Groundwater–Surface Water Interaction

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1 Introduction

Understanding the hydrologic cycle requires not only studying the reservoirs, such as groundwater, surface water, and the atmosphere, but also flux boundaries from one reservoir to another. Groundwater is defined as water within saturated pores in the subsurface. As such, the flux of water between groundwater and surface water (q) is governed by the equation for saturated flow in porous media, Darcy's Law (Eq. 1):

$$q = Ki \tag{1}$$

where K is the hydraulic conductivity of sediments on the bed of a lake, stream, or other surface water feature and i is the head gradient across the interface. Although this equation shows the potential for quantifying interactions, the fluxes between groundwater and surface water are not well characterized [1]. The boundary, by definition, is underwater, which makes measurement difficult. Recharge to and discharge from groundwater is heterogeneous, so point measurements do not sufficiently characterize the nature of these fluxes.

The interface between groundwater and surface water occurs where the water table (top of the saturated zone) intersects land surface. In other words, the groundwater–surface water interface occurs in streams and rivers, lakes, wetlands, and the ocean. Other than point discharge at springs, the discharge is occurring below the surface of the water, largely unseen. Yet we know the discharge is occurring because surface water is a manifestation of groundwater discharge. Many streams would go dry between rain events if they did not receive groundwater discharge (other than streams that receive discharge from human sources or melting glaciers). When the water table drops, wetlands and lakes can go dry. Perched lakes on low permeability sediments can persist above the water table, but most lakes receive groundwater discharge or have a combination of groundwater discharge and recharge. Even for a water body as vast as the ocean, groundwater discharge has been increasingly recognized as an important component of the hydrologic cycle [2]. Thus, groundwater discharge is a key to sustainability of all surface water resources [3].

The interface can involve fluxes in both directions (Figure 1). Lakes and wetlands are known to both gain and lose water to groundwater. Perched lakes can lose groundwater through low permeability lake beds. Groundwater discharge to lakes typically occurs along the shore based on flow system concepts [1, 4] but can be gaining (receive flux from groundwater) on one shore and losing (recharge flux to groundwater) on another. Streams can also be gaining or losing and both can occur along different reaches of a stream. At baseflow (period with no stormflow), the water in the stream is typically derived from groundwater unless there is wastewater discharge from humans.

One type of exchange in streambeds, which will not be covered in this summary, is hyporheic flow. Hyporheic means flow beneath the surface and it refers to stream water that enters the streambed or stream bank and exits back out [5]. This water is not strictly groundwater since it is derived from the stream, but it could be considered to become groundwater for the period that it is flowing in the streambed or stream banks. This flow is typically shallow, penetrating only 10s of centimeters within the streambed, with a shorter flow path, and different geochemical makeup than groundwater. Thus, a distinction needs to be made between groundwater fluxes to streams and the hyporheic water that enters the stream but was originally streamwater (Figure 2). Because of these differences, hyporheic flow is not covered further in this article.

Submarine groundwater discharge (SGD) occurs at the margins of the oceans, driven by the head gradient onshore (Figure 3) as well as density gradients, tides, and

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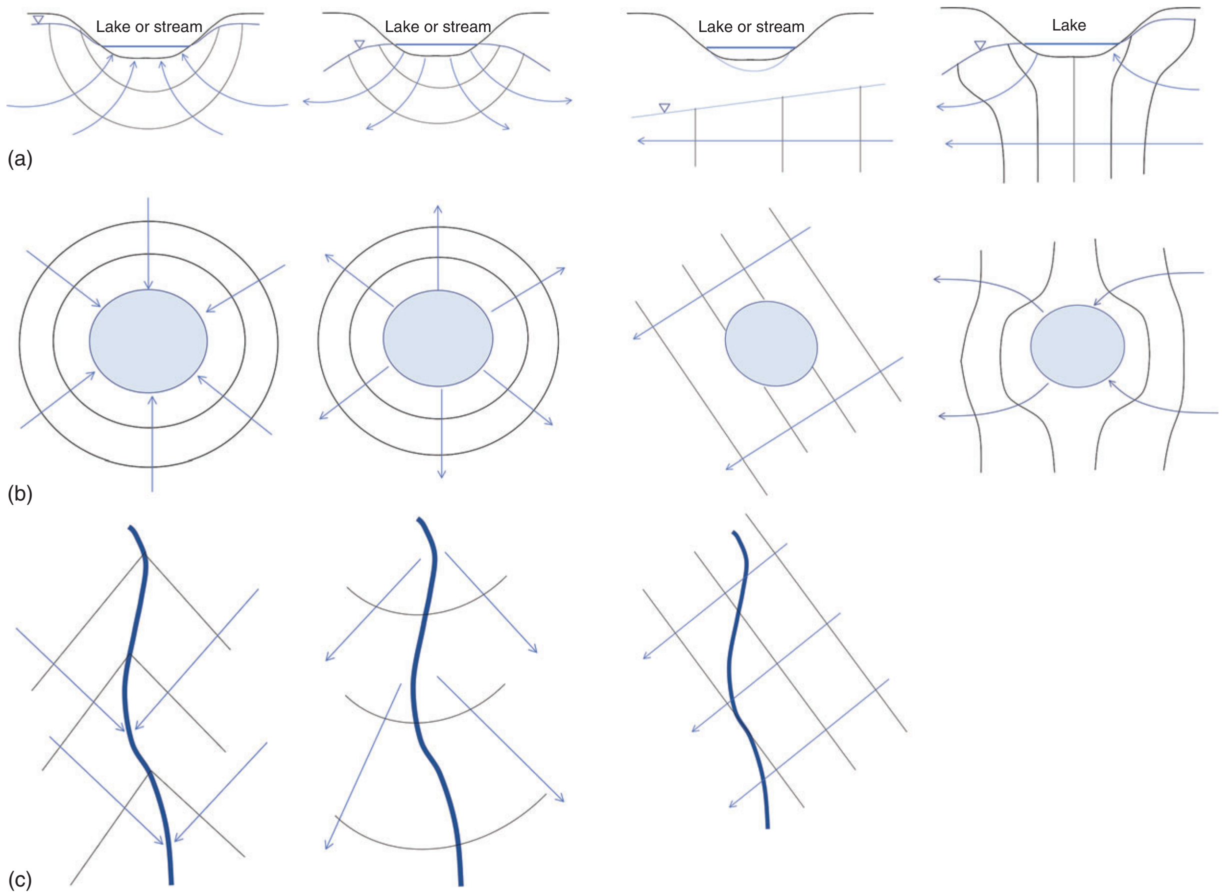

Figure 1 Groundwater–surface water interaction can occur as streams or lakes showing gaining (discharge from groundwater), losing (recharge to groundwater), flow under, or flow through. (a) Top panel shows cross sections with hypothetical contours and flow lines. (b) Middle panel shows plan view for lakes. (c) Bottom panel shows plan view for streams.

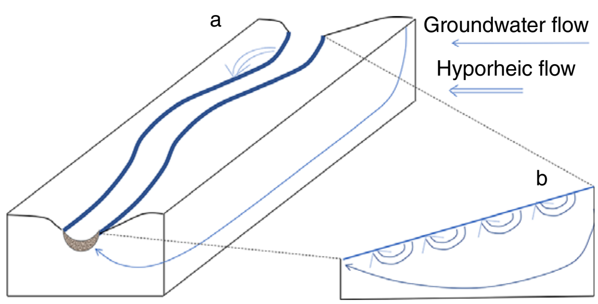

Figure 2 Groundwater discharge to streams occurs along deep flow paths originating from recharge areas. In contrast, hyporheic flow (blue arrows) is water derived from the stream that enters the subsurface, typically travels along short, shallow flow paths in the stream bank (a) or streambed sediments (b) and returns to the stream. Hyporheic water, with its origin in the stream, is distinct from groundwater.

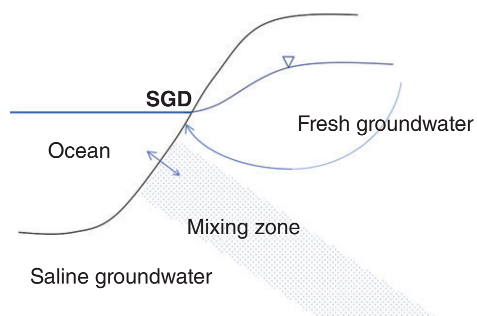

Figure 3 Submarine groundwater discharge (SGD) provides freshwater and dissolved solids to the ocean. SGD occurs along the shore, illustrated here driven by head in aquifers. A mixing zone of freshwater and seawater occurs at the boundary between saline and freshwater, and flow directions can be both into and out from the ocean.

waves [6]. Groundwater discharges to the ocean both offshore and through estuaries [7]. The offshore discharge typically occurs along the freshwater–saltwater interface, but this interface is diffuse and transient, and thus, the spatial extent of discharge is still poorly understood. Seawater intrusion into aquifers is the reverse of SGD (Figure 4) and is driven by changes in head, such as sea level rise or pumping in coastal wells, and changes in flux, such as decline in recharge [8].

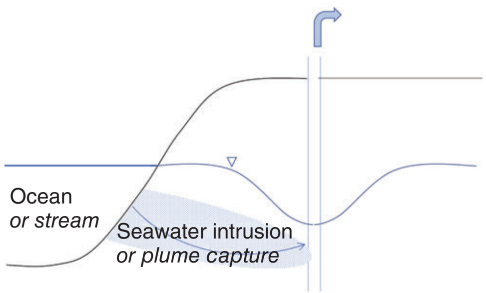

Figure 4 Groundwater–surface water interaction induced by pumping. Seawater intrusion can be driven by pumping wells (shown here) or sea level rise causing flow from saline to freshwater. Pumping can also induce flow from streams to groundwater, with potential to capture plumes and contaminate a well.

2 Importance of Groundwater–Surface Water Interaction

Groundwater–surface water interaction is important for both physical and chemical fluxes in the hydrologic cycle. Groundwater and surface water systems have commonly been studied separately, making it difficult to quantify the fluxes. Nonetheless, there is increasing recognition of the importance of understanding how this interface influences hydrologic budgets, geochemical cycles, ecosystems, and global change.

2.1 Hydrologic Fluxes

The contribution of groundwater to the ocean has been quantified through groundwater tracers and models. Zektser and Loaiciga [9] estimated that while about 6% of freshwater fluxes to oceans comes from groundwater as opposed to surface water, the salt loading was at least 50% from groundwater. More recent studies [10] have confirmed the role of SGD to the ocean and estimated that 70% of SGD occurs in the Indo-Pacific oceans. Modeling coastal systems, as well as direct measurement from boreholes, have shown that discharge can extend at least 100 km offshore [11].

Estimates of groundwater discharge are even more variable for streams and lakes because of the wide variety of settings. Zektser and Loaiciga [9] provided estimates for percent of river runoff as groundwater baseflow across the globe ranging from 35% to 55%. In 24 regions of the USA delineated by Winter et al. [1], groundwater contribution to streamflow ranged from 14% to 90% with a mean of 55%. Rosenberry et al. [12] reviewed data from 102 lakes, primarily in North America, and found that groundwater discharge to and recharge from lakes varied five orders of magnitude. They pointed out that the relative contribution of groundwater to lakes tends to decrease with increasing lake size for lakes of at least 100 ha. These estimates of the influence of groundwater in the hydrologic budget indicate the importance of understanding the fluxes across this interface.

2.2 Geochemical Fluxes: Ecosystem Influence and Contaminant Pathways

With hydrologic fluxes come geochemical fluxes. Groundwater chemistry has a signature distinct from streams, lakes, and oceans. While groundwater is less saline than ocean water, groundwater discharge is increasingly recognized as a source of dissolved solids that impacts geochemical cycling and ecosystems disproportionately to the total fluxes [2, 7].

Seawater intrusion (flux from the ocean to groundwater, Figure 4) is a threat to freshwater supplies in coastal aquifers [13]. Seawater intrusion may be enhanced due to pumping and sea level rise [8, 14, 15]. The contamination of aquifers by saltwater varies from individual wells to regional [15], and unconfined aquifers are more susceptible to intrusion [8]. Most studies of saltwater intrusion have involved laboratory and numerical models due to the difficulty in measuring the dynamic interface [14]. The coastal freshwater–seawater interface is important not only because of the need to protect coastal aquifers but also recently identified offshore aquifers with potential as freshwater resources [11].

Typically, groundwater has more total dissolved solids than streams and lakes (although saline lakes occur in some climates). The contrast between groundwater and surface water geochemistry influences ion transport and redox conditions. As a consequence, groundwater discharge can create zones of nutrient transport, enhance eutrophication (algal blooms), and influence ecosystem habitats. For example, the low redox conditions of groundwater can enhance phosphorous transport to lakes, enhancing eutrophication [16]. In contrast, the steady temperature of groundwater can provide habitats for fish and macrophytes in lakes and streams [17–20]. Higher base cations and metals were associated with groundwater seepage in Adirondack lakes [21], whereas sulfate reduction near zones of groundwater discharge influenced acid-neutralizing capacity in lakes influenced by mining in Germany [22]. These changes in chemistry along the lake bottom affect the overall lake chemistry and habitats for aquatic vegetation.

Geochemical fluxes across groundwater–surface water interface can also provide a mechanism for spreading contamination. Streams and lakes may be contaminated by agricultural activities, septic systems, or spills. In a nationwide study, the USGS detected nutrients, especially nitrate, in baseflow at about 2/3 of the streams in their study and attributed the loadings to groundwater

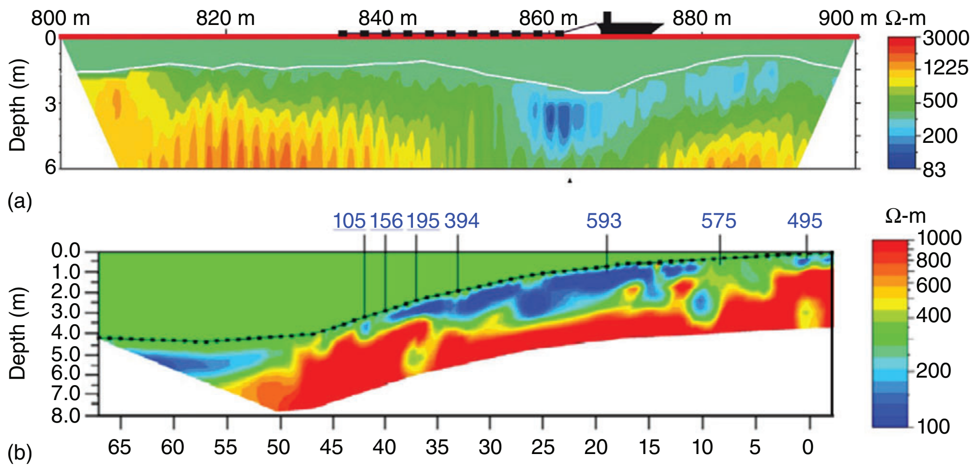

Figure 5 A plume of groundwater contaminated by road salts was revealed beneath Mirror Lake in NH using geophysical surveys. Two perpendicular cross sections are shown. (a) The towed resistivity survey around the circumference of the lake showed a high conductivity region in blue at 860 m. (b) The length of the plume perpendicular to the shore was then delineated by a lake bottom resistivity survey. The plume was confirmed by chloride concentrations in water collected from seepage meters. Chloride concentrations from 105 to 593 are shown in units of μeq L^{-1}. The lake concentration was 104 μeq L^{-1}. The conductive zone from 55 to 65 m in the deep part of the lake is organic sediments.

discharge [23]. The loadings are affected by the degree of buffering in the riparian zone and degradation where organic-rich sediments are available. Management of irrigation practices will not be as effective if groundwater is already contaminated. Another source of surface and groundwater contamination is road salt. An effort to curb road salt to a Mirror Lake, a pristine lake in NH, was thwarted when the salt plume bypassed a surface berm and took a groundwater pathway to the lake ([24], Figure 5). McCobb et al. [25] tracked a plume from a sewage treatment plant 120 m offshore on Ashumet Pond on Cape Cod using 52 drive point samplers; Coleman and Friesz [26] similarly tracked a septic plume on Walden Pond with drive point samplers on the perimeter of the pond. Numerous examples of aquifer plumes contaminating streams from a variety of land uses can be found in the literature, and improved techniques are needed to monitor and quantify their effects [27–29].

Conversely, groundwater can be contaminated by surface water particularly when pumping induces infiltration (Figure 4). Atrazine contamination in aquifers was traced to surface water sources in agricultural areas in Nebraska and Kansas [30, 31]. Pharmaceutical products have also been found in systems for managed aquifer recharge, whereby stream water used for drinking supplies is first filtered through groundwater [32, 33]. In a well-known trial to identify who was liable for groundwater contamination of an aquifer near Woburn, MA, the potential for the town's water supply well to draw contamination from surface water was one of the confounding factors in assigning responsibility [34].

2.3 Global Change

Monitoring fluxes at the groundwater–surface water interface may also provide an indicator of impacts of climate change. The hydrologic cycle is likely to intensify under climate change scenarios, as manifest by stronger storms and longer droughts. However, uncertainties in the hydrologic budget make prediction difficult [35], and the different scales of climate models and hydrologic components makes assessing impacts challenging [36]. Furthermore, land use change and climate impacts have overlapping and potentially amplifying effects [37]. These issues highlight the importance of improving our understanding of fluxes across the groundwater–surface water interface.

Human settlement is often clustered around surface water resources – coasts, rivers, lakes, and springs [38, 39]. Humans, in turn, have impacted the hydrologic cycle wherever they have settled through land use change, climate change, overuse of water, redirection of water, and water quality changes [39, 40]. Our understanding of groundwater–surface water interaction helps us map out not only our history but our future.

3 Monitoring Challenges

There are three challenges that make monitoring groundwater–surface water interaction complex. First, monitoring underwater is difficult. Visual assessment to help identify recharge or discharge is limited. An underwater interface also requires all monitoring equipment be waterproof. Equipment is vulnerable in these environments. Monitoring equipment that is left in the surface water environment, even if waterproof, may be susceptible to human and other animal disturbance. Surface water environments are dynamic; storms can move the equipment or flood batteries that power equipment. Monitoring wells may provide access to the subsurface environment and a place to sequester monitoring equipment. However, many sites are accessible only by boat, which makes drilling wells impossible or expensive. Even in streams that are shallow, finding a platform for well installation can be difficult due to steep slopes or unstable ground, and many streambeds are too rocky for drive point installation.

Second, the recharge and discharge interface is spatially heterogeneous. Geologic heterogeneity can affect fluxes across this interface, and heterogeneity is notoriously difficult to characterize. Order of magnitude (or more) variations in flux can occur on the scale of meters. In addition, surface water systems can have fine layers that exert control on vertical fluxes and are particularly difficult to identify. As detection techniques for measuring fluxes improve, allowing for low rates to be measured, the true heterogeneity of these systems is being revealed.

The third challenge is the transient nature of the interface. Recharge and discharge areas can vary in time, so continuous data are needed to capture this variation. Recharge areas can become discharge areas and vice versa. A number of different factors cause variations in time, from natural variation in precipitation to human-induced alterations in land and water use. The impacts can be delayed, making the need for long-term monitoring an additional challenge. To meet these challenges, new techniques are being developed to extend our knowledge of the groundwater–surface water interface and explore the complexities of this environment.

4 Innovations in Monitoring the Groundwater–Surface Water Interface

Techniques to monitor groundwater–surface water interaction, range from direct measurements to inferences from tracers and remote sensing. In this section, the strengths and limitations of the basic techniques are only briefly mentioned. The emphasis here is on examining recent innovations that are allowing us to see better into this interface rather than a retrospective review of monitoring techniques, which has been reviewed elsewhere [12].

4.1 Point Measurements

Direct point measurement of fluxes at the groundwater–surface water interface typically involves using Darcy's Law (Eq. 1). Wells are installed at different depths to obtain the vertical gradient, and hydraulic conductivity must be measured either by collecting a sample and analyzing in the lab or a field measurement such a slug testing [41, 42]. This technique suffers from inaccuracy in calculating small head gradients and the difficulties in obtaining reliable hydraulic conductivity measurements. Heterogeneity issues are compounded by not only horizontal variations but also vertical variations – fluvial and lacustrine environments often have fine layers of variable sediment grain sizes [43, 44]. These layered systems create sharp differences in hydraulic conductivity and head over fine scales. Recognition of scaling issues with point measurements has led to the development of techniques that integrate over larger scales.

In lakes, wetlands, or the ocean, seepage meters are used for point measurements at a slightly larger scale [45]. A seepage meter typically consists of the cut-off top of a storage barrel that is pressed into a submerged sediment bed to isolate seepage flow across a known surface area. A bag partially filled with water is attached to the barrel to record change in volume over time. Loss of water from the seepage bag records flux from the lake to groundwater, gain in water records flux from groundwater to the lake. Although the technique has been applied for decades, improvements continue to be made such as using bag shelters and weighing bags for more accurate water volumes [46]. Seepage meters suffer from the typical limitations of point measurements in that they may not predict heterogeneity well. However, increasing use of networks of seepage meters has helped reveal patterns in groundwater–surface water interaction (Figures 6 and 7). Schneider et al. [47] instrumented the 88-km circumference of Oneida Lake in New York with 25 seepage meters and found that discharge rates varied both spatially and temporally but did not correlate with bed sediment texture. Michael et al. [48] used a dense network of 40 seepage meters in Waquoit Bay in Massachusetts to reveal discharge patterns that did not decrease away from the shore as traditionally modeled. In a follow-on study, Michael et al. [49] quantified seasonal changes in SGD with higher discharge in the spring in response to precipitation and lower in the summer when evapotranspiration was higher. Toran et al. [50] found variations in discharge from 0 to $-282\,\mathrm{cm\,d^{-1}}$ in a network of 28 seepage meters along the southwest shore of Mirror Lake (Figure 7). The region with the highest seepage correlated with transition from till to an outwash zone identified with geophysical monitoring. In addition, smaller variations were related to permeability

Figure 6 Photograph of an array of seepage meters on the southwest shore of Mirror Lake, NH. The seepage meter is a half-barrel inserted into the bed sediments with a bag attached by a hose. The bag is protected from waves by a bag shelter. Mini piezometers are driven into the lake bed nearby to measure the difference in head between the groundwater and the lake.

of a thin (2 cm) layer of lake bed sediment. Because this layer is thin, sediment disturbance can create significant changes in seepage rates [43]. Logging seepage meters allowed greater temporal resolution in seepage meter data not just seasonally or daily but hourly. Variations from 10% to more than an order of magnitude have been recorded due to waves, tides, and storm events [51, 52].

Seepage meters are not used as much in streams because the flowing water can create a head differential between the seepage meter and the bag, causing significant error in measurement [46]. An adaption of the seepage meter, called a seepage blanket, has been used to calculate transit time and denitrification rates in a stream in North Carolina [53, 54]. The seepage blanket is a low-profile seepage meter (to minimize disturbance of the stream flow) and uses a dilution flow meter rather than a seepage bag. The dilution flow meter works by injecting a salt tracer into a mixing chamber, then monitoring with a conductivity meter. Dilution of the tracer over time provides a flux. With these innovations, seepage meters can better monitor natural pore fluid geochemistry which is another tool for characterizing groundwater–surface water interaction, as discussed in the following section.

4.2 Stream Gauging

A more common technique of estimating groundwater inflow to and outflow from streams is differential stream gauging (sometimes called a seepage run), which provides measurements at the 10s to 100s of meters scale for reach lengths with no tributary or pipe flow. Discharge is measured along two cross sections, and the difference in flow is attributed to groundwater input or stream loss. Some challenges associated with discharge measurements include finding appropriate reaches for accurate measurements and collecting data at high flow to determine relationships under dynamic conditions. High flow conditions and associated fast flowing water are dangerous for monitoring, and errors associated with measurement of larger flows can be larger than the change in flow between gaging locations. Because of these challenges, many stream reaches remain ungauged and new techniques are needed [55]. Profiling streams with acoustic Doppler velocity meters has improved data collection, especially since floatation devices have been developed so that the user does not have to be in the stream [56]. Nonetheless, spatial and temporal resolution is still limited.

Salt dilution gauging has been used to meet another challenge: gauging headwater streams with low flow rates that are difficult to resolve with flow meters. Although the technique was introduced in the 1970s [57, 58], use has been increasing in recent years [59]. For example, Payn et al. [60] used sequential upstream tracer tests to distinguish patterns of gains and losses along 13 reaches of a stream in Montana. The tests were repeated at high, intermediate, and low flows to observe changes in

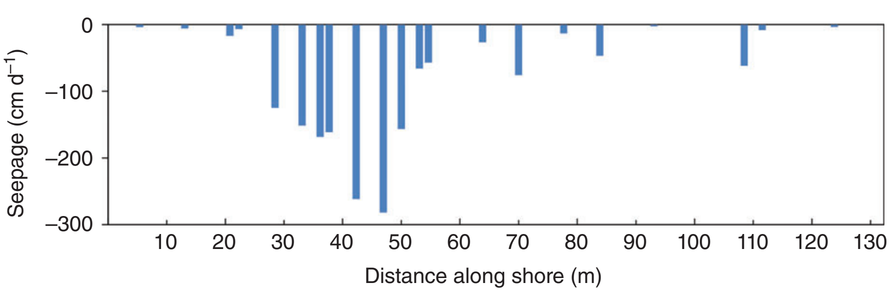

Figure 7 Histogram of seepage variation along the southwest shore of Mirror Lake, NH ranged from 0 to -282 cm d^{-1} (negative for seepage into the lake bed) for 28 seepage meters at approximately 5 m intervals.

groundwater–surface water interaction. They observed reaches with both gains and losses, particularly at intermediate and low flows, and were able to distinguish shallow (hyporheic) and deep (groundwater) flow.

Additional innovations in remote sensing and cameras have been introduced to provide continuous data on stream discharge and other groundwater fluxes [61]. Cardenas et al. [62] evaluated thermal imaging during storms and pointed out the potential interferences from biota, turbidity, image angle, and nearby personnel. Improved computer visualization algorithms have allowed the use of particle tracking in imaging to estimate flow rates [63] although additional refinement is needed to reduce errors estimated at 10–30%. Potential innovations include fluorescent particles to improve velocity tracking [64] and drones to extend coverage to areas typically inaccessible during storms [65, 66].

4.3 Temperature as a Tracer of Fluxes

Because groundwater may have a different temperature than surface water, temperature can be a natural tracer of groundwater flux [67, 68]. Groundwater, with a steady temperature, tends to be cooler than surface water in the summer and warmer in the winter. Vertical temperature gradients can reveal rates of groundwater discharge. In streams, the diurnal temperature signal differs in streambed sediments beneath reaches with upward fluxes (muted diurnal signal) and reaches with downward fluxes (enhanced diurnal signal). Modeling the fluxes can be used to quantify the flux [69–71]. Conant [72] mapped spatial heterogeneity of groundwater discharge in a stream in Canada using hundreds of mini-piezometers to insert temperature probes for point measurements. Essaid et al. [73] used both seepage measurement and 1D heat flow modeling to find reversal from upward to downward fluxes during high stage events. Aerial infrared imagining has been used to detect groundwater discharge in the ocean [74] and can be used to quantify plume areas [75]. Both aerial methods [76, 77] and handheld thermal imaging devices [78, 79] have potential for mapping discharge in lakes and streams. Quantifying discharge by thermal mixing compared favorably to conductivity and discharge measurements of discharge [80] in a headwater stream in Germany, but there are challenges to quantifying the discharge in larger scale systems due to heterogeneity [72]. Diffuse discharge is more difficult to detect and high-temperature contrasts are needed for quantification.

A technical innovation in temperature sensing is distributed temperature systems (DTS) which leverages the relationship between travel time of light and temperature in a fiber optic cable [81, 82]. The fiber optic cable can

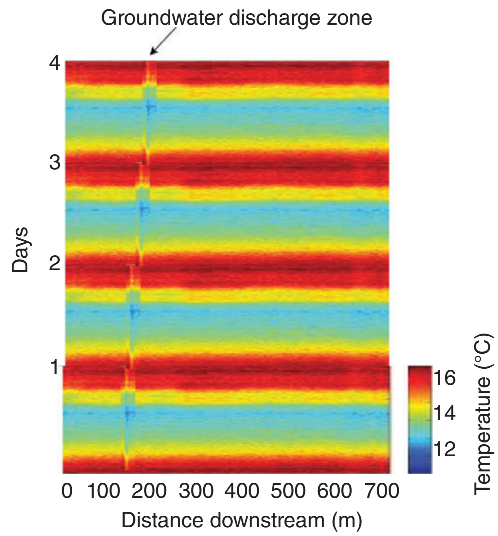

Figure 8 Hypothetical distributed temperature system (DTS) data showing the spatial and temporal resolution. Along the vertical axis is the variation in time over four days, which shows a diurnal temperature signal. Along the horizontal axis is a 700-m-long survey. In this hypothetical diagram, a groundwater discharge zone indicated by a shift in temperature (cooler in day and warmer at night) is shown around 150 m. This zone drifts slightly downstream to 200 m over the course of the monitoring (artificially generated offset in the temperature shift).

provide fine-scale measurements (m) over long distances (km) at a fine temporal resolution (Figure 8) – clearly all benefits for trying to detect groundwater discharge in lakes and streams. The fiber optic cable is relatively easy to deploy, although it requires field calibration using ice baths, and the cable can be susceptible to disruption by animals. Selker et al. [83] used DTS to map spatial variability of discharge in a stream, and Lowry et al., [84] found spatial variability of discharge in a stream was stable over multiple deployments. Krause and Blume [85] found improved resolution when operating in the winter and using the instrument to record both forward and reverse traces of light. Briggs et al. [86] used DTS to map vertical temperature distribution by wrapping the cable around a PVC pipe, providing an even finer spatial resolution (on the order of cm). They related vertical fluxes to streambed morphology and identified temporal variations in discharge. Briggs et al. [87] found greater sensitivity using DTS compared to streamflow and tracer test techniques for detecting lower rates of groundwater discharge to streams.

4.4 Geophysical Monitoring

Geophysics offers a tool to understand heterogeneity at the groundwater–surface water interface because it provides continuous rather than point measurements.

A range of geophysical techniques can be applied at a variety of scales from airborne techniques for basin scale to handheld instruments that map a stream cross section as reviewed in Robinson et al. [88]. In addition, research into characterizing pore scale properties such as hydraulic conductivity is ongoing [89]. This summary will focus on one example technique, electrical resistivity, which measures the ease with which ions migrate between the two current electrodes. This technique has been applied to monitoring groundwater–surface water exchange in the ocean, in lakes, and in streams.

An electrical resistivity survey can map three potential targets; porosity, pore fluid conductivity, and clay content, all of which are relevant in characterizing streambeds or lake beds. Several types of electrical resistivity surveys can be conducted at the groundwater–surface water interface: dynamic surveys with the resistivity cable towed behind a boat, stationary surveys with the resistivity cable floating on the surface, stationary surveys with the cable deployed on the lake or stream bottom, and time-lapse surveys which show the change in resistivity at the same location over time.

Towed surveys have been used to map the distribution of clay sediments hindering the recharge of an aquifer beneath the Ohio River [90], determine the strike of fault zones beneath a river [91], uncover a road salt plume beneath Mirror Lake in NH ([24], Figure 5), and identify freshwater springs and discharge points in saline lakes in Nebraska [92]. The conductivity contrast between groundwater and seawater has also provided a geophysical target [93] and showed preferential pathways where radon-222 indicated freshwater is discharging to a reef [94]. Dimova et al. [95] used resistivity measurements to help identify point source discharge of SGD in Hawaii, which improved rate estimates for natural tracers. Swarzenski and Izbicki [96] used resistivity to map the limit of offshore SGD.

Surveys on stream or lake bottoms help reveal heterogeneity of fluvial and lacustrine sediments and the impact on groundwater–surface water interaction. 2D lake bottom resistivity surveys at Mirror Lake NH showed contrasts in seepage based on a transition between till and outwash, while 3D resistivity surveys were needed to identify smaller scale permeability contrasts [50]. Cardenas et al. [97] distinguished hyporheic flow from deeper groundwater using electrical resistivity along the Colorado River in Texas. Resistivity surveys helped extend point measurements of riverbed hydraulic conductivity in the Great Miami River in Ohio [98] and mapped sediment thickness near a log jam [99]. Nyquist et al. [100] revealed small-scale patterns of ground-water discharge in streams using time-lapse surveys, one conducted during low stage when groundwater discharge dominated, the other during high stage stream conditions when more surface water filled the streambed sediments. Areas that showed little or no change between low flow and high flow surveys correlated with the locations of known seeps. These and other studies have shown the potential for increasing our understanding of spatial variation in groundwater–surface water interactions using geophysics.

4.5 Geochemical Tracers

Geochemical tracers have been important in providing integrated signals, which average over space or time. A geochemical tracer is found at high concentration in one component, such as groundwater, and can be used qualitatively or quantitatively. For example, radon-222 is found at higher concentration in groundwater than in surface water because it is volatile; its presence in streams, lakes, or the ocean is a signature of groundwater discharge. Quantifying the amount of discharge requires some form of a chemical mass balance equation. If the concentration in all but two components is negligible, and the flux is known for one component, then the flux of the final component can be quantified. For instance, the chemical flux in a river could be simplified to show three components, groundwater input, groundwater recharge, and evaporation (or in the case of radon, volatilization):

$$Q_s C_s = Q_{gin} C_{gin} - Q_{gout} C_s - Q_e C_e \tag{2}$$

where Q is the volume per time of stream water (s), groundwater (gin or gout), or evaporation (e) and C is the concentration of the geochemical tracer for each component. Calculation of evaporation using empirical equations allows estimation of net groundwater input ($Q_{gin} - Q_{gout}$) once the stream and groundwater concentration are measured along with streamflow. The above mass balance equation is for baseflow in a stream. Mass balance equations for storm hydrographs also make use of tracers, which is important for contrasting groundwater contributions and overland flow; these contrasts are reviewed elsewhere [101]. However, storm hydrographs are also important to estimate annual baseflow contributions to streams [102]. For the ocean or a lake, the mass balance input and output terms might include groundwater, stream water, precipitation, and evapotranspiration. For a lake system closed to surface water, estimating the evaporation component allows estimation of groundwater fluxes using stable isotopes of water [103]. Cook [104] provides error analysis for estimates of fluxes using different tracers.

Isotope tracers, in particular, radon and radium isotopes, have been a valuable tool to estimate SGD by mass balance methods because they provide large-scale integration (as reviewed in [2, 105]). The radon-222 signature in groundwater contributes measurable quantities in the ocean before volatilizing and reveals variations in SGD, for example higher discharge in the North Atlantic than the South Atlantic [2]. Burnett and Dulaiova [106] recommended combining radon-222 and radium-224 measurements to account for atmospheric losses and mixing. The multiple isotopes of radium provide a sensitive tracer and have provided evidence for seasonal variations is SGD [2]. Kwon et al. [91] used radium-228 and a numerical model to estimate global SGD flux that was greater than mass balance estimates, which suggests a nonfreshwater component may be important. Their global estimates of SGD point to large fluxes in the Indo-Pacific Ocean and suggest greater threats due anthropogenic influences on groundwater there.

Although tracers have been used less frequently in lakes and streams [12], they can complement other techniques for measuring groundwater–surface water interaction. Radon input to lakes has been used to map groundwater inflow much like the coastal systems [107, 108]. Kluge et al. [109] linked areas of high radon to bedrock heterogeneity beneath a lake and Shaw et al. [110] used radon to look for areas of nutrient input to a lake. In streams, radon or helium input from water moving through the hyporheic zone can artificially increase estimates of groundwater discharge [107] and multiple tracers are recommended to better identify different components [111]. Arnoux et al. [112] used radon-222 and water isotopes to distinguish short-term variations in groundwater discharge to a lake from long-term evaporative effects. Groundwater itself can come from multiple pathways that potentially are identifiable by combinations of tracers. Bank storage can be distinguished from groundwater discharge based on 18-O isotopes even when the ion chemistry is similar [113, 114]. Stewart et al. [115] point out that deep groundwater discharge may have a distinct age signature, and found that a component of old, deep groundwater is typically neglected in residence time calculations using only 18-O and 2-H isotopes in storm hydrographs. They recommend using tritium isotopes to improve residence time calculations.

Despite the challenges of untangling multiple components and transient conditions, new measurement techniques are improving our use of tracers as a complementary tool for understanding groundwater–surface water interaction. Faster analyses and smaller sampler sizes have increased use of tracers. Complementary modeling helps interpret transients, such as annual variations in lake inputs [116]. However, models available for coupled flow and transport are limited and calibration is difficult [117]. Passive sensors [118] that provide finer-scale measures in bed sediments show promise for improving our ability to estimate vertical fluxes across the groundwater–surface water interface.

5 Conclusions

Despite the challenges in monitoring, considerable progress has been made in understanding groundwater–surface water interaction. More is known about spatial variation as well as temporal variation in fluxes because of new techniques that are being applied. Geophysics has become an up and coming method for detecting groundwater–surface water interactions because of the benefits of distributed sensing although there are still some challenges in interpreting the signal. Temperature sensing to detect groundwater–surface water interactions has evolved from point measurements to fiber optic distributed temperature sensing systems (DTS) that provide fine spatial and temporal resolution.

Some of the advances have come from using techniques in combination and long-term monitoring. Geophysics, complemented with geochemical tracers and direct flux measurements, can provide a distributed signal that improves hydrologic budget calculations. Transient monitoring with logging seepage meters, DTS, and remote sensing have increased accuracy of measured fluxes and revealed stresses across the interface. These improvements in data provide input to models which will further expand our understanding of the factors that influence fluxes and reveal dynamic responses of groundwater–surface water interaction.

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