difference	Piezo water level (ft, bmp)	Stream level (ft, bmp)	Water level difference	Piezo water level (ft, bmp)	Stream level (ft, bmp)	Water level difference
9-08	2.32	2.24	-0.08	---	---	---	2.35	2.27	-0.08	---	---	---
9-13	---	---	---	3.13	3.11	-0.02	2.33	2.23	-0.10	---	---	---
9-27	2.27	2.20	-0.07	3.14	3.09	-0.05	2.32	2.24	-0.08	---	---	---
10-4	2.28	2.23	-0.05	3.15	3.11	-0.04	2.34	2.24	-0.10	---	---	---
10-12	2.31	2.28	-0.03	3.10	3.03	-0.07	2.27	2.19	-0.08	1.93	2.05	0.12
10-20	2.30	2.27	-0.03	3.08	3.02	-0.06	2.27	2.17	-0.10	1.96	2.08	0.12
10-25	2.31	2.28	-0.03	3.08	3.00	-0.08	2.29	2.22	-0.07	1.96	1.90	-0.06
11-2	2.31	2.26	-0.05	3.05	3.02	-0.03	2.31	2.22	-0.09	1.92	2.08	0.16

Table Box 6-2 - Results of seepage at seepage meter sites near P2 and P3 as shown in Figure Box 6-1. P2, piezometer; mL, milliliters; min, minutes (Dumouchelle, 2001).

Test location	Initial volume (mL)	Ending volume (mL)	Change in volume (mL)	Time (min)	Difference in temporary piezometer-stream levels (feet)
Near P2	200	121	-79	98	-0.04
Near P2	200	124	-76	102	-0.04
Near P3	200	334	134	110	-0.11
Near P3	200	406	206	90	-0.02

Table Box 6-3 - Results of paired temperature monitors as located in Figure Box 6-1 (Dumouchelle, 2002).

Station	Dates of collection (2000)	Tree cover	Streambed composition	Stream depth[1] (in.)	Average temperature[2] (°C)	
					Streambed	Stream water
1	Oct 20-Dec 5	Mostly shaded	Sand	6	8	7.9
2	Oct 20-Nov 2	Unshaded	Fine sand, cobbles	24	12.9	12.5
3	Oct 20-Nov 3	Unshaded	Cobbles, some gravel	6	13.2	13.1
4	Oct 20-Nov 4	Partially Shaded	Cobbles, gravel	9	12.9	12.8
5	Oct 20-Nov 5	Shaded	Sand	9	13.1	12.9

[1]Depth is approximate, based on conditions of October 20, 2000.
[2]Average temperature listed is the average of the entire period of record.

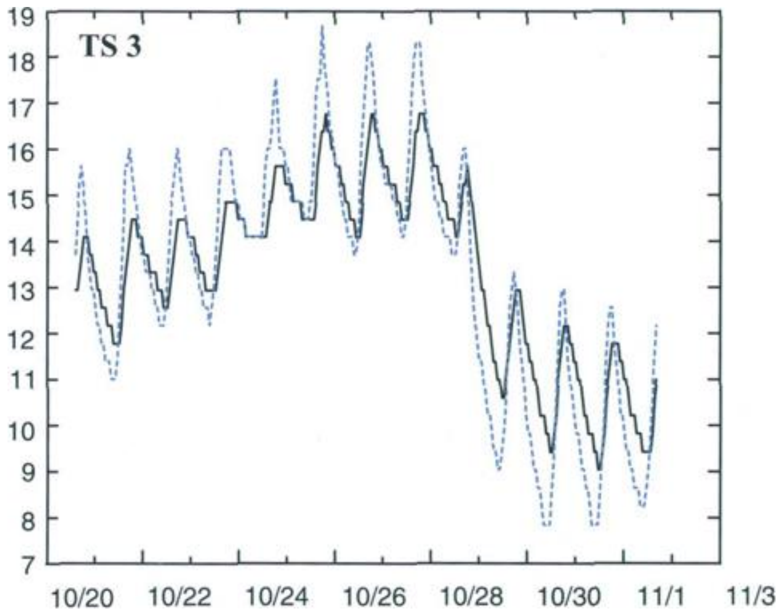


Figure Box 6-2 - The graph (TS3) is an example of a plot of temperature (℃) versus time for hourly measurements over 14 days as shown in Figure Box 6-1. The dashed blue line is the surface-water temperature and the solid line is the groundwater temperature. in, inches; °C, degrees Celsius (Dumouchelle, 2001).

Table Box 6-4 - Results of the seepage run. Locations of steam measuring points are shown on Figure Box 6-1. [trib., tributary; W, wading (current meter) streamflow measurement; V, volumetric streamflow measurement; F, fair (+/-8 percent); G, good (+/-5 percent); E, excellent, (+/-2 percent); ft^3/s, cubic feet per second; ---, not applicable] (Dumouchelle, 2001).

Site Name	Method	Quality rating	Streamflow (ft^3/s)	Change in main-stem flow (ft^3/s)
Chapman Creek at Coffin Station Road	W	F	0.89	---
Unnamed trib., At northwest intersection of Terre Haute, Thackery, and Snyder-Domer Roads	V	G	0.008	---
Unnamed trib., right bank, between Coffin Station and Terre Haute Roads	V	F	0.007	---
Chapman Creek at Snyder-Domer Road	W	F	1.55	0.65
Unnamed trib., left bank, upstream from Knollwood Road	V	G	0.012	---
Unnamed trib., right bank, downstream from Knollwood Road	V	E	0.004	---
Unnamed trib., left bank, upstream from Willow Dale Road	V	E	0.005	---
Chapman Creek at Willow Dale Road	W	F	2.05	0.48
Unnamed trib., left bank, upstream from Hominy Ridge Road	V	F	0.007	---
Chapman Creek at Hominy Ridge Road	W	G-F	2.46	0.40

The author noted that the shallow installations of the instream instruments most likely reflected the circulation of stream water into the bed sediments (hyporheic exchange), conditions at this study site that created mostly losing conditions

(Figure Box 6-2). However, the seepage run data support an interpretation that the reach is an overall gaining stream.

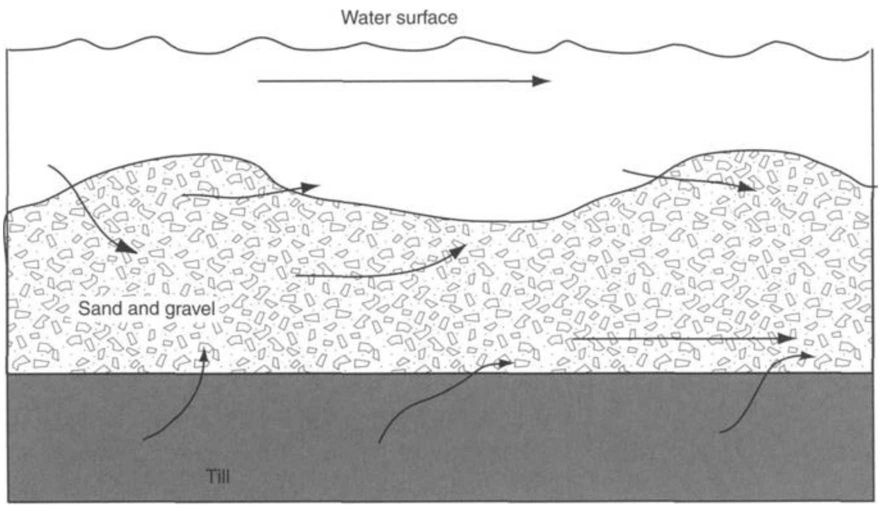


Figure Box 6-2 - Conceptual model suggesting why shallow instrumentation measured losing conditions where stream water infiltrated into the unconsolidated sediments. However, the overall gaining conditions (supported by the seepage run) suggest groundwater is discharging to the stream by upward flow from the underlying till (Dumouchelle, 2001).

The author's conceptual model of the groundwater exchange did not account for the different roles the coarse-grained layer and finer-grained layer played in the exchange process. Mini-piezometers nested in both the sand and gravel hyporheic system and the deeper till were needed to better define exchanges at local and in larger areas. The use of multiple methods to examine the exchange process is recommended.

Return to where text links to Box 6

Box 7 Heat Transport Modeling

Analytical and numerical models are used to estimate flux rates from observed temperatures (e.g. Stallman, 1965; Goto et al., 2005; Hatch et al., 2006; Keery et al., 2007; Schmidt et al., 2007; Lien and Ford, 2014). Modeling requires assigning sediment thermal properties, estimates of porosity, measurements of VHG in the porous media, and assignment of thermal boundary conditions. The model calculates groundwater flux and groundwater velocities based on field-observed heat transport data sets. Stonestrom and Constantz (2004), Anderson (2005), and Constantz (2008) present tables of sediment thermal properties. VS2DH↗ is widely used for such modeling. It is a public-domain, numerical, one-dimensional, energy transport and fluid flow simulator that can evaluate variably saturated flow and solute transport (Healy and Ronan, 1996). Other codes used for this purpose include Ex-Stream (Swanson and Cardenas, 2011) which uses MATLAB (Mathworks↗, Inc., and VFLUX↗ (Gordon et al,. 2012). All three of these models simulate one-dimensional vertical flow. Two- and three-dimensional transport certainly occurs in some settings and the application of a simple one-dimensional model may be inappropriate (e.g., Cuthbert and Mckay, 2013).

Return to where text links to Box 7↥

Box 8 Stream Tracer Breakthrough Models

Cardenas (2015) and Boana and others (2014) offer a comprehensive review of conceptual models and methods to evaluate stream tracer test results. Boana and others (2014) present governing equations representing four models to assess tracer breakthrough data: advection-dispersion and fractional advection-dispersion equation, the space-time fractional advection-dispersion equation, fractional spatial-derivative advection-dispersion equation, and the fractional temporal-derivative advection-dispersion equation. Most approaches attempt to fit model parameters to reproduce the behavior of observed tracer breakthrough. These results are used to describe the magnitude of exchange between the stream and groundwater.

Often modeling is simplified to represent tracer transport in a one-dimensional setting with a storage mechanism (e.g., Hays, 1966; Thackston and Schnelle, 1970; Valentine and Wood, 1979; Bencala and Walters, 1983; Jackman et al., 1984; Kim et al., 1992; Wörman, 1998; Bencala et al., 2011). The basic One-Dimensional Transport with Inflow and Storage (OTIS↗) code developed by the United States Geological Survey is widely used and usually applied before more sophisticated modeling is attempted (e.g., Runkel and Chapra, 1993; Runkel et al., 1998). A limitation is that OTIS allows the adjustment of only a few parameters, whereas the transport process is generally more complicated (e.g., Choi et al. 2000; Runkel, 2002; Bencala et al., 2011). Improvements to this code incorporate additional field characterization to include other components in the exchange analyses (e.g., Haggerty et al., 2000; O'Connor et al., 2010; Worman et al., 2002; Boano et al., 2007). The STAMMTL↗ model adds multi-rate mass transfer to transport models (Haggerty and Reeves, 2002). The Solute Transport in Rivers Model↗ includes the designation of separate storage locations and the timing of hyporheic exchanges (Marion et al., 2008). Refer to Boano and others (2014) for a more complete discussion of tracer modeling methods.

As analysis techniques have advanced, new models account for varying exchange time scales (Haggerty et al., 2002; Gooseff et al., 2003ab) and multiple storage mechanisms (Harey and Fuller, 1998; Haggerty et al., 2009). Some researchers (e.g., Haggerty et al., 2009; Liao and Cipka, 2011; Liao et al., 2013) have developed methods to look at geochemical changes in water cycled in the hyporheic zone. Transient changes of stream channel depths and vegetation have been noted to change stream retention conditions and studies may need to account for changes in physical conditions of the stream (e.g., Harvey et al., 2003).

Return to where text links to Box 8↥

Box 9 Mixing Model Used to Separate Stream Base Flow

Kish and others (2010) examined the use of a geochemical mass balance approach to determine the portion of groundwater contributing to the Hillsborough River in west-central Florida, USA. They used both a geochemical mass balance (GMB) approach that accounted for flows, and anions, cations, and stable isotopes, and a conductivity mass balance (CMB) (Stewart et al., 2007) that relied on continuous and spot measurements of specific conductance. Both streamflow and groundwater water were dominated by similar major ions, a calcium-bicarbonate water type. Stream water became diluted by precipitation and runoff during high flow events. The lack of continuous ion sampling during runoff events and high discharges limited the GMB application to "low-intensity" runoff events. The continuous data sets collected as part of the CMB approach proved more useful in deciphering base flow. Results showed that the average proportion of groundwater (base flow) in stream runoff ranged from 31 to 100 percent (Table Box 8-1).

Table Box 9-1 - Base Flow Analyses for Sites SR39 and HRSP Showing Balance Derived Estimates of Base Flow Using GMB and CMB Approaches, Hillsborough River, Florida, USA (Kish et al., 2010).

Site	Storm event	Percentage of base flow in peak and average discharge determined by CMB and GMB methods							
		Determined from continuous specific conductance		Determined from sampled specific conductance		Determined from calcium concentration		Determined from magnesium concentration	
		Peak	Average	Peak	Average	Peak	Average	Peak	Average
SR39	September 2003	100	100	100	100	94.6	90.9	52.5	45.7
SR39	March 2004	4.0	92.4	93.6	93.7	100	98.7	65.2	85.3
SR39	August-November 2004	<10.0	31.9	nd	nd	nd	nd	nd	nd
SR39	January 2005	100	100	100	100	100	100	100	100
SR39	February 2005	100	100	100	100	100	100	100	98.6
HRSP	July 2004	50.8	87.9	nd	nd	nd	nd	nd	nd
HRSP	September-November 2004	<10.0	36.4	nd	nd	nd	nd	nd	nd
HRSP	January 2005	100	100	100	100	100	100	100	100
HRSP	February 2005	100	100	100	100	51.8	92.0	36.5	89.1
HRSP	September 2009	nd	nd	<10.0	44.7	nd	nd	nd	nd

Return to where text links to Box 9

10 Exercise Solutions

Solution to Exercise 1

1) The conceptual models presented in this book keep conditions simple by using a constant value for hydraulic conductivity, cross sections aligned with groundwater flow lines and cross sections plotted without vertical exaggeration. Examine Figure P1 below. The cross section in Figure Exercise 1a, is constructed parallel to flow, as indicated by red line (A-R-B). Under the conditions illustrated in Figure Exercise 1, think about the consequences of using head data to interpret the flow field from a cross section constructed at right angles to the stream as indicated by the line C-R-D. To make the comparison use A-R and C-R. Explain why the cross section along line A-R correctly represents horizontal and vertical groundwater flow, but head and flow data in a cross section along C-R does not appropriately represent flow conditions.

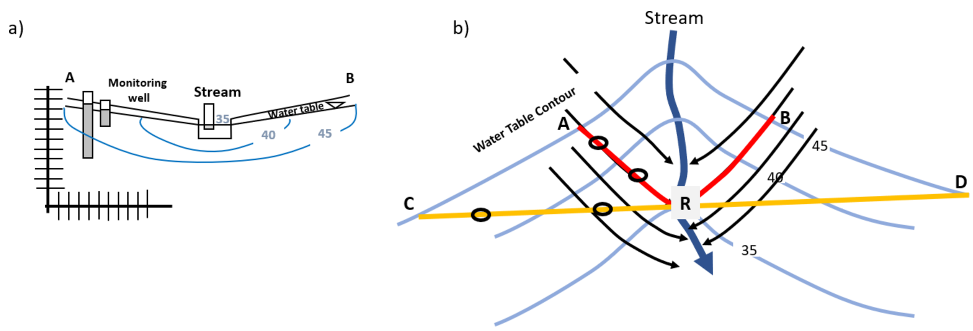
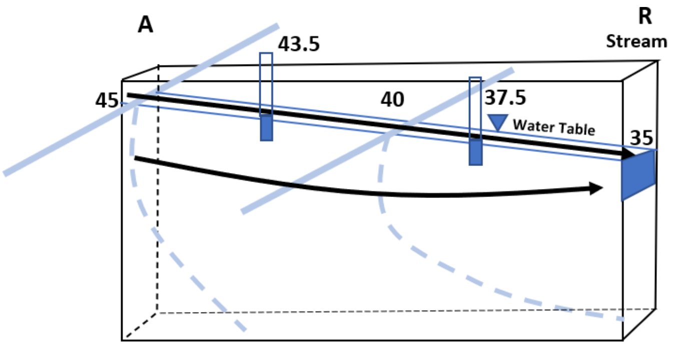
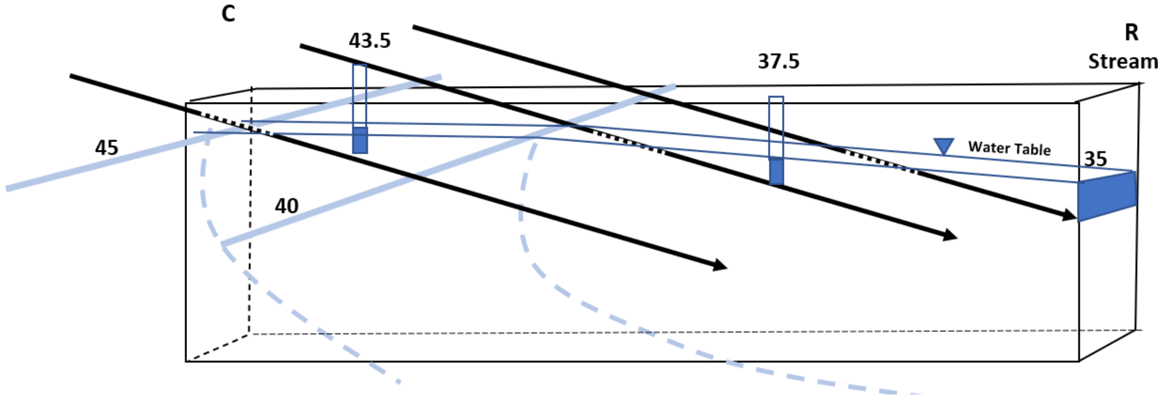


Figure Exercise 1 - Cross section and map view of a stream. Hydrogeologic conditions are isotropic and homogeneous and the groundwater conditions are effluent. The black ovals represent the location of monitoring wells on the cross sections. R is located where the cross sections intersect the stream. a) The cross section has no vertical exaggeration. The section is constructed along groundwater flowlines (red lines A-R and B-R) shown in (b). The left and right boundaries of the cross section coincide with the head contour of 45 (after Healy et al., 2007). b) Map view of flow field for the effluent stream. Water table contours are in blue and groundwater flowlines in black. The red line, A-R-B, shows the location of the cross section in (a). The orange line, C-R-D, is a cross section constructed at right angles to the river (Woessner, 2020).

SOLUTION:

Construction of the cross section along a flow line (A-R of Figure Exercise 1) results in the flow of groundwater within and parallel to the cross section. The cross section is a slice of the groundwater system and for the head and equipotential lines to be accurately represented the flow must be in the plane of the cross section (see A-R below). Thus, under isotropic and homogeneous conditions, flow lines plotted on the map and cross section will cross the equipotential lines at right angles. If flow conditions are anisotropic, they will cross the equipotential lines at the appropriate angle as discussed in the Groundwater Project book titled Hydrologic Properties of Earth Materials and Principles of Groundwater Flow↗.





Cross section A-R is plotted parallel to the flowlines so flow within the cross section is at right angles to the equipotential lines and head values at monitoring wells are evenly spaced.

Cross section C-R is plotted at a right angle to the stream, so it crosses flowlines which means that flow moves in and out of the section (dashed portions of black lines in the cross section). Also, the head change from 45 to 35 occurs over a longer distance than in A-R so the head distribution (lower gradient) is not representative of a section constructed parallel to flow. Since the

flow is not parallel to the cross section (i.e., along line C-R) the distribution of head and the position of the flow lines will not be properly represented. Using the head values obtained along this section will yield gradients that are not representative of flow along a flowline

It some cases it may be appropriate to place wells at right angles to streams to compute gradients and interpret vertical flow conditions. However, if groundwater flow to and from the stream channel occurs at some angle to the stream channel additional monitoring points are needed so that a cross section along a flow line can be constructed. When these conditions are met, vertical and horizontal flow conditions within a cross section will be correctly represented.

Return to Exercise 1↥

Solution to Exercise 2

2) Examine the scaled cross section of Figure Exercise 2 showing two surface-water bodies (solid blue: rivers, lakes or wetlands) in an isotropic and homogeneous setting. Equipotential lines (dashed) were constructed from a monitoring well network containing nests of wells (vertical black lines) open only at the bottom. The water table is indicated by a solid blue line. Equipotential lines represent intervals of 5 (any units can be used).

a) Construct flowlines.

b) Label local, intermediate and regional flow systems if they are present.

c) Identify where stagnation points form and circle the locations on the cross section if they are present.

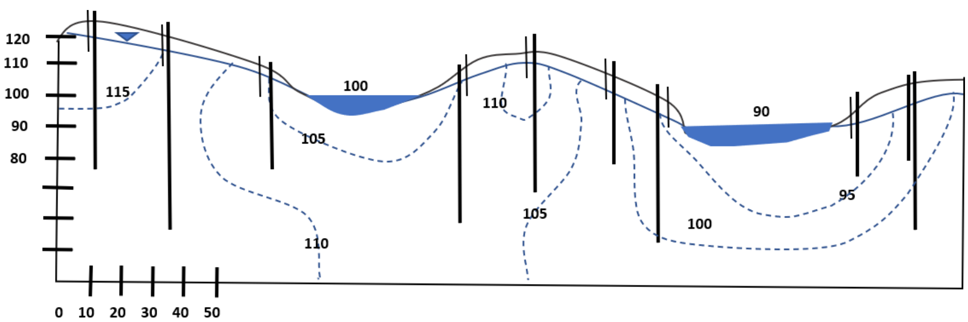
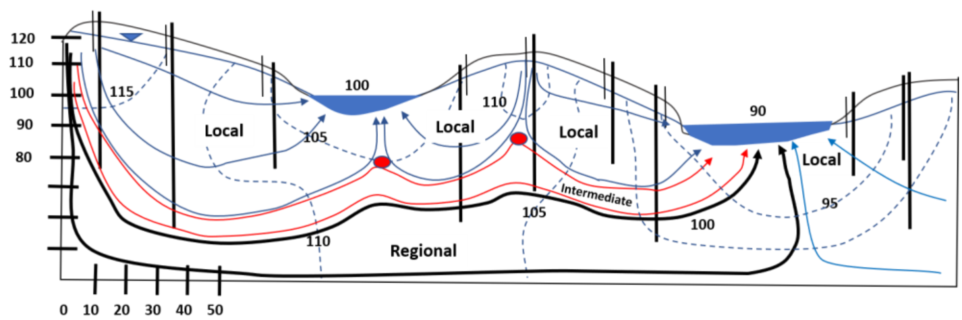


Figure Exercise 2 - Scaled cross section of a hydrologically isotropic and homogeneous system with two water bodies. Elevations are shown on the vertical axis and the stages of the water bodies are relative values. Black vertical lines represent monitoring well locations where head data were collected. Wells are open only at the bottom. Dashed lines are equipotential lines in 5-unit intervals (Woessner, 2020).

SOLUTION:

a) Flow lines (blue arrows, local system; red arrows intermediate system; black arrows, regional system) are constructed at right angles to equipotential lines. A large number of flow lines group in the recharge area of the upper left side.



b) Flow systems are labeled. Local flow systems (blue arrows) originate in recharge areas (water table highs) and discharge at adjacent discharge areas (surface-water bodies in this case). The one intermediate system (red arrows) originates at a recharge area (left) and discharges to a discharge area not immediately adjacent to the recharge area. It incorporates at least one local flow system. It discharges to the surface-water feature with a stage of 90. The regional flow system (black arrows) starts at the left and flows underneath the local and intermediate flow systems discharging to the surface-water feature with a stage of 90.

c) Stagnation zones (red dots) occur at groundwater divides where flow systems converge (local and intermediate). Winter (1976) describes a stagnation point as "… a point on the divide (between flow systems) at which the head is a minimum compared to every other point along the divide." … "It is a point in the flow field at which vectors of flow are equal in opposite directions and therefore cancel. … The stagnation point is a point of diversion of groundwater-flow paths." In this system the stagnation points occur on the flow divide where two local and the intermediate system converge.

Return to Exercise 2↥

Solution to Exercise 3

3) An exchange study of a pond was conducted by installing three seepage meters (A, B, C) and three adjacent mini-piezometers (small black open circles) as shown in Figure Exercise 3. Using the data provided compute the following:

a) The vertical hydraulic conductivity of the bed sediments at location A.

b) The VHG for B.

c) The seepage flux for C.

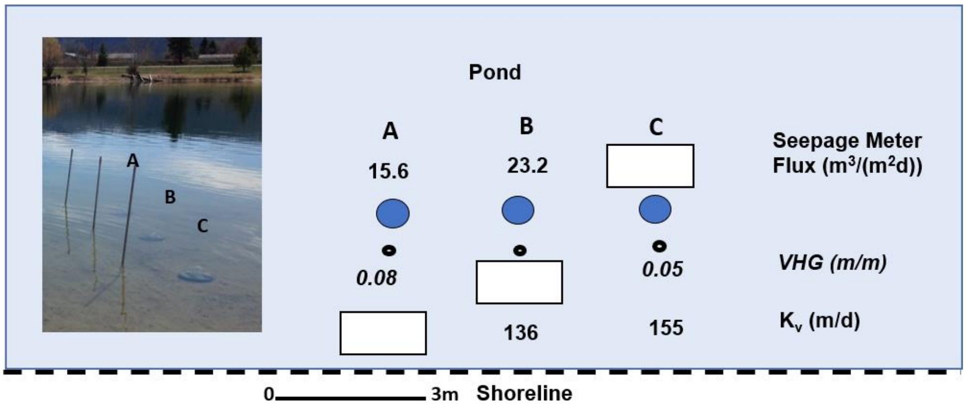


Figure Exercise 3 - An exchange study at a pond constructed in a sand and gravel aquifer. Three seepage meters, A, B and C (blue drums in the picture and blue circles in the map view), were installed about three meters from the shoreline. Adjacent to the meters, mini-piezometers (vertical tubes in the picture and black open circles in the map) were driven about 15 cm into the sand and gravel sediments of the bed. Seepage rates, VHG, and estimates of the bed vertical hydraulic conductivity were obtained for each of the three seepage meter locations (Woessner, 2020).

SOLUTION:

The seepage meters directly measure groundwater seepage, flux (q). Using Darcy's law, the relation among seepage, vertical gradient and vertical hydraulic conductivity is as follows:

$q = Q/A =- K_v \, Dh/DL_v = -K_v(VHG)$

where:

K_v is the vertical hydraulic conductivity (L/T)

Δh is the change in head in the direction of groundwater flow, lowest head minus the highest head, in this case (pond surface elevation) - (groundwater elevation) (L)

ΔL_v is the difference between the pond bottom and the open end of the mini-piezometer (L)

VHG is the vertical hydraulic gradient = $-\Delta h/\Delta L_v$ and is input as a negative gradient when used with Darcy's Law (dimensionless)

The sign of the *VHG* is used to indicate if groundwater is flowing into the pond (positive) or pond water is flowing into the groundwater (negative).

a) The vertical hydraulic conductivity of the bed sediments for A is computed as

$$K_v = -q/VHG = -15.6\ m^3/(m^2d)/-0.08 = 195.0\ m/d$$

b) The corresponding *VHG* for B is

$$VHG = q/K_v = 23.2\ m^3/(m^2d)/136\ m/d = 0.17\ m/m$$

Given that no field measurements of head difference are provided, the *VHG* is positive based on other data given for B. The seepage meter flux is positive so an upward *VHG* (positive) must be occurring.

d) The seepage flux for C is

$$q = \_ -K_v VHG = -155\ m/d\ (-0.05) = 7.8\ m^3/(m^2d)$$

Once again, the *VHG* is input as a negative hydraulic gradient and the flux, q, is positive. The flux at this location is upward from the groundwater to the pond.

Return to Exercise 3↥

Solution to Exercise 4

4) A water budget for a shallow subtropical lake found surface-water drainage canal discharges (34%), direct precipitation (24%) and groundwater discharge (14%) dominated inflows. Water left the lake by sheet flow into a swamp (65%), evaporation (34%), and groundwater outflow (1%). Nitrogen loading to the lake was also a concern and the lake was classified as dystropic, (chronic hypoxia and high concentrations of unionized ammonia). Groundwater was computed to deliver 48% of the load. Restoration efforts were initiated in 1996 and included removal of organic sediment, dredging and planting of native vegetation. Additional nitrogen reduction restoration is needed. The lake is in a depression created by limestone dissolution. A seepage study of the lake was conducted using twenty 208-liter seepage meters and mini-piezometers installed next to each meter (Figure Exercise 4). Meters were installed and seepage rates measured every 14 days from October through mid-May. Meters 1-14 were located in the littoral zone and 15-20 were in open water. Meters equilibrated for three months before study operation. Bags were prefilled with 1000 ml of deionized water for each seepage event. Mini-piezometers were placed in holes excavated into the bed sediment and bedrock and backfilled with sand and bentonite. Water quality samples were taken from the mini-piezometers. Mean seepage rates and total nitrogen concentrations were computed for each of the twenty sites (Table Exercise 4).

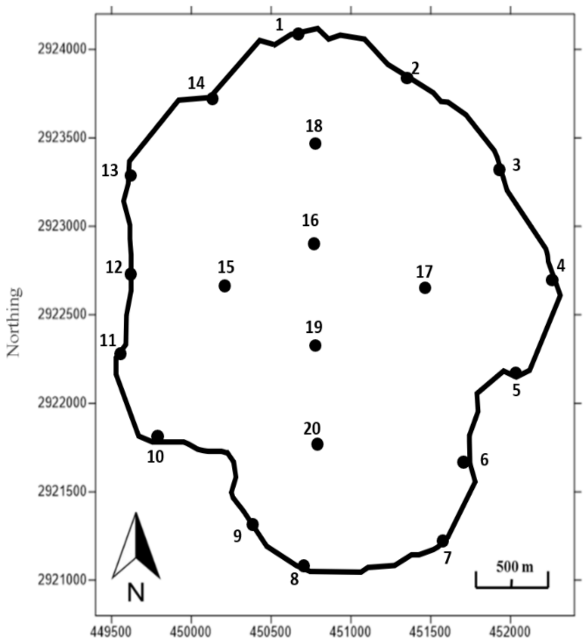


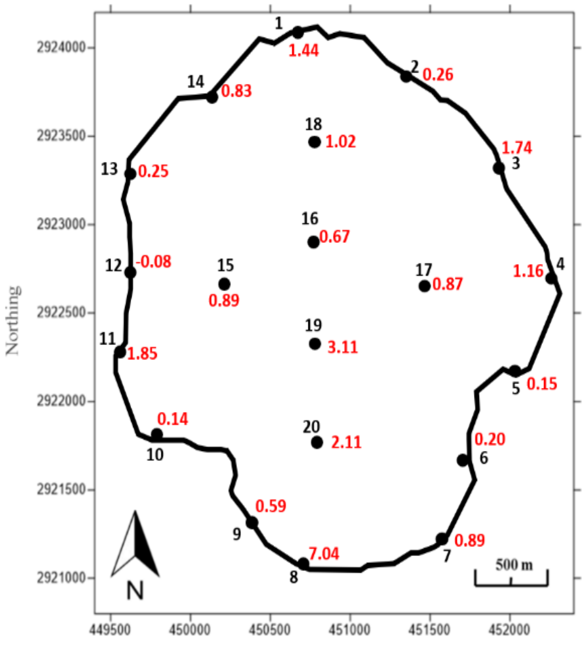
Figure Exercise 4 - Location map of seepage meter and mini-piezometer installations (black dots and numbers) in a lake (modified from Lucius, 2016).

Table Exercise 4 - Mean seepage rates & total nitrogen concentrations (Lucius, 2016).

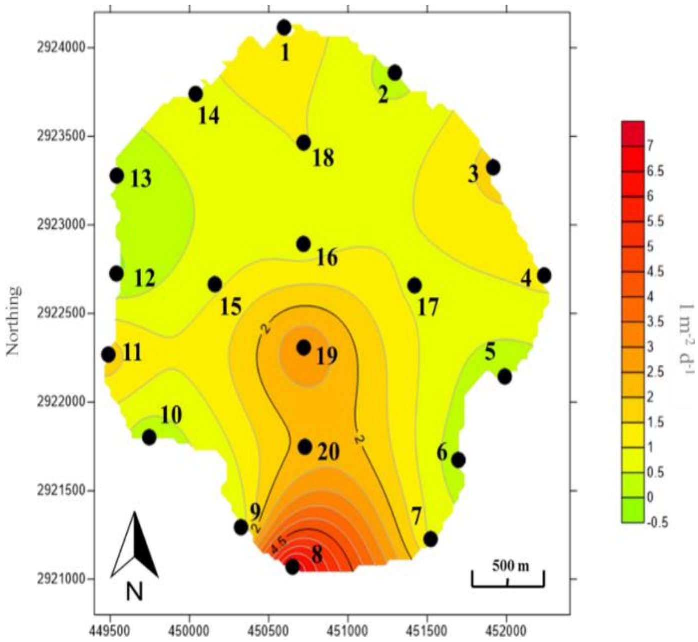
Site	Northing	Easting	Mean Seepage (L/m^2/d)	Mean Total Nitrogen (mg/l)
1	450593.39	2924113.61	1.44	11.84
2	451291.49	2923858.38	0.26	17.7
3	451916.45	2923324.12	1.74	9.06
4	452238.73	2922714.82	1.16	4.38
5	451989.45	2922144.28	0.15	20.32
6	451695.94	2921673.69	0.2	17.1
7	451520.61	2921227.23	0.89	4.55
8	450647.05	2921081.35	7.04	5.45
9	450321.19	2921294.35	0.59	5.64
10	449740.75	2921801.64	0.14	13.02
11	449481.89	2922268.64	1.85	4.16
12	449533.30	2922725.53	-0.08	6.01
13	449535.09	2923278.14	0.25	4.74
14	450033.85	2923738.93	0.83	6.49
15	450155.54	2922664.82	0.89	52.28
16	450716.41	2922891.76	0.67	28.53
17	451418.80	2922658.34	0.87	24.27
18	450717.27	2923464.25	1.02	11.24
19	450717.90	2922308.40	3.11	23.78
20	450725.32	2921747.50	2.11	6.46

SOLUTION:

a) Plot the mean seepage data on the location map and contour it using an interval of 0.5 L/(m^{2}d). Where is the seepage concentrated? Is water seeping from the lake to the groundwater? Would you classify the lake as effluent, influent, flow-through, or mixed? Support your answer.



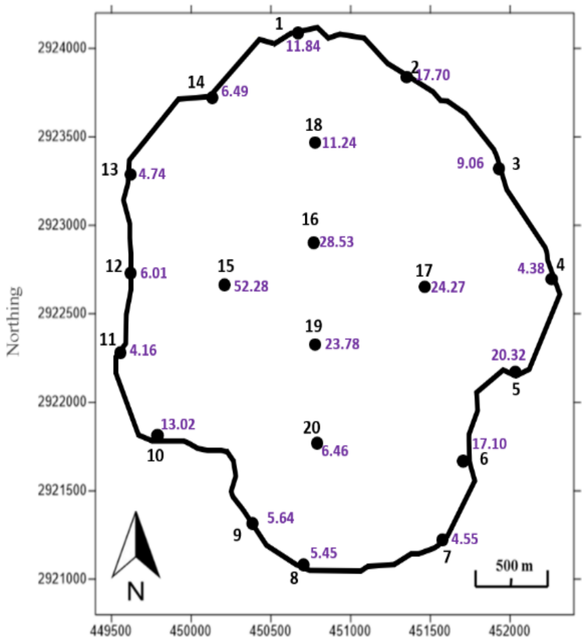
Plot of mean seepage values on base map L/m^{2}d.



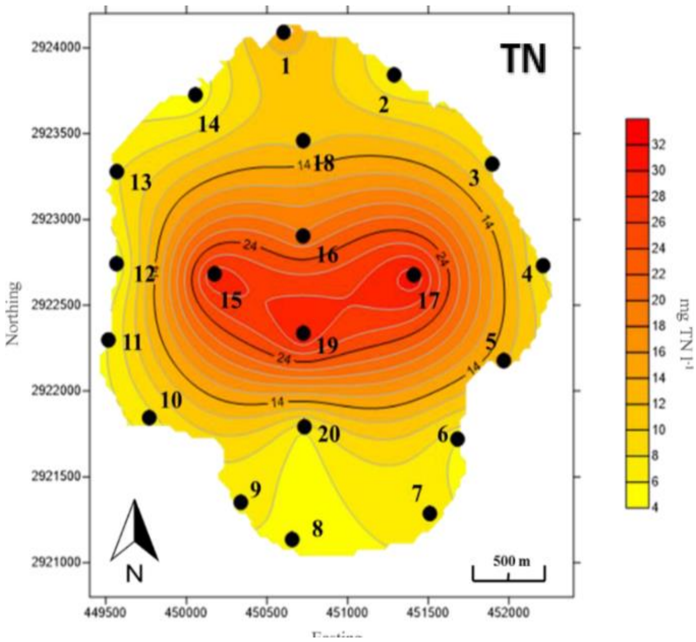
Mean seepage rates in L/m²d contoured using Surfer 12 from Lucius (2016).

The seepage is concentrated in the southern portion of the lake bottom and shore line, 8, 20, 19. The lake is receiving groundwater discharge at all monitored locations except for the region at site 12 where water was found seeping from the pond (negative seepage, low rate). The lake would be classified as effluent or a gaining lake as seepage rates are positive at the shoreline and in the main part of the lake.

b) Plot the mean total nitrogen data on a second copy of the location map. Use a contour interval of 2 mg/l. Where are the concentrations of total nitrogen the highest? Do they correspond with the highest groundwater seepage rates?



Plot of mean total nitrogen values in mg/l.



Mean total N concentrations in mg/l contoured using
Surfer 12 from Lucius (2016).

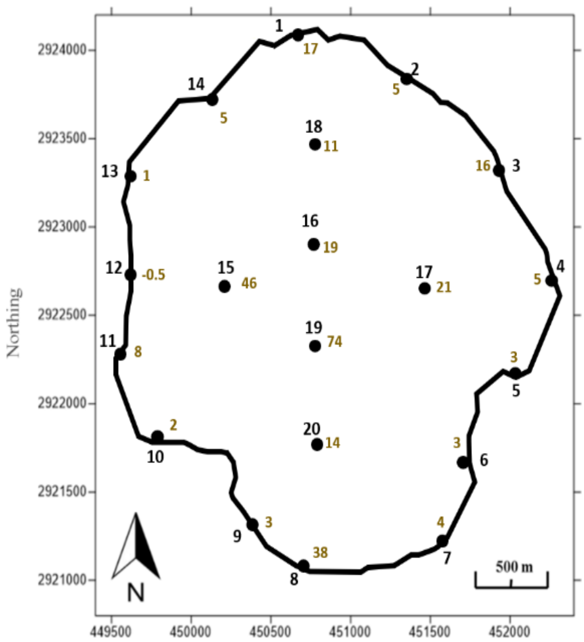
The highest total N concentrations are found in the central area of the lake, 19, 15 and 17. These locations had relatively high seepage rates; however, the total nitrogen concentration was relatively low at location 8 where the highest mean seepage rate was recorded.

c) Compute the mean total nitrogen loading at each location and contour the data. Use a contour interval of 20 mg/(m²d). Discuss how mean seepage meter and mini-piezometer mean water quality data can be used to target nitrogen reduction remediation efforts. Think about what other hydrogeologic information is needed to accomplish remediation goals (big picture).

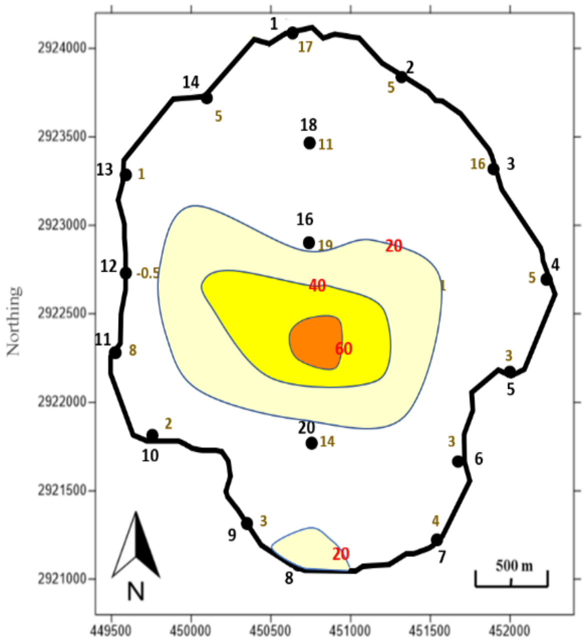
The total loading in units of mg/m²d were computed by multiplying the mean seepage rate and mean total N loading values in the table.

Site	Mean Tot N Load mg/m²/d
1	17
2	5
3	16
4	5
5	3
6	3
7	4
8	38
9	3
10	2
11	8
12	-0.5
13	1
14	5
15	47
16	19
17	21
18	11
19	74
20	14

The table values were rounded to the nearest whole number with the exception of the negative value at site 12. Values were plotted at the locations shown on the map and then hand contoured using linear interpolation.



Mean total nitrogen loading in units of mg/m²/d



Total loading hand contoured using an interval of 20 mg/m²d

Total nitrogen loading is concentrated in the center of the lake where groundwater concentrations of total nitrogen were high and seepage rates were also elevated. Loading also occurred in the vicinity of location 8, principally because groundwater seepage rates were high even though total nitrogen concentrations were only 6.43 mg/l. Nitrogen reduction efforts should focus on the source area of groundwater that seeps into the central part of the lake. Additional monitoring wells are needed to identify the groundwater flow paths in the vicinity of the lake so that pathways from the lake central area can be traced up gradient to the source. Sources of nitrogen encountered between the recharge area and source area can be identified and an evaluation made to see if the total nitrogen entering the system can be reduced. Other approaches might include engineering controls such as reducing the rate of nitrogen rich seepage to the lake by capturing and treating the groundwater before it reaches the lake.

Return to Exercise 4⮥

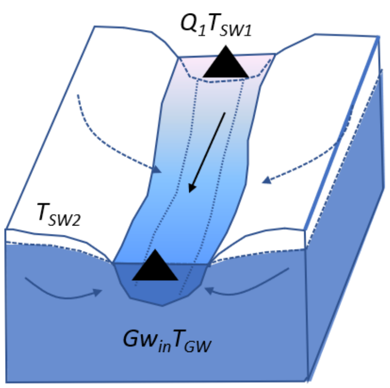
Solution to Exercise 5

5) Under relatively simple conditions, the temperature of stream water can be used to estimate the net gain in groundwater for an effluent reach of stream. Figure 71 presents the concept. If the streamflow at the upstream site Q1 is 2.0 m³/s and the mixed stream water temperature is 12°C, the shallow groundwater system temperature is 8.2°C, and a stream temperature measurement at the downstream site, Q2, is 10.4°C, then:

SOLUTION:

a) Assuming that over this 2 km river reach the other components of the heat budget are small, what is the net quantity of groundwater that discharges to this river reach?

The equation presented in Figure 71 relates stream discharge and temperatures to groundwater input.



$$Q_1T_{SW1} + GW_{in}T_{GW} = Q_2T_{SW2} = (Q_1 + Gw_{in})\,T_{SW2}$$

Based on the information given: $Q_1 = 2.0$ m³/s; $T_{SW1} = 12$°C; $T_{GW} = 8.2$°C and at 2 Km downriver the river temperature is 10.4°C. Thus, the GW discharge at 8.2°C over the 2 km reach is:

$2.0\text{m}^3\text{/s}\ (12°\text{C}) + GW_{in}\ (8.2°\text{C}) = (2.0\text{m}^3\text{/s} + GW_{in})\ (10.4°\text{C})$

$24\ \text{m}^3\text{/s °C} + 8.2\ GW_{in}\ \text{°C} = 20.8\ \text{m}^3\text{/s °C} + 10.4\ GW_{in}\ \text{°C}$

$3.2\ \text{m}^3\text{/s °C} = 10.4\ GW_{in}\ \text{°C} - 8.2\ GW_{in}\ \text{°C} = 2.2\ GWin\ \text{°C}$

$3.2\ \text{m}^3\text{/s °C}/2.2 = GWin\ \text{°C} = 1.45\ \text{m}^3\text{/s}$

Over the 2 km reach 1.45 m³/s of groundwater entered the stream.

b) How could individual temperature monitors installed in the stream and stream bed be used to verify the stream is gaining in this reach?

A small diameter monitoring well or mini-piezometer could be installed at a number of locations in the 2 km long stream channel. Individual temperature recorders isolated in the well bore using baffles would provide vertical temperature readings. A second approach would be to insert a metal temperature probe into the bed and record temperature variations at multiple vertical locations. In either case, when groundwater is discharging to the stream the temperature in the stream bed will reflect the groundwater temperature. The temperature will remain constant as the sampling depth below the bed increases. Probes may only need to be inserted a few 10's of centimeters into the bed. If a strong hyporheic exchange is also occurring in portions of the bed, stream water temperatures may influence the readings until probes are inserted to depths where the upwelling groundwater dominates.

Return to Exercise 5↥

Solution to Exercise 6

6) Heat tracing in the bed of surface-water bodies is an inexpensive and valuable tool for tracing the direction of exchanges, estimating hydraulic properties of bed material, and estimating exchange rates. USGS (2003) Circular 1260 is an excellent resource that explains the relevant principles, methods and modeling approaches. Review the document: https://pubs.usgs.gov/circ/2003/circ1260/pdf/Circ1260.pdf↗

SOLUTION:

a) Of the seven case studies presented, choose one and summarize the goals, methods and results. Include two key figures supporting your summary.

The table of contents of the report is presented here.

As an example, here is a summary of the report presented in Chapter 6. Trout Creek (Allander, K.K., 2003).

Purpose: To develop information on the magnitude and timing of groundwater discharge and the corresponding groundwater nutrient load to tributaries of Lake Tahoe. Trout Creek was chosen for groundwater exchange evaluation because creek flows provide the second largest stream nutrient sediment load to the lake. It is urbanized in the lower reach, has on-going monitoring, and has undergone channel restoration in the urban reach.

Methods: Along selected study reaches: Streamflow discharge measurements (synoptic surveys); monitoring well water level measurements in shallow monitoring wells installed in the channel and fitted with transducers; streambed temperature measurements using thermocouple sensors at two sites at 5 depths below the stream bottom (to 2.1 m).

Results: Water level differences between the monitoring well and stream stage varied seasonally (see Figure 7 copied below).

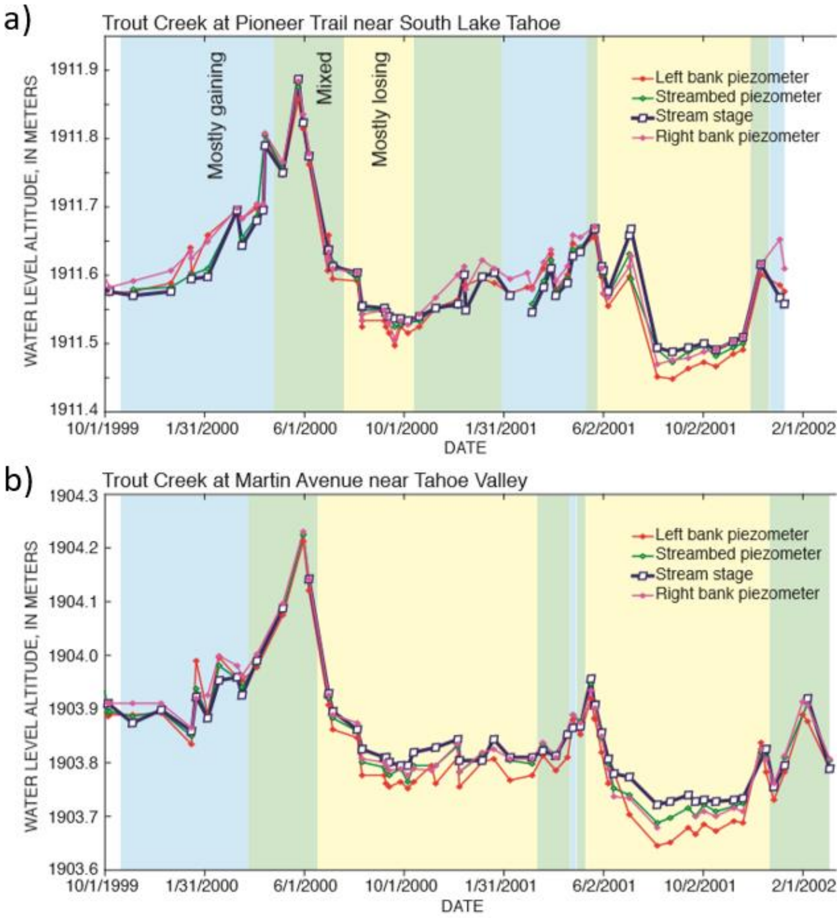


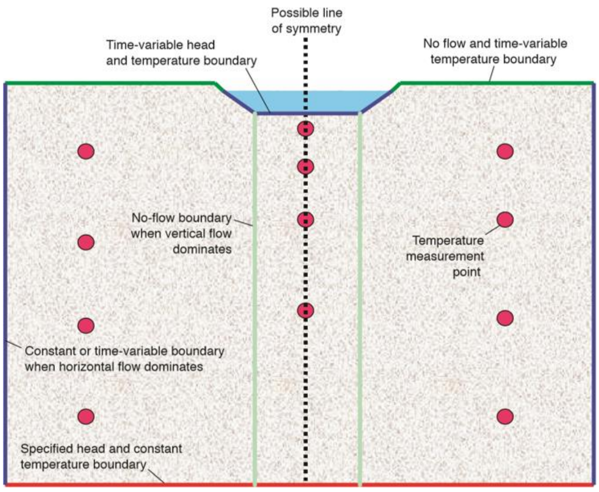
Figure 7. Water-level hydrographs of Trout Creek piezometers and stream stage.

The exchange process varied seasonally as indicated by VHGs measured in piezometers.

Summary: Study results showed a dynamic exchange process. Groundwater generally contributes to streamflow in the winter and early spring (blue on Figure 7), and the stream mostly loses flow to the groundwater in the summer and fall (yellow on Figure 7). The next step is to combine groundwater exchange with nutrient concentrations to produce loading information.

b) The publication also provides information on modeling of heat flow in Appendix B: "Modeling heat as a tracer to estimate streambed seepage and hydraulic conductivity" by Richard G. Niswonger and David E. Prudic. It is useful to study the appendix prior to designing field instrumentation so that field efforts will generate the required data for modeling. Read the appendix and list the parameters and boundary conditions needed when simulating heat transport from a river into the riverbed.

Example of Model Set up, boundary conditions and features:



Surface Boundary

Specified head and time variable temperature

Bottom Boundary

Specified head and constant temperature

Initial Conditions

Initial estimates of heads and temperatures at the beginning of the modeling

Model parameters

Example of parameters in Table 1 (Appendix B).

Table 1. Parameters used in VS2DH to model heat as a tracer through fluvial sediments

Parameter	Sensitivity	Range in values
Parameters for saturated flow through fluvial sediments		
Saturated hydraulic conductivity[1] (m/s)	High	10^{-7} to 10^{0}
Horizontal and vertical hydraulic conductivity ratio[1]	High	3 to 100
Porosity[1] (m^3/m^3)	Moderate	0.25 to 0.5
Dispersivity[2] (m)	Moderate	0.01 to 1
Heat capacity of dry sediments[3] (J/m^3 °C)	Moderate	1.1×10^6 to 1.3×10^6
Thermal conductivity of saturated sediments (W/m °C)[3]	Moderate	1.4 to 2.2
Heat capacity of water at 20 °C [4] (J/m^3 °C)	Low	4.2×10^6
Additional parameters for variably saturated flow through fluvial sediments		
Unsaturated hydraulic conductivity parameters in van Genuchten retention model[5]		
α (per meter)	Moderate	1 to 500
n (dimensionless exponent)	Moderate	1.1 to 2.8
Thermal conductivity at residual water content[3] (W/m °C)	Moderate	0.18 to 0.26
Residual water content[5] (m^3/m^3)	Low	0.00 to 0.10

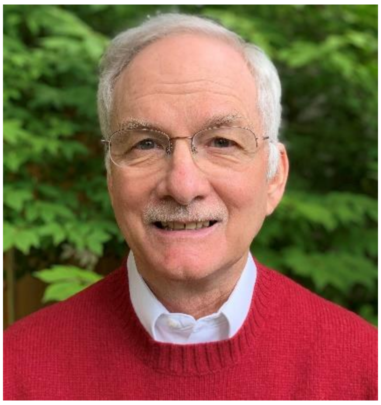
[1] Values are from Freeze and Cherry (1979, p. 29, 32–34, and 37) for silty sand, clean sand and gravel.
[2] Thermal dispersivity is assumed analagous to solute dispersivity. Solute dispersities are from Fetter (1993, p. 71–77) for observation scales between 1 and 10 m.
[3] Values are for sandy, loamy, and clayey soils – see Appendix A, table 1B.
[4] See Appendix A, table 1A.
[5] Values are from Kosugi and others (2002, p. 743) for soils, and from Fayer and others (1992, p. 693) for gravel.

Observation Data

Spatial and temporal distributions of head and temperatures to be used as initial conditions and in model calibration.

Return to Exercise 6↑

11 About the Author



Dr. William W. Woessner is a Montana University System Emeritus Regents' Professor of Hydrogeology at the University of Montana. He taught introductory and graduate courses in the hydrogeological sciences; is the Past Acting Director of the Center for Riverine Science and Stream Re-naturalization, and President of Woessner Hydrologic. With over 40 years of experience he has mentored over 60 graduate students, many of whom have founded or work for environmental consulting firms; consulted for local, state and federal agencies, corporations, and non-profits applying field and modeling approaches to solve groundwater development and management issues including production well design and installation, and water use. He has proposed resolutions to groundwater contamination events, and identified how groundwater-surface water interactions affect ecological systems. Dr. Woessner is the coauthor of two editions of "Applied Groundwater Modeling: Simulation of Flow and Advective Transport", a 2011 Fulbright Scholar-NAWI Austria, a Fellow of the Geological Society of America, the Birdsall-Dreiss Lecturer in 2005, and the 2020 recipient of the O. E. Meinzer Award. In addition, he is a Fellow and Life Member of the National Groundwater Association and was awarded the John Hem Excellence in Science and Engineering Award in 2008.

THE GROUNDWATER PROJECT