Interactions between groundwater and surface water: the state of the science

Marios Sophocleous

Abstract The interactions between groundwater and surface water are complex. To understand these interactions in relation to climate, landform, geology, and biotic factors, a sound hydrogeoecological framework is needed. All these aspects are synthesized and exemplified in this overview. In addition, the mechanisms of interactions between groundwater and surface water (GW–SW) as they affect recharge–discharge processes are comprehensively outlined, and the ecological significance and the human impacts of such interactions are emphasized. Surface-water and groundwater ecosystems are viewed as linked components of a hydrologic continuum leading to related sustainability issues. This overview concludes with a discussion of research needs and challenges facing this evolving field. The biogeochemical processes within the upper few centimeters of sediments beneath nearly all surface-water bodies (*hyporheic zone*) have a profound effect on the chemistry of the water interchange, and here is where most of the recent research has been focusing. However, to advance conceptual and other modeling of GW–SW systems, a broader perspective of such interactions across and between surface-water bodies is needed, including multidimensional analyses, interface hydraulic characterization and spatial variability, site-to-region regionalization approaches, as well as cross-disciplinary collaborations.

Résumé Les interactions entre les eaux souterraines et les eaux de surface sont complexes. Pour comprendre ces interactions, qui dépendent du climat, des paysages, de la géologie et de facteurs biotiques, il est nécessaire de bien connaître le cadre hydro-géo-écologique. Tous ces aspects sont synthétisés et donnés en exemple dans cette revue d'ensemble. En outre, dans la mesure où ils affectent les processus de recharge–décharge, les mécanismes des interactions entre les eaux souterraines et les eaux de surface sont largement ébauchés, et la signification écologique et les impacts humains de telles interactions sont mises en avant. Les écosystèmes des eaux de surface et souterraines sont conçus comme étant des composantes liées appartenant à un continuum hydrologique conduisant à des questions sur le développement durable. Cette revue d'ensemble conclut par une discussion sur les besoins de recherche et des problèmes posés par ce thème en développement. Les processus biogéochimiques dans les quelques centimètres de sédiments immédiatement sous toutes les eaux de surface (*la zone hyporhéique*) ont un effet essentiel sur le chimisme des échanges d'eau, et c'est sur ce point que se sont concentrées la plupart des recherches récentes. Cependant, pour faire progresser la modélisation conceptuelle et les autres modélisations des systèmes eau souterraine–eau de surface, une perspective plus large de ces interactions à l'intérieur et entre les hydrosystèmes de surface est nécessaire, en prenant en compte des analyses multidimensionnelles, la caractérisation hydraulique de l'interface et la variabilité spatiale, les approches par régionalisation du site local à la région, aussi bien que des collaborations transdisciplinaires.

Resumen Las interacciones entre aguas subterráneas y superficiales son complejas. Para entenderlas en relación con factores climáticos, de relieve del terreno, geológicos y bióticos, se necesita un marco hidrogeoecológico robusto. Este artículo resume y presenta ejemplos de todos estos aspectos. Además, se describe con profusión los mecanismos de interacción entre las aguas superficiales y subterráneas que afectan a los procesos de recarga y descarga, haciendo hincapié en la importancia ecológica y en los impactos humanos de tales interacciones. Los ecosistemas de aguas superficiales y subterráneas son considerados como elementos unidos de un continuo hidrológico que llevan a plantear su sustentabilidad. La revisión concluye con una discusión de las necesidades de investigación y de los retos que afronta este campo tan dinámico. Los procesos biogeoquímicos que se producen en los primeros centímetros de los sedimentos en la mayoría de los cursos y reservorios de aguas superficiales (zona hiporreica) tienen un profundo efecto en la química del intercambio de agua, y es aquí donde incide la mayoría de la investigación más reciente. Sin embargo, se

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requiere una perspectiva más amplia de las interacciones entre aguas superficiales y subterráneas con el objeto de avanzar en la modelación de estos sistemas, incluyendo análisis multi-dimensionales, caracterización de la hidráulica en la interfaz y de la variabilidad espacial, métodos de regionalización, y colaboraciones interdisciplinares.

Keywords Groundwater recharge · Hydraulic properties · Hyporheic zone · Hydroecology · Water sustainability

Introduction

Groundwater and surface water are not isolated components of the hydrologic system, but instead interact in a variety of physiographic and climatic landscapes. Thus, development or contamination of one commonly affects the other. Therefore, an understanding of the basic principles of interactions between groundwater and surface water (GW–SW) is needed for effective management of water resources.

In recent years, as Winter (1995) points out, studies of GW–SW interactions have expanded in scope to include studies of headwater streams, lakes, wetlands, and estuaries. The interaction between groundwater and lakes has been studied since the 1960s because of concerns related to eutrophication as well as acid rain. Interest in the relationship of groundwater to headwater streams increased greatly in the past two decades because of concerns about acid rain. Interest in the relationship of groundwater to wetlands and to coastal areas has increased in the past 20 years as these ecosystems are lost to development (Winter 1995). Recently, attention has been focused on exchanges between near-channel and in-channel water, which are key to evaluating the ecological structure of stream systems and are critical to stream-restoration and riparian-management efforts. The teaming of geologists and hydrologists with ecologists is resulting in a more comprehensive conceptualization of GW–SW interactions. This work attempts to synthesize this broader, multidisciplinary perspective of GW–SW interactions, starting with some underlying prerequisites for comprehending environmental systems.

Principal Geomorphologic, Hydrogeologic, and Climatic Controls on Groundwater Flow Systems and GW–SW Interactions

To understand GW–SW interactions, it is necessary to understand the effects of what Tóth (1970) calls the "hydrogeologic environment" on groundwater flow systems – that is, the effects of topography, geology, and climate. Differences in surface topography are often mirrored by differences in potential. As Hubbert (1940) shows, given an areally uniform precipitation and infiltration rate over an undulating surface, a groundwater flow system will develop driven by a water-table surface that is a subdued replica of the land surface. The resulting groundwater flow pattern is not only controlled by the configuration of the water table but also by the distribution of hydraulic conductivity in the rocks. In addition to topographic and geologic effects, groundwater flow is affected by climate (precipitation being the source of recharge). Groundwater moves along flow paths that are organized in space and form a *flow system*. In nature, the available subsurface flow domain of a region with irregular topography contains multiple flow systems of different orders of magnitude and relative, nested hierarchical order. Based on their relative position in space, Tóth (1963) recognizes three distinct types of flow systems – local, intermediate, and regional – which could be superimposed on one another within a groundwater basin. Water in a *local flow system* flows to a nearby discharge area, such as a pond or stream. Water in a *regional flow system* travels a greater distance than the local flow system, and often discharges to major rivers, large lakes, or to oceans. An *intermediate flow system* is characterized by one or more topographic highs and lows located between its recharge and discharge areas, but, unlike the regional flow system, it does not occupy both the major topographic high and the bottom of the basin. Regional flow systems are at the top of the hierarchical organization; all other flow systems are nested within them. Detailed aspects of complex systems and scaling with the encompassing hierarchy theory and its applications are described in Allen and Starr (1982), Klemes (1983), O'Neill et al. (1986), Grimm and Fisher (1991), Wu and Loucks (1995), Stanley et al. (1997), Fisher et al. (1998), Marceau (1999), Marceau and Hay (1999), and Wu (1999), among others.

Flow systems depend on both the hydrogeologic characteristics of the soil/rock material and landscape position. Zones of high permeability in the subsurface function as drains, which cause enhanced downward gradients in the material overlying the upgradient part of the high-permeability zone (Freeze and Witherspoon 1967). Areas of pronounced topographic relief tend to have dominant local flow systems, and areas of nearly flat relief tend to have dominant intermediate and regional flow systems.

In topography-controlled flow regimes, groundwater moves in systems of predictable patterns, and various identifiable natural phenomena are regularly associated with different segments of the flow systems. The interactions of streams, lakes, and wetlands with groundwater are governed by the positions of the water bodies with respect to groundwater flow systems, geologic characteristics of their beds, and their climatic settings (Winter 1999). Therefore, for a thorough understanding of the hydrology of surface-water bodies, all three factors should be taken into account. As Tóth (1999) points out, such recognition was not appreciated until the 1960s (Tóth 1962, 1963; Freeze and Witherspoon 1967), when the systems-nature of groundwater flow became understood. This recognition of the systems-nature of subsurface water flow has provided a unifying theoretical back-

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ground for the study and understanding of a wide range of natural processes and phenomena and has thus shown flowing groundwater to be a general geologic agent (Tóth 1999). For a comprehensive outline of groundwater recharge processes from the systems perspective as well as estimation methodologies, the reader is referred to Scanlon et al. (2002), Sophocleous (2002), and other articles in this theme issue.

The spatial distribution of flow systems also influences the intensity of natural groundwater discharge. The main stream of a basin may receive groundwater from the area immediately within the nearest topographic high and possibly from more distant areas. However, as the works of Tóth (1962, 1963, 1966, 1999), Meyboom et al. (1966), Meyboom (1966, 1967), and others have shown, groundwater discharge is not only confined along the stream channel but also extends throughout the discharge area downgradient from the basin *hinge line* [i.e., the imaginary line separating areas of upward (discharge) from downward (recharge) flow]. Therefore, if baseflow calculations are used as indicators of average recharge, significant error may be introduced, because baseflow would represent only a relatively small part of the total discharge occurring downgradient from the hinge line. Hence, baseflow analysis based on lumped-parameter concepts may give numerical results that are of little practical use unless examined in the light of spatial flow characteristics (Domenico 1972).

A geomorphologic perspective is also helpful in characterizing larger-scale GW–SW interactions and in estimating the extent and location of such interfaces. For example, Larkin and Sharp (1992) classify stream–aquifer systems (based on the predominant regional groundwater flow component) as (1) *underflow-component dominated* (the groundwater flux moves parallel to the river and in the same direction as the streamflow); (2) *baseflow-component dominated* (the groundwater flux moves perpendicular to or from the river depending on whether the river is effluent or influent, respectively; see the next section); or (3) mixed. They conclude that the dominant groundwater flow component, baseflow or underflow, can be inferred from geomorphologic data, such as channel slope, river sinuosity, degree of river incision through its alluvium, the width-to-depth ratio of the bankfull river channel, and the character of the fluvial depositional system (Larkin and Sharp 1992). The underflow component is demonstrably predominant in systems with large channel gradients, small sinuosities, large width-to-depth ratios, and low river penetrations; and, in fluvial depositional systems of mixed-load to bed-load character, in upstream and tributary reaches and valley-fill depositional environments. Baseflow-dominated systems have characteristics typical of suspended-load streams with the opposite to the aforementioned geomorphic attributes for systems dominated by the underflow component. Mixed-flow systems occur where the longitudinal valley gradient and channel slope are virtually the same and also where the lateral valley slope is negligible (Larkin and Sharp 1992).

Mechanisms of GW–SW Interactions

Basic Concepts

Hydrologic interactions between surface and subsurface waters occur by subsurface lateral flow through the unsaturated soil and by infiltration into or exfiltration from the saturated zones. Also, in the case of karst or fractured terrain, interactions occur through flow in fracture/solution channels. Water that enters a surface-water body promptly, in response to such individual water input events as rain or snowmelt, is known as *event flow*, *direct flow*, *storm flow*, or *quick flow*. This water is distinguished from *baseflow*, or water that enters a stream from persistent, slowly varying sources and maintains stream flow between water-input events. Although some baseflow is derived from drainage of lakes or wetlands, or even from the slow drainage of relatively thin soils on upland hill slopes, most baseflow is supplied from groundwater flow. Subsurface flow can also enter streams quickly enough to contribute to the event response. Such flow is called *subsurface storm flow* or *interflow*. Beven (1989) defines interflow as the near-surface flow of water within the soil profile resulting in seepage to a stream channel within the time frame of a storm hydrograph. Interflow involves both unsaturated and saturated flows, the latter being in zones of limited vertical extent caused by soil horizons impeding vertical percolation. If interflow encounters a seepage face, the interflow process may grade into *return flow* by which subsurface water can contribute to *overland flow* (Dunne and Black 1970). Results from environmental-isotope studies (Sklash and Farvolden 1979) indicate that interflow may be primarily a displacement process in which the storm rainfall induces the displacement of subsurface-stored water (pre-event water).

In general, subsurface flow through porous media is sluggish. The mechanisms by which subsurface flow enters streams quickly enough to contribute to streamflow responses to individual rainstorm and snowmelt inputs (storm hydrograph), although still not fully understood, are summarized in various publications (including Ward 1984; Beven 1989; Dingman 1994). Beven (1989) identifies four mechanisms to account for fast subsurface contributions to the storm hydrograph: (1) translatory flow, (2) macropore flow, (3) groundwater ridging, and (4) return flows.

Translatory flow (Hewlett and Hibbert 1967), also known as plug flow or piston flow, is easily observed by allowing a soil column to drain to field capacity in the laboratory and slowly adding a unit of water at the top. Although some water flows from the bottom almost immediately, it is not the same water that was added at the top. Rapid subsurface responses to storm inputs may be the result of fast flow through larger noncapillary soil pores, or *macropores* (Beven and Germann 1982). Normally, the assumption is made that water does not enter a large noncapillary pore unless it is at or above atmospheric pressure (Taylor and Ashcroft 1972). Such conditions only occur either below the water table or after ponding during rainfall at the soil surface.

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A third concept used to account for rapid subsurface responses is *groundwater ridging* (Sklash and Farvolden 1979), which describes the large and rapid increases in hydraulic head in groundwater during storm periods. Rapid changes in near-stream water-table levels are also described by Ragan (1968), O'Brien (1980), and Bonell et al. (1981) and are attributed to the conversion of a tension-saturated zone or capillary fringe overlying the pre-storm water table to a zone of positive potentials. When even a small amount of water percolates to the top of this zone, the menisci that maintain the tension saturation are obliterated and the pressure state of the water is immediately changed from negative to positive (Gillham 1984). This phenomenon thus produces a disproportionately large rise in the near-stream water table (sloping toward the stream). As a result, an increase occurs in the net hydraulic gradient toward the stream and/or the size of the seepage face, thus enhancing groundwater fluxes to the stream. The streamflow contribution induced thereby may greatly exceed the quantity of water input that induced it.

If the water table and capillary fringe are close to the soil surface, then only small amounts of applied water are necessary to saturate the soil profile completely. This saturation might lead to the discharge of subsurface water onto the surface as *return flow* (Dunne and Black 1970). The contributing area of return flow could expand rapidly in an area where the capillary fringe is close to the surface. The contributing area would also be expected to serve as an area of saturation-excess surface-runoff production, so that discharge into the stream would be expected to be a mixture of both event and pre-event water (Beven 1989).

The response of any particular catchment may be dominated by a single mechanism or by a combination of mechanisms, depending on the magnitude of the rainfall event, the antecedent soil-moisture conditions of the catchment, and/or the heterogeneity in soil hydraulic properties (Sklash 1990; Wood et al. 1990). Thus, during any particular storm, different mechanisms generate runoff from different parts of a catchment. Surface runoff from these (partial) contributing areas is generated either (1) by the *infiltration excess* mechanism (Fig. 1a,b), where the rainfall rate exceeds the infiltration capacity of the soil; or (2) from rainfall in areas of soil saturated by a rising water table, even in high-permeability soils. Such *saturation-excess overland flow* is represented as mechanisms c and d in Fig. 1.

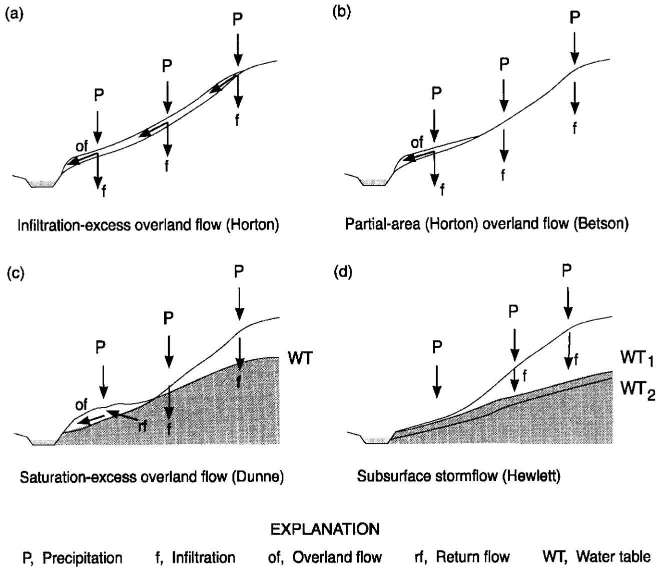

Fig. 1 Mechanisms of runoff production. (Adapted from Beven 1986)

Larger-Scale Interactions

The larger-scale hydrologic exchange of groundwater and surface water in a landscape is controlled by (1) the distribution and magnitude of hydraulic conductivities, both within the channel and the associated alluvial-plain sediments; (2) the relation of stream stage to the adjacent groundwater level; and (3) the geometry and position of the stream channel within the alluvial plain (Woessner 2000). The direction of the exchange processes varies with hydraulic head, whereas flow (volume/unit time) depends on sediment hydraulic conductivity. Precipitation events and seasonal patterns alter the hydraulic head and thereby induce changes in flow direction. Two net directions of water flow are distinguished: (1) the influent condition, where surface water contributes to subsurface flow; and (2) the effluent condition, where groundwater drains into the stream. On the other hand, variable flow regimes could alter the hydraulic conductivity of the sediment via erosion and deposition processes and thus affect the intensity of the GW–SW interactions.

Brunke and Gonser (1997) comprehensively summarize the interactions between rivers and groundwater. Under conditions of low precipitation, baseflow in many streams constitutes the discharge for most of the year (effluent condition). In contrast, under conditions of high precipitation, surface runoff and interflow gradually increase, leading to higher hydraulic pressures in the lower stream reaches, which cause the river to change from effluent to influent condition, infiltrating its banks and recharging the aquifer. During flooding, the river loses water to bank infiltration, which reduces the flood level and recharges the aquifer. The volume of this *bank storage* depends on duration, height, and shape of the flood hydrograph, as well as on the transmissivity and storage capacity of the aquifer. During a dry season, the release of stored water compensates for a decrease in stream discharge. In some river reaches, the water released to the river from bank storage originating from flood runoff exceeds groundwater discharge under baseflow conditions. Thus, successive discharge and recharge of the aquifer has a buffering effect on the runoff regimes of rivers (Brunke and Gonser 1997).

Groundwater exfiltration occurs diffusely or at discrete locations. Perennial, intermittent, or ephemeral stream-discharge conditions depend on the regularity of

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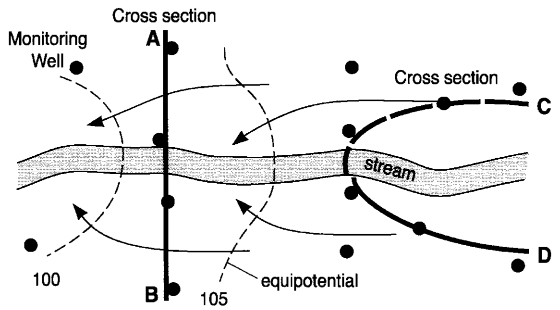

Fig. 2 Map of a portion of a fluvial plain and stream channel instrumented with monitoring wells (*black dots*). Cross section *C–D* is located along a flow line, whereas *A–B* is not. (Woessner 2000)

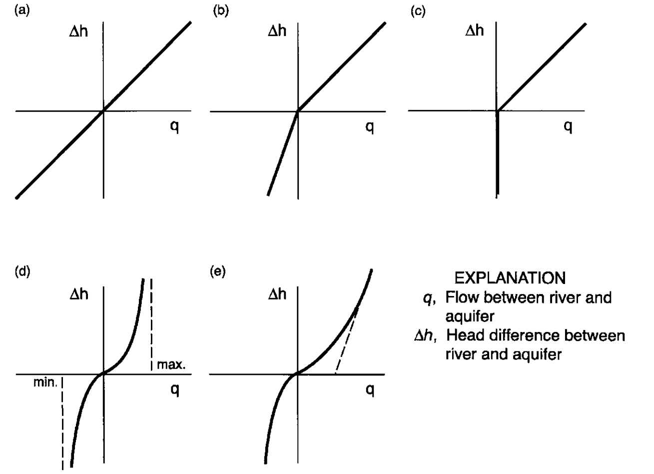

Fig. 3 Illustrations of the various mechanisms describing flow between the river and aquifer (q) as a function of the difference between the river and aquifer heads (Δh). **a–d** represent various flow conditions; see text for additional explanations. (Adapted from Rushton and Tomlinson 1979)

baseflow, which is determined by the groundwater level. In *perennial* streams, baseflow is more-or-less continuous, whereby these streams are primarily effluent and flow continuously throughout the year (Gordon et al. 1992). *Intermittent* streams receive water only at certain times of the year and are either *influent* (losing) or *effluent* (gaining), depending on the season. In *ephemeral* streams the groundwater level is always beneath the channel, so they are exclusively influent when they are flowing (Gordon et al. 1992).

When the stream channel is generally oriented parallel to the alluvial plain, *gaining*, *losing*, and *parallel-flow channels* are most likely to occur. Parallel-flow channels occur when the channel stage and groundwater head are equal. *Flow-through reaches*, which occur where the channel stage is less than the groundwater head on one bank and is greater than the groundwater head at the opposite bank, most often exist where a channel cuts perpendicular to the fluvial-plain groundwater flow field (Hoehn 1998; Huggenberger et al. 1998; Wroblicky et al. 1998; Woessner 2000).

As Woessner (2000) points out, the construction of alluvial-plain and stream cross sections to show groundwater flow and quality along a flowpath to the stream requires careful consideration (e.g., Harvey and Bencala 1993; Wondzell and Swanson 1996). Figure 2 (Woessner 2000) illustrates the proper location of wells (cross section C–D) to accomplish these purposes for a section of gaining stream. Cross section A–B may be used to illustrate the geology; however, within the alluvial plain flow system, it is not parallel to a flow line. Proper conceptualization and measurement of the flow field in the near-channel area results in appropriate locations of hydrogeologic cross sections (Woessner 2000).

Quantitative Analysis

For hydraulically connected stream–aquifer systems, the resulting exchange flow is a function of the difference between the river stage and aquifer head. A simple approach to estimate flow is to consider the flow between the river and the aquifer to be controlled by the same mechanism as leakage through a semi-impervious stratum in one dimension (Rushton and Tomlinson 1979). This mechanism, based on Darcy's law, where flow is a direct function of the hydraulic conductivity and head difference, can be expressed as

$$q=k\Delta h, \tag{1}$$

where $\Delta h=h_a-h_r$, (h_a is aquifer head, and h_r is river head); q is flow between the river and the aquifer (positive for baseflow – for gaining streams; and negative for river recharge – for losing streams); and k is a constant representing the streambed leakage coefficient (hydraulic conductivity of the semi-impervious streambed stratum divided by its thickness).

The simple mechanism described by Eq. (1) can be used to represent both baseflow and river recharge, depending on the sign of Δh. Figure 3a implies that the mechanisms for flow from the aquifer to the river (baseflow) and from the river to the aquifer (river recharge) are the same, although, in practice, the mechanisms representing the two processes can be different. Figure 3b illustrates the situation where the rate of flow from the river to the aquifer is slower than the rate of flow from the aquifer to the river, and Fig. 3c illustrates the situation where no flow can occur from the river to the aquifer.

The assumption of a linear relationship between q and Δh is often too simplistic. Several publications, including Rushton and Tomlinson (1979), note that total leakage (baseflow) during streamflow recession is largely independent of the leakage coefficient, k. Also, at times of very high recharge, the leakage calculated from Eq. (1) is much greater than would occur in practice and takes no account of the increased resistance to the passage of water as its volume increases. Thus Rushton and Tomlinson (1979) propose that a nonlinear relationship

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represented in Fig. 3d might be a more appropriate model of this increased resistance at high flows:

$$q=k_1[1-\exp(-k_2\Delta h)], \quad (2)$$

where k_1 and k_2 are constants. This relationship permits a rapid increase in the flow for small head changes when the head difference is small, but postulates maximum flows that cannot be exceeded as long as the head difference becomes larger.

The linear relationship described by Eq. (1) and the nonlinear relationship described by Eq. (2) have different advantages. However, in cases where the suggestion that a maximum flow rate exists is not acceptable, Rushton and Tomlinson (1979) propose a combination of linear and nonlinear relationships:

$$q=k_1\Delta h+k_2[1-\exp(-k_3\Delta h)], \quad (3)$$

where k_1, k_2, and k_3, are constants. This relationship is illustrated in Fig. 3e. Because the exponential term is relatively large for small values of Δh, the nonlinear relationship dominates for small head differences, whereas for larger head differences the linear relationship becomes more important. However, when the aquifer head is lower than the river head, an exponential relationship with a maximum flow is used (Rushton and Tomlinson 1979; Fig. 3e).

In areas of low precipitation, the water table is usually well below the base of the channel; as a result, channel seepage is often the largest source of recharge (Stephens 1996). The magnitude of the infiltration depends upon a variety of factors, such as vadose-zone hydraulic properties, available storage volume in the vadose zone, channel geometry and wetted perimeter, flow duration and depth, antecedent soil moisture, clogging layers on the channel bottom, and water temperature. If the value of the depth of the water table below the stream stage is greater than twice the stream width, the seepage begins to rapidly approach the maximum seepage for an infinitely deep water table (Bouwer and Maddock 1997).

Key Theoretical and Field Studies

Only a few field investigations detail the pathways of water migration from ephemeral stream channels or from canals to the water table (Stephens 1996). The problem has been addressed by mathematical modeling by Riesenauer (1963), who used a variably saturated finite-difference model to study seepage from an unlined irrigation canal (Fig. 4). The most interesting feature of his simulation is the distributions of steady-state moisture content and pressure head, which reveal incomplete saturation through the vadose zone beneath the edge of the channel, even at steady state. Owing to the relatively great depth to the water table, no groundwater mound would rise through the vadose zone to intersect the channel. This would be true at all times, even if the flow duration were sufficiently long for the vadose zone to reach steady-state moisture distribution, as long as the aquifer can transmit the recharge away from the area (Stephens 1996). For relatively deep water-table conditions, saturated zones do occur beneath the channel, but only to a limited depth. The base of the saturated zone beneath the channel would be regarded as an inverted water table. Unsaturated flow would occur between the inverted water table (0.33 cm^3/cm^3 contour in Fig. 4) and the regional water table. Where the water table is relatively shallow, however, complete saturation may exist between the channel and the regional water table.

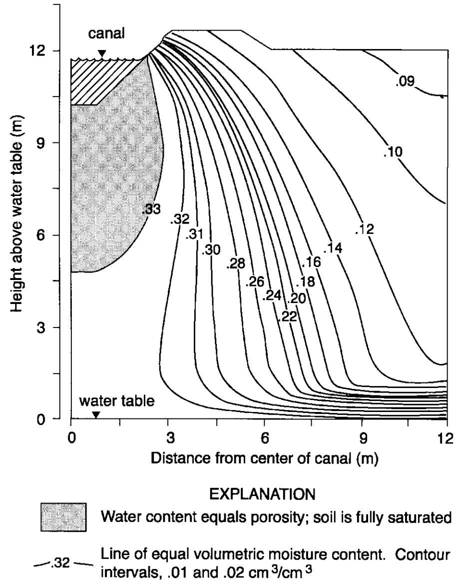

Fig. 4 Distribution of steady-state moisture content below a canal in a homogeneous soil. (Adapted from Riesenauer 1963)

Transient numerical simulations by Peterson and Wilson (1988) demonstrate the importance of recognizing unsaturated flow when predicting the increase in recharge from stream infiltration that occurs when water tables are lowered by groundwater pumping. This unsaturated-flow condition usually occurs where a relatively low-permeable clogging layer is present on the channel bottom. If the free surface on a groundwater mound rises from the shallow regional water table to intercept the water level in the channel, the stream–aquifer system is *hydraulically connected* (Fig. 5). On the other hand, if unsaturated sediments exist between the channel and the regional water table, then the system may be *hydraulically disconnected* (Stephens 1996). However, the simulations of Peterson and Wilson (1988) show that even when the unsaturated condition is present, the stream and aquifer may in fact be connected, in the sense that further lowering of the regional water table could increase channel losses. At some critical depth to the water table, however, further lowering has no influence on channel losses, as previously mentioned (Bouwer and Maddock 1997). At this depth, which depends mostly on soil properties and head in the channel, the aquifer becomes hydraulically disconnected from the stream.

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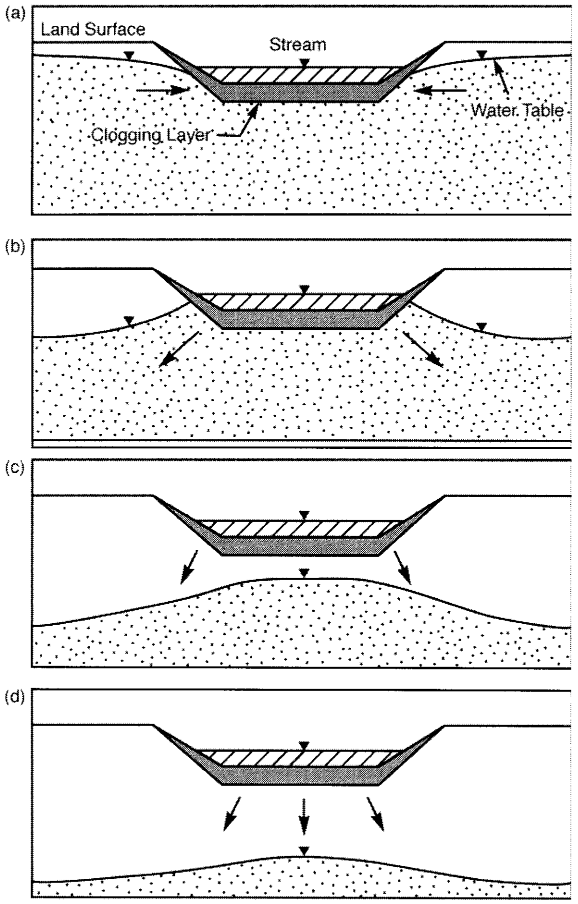

Fig. 5 Stream–aquifer relationships for the case of a clogged streambed: **a** connected gaining stream; **b** connected losing stream; **c** disconnected stream with a shallow water table; **d** disconnected stream with a deep water table. (From Peterson and Wilson 1988)

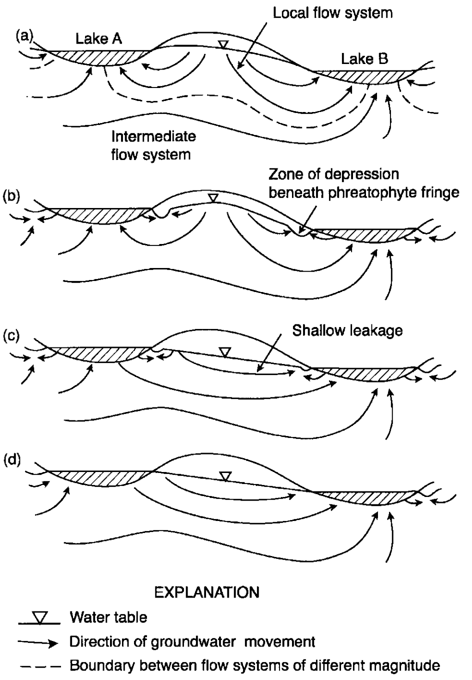

Fig. 6 Flow conditions near two permanent lakes with **a** a spring condition of discharge from local and intermediate systems, **b** a summer condition of seepage toward the phreatophyte fringe, **c** deterioration of local flow in the absence of recharge, and **d** a fall and winter condition for the deteriorated system, where shallow movement is superimposed on the intermediate system. (Meyboom 1967)

The effect of climate on seepage distribution in surface-water bodies is reviewed by Winter (1999). The most dynamic boundary of most groundwater flow systems is the water table. The configuration of the water table changes continually in response to recharge to and discharge from the groundwater system. Winter (1983) evaluates the effect of the distribution of recharge on the interaction of surface water and groundwater, using a variably saturated subsurface-flow model. The principal results of that study indicate that recharge is focused initially where the unsaturated zone is thin relative to adjacent areas. Recharge then progresses laterally over time to areas that have thicker unsaturated zones. This process has significant implications for the interaction of groundwater and surface water, because the unsaturated zone in most landscapes is thin in the vicinity of surface water, and, in fact, has zero thickness at the shoreline (Winter 1999). The changing volumes and distribution of recharge results in dynamic growth and dissipation of transient, local, groundwater flow systems directly adjacent to surface water, which causes highly variable seepage conditions in the near-shore beds of surface water.

Because of the shallow depth of groundwater near surface water, transpiration from groundwater by near-shore vegetation often intercepts groundwater that would otherwise discharge to surface water. Furthermore, transpiration from groundwater commonly creates cones of depression that cause surface water to seep out through the near-shore parts of its beds (Meyboom 1966; Winter and Rosenberry 1995).

Field studies have resulted in increased understanding of groundwater flow processes associated with surface water (Domenico 1972). For example, in areas of hummocky terrain, ephemeral water bodies have been observed to function as recharge points during spring and early summer, and as discharge points during summer and autumn. On the other hand, permanent lakes are usually areas of permanent groundwater discharge (Meyboom 1966, 1967). Four typical flow conditions near permanent lakes are shown in Fig. 6, which demonstrate (a) a spring condition of discharge from local and intermediate flow systems; (b) a summer condition of seepage toward the phreatophyte fringe surrounding the lake; (c) a deterioration of local flow owing to insufficient recharge, which produces shallow movement from lake A to lake B; and (d) an autumn and winter condition for the deteriorated system, where shallow movement occurs from lake A to lake B, superimposed on the intermediate flow system. These studies demonstrate that lakes are dynamic bodies, and the movement of groundwater in their vicinity cannot

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be described in terms of static analysis. A set of hydraulic-potential measurements gives information about movement only at a particular moment in time (Domenico 1972). An identical attitude applies also to the chemical character of lakes, where a chemical analysis of a single water sample applies only to a specific set of circumstances (Livingstone 1963; Garrels and Mackenzie 1967).

Combined field and theoretical modeling studies have further contributed to our understanding of GW–SW processes. For example, as Winter (1999) points out, upward breaks in slope of the water table result in upward components of groundwater flow beneath the area of lower slope, and downward breaks in slope of the water table result in downward components of groundwater flow. These flow patterns apply to parts of many landscapes. For example, drainage basins with concave hill-slope profiles result in subsurface flow lines converging in such "hill-slope hollows" and concave slope breaks. At these sites, the hydraulic gradient inducing subsurface flow from upslope is greater than that inducing downslope transmission, resulting in saturated areas from below. The upward-moving groundwater near upward breaks in slope of the water table commonly results in: (1) groundwater discharge to surface water, because water tables generally have a steeper slope relative to the flat surface of surface-water bodies; (2) the presence of wetlands at the edges of river valleys and other flat landscapes adjacent to uplands; and (3) the formation of saline soils, especially in semiarid and arid landscapes (Winter 1999). The groundwater flux through a surface-water bed or to land surface associated with these breaks-in-slope is not uniformly distributed areally (Winter 1999). Where groundwater moves to or from a surface-water body underlain by isotropic and homogeneous porous media, the flux is greatest near the shoreline and decreases approximately exponentially away from the shoreline (McBride and Pfannkuch 1975).

Ecological Significance of GW–SW Interactions

Water flows not only in the open stream channel but also through the interstices of stream-channel and bank sediments, thus creating a mixing zone with subsurface water. The region of mixing between subsurface water and surface water is the *hyporheic zone*, HZ, which is a region of intensified biogeochemical activity (Fig. 7; Grimm and Fisher 1984; Duff and Triska 1990; Triska et al. 1993a, 1993b). Subsurface exchanges affect the type and increase the rate of material transformation as water moves downstream. For example, as Findlay (1995) points out, the time-of-travel estimated for water in the stream channel might be too short to permit significant mineralization of organic nutrients. However, if hyporheic exchange is an important process, residence time within a reach and contact with subsurface sediments may result in dramatic alterations in material transported from the catchment to the receiving body of water. Therefore, an important aspect of GW–SW interchange

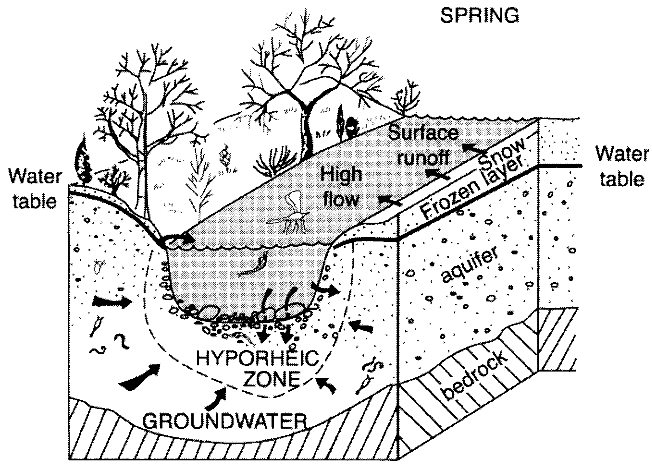

Fig. 7 Descriptive model of the dynamics of the hyporheic zone and surrounding surface water and groundwater. Direction of water movement is indicated by *arrows*, and their size indicates the relative magnitude of flow. Stages in the life cycle, location, and size of representative invertebrates are also shown, as are features of riparian vegetation. (Adapted from Williams 1993)

is that surface water in streams, lakes, and wetlands repeatedly interchanges with nearby groundwater (Winter et al. 1998). Thus, the length of time water is in contact with mineral surfaces in its drainage basin is extended after the water first enters a stream, lake, or wetland. An important consequence of these continued interchanges between surface water and groundwater is their potential to further increase the contact time between water and chemically reactive geologic materials.

Maddock et al. (1995) reviewed various stream–aquifer studies that show how flow paths within the bed are primarily a function of the surface morphology of the bed and hydrologic features. Laboratory flume experiments (Vaux 1968) indicate that stream-water downwelling occurs where the longitudinal bed profile is convex or where an increase in stream-bed elevation exists, as in the transition from an upstream pool to a riffle. Where the shape of the bed is concave or where a decrease in streambed elevation exists, as in the transition from a riffle to a downstream pool, upwelling in the substratum occurs. In summary, water is likely to enter the stream bed at the crest of riffles and re-emerge at the downstream end. The downstream end is also the place where deeper groundwater is likely to emerge. Certain species of stream fauna are reported to rely on the upwelling of groundwater for their survival (Creuzé des Châtelliers and Reygrobellet 1990).

As Woessner (2000) points out, the stream-bed topography and the corresponding water exchange causes localized flow systems within the beds of overall gaining and losing stream reaches. Thus, obstacles and stream-bed roughness tend to cause an influx of channel water into the hyporheic zone, even in effluent (gaining) stream reaches (Brunke and Gonser 1997).

The greater penetration of surface water at gravel riffle sites is reflected in the higher hyporheic tempera-

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tures of the riffle gravel during the summer months. A thermally induced, density-dependent mechanism that causes convection of surface water into the interstices is proposed by Whitman and Clark (1982). Cooler stream water tends to displace warmer interstitial water seasonally during winter and diurnally during the night in summer and autumn. In spring and summer, warming of the surface water during the day inhibits this mixing process (Brunke and Gonser 1997).

Because of their high hydraulic conductivity and short residence times, preferential subsurface flow paths, such as paleo-channels, transport water with physico-chemical properties similar to the surface water into deeper alluvial layers beneath the flood plain. These subsurface flows extend direct connections between rivers and groundwater into the subterranean landscape and may sustain a high interstitial biodiversity and biomass by delivering resources. Ward et al. (1994) propose that paleochannels in the alluvium of the Flathead River in Montana, USA, are a significant factor influencing the spatial distribution of crustaceans. Sophocleus (1991) indicates that some buried channels, when in contact with active surface channels, are avenues of fast transmission of pressure pulses resulting from surface-channel flooding, causing water-level fluctuations in distant wells screened in these buried channels.

In conclusion, as Brunke and Gonser (1997) point out, ecological studies concerning the faunal composition, distribution, and abundance of the GW–SW interface reveal an extraordinary patchiness and variability, owing to the inherent heterogeneity of the physical parameters. The main determinants of the interstitial habitat of rivers are the usable pore space, dissolved-oxygen concentrations, temperatures, organic matter, and nutrient contents, all of which are influenced on a higher hierarchical scale by the sediment facies, the hydrology, and climate (Brunke and Gonser 1997). A large body of literature exists on biochemical and water-quality impacts on GW–SW interactions (for example, Schwarzenbach et al. 1983; Von Gunten et al. 1991; Bourg and Bertin 1993; Brunke et al. 1998; Dahm et al. 1998; Hedin et al. 1998; and references therein), but due to space limitations these aspects are not covered here.

Human Impacts and Water-Resource Depletion and Sustainability

Despite its general abundance, water does not always occur in the place, at the time, or in the form desired. People strive to grow crops and other water-consuming products in semiarid regions, and they attempt to use water simultaneously as a pure source and, deliberately or inadvertently, as a dump for waste. Consequently, society faces increasingly serious water-management problems (National Research Council 1981; Sophocleus 1997, 1998, 2000a, 2000b).

The decline of groundwater levels around pumping wells near a surface-water body creates gradients that capture some of the ambient groundwater flow that would have, without pumping, discharged as baseflow to the surface water. At sufficiently large pumping rates, these declines induce flow out of the body of surface water into the aquifer, a process known as *induced infiltration*, or *induced recharge*. The sum of these two effects leads to *streamflow depletion*. Quantifying the amount of induced infiltration, which is a function of many factors, is an important consideration in *conjunctive water use* as water demand increases and the reliability of surface supplies is threatened by streamflow depletion. As discussed in the previous section, stream–aquifer interactions are also important in situations of groundwater contamination by polluted surface water, and in situations of degradation of surface water by discharge of saline or other low-quality groundwater. Because of the potential for pollution of both groundwater and surface water from varied sources and by varied pollutant species, quantifying the amount of induced infiltration is also an important factor in evaluating the reliability of well-water quality.

Human Impacts

The ecological integrity of groundwater and fluvial systems is often threatened by human activities, which can reduce connectivity, alter exchange processes, and lead to toxic or organic contamination. Brunke and Gonser (1997) reviewed human impacts on alluvial hydrosystems, and Fig. 8 summarizes human-induced disruptions of hydrologic-exchange processes and their ecological consequences. The following draws on their review.

Organic and toxic contamination in surface water can be transferred to the groundwater in influent reaches. The quality of the downwelling surface water is normally altered during its passage through the first few meters of the infiltrated sediments. However, this may not be the case for persistent organic compounds, such as chloroform and inorganic pollutants, which may contaminate extensive areas of groundwater (Schwarzenbach et al. 1983; Santschi et al. 1987; Whittemore et al. 2000).

Increased sewage loading often leads to clogging by promoting the development of dense algal mats, or by causing sedimentation of an organic layer on the river bed. The extent of these processes is related directly to land-use practices that increase suspended particulate matter (*seston*) and sediment loading (Karr and Schlosser 1978). In many streams, gradual clogging (*colmation*) occurs naturally through the siltation of fine material during low discharge, alternating with a reopening of the interstices during flooding or exfiltration (*decolmation*). Although increased current velocity usually flushes fine material out of the upper layers, only bed-load movement opens deeper interstices. A balanced relationship between clogging and streambed scouring can be disturbed by increased organic and fine sediment inputs, hydroengineering, and increased river-bank filtration for the supply of drinking, industrial, and irrigation waters. Each of these factors is capable of causing permanent

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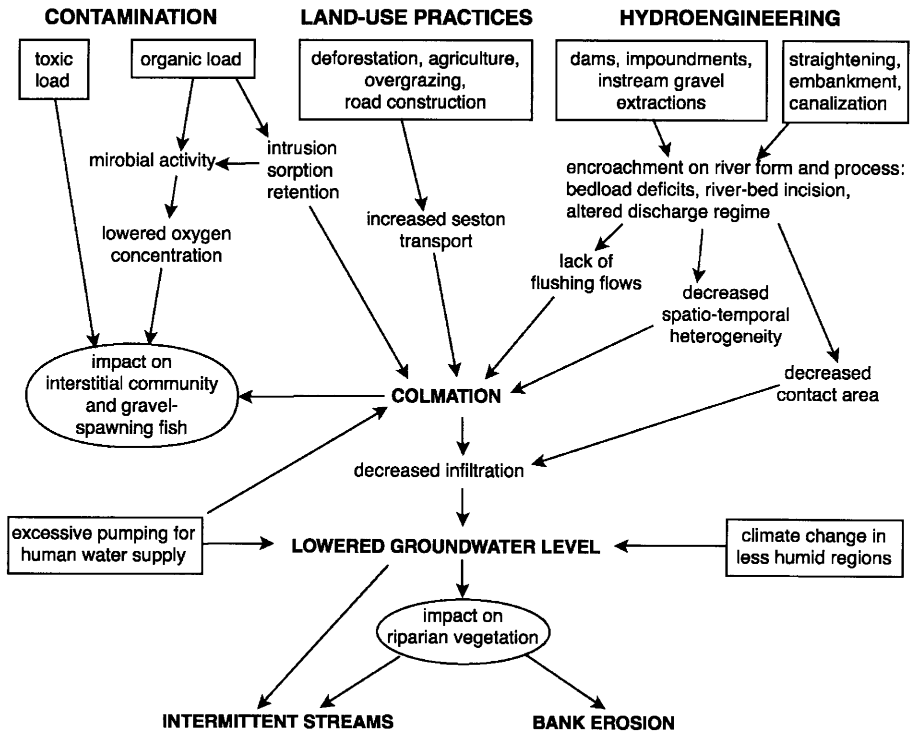

Fig. 8 Human-induced impacts that promote clogging of stream-bed sediments, and their ecological consequences. (Brunke and Gonser 1997)

clogging. As Brunke and Gonser (1997) point out, clogging exerts severe impacts on the renewal of groundwater through river-bank filtration and the development and colonization of invertebrates and fish. Furthermore, alterations of the fluvial temperature regime are possible, with wide-ranging implications for the biota. The same authors also refer to a case study where the mechanical opening of a clogged section of the stream bed of the Rhine River, Germany, near a drinking-water bank-filtration site induced a 1-m rise in the water table near the river, but after a few weeks, the opened section had become sealed again. Conversely, a clogged bed may act as an intrusion barrier that prevents the contamination of groundwater by polluted surface water (Younger et al. 1993).

As Brunke and Gonser (1997) also indicate, river-bed incision results from bed-load deficits due to sediment retention by impoundments and from increased transport capacity following channel straightening. Such incision determines the dominant subsurface flow direction and lowers the adjacent groundwater level (Galay 1983; Golz 1994). Desiccation of the floodplain endangers aquatic and riparian vegetation, reduces the connectivity and spatio-temporal heterogeneity of former channels, and ultimately alters biodiversity patterns (Dister et al. 1990; Allan and Flecker 1993; Bornette and Heiler 1994). The vegetation contributes to the resisting forces by stabilizing the bank material with roots and decreasing the velocity of floodwaters. Thus, riparian vegetation that has been impacted by a lowered water table enhances the danger of stream-bank erosion during flooding (Keller and Kondolf 1990). Changes from perennial to intermittent flow may alter bank vegetation and moisture content, and hence fluvial geomorphology (Keller and Kondolf 1990).

Human impacts on terrestrial and aquatic systems may lead to reductions in exchange processes that connect running waters to their surroundings, and thus diminish the ecological integrity of subsurface and surface-water ecosystems. By preventing communication between these systems, cascading effects in ecosystem structure and function occur (Fig. 8; Brunke and Gonser 1997), with consequences on water-resource depletion and water sustainability.

Water-Resource Depletion and Sustainability

The topics of water-resource depletion, GW–SW interactions, and water-resource sustainability were recently reexamined by Sophocleous (1997, 1998, 2000a, 2000b). To understand this depletion, a thorough knowledge of the hydrologic principles, concisely stated by Theis (1940), is required. Under natural conditions, prior to development by wells, aquifers approach a state of dynamic equilibrium: over hundreds of years, wet years, when recharge exceeds discharge, are offset by dry years, when discharge exceeds recharge. Discharge from wells upsets this equilibrium by producing a loss from aquifer storage; a new state of dynamic equilibrium is approached when there is no further loss or minimal loss from storage. This state is accomplished either by an increase in recharge, a decrease in natural discharge, or a combination of the two.

Consider a stream–aquifer system such as an alluvial aquifer discharging into a stream, where the term "stream" is used in the broadest sense of the word to in-

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clude rivers, lakes, ponds, and wetlands. A new well drilled at some distance from the stream and pumping the alluvial aquifer forms a cone of depression. The cone grows as water is taken from storage in the aquifer. Eventually, however, the periphery of the cone arrives at the stream. At this point, discharge from the aquifer to the stream appreciably diminishes or ceases, or water starts to flow from the stream into the aquifer. The cone continues to expand with continued pumping of the well until a new equilibrium is reached, in which induced recharge from the stream balances the pumping.

The length of time, *t*, before an equilibrium is reached depends upon (1) the *aquifer diffusivity* (expressed as the ratio of aquifer transmissivity to storativity, *T/S*), which is a measure of how fast a transient change in head is transmitted throughout the aquifer system; and (2) the distance, *x*, from the well to the stream. For radial flow of groundwater, a tenfold increase in distance from the surface-water body causes a 100-fold delay in the response time, whereas a change in diffusivity is linearly proportional to the response time (Balleau 1988). Generally, if the wells are distant from the stream, tens or hundreds of years must pass before their influence on streamflow is felt.

Once the well's cone has reached an equilibrium size and shape, all of the pumping is balanced by flow diverted from the stream. In that case, a water right to withdraw groundwater from the well, as described, becomes a water right to divert from the stream at the same rate. A crucial point, however, is that before equilibrium is reached (that is, before all water is coming directly from the stream), the two rights are not the same (DuMars et al. 1986). Until the perimeter of the cone reaches the stream, the volume of the cone represents a volume of water that has been taken from storage in the aquifer, over and above the subsequent diversions from the river. It is this volume that is called *groundwater depletion*. Thus, groundwater sources include groundwater (or aquifer) storage and *induced recharge* of surface water. The timing of the change from groundwater depletion (or mining) to induced recharge from surface-water bodies is key to developing sound water-use policies (Balleau 1988).

The shape of the *transition* or *growth curve* for an idealized, two-dimensional, homogeneous, and isotropic system is shown in Fig. 9 in nondimensional form, based on Glover's (1974) analytical solution and tabulation. In Fig. 9, the percent of groundwater withdrawal derived from groundwater storage is plotted on the *y*-axis against dimensionless time [or normalized time, $t^*=\{4(T/S)/x^2\}t$] on the *x*-axis. For example, if groundwater storage is 85% of the water source after 1 month (or 1 year) of pumping, it ends up being only 5% of the water pumped coming from aquifer storage after 1,000 months (or 1,000 years) of pumping. The general shape of the transition curve is retained in systems with apparently different boundaries and parametric values (Balleau 1988). The rate at which dependence on groundwater storage (as shown in the left portion of the graph) converts to dependence on surface-water depletion (as shown in the right portion of the graph) is highly variable and is particular to each case.

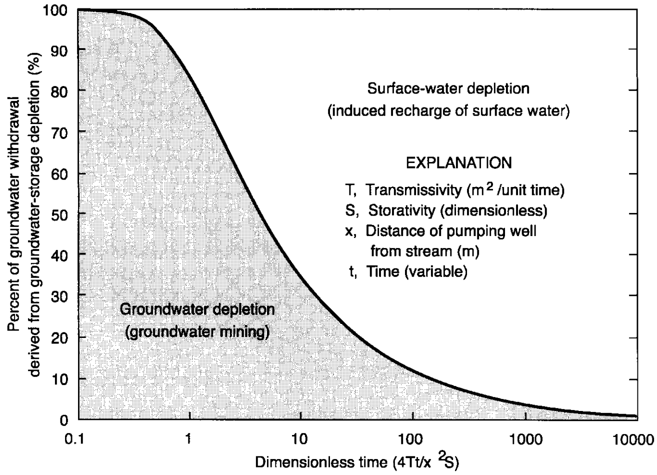

Fig. 9 Transition from reliance upon groundwater storage to induced recharge of surface water. (Adapted from Balleau 1988)

The initial and final phases of the transition curve (Fig. 9) are separated in time by a factor of nearly 10,000. As the example above shows, full reliance on indirect (induced) recharge takes an extremely long time. The distinct category of groundwater mining depends entirely upon the time frame. Initially, all groundwater developments mine water, but ultimately they do not (Balleau 1988).

Aquifer drawdown and surface-water depletion are two results of groundwater development that affect policy. Both are fundamentally related to pumping rate, aquifer diffusivity, location, and time of pumpage. The natural recharge rate is unrelated to any of these parameters. Nonetheless, policy makers often use natural recharge to balance groundwater use, a policy known as *safe yield*. However, this policy completely ignores natural groundwater discharge, and eventually it leads to the drying of springs, marshes, and riverine–riparian systems that constitute the natural discharge areas of groundwater systems, as has already happened in many parts of the world (Sophocleous 1997, 1998, 2000a, 2000b). As Balleau (1988) points out, public purposes are not served by adopting the attractive fallacy that the natural recharge rate represents a safe rate of yield.

To illustrate the influence of the dynamics of a groundwater system in response to development, Bredehoeft et al. (1982) chose a simple, yet realistic, system for analysis – a closed intermontane basin of the sort common in the western states of the US (Fig. 10). Under predevelopment conditions, the system is in equilibrium: phreatophyte evapotranspiration in the lower part of the basin (the natural discharge from the system) is equal to recharge from the two streams at the upper end. Pumping in the basin is assumed to equal the recharge. This system was simulated by a finite-difference approximation to the equations of groundwater flow

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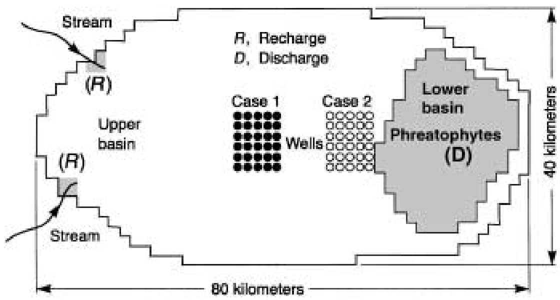

Fig. 10 Schematic map of an intermontane basin showing areas of recharge (*R*), discharge (*D*), and two hypothetical water-development schemes, case 1 and case 2, described in the text. (Adapted from Bredehoeft et al. 1982)

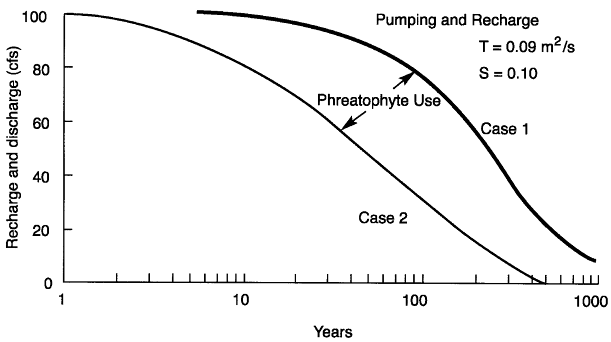

Fig. 11 Plot of the rate of recharge, pumping, and phreatophyte water use for an intermontane basin under two hypothetical water-development schemes, case 1 and case 2 (see Fig. 10). To convert from cfs to L/s, multiply by 28.3. (Adapted from Bredehoeft et al. 1982)

(Bredehoeft et al. 1982) for 1,000 years. Stream recharge, phreatophyte water use, pumping rate, and change in storage for the entire basin were graphed as functions of time. Two development schemes were examined: case 1, in which the pumping was approximately centered within the valley, and case 2, in which the pumping was adjacent to the phreatophyte area (Fig. 10).

The system does not reach a new equilibrium until the phreatophyte water use (i.e., the natural discharge) is entirely salvaged or captured by pumping (Fig. 11). In other words, phreatophyte water use eventually approaches zero as the water table declines and plants die. In case 1, phreatophyte water use is still approximately 10% of its initial value at year 1,000 (Fig. 11). In case 2, it takes approximately 500 years for the phreatophyte water use to be completely captured. These curves are similar to the transition or growth curves referred to earlier (Fig. 9), where initially most of the water pumped comes out of aquifer storage, whereas at later times it comes from capturing groundwater discharge.

This example illustrates three important points (Bredehoeft et al. 1982): (1) the rate at which the hydrologic system can be brought into equilibrium depends on the rate at which the discharge can be captured; (2) the placement of pumping wells changes the dynamic response and the rate at which natural discharge can be captured; and (3) some groundwater must be mined before the system can approach a new equilibrium. Steady state is reached only when pumping is balanced by capturing discharge and, in some cases, by a resulting increase in recharge (induced recharge). In many circumstances, the dynamics of the groundwater system are such that long periods of time are necessary before any kind of an equilibrium condition can develop. In some circumstances, the system response is so slow that mining continues well beyond any reasonable planning period.

A suitable hydrologic basis for a groundwater plan that is aimed at determining the magnitude of possible development would be a curve similar to the transition curve shown earlier (Fig. 9), coupled with a projected pattern of drawdown for the system under consideration. Since the 1980s, three-dimensional numerical models of the complete stream–aquifer hydrogeologic system have been used for water-rights purposes (Balleau 1988). These models provide a predictive tool that explains the connection between well-field withdrawal and surface-water depletion at particular sites. Groundwater models are capable of generating the transition curve for any case by simulating the management or policy alternatives in terms of the sources of water from groundwater storage and from surface-water depletion throughout the area of interest. Specified withdrawal rates, well distribution, and drawdown of water levels to an economic or physical limit are used in the model for such projections (Balleau 1988). However, a *planning horizon* must be defined to assess which phase of the transition curve will apply during the period of the management plan.

Conclusions and Needed Research

As Stanley and Jones (2000) note, the growth in research related to surface–subsurface exchange processes has mushroomed during the 1990s, particularly with respect to physical (hydrological) and biogeochemical processes. The frontier in GW–SW interactions seems to be the near-channel and in-channel exchange of water, solutes, and energy; an understanding of these processes is the key to evaluating the ecological structure of stream systems and their management.

Boulton et al. (1998) conclude that the relative importance of variables affecting the activity of the hyporheic zone (HZ) at sediment and reach scales over time is unclear, whereas Dahm et al. (1998) conclude that the spatial and temporal dynamics of groundwater discharge and recharge along active channels in varying geomorphic settings needs further elucidation. Quantification of the temporal dynamics of water and chemical fluxes through these boundaries is essential. Identification of stream reaches that interact intensively with groundwater would lead to better protection strategies of such systems. However, quantification of water fluxes in general, and specifically between groundwater and surface water,

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is still a major challenge, plagued by heterogeneity and scale problems.

The HZ is both complex hydrologically and relatively inaccessible and difficult to manipulate. Palmer (1993) outlines major obstacles that must be overcome in order to make significant progress in experimentation in the HZ and suggests corollary experiments or technical developments that should lead to major breakthroughs in the understanding of HZ processes. The choice of proper temporal and spatial scales for conducting such experiments is critical, because the particular site and time of year in which experiments are performed are likely to dramatically influence results. Different geomorphologies of sites selected for study could lead to evaluation of different processes, particularly because groundwater inputs and subsurface flows often vary dramatically within and between stream reaches.

Understanding GW–SW interactions presents unique challenges. The biogeochemical processes within the upper few centimeters of sediments beneath nearly all surface-water bodies have a profound effect on the chemistry of groundwater entering surface water, as well as on the chemistry of surface water entering groundwater. Knowledge of biogeochemical processes occurring within the sediments depends on understanding GW–SW hydrologic interactions and on gaining a better understanding of subsurface microbial processes. Jones and Holmes (1996) conclude that whereas surface–hyporheic exchanges and water residence times are known to be important regulators of subsurface biochemical transformations, the manner in which these parameters vary across streams is not yet known. They emphasize that this broader perspective is important not only for generalizations about subsurface processes but more fundamentally for advancing conceptual models of streams. Thus, studies of the interaction of groundwater and surface water should emphasize broader perspectives through cross-disciplinary collaborations. Also, as in most sciences, methods are needed for extrapolating results from small instrumented stream reaches to stream-network or basin scales.

The hydraulic properties of stream and lake beds control the interactions between these surface-water and groundwater systems, but these properties are normally difficult to measure directly. The primary limitation to date has been the difficulty of spatially defining the hydraulic properties and spatial heterogeneities of a stream or lake bed. In a stream–aquifer study, Sophocleous et al. (1995) rank stream-bed clogging, stream partial penetration, and aquifer heterogeneity as the three most significant factors in stream–aquifer problems. All these factors relate directly to the multidimensional nature of the stream–aquifer process. Yet most analytical treatments of GW–SW interactions ignore these factors.

Because streams and aquifers exchange water horizontally and vertically, flow dynamics are inherently three-dimensional. However, most hydrologic modeling studies have used one-dimensional or two-dimensional models. Analysis and simulation of the three-dimensional nature of the problem is needed for a better understanding of the stream–aquifer process (Sophocleous et al. 1988, 1995). Despite the current emphasis on near-stream and in-stream processes, most models today [e.g., the widely used MODFLOW model (McDonald and Harbaugh 1988) and later upgrades] are not well equipped to deal with local phenomena related to flow near domain boundaries. To properly handle the physics of stream–aquifer interaction, close attention must be devoted to the mechanisms operating at the GW–SW interface. This would involve, among other things, addressing the dynamics of seepage-face boundary conditions in detail. Because stream–aquifer seepage flows are driven by the head differential at the interface of the two systems, inaccuracies in the determination of aquifer heads on the seepage face would affect seepage fluxes; this would in turn impact channel flow rates and stream stages and thus again affect the head differential. Thus, it is important to compute rapidly changing stream stages accurately. This effort involves modeling of wave diffusion and bank storage on a physical basis – that is, by taking into account streamflow kinematics. In evaluating GW–SW interactions, both analytical and numerical methods need to be continually improved by more realistically simulating observed field conditions.

Longitudinal flow paths along a riffle-pool sequence and lateral flow paths into the stream bank create three-dimensional physicochemical patterns that are thus controlled by the flow patterns (Brunke and Gonser 1997). Hydrologic exchange between the stream surface and underlying sediments is characterized by using models and by direct measurement of hydrologic parameters of subsurface flow velocities. As Jones and Holmes (1996) point out, a key step for advancing understanding is the integration of hydrologic models with biochemical transformations, and in general linking ecology and hydrology.

Understanding the hydrologic and biologic processes that define the relationships between surface and subsurface waters, the landscape connectivity of riverine or aquatic habitats, and human-induced changes and associated responses of floodplains is essential if one is to understand the ecological effects of water-resources management decisions in a basin. Despite the fact that hierarchy theory (for overviews, see Marceau and Hay 1999; and Wu 1999) offers a useful conceptual framework for linking processes at multiple scales, the development of operational hierarchies and upscaling from reaches to watersheds remains a major research challenge today. The present inability to characterize subsurface heterogeneity exacerbates the upscaling problem and leads to great uncertainties in data interpretation. In the face of such uncertainties, multiple techniques for quantifying GW–SW exchanges need to be pursued, utilizing both in-situ and remote-sensing observations coupled with GIS technological advances, numerical models, and statistical analyses to study these processes in a multidisciplinary and multiscale approach.

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