A GROUNDWATER-SURFACE WATER PARTITION FOR THE CONTIGUOUS UNITED STATES AND SELECT CASE STUDIES

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ABSTRACT OF THE THESIS:

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The terrestrial water cycle is a highly effective, yet incompletely understood agent for the distribution of continental energy, and hence, the continental energy and water budgets are closely linked. The spatial organization and temporal memory of the groundwater reservoir, and its interaction with the surface water has an integral role in the lateral transport of water and energy, affecting soil moisture distributions, evapotranspiration, precipitation and stream discharge across the continent. The current climate models are unable to account for this lateral component, and consequentially are inadequate at predicting future hydrologic conditions; hence, a separation of groundwater flow from surface water flow is necessary to asses the relative importance of each reservoir across the land surface. Here we present the results of such a groundwater-surface water partition, where 39 years of surface recharge, derived from VIC simulation, are separated from USGS HCDN annual mean observed (naturalized) stream discharge from 1555 basins across the continental U.S. It was found that stream discharge (Qr) may account for 2% to 891% of the total surface recharge (R) across the 1555 basins, suggesting that individual drainage basins export or import significant amounts of water to or from the

groundwater reservoir (e.g., a Qr/R value of 2 (200%) for a basin indicates that half the river discharge from that basin is derived from groundwater input from other basins). Detailed investigations of individual basins across the continent in terms of this partition indicate that the control over lateral transport of subsurface water is primarily a function of the subsurface geology. Further, a marked incongruity between the surface drainage flow direction and groundwater flow direction is apparent in several cases – particularly where regional groundwater flow has developed – suggesting that surface drainage as a result of elevation is only partially indicative of subsurface flow regimes. The modulation of surface drainage by the groundwater system suggests that groundwater flow is a significant portion of the continental water cycle. Hence, this wide range of effects attributable to groundwater flow implies that the groundwater reservoir should be included in climate modeling efforts, particularly if estimates of future water resource availability are a goal of such efforts.

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1. Introduction

The terrestrial water cycle is a highly effective, yet incompletely understood agent for the distribution of continental energy, and hence, the continental energy and water budgets are closely linked. The groundwater component of this terrestrial water cycle plays a large role in the transfer of energy across the landscape, and is essential to balancing a continental water budget. Furthermore, the interaction of groundwater and surface water on any order of scale is the fundamental exchange process influencing land-surface and atmosphere energy fluxes. Information concerning the degree of differentiation between the groundwater and the surface water systems is essential to our understanding of these processes, particularly in terms of the current climate modeling systems.

Toth (1963) surmised that there are three potential groundwater flow systems which laterally redistribute continental waters: local, intermediate and regional flow. Although not specifically discussed by Toth, these flow systems are bounded by the substrate geology and the overall climate, which, on the macro-scale, are generally regionally persistent. Hence, we expect that the continental groundwater system will have a high degree of spatial organization, and its large storage capacity will impart a long temporal memory. This spatial organization, and ultimately macro-scale memory make the groundwater system a more static reservoir in the terrestrial water cycle, buffering event scale and seasonal surface influences over the longer term climatologic equilibrium state. This buffering effect profoundly influences the distribution of soil moisture throughout the landscape, thus rates of evapotranspiration, and surface water characteristics.

The groundwater system interacts with the surface water system through its spatial organization and memory, particularly through its lateral subsurface transport of water through the theoretical flow systems proposed by Toth (1963). The flow system with the most temporal and spatial influence is arguably regional flow, which transports (and stores) the largest volume of water over the largest spatial scales, and responds the slowest to climatologic forcing. Further, the regional flow system effects the spatial distribution of groundwater recharge and discharge zones, and controls the development of a hinge-line – the delineation between groundwater recharge and discharge zones. For a river basin in a given geologic and climatologic situation, zones of recharge to the regional flow system are characterized by losing basins, and zones below the hinge line in areas of regional groundwater discharge are characterized by gaining basins. Thus, small basins within a larger drainage basin may be classified in terms of their position relative to the hinge line, and thus be either gaining or losing basins. Even in areas of uniform surface input from precipitation, the characteristics of a given set of adjacent basins need not reflect this uniform distribution of moisture.

The current General Circulation Models (GCM's) of global climate have very thin soil and only a rudimentary representation of the terrestrial water cycle, such that precipitation minus evapotranspiration (the atmospheric surplus) is routed directly to stream discharge, and only buffered slightly by the soil column. Thus, the persistent spatial organization and long temporal memory of the hydrologic system is not included. The implication of this exclusion is that the lateral transport of water, and hence energy through the landscape, cannot be accounted for by the model. The failure to differentiate between surface and subsurface flow pathways means that the climate models are inadequate in the horizontal dimension, which in turn alters flux in the vertical dimension.

Since precipitation can leave a basin either through stream discharge, or the groundwater system, the next logical question is how much leaves via either pathway? It is possible to quantify the flux across this groundwater-surface water flow partition through and simple, first-order water budget analysis, as follows. For a given basin, regardless of scale:

(1) $P - ET = Qr + Qg$

Where precipitation (P), minus evapotranspiration (ET), represents the atmospheric surplus over a basin, which is partitioned into river discharge (Qr) and groundwater discharge (Qg). The P-ET parameter of Equation 1 is herein referred to as surface recharge (R), which includes recharge to water table as well as runoff to the surface water system. The river discharge (Qr) parameter represents the combined effects of surface runoff (Qs) and baseflow (Qb) to rivers, where $Qr = Qs + Qb$. The purpose of this study is to assess the relative importance between the two outflow pathways (Qr and Qg) in Equation (1) across the continent. This groundwater surface water partition is completed in terms of the drainage basin as the fundamental hydrologic unit, across several orders of scale.

This analysis will help to better understand the distribution of drainage basins in terms of their relative contributions to the groundwater or surface water flow pathways, which is essential to balancing the water budget in current climate models. Inclusion of the water table and known zonation of surplus or deficits in continental surface waters allow more accurate simulation of evapotranspiration and river discharge variables, and will greatly improve the atmospheric water and energy balance of GCMs.

From a hydrogeologic standpoint, this surface-water groundwater partition and its distribution can only be accurately explained with consideration of the subsurface geology controlling the groundwater flow system. In lieu of a single continental data set of subsurface geology, a second and more critical portion of this analysis evaluates the groundwater-surface water partition on a case-by-case basis. Selected basins are analyzed in depth with respect to geologic characteristics to better explain the spatial distribution of groundwater exporting or importing basins. Thus the first-order separation of the fundamental groundwater and surface water reservoirs is partially validated by a second-order consideration of geologic controls.

2. Background

2.1. Groundwater in the Climate Models

Treatment of soil and surface waters by the current climate general circulation models (GCMs) is largely by means of the free-drain approach, represented schematically in Figure 2-1. This approach allows surplus soil moisture to drain out of the model column at a rate prescribed by the hydraulic conductivity of the substrate and the saturation of the bottom soil layer. Since the model soil layer is very thin (1-3 m), water from this leaky bottom layer is transferred directly to the rivers as runoff; and as a result, the modeled land surface has very little memory (Coe, 2000). A potential downside of the method is that this water is no longer available for evapotranspiration during dry periods, which ultimately changes the energy balance of the atmosphere (York, et al., 2002).

If a groundwater reservoir is included in the subsurface scheme, soil moisture is incorporated into the groundwater reservoir and does not simply drain directly to rivers.

Thus, the soil column remains wetter for a longer period between precipitation events than by soil drainage alone (Chen and Kumar, 2001). This longer memory has the potential to greatly alter the atmospheric energy budget by sustaining dry-period evapotranspiration, which can lead to localized convective precipitation (most recently, Bierkens and van den Hurk, 2007, discussed below). This localized convective precipitation leads to wetter soils, implying that a single rain event may lead to a positive feedback of increased long-term soil moisture.

Because precipitation over a given basin may leave laterally via groundwater pathways, GCM's that route precipitation directly to river discharge tend to overestimate streamflow if a basin is a net groundwater exporter, or underestimate streamflow if a basin is a net groundwater importer. A common solution to this problem is to modulate the surficial evapotranspiration until modeled streamflow values match those observed (Liang and Xie, 2003). However, this method simply transfers an excess or deficit in one variable to another, so that evapotranspiration values are incorrect. A consequence of this exchange is that the hydrologic flux across the land surface-atmosphere interface is also incorrect, and hence latent heat absorption or loss by the land surface is inaccurately calculated. This affects not only the water content of the atmosphere, but the energy balance through latent heat of condensation or precipitation, etc. Higher atmospheric water content leads to more clouds, which alter the radiative energy budget to the land surface, and latent heat of condensation high in the atmosphere changes the thickness of the convective layer by feeding thermal updraft (Yeh, et al., 2005), thus altering wind patterns.

This implies that the groundwater system is not a trivial participant in the continental water cycle. By supplying dry season soil moisture and accepting surface water surplus during wet periods, the groundwater system acts as a low-pass filter for the land surface as well as the atmosphere, buffering the intensity and magnitude of flux between the two (Fan, et al., 2007). In this way, water is an essential conduit for thermal and mechanical energy through the landscape; vertically, horizontally and through time.

With such an incomplete terrestrial water budget, GCM's cannot hope to predict future hydrologic conditions over land. Estimates of future water availability for drinking or agriculture are great concerns to humanity, but are largely unattainable with the current modeling techniques (Yeh and Eltahir, 2005). The intensity and duration of precipitation events is also often cited as a result of climate change (IPCC, 2007), but how can these fluxes be estimated without first balancing the water budget?

2.2. Toth's Groundwater Flow

The theoretical basis of groundwater flow has long been recognized (Hubbert, 1940; Toth, 1963). In his 1963 pioneering paper, Toth asserts that there are essentially three scales of groundwater flow that exist in a given basin, whose organization depends primarily on the topography and land surface relief. These three flow regimes are local, intermediate and regional, all of which may theoretically exist simultaneously in a nested arrangement within a basin of homogeneous medium, but may exist separately as well. Figure 2-2 describes the nested relationship of Toth's (1963) flow regimes for a theoretical basin.

According to Toth (1963), a local flow system is characterized by groundwater recharge at a topographic high and discharge at an adjacent topographic low. An intermediate system is very similar to that of the local system, but the areas of recharge at topographic highs need not be adjacent to areas of discharge at topographic lows. A local flow system may exist between (within) the locations of recharge to the intermediate flow regime, and discharge from it (Figure 2-2). A regional groundwater flow system, according to Toth, can be distinguished by having a recharge area within the topographic groundwater divide, and a discharge area at the lowest point of the basin, which may lie beneath the base level of the large streams in the basin (Figure 2-3).

Toth evaluates the implications of his theoretical model based on different scenarios of topography, relief, and outlines the potential chemical effects on the groundwater. In extended areas of low relief, he asserts that neither local nor regional flow systems will develop, and the primary mechanism of groundwater discharge in such a situation is evapotranspiration. Assuming a relationship between groundwater flow velocity and solute concentration, the slow moving to stagnant groundwater in these regions will have high solute concentrations. Toth (1963) then suggests that in areas where local relief is low but the basin has a general slope, a regional flow system will develop and is dominant (Figure 2-3). This situation may cause a gradual increase in the concentration of dissolved solutes with depth, largely because of the relatively low flow velocity of the regional system.

In the contrary relationship, Toth proposes that where the topography has high local relief, local flow systems are likely to develop. The depth of these local flow systems is proportional to the height of the relief. Furthermore, he presumes that the pronounced development of local flow systems implies that the development of regional flow systems will be suppressed (Figure 2-3).

Among his final remarks, Toth (1963) concludes that the major stream channel in a given basin receives groundwater input from adjacent topographic highs in the local flow system, and from regional flow – he surmises that regional flow probably makes a less significant contribution than that of local flow because of its low flow rate. Toth goes further to say that the shallow groundwater systems are most responsive to climate and seasonal changes in recharge and discharge, and that the larger regional systems do not fluctuate as extensively as their local counterparts. This leads him to the conclusion that only a small percentage of the water stored in a given basin participates in the active hydrologic cycle (Figure 2-2), and that the regional system is responsible for storing this large portion of relatively inactive water.

While Toth's analysis is valid in terms of the general groundwater flow regimes, he has based his model on a basin and aquifer of highly idealized characteristics. With the understanding that such a rudimentary approach is necessary for a first order attack at any problem, Toth's theoretical investigation failed to explicitly consider two key factors: the dynamic effects of geology and climate on the groundwater flow system. A primary assumption of Toth's groundwater flow regimes is based on Hubbert's (1940) assertion that the water table is generally a subdued version of the surface topography, assuming uniform precipitation. This may indeed be the case when we consider a single homogeneous layer of unconsolidated sand, but is only partially representative of reality.

In the case of climate, Toth reasoned that the higher the topographic relief, the more prevalent local groundwater flow becomes. This may be correct in areas where the water table is shallow and recharge is not limiting, but is most likely not the situation in examples such as the inter-mountain valleys of the Western US or Gulf coastal plain, since in these regions the arid climate has brought the water table below the local relief. Along the same lines, Toth has also not sufficiently considered the effects of anisotropy in the vertical vs. horizontal conductivity of the flow media in his theoretical analysis. Greater permeability in the horizontal direction than the vertical, or vice versa, will greatly alter the potential development of a particular flow system.

Geology is by far the most crucial factor for the development of any sort of groundwater flow. The substrate comprises the flow media; changes to its character deviant from that of homogeneous sand may define the groundwater flow regime (Modica, et al., 1997). A regional flow system is not likely to develop as readily as local flow systems in a fold and thrust belt where geologic units are sandwiched by confining layers, despite the presence of moderate local relief superimposed over general basin slope. Conversely, a basin sitting in an unconfined and highly permeable unit with very gentle surfacial slope – but steeply dipping base – will likely favor a regional flow regime over Toth's suggested stagnant conditions (Haria and Shand, 2004). Similarly, fractured basalt has extremely high permeability and water rapidly flows through joint planes, while massive basalt with no jointing (or scoria/vesiculation) may be nearly impermeable. (Reference for basalt??)

That being said, Toth's framework is the best conceptual model by which to compare and evaluate the deviations of realistic examples. A final postulate of Toth's is that the small basins within the larger basin are the most important unit in the groundwater regime, in part because uncertainty increases with basin size. The characterization of adjacent small basins makes the large basin characterization more straightforward because smaller basins are less complicated by geology and topography. Further, the most effective way to recognize broad spatial trends within a basin is to compare it to its internal constituents; this will remain a theme throughout the analysis.

2.3. Examples of Potential Insufficiency in the Climate Model Water Budget

Following Toth's (1963) theoretical logic, we must return to the evaluation of the current climate models and their treatment of the terrestrial water cycle in a less idealized manner. A number of groundwater situations that elude mention by Toth may have an even greater impact on the treatment of water in the climate models. One example of such a groundwater flow regime is that of the High Plains Aquifer system, which extends from Southern South Dakota to Northern Central Texas along the Rocky Mountain front range of the Western U.S. (Figure 2-4).

The High Plains Aquifer is classified by the USGS as a blanket sand and gravel aquifer that spans approximately 450,600 km^2, and averages ~60 m deep, with a maximum depth of about 305 m (Gutentag, et al., 1984). Groundwater flow in the High Plains Aquifer system is characterized by a west to east movement along a gradient from the Rocky Mountains eastward, where ephemeral streams slowly give way to perennial streams and eventually major channels on the central to eastern side of the aquifer system (Luckey, et al., 1986). Much of the groundwater is sourced from the Rocky Mountains to the west, often from seasonal snow accumulation, and travels downgradient through the highly permeable sand composing the aquifer (Weeks, et al., 1988). When the water table is high, such as in late spring following snowmelt in the mountains, the aquifer feeds small channels and ephemeral streams through local flow, while a large scale and persistent regional flow transports water to the major channels in the eastern portion of the aquifer. This regional flow is particularly important during the more often dry periods when the water table is below the local relief and flows almost exclusively to major channels (Gutentag, et al., 1984).

Figure 2-4 shows a map view of the High Plains Aquifer's extent and general groundwater flow directions along with stream channel locations. Figure 2-5 shows an idealized cross section through the High Plains Aquifer system, with depth to the water table, generalized flow directions, and stream channels denoted. A comparison between Figure 2-5 and Figure 2-1 reveals a deficiency in the ability of the climate model to accurately represent the water budget in such a region. As explained earlier, precipitation over the land surface in the climate model is sent directly to the stream channels for drainage (Figure 2-1). This is a particularly unsuitable scenario for precipitation over the High Plains Aquifer, where precipitation that does not return to the atmosphere as much by evapotranspiration as it infiltrates the ground-surface and feeds into the water table far below (Gutentag, et al., 1984) (Figure 2-5). The water is added to the groundwater reservoir and eventually resurfaces through lateral flow, to a stream channel far downstream within the basin at a considerable distance from its original source (Gutentag, et al., 1984). The climate model's inability to account for this effective lateral redistribution of surface energy and water greatly biases its results and exemplifies the interplay between geology and climate.

The mountainous valleys of the Western US, specifically in the Basin and Range region, are characterized by closed hydrologic systems, where low elevation valley floors are bounded by extremely high elevation mountain ranges on their flanks (Figure 2-5). In a system such as this, the climate gradient, in part due to elevation and air-mass location, allows the accumulation of a significant snow pack in the high mountainous regions. During the spring melt of this snow accumulation, a large amount of stored water is transported first as surface flow in the mountains, where the regolith is relatively thin, to the sandy, gravely alluvial fans where it becomes subsurface flow. This subsurface flow infiltrates the areas at the base of the low, flat valley floors where it eventually seeps to the surface in the form of ephemeral playa lakes (Fan, et al., 1997). Since the precipitation received at the valley floors is extremely low (<10 mm yr^{-1}), the playa lake systems that occupy these areas would not exist without the lateral transport of groundwater through the alluvial fan systems (Prudic, et al., 1995).

Furthermore, the transition from the cool and moist high elevation mountains to the adjacent very hot and very dry playa valley floors represents a dramatic energy gradient across the land surface. The effects on the terrestrial energy balance are obvious in a very qualitative sense; water that was once stored frozen as snow eventually becomes a constituent of the latent heat process in the valleys, as evidenced by evaporite deposits on valley floors and high salinity of playa lake systems (Prudic, et al., 1995).

Figure 2-6 details a schematic representation of an intermountain valley and its associated groundwater flow fields. Again, comparison between Figure 2-6 and Figure 2-1 makes obvious the climate model's inability to account for such lateral transport processes. Without inclusion of lateral groundwater transport in the climate model, there is no representation of this high contrast, efficient water/energy transfer, or very long temporal memory of the land surface and its water balance.

2.4. The Spatial Organization and Temporal Memory of the Groundwater Reservoir

As touched upon by the previous two examples, groundwater systems typically maintain a high spatial structure and a long temporal memory of water and energy fluxes across the continent. In both the examples above, the spatial structure imparted by the groundwater system maintains a regionally persistent hydrologic regime based on climate, geologic structure and topography. The long temporal memory of the groundwater flow system, particularly at the regional scale, supports dry-season flow in the lowland streams on the east side of the High Plains Aquifer system, and maintains relatively wet valley bottoms in the intermountain valley regions further west, despite the highly evaporative regime. Because the past two examples displayed such influence from the spatial structure and temporal memory of the groundwater system, these two characteristics warrant a more thorough review.

2.4.1. Spatial Organization

Because the water table depth and groundwater reservoir are products primarily of geology and secondarily of climate regime, a persistent spatial organization of the continental groundwater system comes as no surprise. A simulation conducted by Fan, et al. (2007) partially accounts for the geologic heterogeneity of the continent by using mean annual water table depths based on USGS observations, and has produced a detailed representation of the continental water table depth and its spatial structure (Figure 2-7).

Fan, et al. (2007) term this output the Equilibrium Water Table (EWT), which is the climatologic mean water table position within the continent (Figure 2-7). They clearly indicate that the controlling factor over the hydrologic equilibrium is a balance between vertical, climate induced flux, and lateral hydrologic divergence/convergence, which is geologically controlled. With this in mind, they make a few key observations concerning the spatial organization of the water table across the continent. They note that a shallow water table can exist in regions with a humid climate, such as the southeastern U.S., and in arid regions, such as the earlier example of the intermountain valleys of the western U.S. Fan, et al. (2007) also observe that in arid and semi-arid climates, such as the northern U.S. Great Plains, the water table can be deep. Likewise, they note that the water table may also be deep in humid climates, such as the Appalachians. In the case of the Appalachians, the water table is deep despite the large climate-induced contribution because of undulating topography, shallow soil and a dense river network capable of quickly transporting surface input downgradient.

Outside of the watershed scale, Fan et al. (2007) notice a significant shallowing trend of the water table from the higher to lower drainage regions of the Mississippi River, due to topographic gradient. They also note this shallowing from the west to the east of these two regions, due to both climatic and geologic gradients. Over the Atlantic Coastal Plains region, the water table becomes much shallower approaching the ocean, which is largely geologically controlled. However, since their model did not explicitly account for the geologic heterogeneity of the groundwater system, these geologically based inferences are purely qualitative. The next logical link made by Fan, et al. (2007) is between the groundwater spatial distribution and the continental soil moisture fields, a second part of their simulation, which will be discussed in section 2.6.

2.4.2. Temporal Memory

There are a myriad of examples attesting to the long temporal memory of the groundwater system, from million year old water being pumped from the Sahel and Northern Africa (Bierkens and van den Hurk, 2007), to the intra-decadal response of the Ogallala aquifer to high snow events during El Nino years (Woodhouse and Overpeck, 1998; Gurdak and Hanson, 2005). For a moment, however, the reader is referred back to Toth's (1963) flow systems, each of varying scale within an unconfined, homogeneous basin (Figure 2-2). Local groundwater flow has response time on the event level scale, the intermediate flow system has a seasonal to inter-annual response time, and regional flow systems respond to recharge and discharge changes on the inter-annual, decadal, or much longer timescale.

Toth's three flow systems exist within a hierarchical relationship that is proportional to their respective residence times (Pulido-Velazquez, et al., 2005). The local flow system has a small volume and short residence time when compared to the regional system, which has a much larger volume with a longer residence time. Therefore, in terms of temporal memory, the regional groundwater flow system is the more influential of the three. It acts as a low-pass filter for hydrologic events from the land surface, responding the slowest to climatologic changes but 'remembering' them the longest. The regional system contains the largest volume potential moving the furthest through space, across large portions of the continent, and thus creates the greatest prospective imbalance for the long-term climate model. Hence, with the shortcomings of Toth's (1963) flow model in mind, it seems pertinent to review some examples of observed regional flow systems while examining their long-term memory potential.

Aside from detailed basin characterization and subsequent simulation with particle tracking, determining residence time and source of water in a particular aquifer is best accomplished using geochemical tracers such as stable or radiogenic isotopes (Clark, et al., 1998). Because groundwater accumulates many of the chemical properties and byproducts of the medium it flows through, and retains the characteristics of the precipitation from whence it originated, dissolved solutes, radiogenic and stable isotopes measured from the water often aids in determining its source in space and time. Stable and radiogenic isotopic studies, for example, have been used to show that high salinity groundwater from central Missouri originated as meteoric recharge at the Colorado Front

Range, over 1000 km to the west (Banner, et al., 1989). Bentley (1986) used isotopic data to map continental-scale groundwater flow in the Great Artesian Basin of Australia. His data suggests deep regional flow from the elevated northeast margin near the Great Divide Range to distal discharge regions ~1000 km away near Lake Eyre, and Chlorine-36 dating indicate a residence time of about 2 My across this span.

More recently, Clark, et al. (1998), used noble gases, stable isotopes and radiocarbon tracer data to investigate the recharge sources of the Dakota Aquifer system, in the central Midwestern U.S. Their well data suggest that a portion of the groundwater in southeastern Colorado, where the Dakota Aquifer is unconfined, may be a relict of the last glacial period. Oxygen isotopes and radiocarbon ages based on dissolved CO_2 in central Kansas wells, downgradient from the Colorado group, suggest a residence time of the lower water mass on the order of 10,000 to 100,000 yrs, with recharge sourced in the Colorado-Kansas border (~350 km to the west). A geochemical and modeling study conducted in Trans-Pecos, an arid closed basin aquifer in western Texas, suggests that baseflow spring discharge is supported primarily by regional flow to the valley floor, recharged some 40-60 km to the northwest (Uliana and Sharp, 2001).

Bakker, et al. (1999) used an analytic element model to simulate the Death Valley regional flow system, with particular attention paid to the proposed nuclear waste repository beneath Yucca Mt., NV. The model confers well with previous research and models of the southern Great Basin (see references therein), suggesting that the Death Valley system is recharged up to 500 km to the northeast, near the White River system and areas to the northwest of Las Vegas, NV (Spring Hills and Pahute Mesa). Moran and Rose (2003) used chlorine isotopes, among others, to map groundwater recharge and discharge areas in the Death Valley regional flow system of the southern Nevada Great

Basin, specifically around the Nevada bomb test site. Using bomb test dates to bracket chlorine isotopic ratios of well-water from various sites, they contend that the groundwater in the vicinity of the test site was recharged some ~150 km to the northeast in the White River regional flow system during the last pluvial period. Radiocarbon, oxygen and hydrogen isotopes of spring waters from Death Valley, measured by Anderson, et al. (2006) suggest that a large component of the discharge was recharged during the late glacial period (~16-12 ka) when the regional climate was much cooler. These data confer well with other isotopic data around the Death Valley region (Anderson, et al., 2006, and references therein), and the ^{14}C model ages suggest the waters contain a significant component that is between 14.5 and 5.5 kyr in age. However, they contend their data does not support the paradigm of an interbasin flow system recharged far to the northeast at Ash Meadows, NV (Anderson, et al., 2006).

In a meso-scale alluvial aquifer system in the interior of Oman, Matter, et al. (2005) identify a regional gradient in well-water source age, from the mountainous highlands to piedmont and flat lowlands, based on carbon, oxygen and chlorine isotopic evidence (among others). They did not venture to speculate on a residence time of water in the groundwater system, but commented that a recharge source could not be identified in the most distal lowlands of the basin. This suggests that the water pumped from the basin floor was recharged during a much wetter climatic regime. In the arid closed system of Independence Basin, Mexico, Mahlknecht, et al. (2006) identified a similar phenomenon using a suite of geochemical evidence sampled from wells and natural springs. The authors note that the groundwater sampled at the center of the ~140 km wide basin dated at ~11 ka. They determined that the age decreased concentrically out of the basin, suggesting a regional flow system exists where recharge is transported laterally through the subsurface from the more distal mountain source.

Modica et al. (1997) noted first through simulation, and subsequently through geochemical analysis (1998) that streams on the Atlantic Coastal Plain of the Eastern U.S. show an age distribution where baseflow waters appear to be progressively older with distance downstream. Modica, et al. (1998) demonstrated that, when the stream channel was aligned with the regional groundwater flow direction, such as the Cohansey River Basin in the New Jersey Coastal Plain, the stream discharge could contain water as much as 50 years old by the time it reached the ocean. The homogeneity of the regional aquifer system promotes regional groundwater flow towards the Atlantic Ocean, sourced far inland, and streams aligned with this flow continually receive groundwater input downstream. It has also been noted by Martin (1998) that groundwater pumped from the deep Magothy Aquifer in the New Jersey Coastal Plain, the source of municipal water for Atlantic City, was recharged during the last glacial maximum (20 ka).

2.4.2.1. Interbasin Groundwater Flow

The majority of the previous examples, with the exception of the New Jersey Coastal Plain aquifer, have attested to regional groundwater flow conditions in recharge limited or arid provinces. However, long distance lateral groundwater flow has been reported in humid tropical regions, such as lowland Costa Rica (Genereux et al., 2002). Genereux, et al. (2002) identified two distinct water masses in lowland rainforest watersheds at La Selva Biological Station (Costa Rica), using dissolved chloride, among other chemical tracers measured from baseflow discharge in riparian wetlands. They recognized a high-solute water mass, which they associated with groundwater in contact with bedrock, and low-solute water drained from more local sources near the watershed. Through sampling adjacent lowland basins, Genereux et al. (2002) found that perhaps half of the water in some streams and 84% from some riparian seeps contains major ions suggesting the water is transferred by deep interbasin groundwater flow.

Genereux and Jordan (2006) compared major ion content samples from riparian wetlands in two adjacent watersheds, one hypothesized not to be effected by interbasin transfer, and the other likely receiving interbasin flow. They found that deep interbasin groundwater flow likely accounts for two thirds of the total discharge in the latter watershed. They also cite oxygen isotopic data which support these major ion measurements, as many lowland basins appear to receive much of their input from interbasin groundwater transfer from distal recharge sources. Thus, it seems that low-recharge (arid) conditions are not necessary for the development of regional flow or groundwater transfer between adjacent basins.

Despite the fact that the inflow and the outflow are proportionally much smaller than the storage of a regional flow system, it is a highly effective conduit for terrestrial waters through the landscape, and imparts an extensive and un-ignorable fourth dimension of time to any investigation. clearly, a climate model lacking a water table, or any of the aforementioned groundwater structure or flow dynamics, will be very poorly suited to simulate continental water and energy balances.

In this discussion of the temporal memory of the groundwater reservoir and regional groundwater transport, another fundamental relationship within the terrestrial water balance has been alluded to: the interaction between groundwater and surface water. This relationship warrants further discussion with particular attention paid to the temporal and spatial aspects of the association.

2.5. Groundwater-surface water interaction

The connection between the groundwater and the surface water system is not nearly as simple as is suggested by its treatment in the climate models, where incident precipitation is piped directly to stream discharge (Figure 2-1). Groundwater and surface water interplay in a dynamic relationship that is highly effective at both energy and water transport laterally across the landscape.

Indeed, it is understood that the surface water systems of a continent are to some degree an expression of their positions with respect to the groundwater flow system, and essentially controlled by the underlying geology, the climatic regime, and the local/regional topography (Winter, 1999). The surface expression of groundwater systems is further complicated by basin slope, scale and depth of the flow medium (Garven, 1995; Sophocleous, 2002). Toth's (1963) three scales of fundamental groundwater flow are a baseline conceptualization of groundwater-surface water interaction for comparison. This is under the hypothesis that the surface water expression of the groundwater system is a function of a variably controlled groundwater flow regime (Wood, 1999).

A primary assumption of Toth's (1963) groundwater flow regimes is based on Hubbert's (1940) assertion that the water table is generally a subdued version of the topography. This depends on uniform recharge across the basin; since precipitation is the source of recharge, groundwater flow is affected by climate, and surface water volume is duly affected.

The spatial distribution of flow systems affects the intensity of natural groundwater discharge. For example, the main stream of a basin may receive discharge from both adjacent topographic highs and more distal recharge areas (Toth, 1963;

Sophocleous, 2002). On the other hand, groundwater discharge is not confined along the stream channel, and can extend through the discharge area downgradient of the 'hinge line' (a line of conceptual separation between areas of recharge and discharge) (Mayboom, 1966, 1967). This implies that baseflow to streams represents only a small portion of the discharge downgradient of the hinge line, and that baseflow measurements may not be representative of recharge conditions for a given stream system (Domenico, 1972).

2.5.1. Observational Studies

Larkin and Sharp (1992) recognized two end-member relationships between groundwater and stream interaction in alluvial aquifers: baseflow dominated and underflow dominated. The baseflow-component dominated system is characterized by potentiomentric groundwater contours parallel to the direction of river flow, with groundwater flow into the channel. The underflow-component dominated system is distinguished by potentiometric surfaces perpendicular to the channel flow direction so that groundwater flow is parallel to stream discharge. Sophocleous (2002) further points out that underflow component systems may exist above the regional hinge line, and baseflow systems below.

Larkin and Sharp (1992) conclude that several factors dictate the development of a baseflow vs. underflow system in alluvial aquifers, namely, the channel slope, river sinuosity, degree of river incision, width-to-depth ratio of the bankfull channel and the specifics of the fluvial deposition system. They noted that underflow was the predominant flow regime in systems where the channel gradient and width to depth ratio were large, and the sinuosity was low. Whereas baseflow-dominated systems occur in situations with archetypal characteristics of suspended load streams (generally opposite those listed above). Mixed baseflow-underflow systems may occur where lateral valley slope is negligible, and the channel slope and valley gradient are very similar (Larkin and Sharp, 1992).

On the continental scale, hydrologic exchange between groundwater and surface water are governed by the larger scale effects of topography, climate regime and geology (Garven, 1995). The direction of groundwater flow to or from rivers varies with hydraulic head (the effects of topography and climate), while the flow volume is largely conductivity dependant (the effect of substrate geology) (Sophocleous, 2002). In this way, surface water may either contribute to subsurface flow, as the case in a losing stream, or receive groundwater input, as the case in a gaining stream (Toth, 1963; Winter, 1999). Losing streams are often considered to be an effect of aridity and hence are associated with a deep water table, where any surface runoff percolates immediately to the water table and does not flow significantly on the surface (Wood, 1999). Along the same regards, a shallow water table is generally considered to support streams which accept input from the groundwater system.

The effects of the relationship can be quite variable: In humid to semi-humid climates, where the water table is shallow, groundwater sustains baseflow, augmenting river discharge, and in more arid conditions receives river seepage (de Vries, 1995; Winter, 1999). Moreover, recharge received by deep disconnected aquifers in more arid areas may indeed contribute to surface water generation lower within a given basin, below the hinge line. That is, the lateral transport of water through groundwater flow in the subsurface may be important factor when considering the spatial extent of land surface water distribution (Wood, 1999; Fan, et al., 2007).

Thus, the groundwater system buffers stream discharge intensity and cushions the continental energy balance. Moreover, the effect can be observed on both the seasonal and climatologic timescales. This buffering effect occurs on the seasonal scale in the High Plains Aquifer, where the groundwater reservoir is a source for ephemeral stream discharge during the spring snow melt, and a sink during dryer times of limited precipitation (Gutentag, et al., 1984). Following the logic of Toth's (1963) flow regimes, local flow becomes an important component of surface water discharge during the spring melt, while regional flow dominates the rest of the season. The net effect is maintenance of discharge to the major stream in a given basin. This seasonal expansion and contraction of stream networks has been demonstrated in shallow groundwater systems in the Netherlands (de Vries, 1995), with particular reference to changes in the regional drainage network.

The same effect may be observed on the paleo-timescale when the climate shifts from humid to arid. If we use the High Plains Aquifer as a hypothetical model, the current regime favors ephemeral streams during spring melt off. However, if the climate shifted to a regime less favorable to this seasonal recharge effect, we would presumably no longer see the development of ephemeral streams, and regional flow may dominate the groundwater flow system. In this sense, the long residence time and memory of the regional system may sustain baseflow discharge to major channels long into the arid regime.

The seasonal and interannual response of the Illinois Aquifer system has been evaluated in terms of variations in precipitation, evapotranspiration and incoming solar radiation by Eltahir and Yeh, (1999). Their statistical analysis of surface and subsurface variables suggests that variations in solar radiation are influential over the water table on the seasonal scale, while precipitation dominates the signal on an interannual scale. They also note that periods of drought have a larger and more pervasive impact on the water table than do floods. They attribute this phenomenon to the efficiency of baseflow at dissipating excess water from the aquifer; when the water table rises, the drainage density increases (more ephemeral drainage systems are activated), and less water is stored during floods than is lost during droughts.

Hood, et al. (2006) observed the response of an alpine headwater lake in British Columbia where they were able to identify two major groundwater sources discharging to the lake through flow measurements and geochemical tracers. Their observations suggest the lake is fed by a shallower local flow component responding on the event scale, and a chemically distinct regional flow component that does not fluctuate with event-scale forcing. The data suggest groundwater inflow is responsible for between ~30 – 67% and 35 – 74% of the total lake outflow over two field seasons, and thus contributes significantly to stream discharge from the lake.

2.5.2. Modeled Groundwater-Surface Water Interactions

Innumerable modeling studies have been undertaken to characterize groundwater-surface water interaction. Since the eventual reach of this study is to investigate continental water budgets, only a few pertinent simulations to that effect are subsequently outlined. Habets, et al. (1999) used a hydrologic model including groundwater-stream interactions and water table considerations, coupled to a land surface parameterization (LSP) scheme with prescribed atmospheric forcing, in an attempt to simulate the water budget and stream discharge conditions in the Rhone River basin. Their simulation suggests that when sub-model grid variability in precipitation, runoff, and vegetation are accounted for, surface fluxes can be computed with a less than 5% error. Their simulated discharge for one year agreed with that of nearly 100 river gauges throughout the region. They attribute this agreement to reflect the dynamic linkage between groundwater and surface water budgets.

Chen and Kumar (2001) simulated the water balance of North American drainage basins for the period between 1987 and 1988. They noted that subsurface redistribution of soil moisture by groundwater flow, controlled by anisotropy in hydraulic conductivity in the horizontal direction, is important in terms of seasonal variability in streamflow. This observation follows model results found to contain enormous and thoroughly unexplainable errors in the distribution of surface energy fluxes and streamflow. Streamflow was also found to be closely related to water table depth when validated using the Mississippi and Ohio Rivers for calibration.

Gusev and Nasonova (2002) combined the interplay between water table dynamics and river flow with their land surface scheme and modeled the water and energy budgets of the Valdai Hills boreal grasslands, Russia. The concluded as well that a shallow water table may have a significant influence on surface energy and water fluxes in the Valdi Hills region.

Liang and Xie (2003) demonstrated that surface water and groundwater interactions can be successfully characterized and simulated using the Variable Infiltration Capacity (VIC) model in a small Pennsylvania watershed. Their model accurately predicted both the position of the water table and the surface runoff over a three year period.

Yeh and Eltahir (2005) demonstrated that on a monthly basis, surface runoff to streams is more closely related to water table depth than to precipitation in a study conducted in Illinois, which is somewhat in agreement with the earlier results of Eltahir and Yeh (1999), discussed above. Krause and Bronstert (2007) also conclude that lateral flow processes and interactions between the water table and surface runoff have a significant impact on the streamflow and water budget of a lowland river catchment in Germany.

2.6. Effects of Groundwater Flow on Soil Moisture and Evapotranspiration

Perhaps the groundwater reservoir's most effective means of continental energy transfer is by way of soil moisture (Liang, et al., 2003). Soil moisture and its respective distribution across the landscape dictates latent heat exchange between the land surface and atmosphere through evaporation and evapotranspiration. Evaporation is most influential in areas where water is exposed to the land surface-air interface, such as surface waters in lakes, streams and wetlands. However, vegetated land surfaces act as direct conduits for soil moisture to the atmosphere (Sankarasubramanian and Vogel, 2003) through complex networks of cellulose and lignin pipettes (i.e., roots), which are arguably more efficient at this process.

In addition to precipitation, the groundwater reservoir is an important source of root-zone soil moisture, as suggested earlier. Despite its disconnect from the land surface in arid climate conditions, the groundwater reservoir may supply the land surface with moisture when precipitation is minimal, and accept surplus moisture during wetter conditions when the groundwater table is sufficiently shallow (Fan, et al., 2007). If the groundwater table is able to interact with the soil root-zone, it may likewise affect the water flux between the land surface and the atmosphere by supplying moisture for evapotranspiration (Liang, et al., 2003; Chen and Hu, 2004; Fan, et al, 2007; Niu, et al., in press).

Indeed, it has been demonstrated that the groundwater system can sustain up to one third of the monthly evapotranspiration during dry periods (Gutowski, et al., 2002) and support between 5 and 20 percent of the annual evapotranspiration for a basin in northeastern Kansas (York, et al., 2002). The latter two analyses attached a single column atmospheric model to a land surface model, calculating water and heat fluxes through soil and vegetation, subsequently coupled to a detailed groundwater model, including shallow subsurface flow and its interaction with stream channels. By demonstrating the potential feedbacks between the atmosphere, land surface, and subsurface reservoirs, these findings highlight the necessity of including the groundwater reservoir in order to balance the terrestrial water and energy budgets.

Water table dynamics have also been included in Liang et al.'s (2003) Variable Infiltration Capacity (VIC) model, which was used to simulate land surface moisture in two small-scale watersheds in Pennsylvania. The inclusion of the groundwater table in this simulation showed that near-surface soil moisture flux and distribution are significantly different (a dryer upper soil layer and wetter lower soil layer), than the output of model runs without water table consideration. The results of their stream discharge simulation from an earlier model are discussed in the previous section.

Although groundwater dynamics are not explicitly represented in their model, Chen and Hu (2004) allowed exchange between the unsaturated soil zone and variation in groundwater table depth in a Nebraska Sand Hills watershed. By comparing model runs where groundwater interaction with the soil zone was not permitted, to model runs with groundwater, their findings suggest that the water table can maintain and augment soil moisture and hence evapotranspiration. With groundwater consideration, their model predicted a 7 to 21% increase in evapotranspiration and a 21% increase in surface soil moisture, compared to runs without groundwater interaction.

Miguez-Macho, et al., (2007) use the equilibrium water table results of Fan et al. (2007) (part 1 of the simulation) to link the groundwater-surface water reservoirs across the contiguous U.S. to the land surface parameterization scheme of the Regional Atmosphere Modeling System (RAMS). The model was used to simulate the spatial and temporal variability in soil moisture fields across the continent with a complete water budget. Their results show a strong spatial and temporal link between the water table and soil moisture fields, particularly in low lying areas, where the water table interacts with the soil root-zone. By comparison of the simulation using the water table, to that using the free-drain approach, they note the largest differences between the two methods during the dry season, where soil moisture is maintained persistently by the groundwater system. They assert that the water table interaction with the soil zone represents a control nearly as important as surface precipitation in humid regions, but particularly in arid regions.

2.6.1. Soil Moisture and Precipitation

Soil moisture and its effect on the atmosphere's energy budget and water content have further been proposed as influential over continental precipitation regimes. In a multi-model effort, the GLACE team attempted to quantify the degree to which the land surface and atmosphere were coupled over earth's surface (Koster, et al., 2004). The work was conducted under the conceptual hypothesis that soil-moisture zones may be likened to oceanic “hot-spots,” such as the eastern equatorial Pacific with its role in El Nino. Their model results suggest that indeed, certain “hot-spots” of precipitation influencing soil moisture values could be predicted, most of which existed in transition zones between arid and humid climates (such as the Midwestern U.S. and the Sahel). They stress that the effects of soil moisture induced precipitation are predominantly local to meso-scale regional, and that soil moisture charged to the atmosphere through evapotranspiration may not substantially affect areas remote from the source.

It has also been suggested that, given the correct climate and atmospheric conditions, groundwater convergence may affect precipitation regimes. Bierkens and van den Hurk (2007) constructed a simple feedback model to represent the precipitation, evapotranspiration, soil moisture and groundwater components of seasonal rainfall areas lying between semi-arid and semi-humid climate zones, such as the Sahel and the Central US (identified as “hot-spots” by Koster, et al., (2004)). It is hypothesized by the authors that evapotranspiration and soil moisture in these areas may be sustained during the dry season by groundwater flow convergence from adjacent regions on the meso-scale. This increased soil moisture and enhanced dry-season evapotranspiration might sustain increased precipitation conditions in these regions during the dry season (Bierkens and van den Hurk, 2007).

The model results of Bierkens and van den Hurk (2007) suggest that the groundwater reservoir does not significantly increase the amount of annual precipitation; however, it does modulate the amplitude of year-to-year variation in those precipitation regimes, particularly over long timescales. Their findings also suggest that groundwater strongly affects the year-to-year persistence of rainfall in these regions. The model developed by Bierkens and van den Hurk (2007) is admittedly rudimentary, but the authors stress the need for a more realistic representation of terrestrial water budget, including full groundwater flow dynamics.

These studies more than adequately establish the influence of the groundwater reservoir on soil moisture distribution and the potential consequences of this distribution. A shallow water table sustaining evapotranspiration is an example of a direct conduit between the subsurface and atmospheric water reservoirs. In the context of regional flow, water sourced from afar may sustain locations of high evapotranspiration, which has the potential to influence precipitation, and river discharge in those regions. General circulation and other climate models cannot dynamically represent such a complex redistribution of surface and subsurface waters, or the partition between the land and atmosphere reservoirs. A fundamental first step in rectifying this deficiency is quantifying continental lateral groundwater movement.

3. Hypotheses and Objectives

The ensuing analysis is motivated by the following research questions:

1. How much water leaves a given drainage basin without ever passing through the surface outlet of that basin?
2. What is the relative importance of this water in terms of the surface/subsurface reservoirs and the continental water balance?
3. What is the relationship between surface water flow and areas of recharge and discharge to the water table? Are sources and sinks of water to surface basins identifiable in the subsurface flow system?
4. What is the ultimate control over the distribution of gaining and losing basins in terms of their surface water expressions?

To answer these questions, the hypotheses developed below are based in part on the theoretical flow systems proposed by Toth (1963). Determining the amount of net surface flux that becomes a component of the groundwater system becomes a problem of differentiating surface water discharge from its groundwater counterpart within any given basin. The extent of surface water and groundwater reservoirs is essentially controlled by recharge (**R**), which is the difference between incident precipitation (**P**) and evapotranspiration (**ET**) over a basin, as outlined in Equation 2 (below), (also refer to Equation 1 in the section 1).

(2) $$\mathbf{R = P - ET}$$

The sum of the river discharge ($\mathbf{Q_r}$) and the groundwater flow ($\mathbf{Q_g}$) from a given basin balance with the net **R** over that basin, as in Equation 3 (below).

(3) $$\mathbf{R = Q_r + Q_g}$$

The relationship between equations 2 and 3 implies that by accounting for precipitation and evapotranspiration over a basin, and the river discharge out of said basin, we may arrive at the resulting groundwater flow through the basin in question. The latter assertion becomes the initial underlying assumption of the ensuing analysis; *by balancing the net surface input over a basin with the surface output from a basin, the difference may indicate the residual inflow or outflow attributable to groundwater flow.*

This simple analysis may be the first step towards understanding the magnitude or importance of the lateral groundwater flow component within a given basin. It is well established (see section 2.5, and references therein) that groundwater and surface water interact as part of an intricate terrestrial water cycle, redistributing continental energy along dynamic pathways. This leads to a second assumption; *that groundwater within the saturated zone can travel hundreds, perhaps thousands of kilometers, redistributing continental water, often concentrating water into large river valleys.*

If we take the example of two nested hypothetical basins (Figure 2-2), where stream discharge is measured at the outlet of the smaller, internal basin, and again at the outlet of the larger basin, the R – Qr = Qg relationship of the larger basin incorporates the same relationship from the smaller basin. In this case, some component of the Qr measured at the outlet of the larger basin contains groundwater exported from the smaller internal basin. This line of reasoning leads the first hypothesis:

> *Hypothesis 1: For a drainage basin with a regional topographic or hydrologic gradient, there exists a hinge-line, above which a basin, and its internal constituents, is a net groundwater exporter, and below which a basin is a net groundwater importer.*

This additionally implies that a larger component of the recharge feeds into the groundwater system further upstream within a basin system (above the regional hinge-line), and no recharge is incorporated in the groundwater system downstream of the hinge-line.

The degree to which a basin is a net importer or exporter of groundwater is quantifiable by calculating the ratio of surface-water discharge to local rivers, Qr, to the surface recharge over a basin, R. In general, Qr/R $\neq$1, except in self contained large basins; that is, the surface water out-flow should not exactly balance surface input except in macro-scale (continental) basins where the effects of climate and geology are homogenized. We can also surmise that the Qr/R ratio may be used as an indicator of the relative importance of the groundwater flow pathway within a basin; the smaller the magnitude of the Qr/R ratio, the larger the magnitude of groundwater flow.

With this basic conceptual framework and underlying hypothesis in mind, several other hypotheses arise when considering the magnitude and location of a hinge line in the groundwater flow system. A useful concept is to consider basins as large scale analogues to stream channels in terms of gaining or losing subsurface input: groundwater exporting basins can be considered 'losing basins,' while their importing counterparts are 'gaining basins.' Thus, a second hypothesis:

> *Hypothesis 2: For a basin of given geology and climate, containing a given stream order, elevation affects the relative position of gaining and losing basins (and hence the hinge-line); losing basins (where Qr/R<1) should be found at higher elevations, and gaining basins (where Qr/R>1) should be found at lower elevations. If both gaining and losing basins are present within the macro-scale basin, then we expect to see a correlation between the Qr/R ratio and basin elevation.*

In this context, a third hypothesis is apparent:

Hypothesis 3: For a basin in a given geology and climate, both internal and external basin scale affect the position of a hinge-line; Qr/R approaches 1 from both above and below with increasing stream order. Hence, the larger the basin, the greater the degree to which R is balanced by Qr; in this way, stream discharge accounts for a progressively larger portion of the total surface recharge over a basin with increasing basin size.

Following this assertion, it seems that key to unraveling this issue will be the investigation of basins in a nested configuration where the dynamic relationships within the smaller internal basins may be placed in the context of the larger enclosing basin. This approach may allow a clearer separation of regional from local flow using stream discharge, and facilitate a conceptual, perhaps quantitative evaluation of Toth's (1963) flow regimes. A fundamental consequence of this assumption is that stream discharge (Qr) may be used as a proxy for local flow systems in small basins, where baseflow sustenance of streams may be more detectable.

As reviewed in the previous section, three primary agents are thought to control the development of groundwater flow systems in a given basin, climate, geology and topography, each with its own range of variability. Augmenting the framework above, some potential combinations of these variables will be examined in terms of the following hypotheses:

Hypothesis 4: For a given geology, the climate affects the development and position of a hinge line; the more arid the climate, the farther Qr/R deviates from 1. In addition, the more arid the climate, the higher the maximum stream order of a *losing basin, and hence, the larger the overall size of the losing basin. Losing or gaining basins may be found over a wide range of climate; however, gaining and losing basins tend to be more pronounced in more arid climates. It follows that losing or gaining basins may be found over a wide range of scale, although in arid climates losing basins tend to be larger and gaining basins tend to be smaller.*

Hypothesis 5: For a given climate, the geology – which controls the topography – affects the development and position of a hinge-line; the steeper the regional gradient vs. the local gradient, the farther Qr/R deviates from 1; the deeper and more permeable the substrate, the farther the Qr/R deviates from 1; the higher horizontal/vertical anisotropy in conductivity, the farther Qr/R deviates from 1.

Hypothesis 6: It is expected that Qr/R approaches 1 at large scales where the combined influences of dynamic subsurface geology and climate regime may interact both constructively and destructively in terms of the net effect; hence, larger scales tend to homogenize any internal differential losses or gains of the large-scale basin. Further, over progressively larger macro-divisions of the continent, the overall hydraulic influx must eventually become out-flux, assuming continental storage is at equilibrium. Thus, an excellent first-order test of the methodology is whether or not the 'ins' do indeed balance with the 'outs' over the largest basins.

4. Methods and Data Source Considerations

4.1. Hydro-Climatic Data Network (HCDN) River Discharge

The United States Geological Survey (USGS) produced the Hydro-Climatic Data Network (HCDN) compellation of high spatial/temporal fidelity river discharge measurements over the contiguous U.S. (Slack et al., 1993). The HCDN streamflow project assembled data from ~1,659 river gauging stations across the U.S., after applying a strict rubric for the characterization of each station as suitably free from anthropogenic complications, with flow naturalized to remove the effects of river diversion or dam storage, for inclusion in the report (Slack and Landwehr, 1992). Included in this compendium is the elevation above sea-level, latitude and longitude of the stream gauge, the area of coverage by the corresponding watershed, channel slope, and basin aspect/topography, over the respective record periods of each station. Additional qualitative information characterizing each station and its coverage area are included as well.

River discharge measurements over the period of water years 1874 through 1988 from each station (depending on the availability at each station) are summarized into daily, monthly, and annual mean discharge figures per station. As a result of this analysis, the HCDN streamflow data have the highest fidelity possible for any given gauging station, and the data set as a whole has high temporal variability. Thus, these data were employed in the ensuing analysis.

4.2. Variable Infiltration Capacity (VIC) Model Simulation and Output

The evapotranspiration data set utilized here is the result of a Variable Infiltration Capacity model (VIC) retrospective simulation (Maurer, et al., 2002). The VIC hydrologic model balances surface water and energy over a macro-scale model grid, and can be applied to a continent as a whole, or its constituents. The specific data used in this study were the result of a 50 year model run (January 1950 to July 2000) at 0.125 º increments (longitude and latitude) at three-hour time steps (Maurer et.al., 2002). The VIC model was calibrated using shorter time-span runs where the modeled runoff values were compared to observational river discharge data at selected basin outlet points. More importantly, each time step in the model closes both the water and surface energy budgets within that time step, and there is no assimilation of observed land surface conditions, aside from the initial test calibration (Maurer et.al., 2002).

The VIC model is unique in its ability to differentiate between the surface water which goes directly to runoff, and that which infiltrates the substrate. This is accomplished by the parameterization of vegetation, topography and spatial variability in soil characteristics, based on a suite of data sets and in part, the simulated outputs of other spatial models, which are discussed below (Maurer, et al., 2002). Also, the VIC model implicitly calculates a surface energy balance as part of each model iteration. Thus, the VIC model is able to more accurately simulate observed surface-water runoff conditions than other macro scale hydrologic models.

The soil attributes used in this VIC model run were taken from the Land Data Assimilation System (LDAS) of Mitchell, et al. (1999) at 0.125º resolution, with parameters such as field capacity, texture and saturated hydraulic conductivity. A three layer soil profile was presubscribed, and the depths of each layer were derived during the model calibration phase. Land-cover assignments were based on the 1km global vegetation classification into 14 classes, as described by Hanson, et al. (2000). Each 0.125º grid cell was assigned a leaf area index (LAI), including seasonal variations, a root depth parameter, based on vegetation type, all of which effect the efficiency and extend of evapotranspiration within each grid cell (Maurer, et al., 2002).

The VIC simulation is bounded by observed atmospheric and radiative forcing wherever possible; most importantly, precipitation, wind, and temperature. Since wind observations are often biased and sporadic, NCEP-NCAR (National Center for Environmental Prediction – National Center for Atmospheric Research) daily wind field product of Kalnay, et al. (1996) was linearly interpolated from the original 1.9º grid to determine values within the 0.125º VIC model grid. Daily total precipitation data from the National Oceanic and Atmospheric Administration (NOAA) were gridded to the 0.125º resolution, and correlated to the appropriate timestep for inclusion in the model. It is noted that areas where snow precipitation adjustments were not made may underestimate actual precipitation values. Using the same gridding algorithm, temperature data, also attained from the NOAA measurement stations, was rescaled to fit the model grid and adjusted for elevation using an adiabatic lapse rate of – 6.5ºC km^{-1} (Maurer, et al., 2002).

Though, it must be noted that the VIC simulation did not consider the influence of the water table, the presence of which could feedback to the infiltration and recharge rate calculated in the model. However, drainage from the bottom layer was carefully constrained using the Arno model (Liang, et al, 1994), and the simulated river discharge was compared to observed discharge using hydrographs from basin outlet points for 12 macro-scale basins over the 10 year test period (Maurer, et al., 2002). A high degree of agreement between the modeled and observed river discharge measurements over this timescale suggest that evapotranspiration is modeled equally as successfully, particularly since observational precipitation data were incorporated into the model (Maurer, et al., 2002). This coupled with the fact that evapotranspiration is not easily or historically measured makes the VIC simulation output the best dataset for analysis in this study. The evapotranspiration variable is the only model product utilized in the ensuing analysis.

4.3. Data manipulation

In order to satisfy one of the primary objectives of this study, separating river discharge from groundwater flow, the data need to be extracted and formatted for input into equations 2 and 3:

(2) $\mathbf{Q_g = Q_r - R}$, **and (3)** $\mathbf{R = P - ET}$

Where: $\mathbf{Q_g}$ is the groundwater underflow, $\mathbf{Q_r}$ is the stream discharge, **R** is the surface recharge parameter, **P** is precipitation, and **ET** is evapotranspiration.

Since groundwater underflow ($\mathbf{Q_g}$) is the unknown, the HCDN observational river discharge product is the most accurate source of the necessary $\mathbf{Q_r}$ surface flux parameter. For the purposes of this study, which is limited to the contiguous U.S., excluding the states of Alaska and Hawaii, 1555 HCDN stations were selected for analysis. The discharge data from these 1555 stations were isolated by USGS station ID and associated with the USGS Hydrologic Unit Code (HUC) of the basin corresponding to each station(s). This information was used to compile a data set where the discharge from each station was averaged over the area of its corresponding basin, yielding the correct $\mathbf{Q_r}$ parameter for the proposed analysis. The discharge, converted to m^3s^{-1} was divided by the basin area, converted from km^2 to m^2, yielding ms^{-1}, which represents the discharge ($\mathbf{Q_r}$) flux out of a given basin.

The VIC modeled P and ET data, discussed below, begin at 1950 and span to the year 2000. Since the HCDN data includes only up to the water year 1988, the discharge values extracted from this data set are limited to the time span between 1950 and 1988. Thus, a total of 39 years of streamflow data are employed, with year (timestep) ‘00’ corresponding to the year 1950 and timestep ‘38’ corresponding to the year 1988.

The VIC simulation output data for monthly precipitation (P) in $kgm^{-2}s^{-1}$ and evapotranspiration (ET) in $kgm^{-2}s^{-1}$ (at monthly timesteps) was first averaged over every 12 timesteps (from October to October, to account for the use of water years in the HCDN discharge product), and converted from NetCDF format to ASCII. The Generic Mapping Tools (GMT) and netCDF Operators (NCO) suite of programs were used to subtract the evaporation data set from precipitation, arriving at a recharge (R) value. This R value was converted from $kgm^{-2}s^{-1}$ to ms^{-1} by dividing by an assumed density of water as $1000kg/m^3$. The VIC P, ET and R data were then converted to ArcGIS readable raster layers for manipulation via a nested drainage basin file corresponding to each of the 1555 HCDN drainage basins in the river discharge data set.

The nested drainage basins were used as a mask to extract each basin’s ET, P and R values so they could be evaluated in terms of a basin-averaged discharge value (Qr) from the HCDN product (discussed above). Since the distribution of basin sizes in the HCDN set included many basins smaller than the 0.125º x 0.125º grid size of the VIC model output, the VIC data were split into 0.0625º x 0.0625º grid cells.

Since the resolution of the Qr data is such that each basin has only one discharge value associated with it, it was necessary to average the R, P and ET values over the span of each basin, so that each basin is represented by a single value for each. This method was deemed appropriate because equations above (Equations 2 and 3) suggest that the spatial variability within the P, ET and R data sets is implicitly included in the stream discharge for a given basin. Concordantly, the product of this procedure is one data value for each variable, Qr, ET, P and R, corresponding to each basin over the 38 years of record, with were then averaged to arrive at the climatologic mean over the 38 years. This climatologic mean was converted from ms^{-1} to $mmyr^{-1}$ for human readability.

4.4. Comparisons between Results

Because the definitive product of Equation 2 is Qr, or the contribution of recharge (R) that ultimately delivered to the stream system, the ratio of Qr to R becomes the surface discharge per-unit recharge to exit a given basin. This Qr:R ratio is calculated for each of the 1555 HCDN basins used in this analysis, and provides a relative basis for comparison of the recharge contribution to the groundwater system vs. the river system within and between basins.

This ratio is used for comparison to other factors, following the hypothesis outlined in section 3. Under hypotheses 2 and 3, that climate affects the distribution of Qr:R ratio with area and elevation, data for the 1555 HCDN basins are binned by climate zones based on recharge as the following: from 0-200 mm/yr, 201-400 mm/yr, 401-600mm/yr, 601-800 mm/yr and above 800 mm/yr. The Qr:R data binned thusly is compared to both basin drainage area and mean basin elevation. The only sure way to test the hypothesis that regional geology may control the distribution of the Qr:R ratio with elevation and area is to examine specific cases where the geology is well characterized. Thus, based on general basin characteristics, and the number of internal basins with flow into one-another, basins are selected from the contiguous U.S. for further analysis.

5. Results and Discussion

5.1. Results: General Features of the Groundwater-Surface Water Partition

Figure 5-1 spatially represents the results of the analysis separating Qr from Qg over the 1555 HCDN basins in terms of the stream discharge per-unit recharge ratio (Qr:R). Basins nested within larger basins are overlaid above those larger basins so the values can be seen; the values for each Qr:R interval are noted in the color scale to the right. Under the hypothesis that climate affects the distribution of Qr:R ratio with area and elevation, data for the 1555 HCDN basins were binned by climate zones based on recharge as the following: from 0-200 mm/yr, 201-400 mm/yr, 401-600mm/yr, 601-800 mm/yr and above 800 mm/yr.

In those basins receiving less than 200 mm/yr of recharge, a comparison of Qr:R to basin drainage area (Figure 5-2) shows a bimodal distribution where those basins with Qr:R values above the Qr:R = 1 line decrease with increasing basin area, and those below increase with increasing area ($r = 0.17$ and 0.16 respectively). Evaluation of the Qr:R ratio against basin elevation (Figure 5-3) in this climate zone reveals increasing stream discharge per-unit recharge with increasing basin elevation. The $200 < R < 400$ mm/yr climate bin shows a similar relationship to the $R < 200$ mm/yr bin between the Qr:R ratio and both basin area and elevation. In both cases, however, the relationship is weaker. In the $400 < R < 600$, $600 < R < 800$, and $R>800$ mm/yr bins, Qr:R shows very little change with area, and a weak trend of increasing Qr:R with increasing elevation, with the exception of Qr:R vs. elevation in the 600-800 mm/yr bin, where $r = 0.47$.

5.1.1. Spatial Distributions of Qr:R Ratio

A few broad observations can be made based on Figure 5-1 (Qr:R distribution across the contiguous US), particularly in terms of the position of a regional hinge line by making broad assumptions about geology and climate, following the hypotheses outlined previously. The coastal plains from Delaware to Texas are often characterized by deep unconsolidated sediments (USGS Groundwater Atlas, see http://capp.water.usgs.gov/gwa/gwa.html). Here, the position of the hinge line is close to the coast across a wide range of climates primarily due to highly efficient aquifer drainage in these regions. Geochemical tracer studies have identified significant submarine discharge of groundwater offshore of these regions (Burnett, et al., 2006). The basins of these coastal plain regions are primarily groundwater exporters, while limited numbers of purely coastal basins are included in the HCDN data set, it can be surmised that these basins would be groundwater net importers if the hinge line were indeed located onshore. Here, geology is the controlling factor on the distribution of gaining and losing basins through several climate regimes.

When considering the position of the hinge line in terms of climate, the distribution of gaining and losing basins in Figure 5-1 suggests that regional hinge line is lower in more arid climates and the size of the losing basins is larger. A comparison may be made between the Arkansas River and the Ohio River: The Arkansas River lies above the regional hinge line, containing largely exporting basins, while the Ohio River lies below the hinge-line, and is for the most part characterized by groundwater importing basins. Also, the regional hinge line is enclosed when larger basins are considered, such as in the Ohio River, Missouri River and Upper Mississippi River basins. Here the size of the basins encompasses the hinge line and we see that many of the smaller internal basins are groundwater exporters while the larger basins are importers, or fall close to the Qr:R = 1 line. In these cases, both climate and basin scale effect the location of the regional hinge line.

Further, a comparison between the Clearwater River basin in Northern Idaho and the Colorado River basin in western Texas leads to some key observations. The Clearwater River basin is composed entirely of groundwater net-importers, where as up the elevation gradient are the primarily losing basins of the Pend Oreille River, suggesting that a hinge line exists somewhere between the two. In contrast, the Colorado River basin consists of entirely of losing basins, with no adjacent gaining basins, such that the regional hinge line must exist somewhere far removed. Here, the climate effects the distribution of gaining and losing basins, and hence the position of the hinge line, as well as the size of losing basin, where the arid Colorado River basin is much larger than the more humid Clearwater River basin.

5.1.2. Relationships with Drainage Area

The relationship of decreasing Qr:R ratio with increasing area in the R<200 mm/yr climate zone is consistent with the hypothesis that a convergence towards the Qr:R = 1 line should occur with increasing area (Figure 5-2). The implication of this hypothesis is that the surfacial input over a basin is better balanced by stream discharge over larger basins, which may include both gaining and losing basins. However, the relationship exists, but is weaker in the 200 < R < 400 mm/yr bin, and all but falls apart in the increasingly humid climates. This observation suggests that regional groundwater flow may be less influential in humid climates, an assertion also made by Toth (1963).

Also related to basin scale, we find that the basins in the R<200mm/yr bin have the most variability in Qr:R with area. Thus, it appears that the dryer the climate, the further the Qr:R ratio deviates from Qr:R = 1 (Figure 5-2). This is consistent with the hypothesis that regional groundwater flow, and hence the degree and size of losing basins, are more pronounced in arid climates. In the progressively more humid bins, the Qr:R ratio is more tightly clustered around the Qr:R = 1 line, and though the overall scatter is in part due to regional flow, local flow may be progressively more influential with decreasing aridity. This assertion is similar to that proposed by Toth (1963), where local flow may become more significant in humid climates where the water table is able to interact with the land surface to a greater degree.

In the interest of considering outliers in the Qr:R vs. basin drainage area relationship, we find that the highest Qr:R value in basin 09419610, a small basin (area = 24 km^2) in Charleston, Nevada (inset discharge plot, Figure 5-2). The Charleston, Nevada basin sits at the foothills of the Sierra Nevada Mountain range, near Death Valley. It is an ephemeral stream, with a strong seasonal discharge regime, and likely receives groundwater discharge from the adjacent mountain range. In this way, the stream discharge value is nearly 820% of the recharge received over the basin, suggesting that groundwater discharge, perhaps from a spring, is a dominant portion of the total stream discharge from the basin.

Further observation indicates that in humid climates the Qr:R ratio is more closely clustered around the Qr:R = 1 line. This observation, and the high discharge to recharge ratio found in the Charleston, NV basin, are both consistent with the hypothesis that gaining and losing basins may be more pronounced in arid climates. An implication of this observation is that the groundwater-surface water reservoirs in these regions may be more tightly coupled than in arid climates, further supporting Toth's (1963) assertion that local flow may be dominant in humid climates. However, this may also suggest that the subsurface geology plays a more important role in regions not limited by recharge to the groundwater system. Thus, elucidating the details of this relationship require a closer look at the geology in such regions.

5.1.3 Relationships with Basin Elevation

In each recharge zone, the Qr:R ratio shows an increasing relationship with increasing basin elevation (Figure 5-3). This suggests that higher elevation basins are overall groundwater net importers; a relationship which is not consistent with the hypothesis that high elevation basins should be groundwater net exporters. A further observation is that, across all recharge zones except the 600-800 mm/yr bin, the Qr:R ratio shows less of a significant relationship with elevation with increasing recharge. This suggests that the relationship is most pronounced in arid climates, and increasingly less prominent in more humid climates.

A potential problem with comparison to basin mean elevation across the data set in its entirety is the inclusion of exceedingly large basins whose average elevation includes both high and low elevations within the basin. The high or low elevation values included in this average may represent smaller basins that are also delineated within the data set. However, a basin where drainage area covers a transition from high relief mountain valleys, for instance, to low relief alluvial plains has a wide range of elevations included in the mean basin elevation. The large area basins may bias the data in this way.

Another potential confounding factor is that the HCDN basin data set is somewhat spatially intermittent, and many basins are not represented. The high elevation gaining basins may not be representative of the highest elevation drainage basins, and thus may be gaining from higher elevation losing basins not included in the analysis. However, this assertion remains purely speculative. The fact that comparison of Qr:R ratio to elevation yields the opposite relationship than that hypothesized, even when binned by recharge, suggests that we need to look at the data on a more case by case basis and consider the effects of the substrate. The seeming lack of correlation between Qr:R ratio and area also suggests that the a significant trend may be washed out by the mass of data and a more in depth approach, using individual basins in a nested relationship, is necessary.

Since climate is only one of the two hypothesized controlling factors over the groundwater to surface water partition, the underlying geology must be considered in order to better explain the relationships presented above. As stated in the previous section, the most accurate way to test the hypothesis that geology exercises great control over the Qr:R distribution with both area and elevation, is to examine specific regions in further detail where the geology is well characterized.

5.2. Select Case Studies

Individual cases of nested basins from specific climatologic and geologic regimes have been selected for further review. These case studies are meant to assess the potential factors influencing the groundwater/surface water partition in detail, particularly with respect to subsurface geology, to test the general hypothesis;

Hypothesis 5: For a given climate, the geology – which controls the topography – affects the development and position of a hinge-line; the steeper the regional gradient vs. the local gradient, the farther Qr/R deviates from 1; the deeper and more permeable the sediment, the farther the Qr/R deviates from 1; the higher horizontal/vertical anisotropy in conductivity, the farther Qr/R deviates from 1.

Such a test is not feasible when examining the data in its entirety. This approach may better isolate some fundamental characteristics of the groundwater-surface water partition, and allow more specific testing of hypotheses 1, 2, 3 and 4, outlined in section 3. The results from each basin analysis are summarized with discussion following each section of results.

5.2.1. Basin 08158000 – Colorado River headwaters to Austin, Texas

Basin 08158000 sits in south-central to southwest Texas and covers 101,033 km^2, with an average annual recharge of 76.36 mm/yr over the analysis period. The main channel in the basin is the Colorado River, and basin 08158000 covers the reach extending from the headwaters at the TX-NM border to Austin, TX (Figure 5-4). The Colorado River basin was chosen for further analysis because of its relatively wide distribution of small and large internal basins with varying elevation and stream order. This basin is ideally configured to test the hypothesis that the maximum order of losing streams should increase with aridity and, that for a given climate, the geology effects the position of a hinge line.

A comparison of the average discharge from the Colorado River basin and its internal constituents as a portion of total calculated recharge (Qr:R ratio) for the analysis period suggests that surface water discharge ranges from 5% - 51% of the total recharge over the internal basins. A correlation of these results to the internal area of each respective component basin suggests that the portion of recharge becoming river discharge decreases linearly with increasing basin area (r = -0.52) (Figure 5-5).

In this case, it is more useful to consider the ratio of Qg:R – that is, the proportion of recharge that becomes groundwater discharge (Qg) from a given basin (see equations 1 and 2) – because these values are closer to 1, the point where Qg = R. The Qg:R values for the Colorado River basin range from 0.49 – 0.95; groundwater discharge comprises 49 -95% of the total R over the basin. When correlated to basin area (Figure 5-6), a linear trend of increasing Qg:R ratio with increasing area (r = 0.52) suggests that groundwater discharge (Qg) becomes a larger component of the total basin recharge (R) as basin size increases. Further, with increasing basin area, the value of Qg:R approaches the Qg:R = 1 line (where groundwater discharge is equal to recharge over the basin).

Further investigation of the groundwater discharge to recharge ratio reveals that Qg:R also increases as a function of basin elevation. A separation of the component basins based roughly on stream order shows an increasing trend in Qg:R ratio with elevation for first and second-order streams (Figure 5-6), while the correlation of all basins to elevation yields an r value of 0.47.

5.2.2. Texas lowlands – Basins 08176500, 08210000, Guadalupe and Nueces River Basins.

The Guadalupe and Nueces River basins lie southeast of the Colorado River basin, in the transition zone between the Texas interior highlands and the coastal plain (Figure 5-7). The Nueces River basin drains a total area of 39,956 km^2 and has an average recharge of 85.62 mm/yr, while the Guadalupe River drains 13,463 km^2 with an average recharge of 184.41 mm/yr over the analysis period. These two basins were selected by their proximity directly adjacent and down-slope of the Colorado River basin to test the hypothesis that elevation and topography, controlled by the geology, effect the distribution of gaining and losing basins.

Since the number of basins within the Guadalupe and Nueces River basins is relatively low, the two basins were grouped for comparison. The calculated Qr:R implies that river discharge accounts for 21% of the recharge in the Nueces River Basin, and 52% in the Guadalupe River Basin; Qr:R values 0.21 and 0.52 respectively. When Qr:R is evaluated against basin drainage area it shows a decreasing linear trend ($r = -0.42$) with increasing basin area (Figure 5-8). Conversely, Qr:R shows a weak linear increase when compared to elevation ($r = 0.30$) (Figure 5-8). The two inset plots in Figure 5-8b show the distribution of Qr:R with respect to elevation before the two basins were grouped; no appreciable trend was weakened by combination, in either case.

It must be noted that a general gradient of decreasing precipitation exists from east to west throughout the region, where the Gulf of Mexico is the primary source of moisture (Barker and Ardis, 1996). Thus, the basin of the Nueces and Frio Rivers in southwest Texas receive less precipitation than the Guadalupe River basin further to the east.

5.2.3. Select Texas coastal basins

The HCDN data set did not include the lower basins of the southeastern Gulf Coast of Texas, but a collection of 17 upper coastal plain basins were available for analysis. The average drainage area of the basins studies was 1150 km^2, with a range between 272 km^2 and 2139 km^2 over the 17 basins (5-7). The basins were selected because they lie down the topographic gradient towards the Gulf of Mexico from the coastal uplands and Colorado River basins, and may be used to evaluate the validity of the hypothesis that regional groundwater discharge occurs below the regional hinge line.

The computed Qr:R values over the 17 coastal basins reveal a mean 71% of the incident recharge can be accounted for by river discharge, with a range of 7% to 138%; Qr:R values over the basins were 0.07 – 1.38 and averaged 0.72. The computed Qr:R ratios for the 17 basins compared to basin drainage area (Figure 5-9) show no significant correlation between the variables. When the Qr:R values over the 17 basins are compared to elevation (Figure 5-9), a linear trend of decreasing Qr:R ratio with increasing elevation becomes apparent (r = -0.54). The elevations of the selected basins range between 50 m and 120 m above sea level.

5.2.3.1. Groundwater Flow in the SW to SE Central Texas basins

At first approach, it would seem the comparison of the Qg:R ratio to basin elevation and basin area are somewhat contradictory within the Colorado River basin. Why would the relative size of groundwater discharge to recharge be increasing with both basin area and elevation? This relationship suggests that perhaps a skew has been introduced by the relatively large, high elevation basin 08126500 (Colorado River headwaters, see Figure 5-4). However, removing this basin from either correlation does little to significantly weaken the relationship between either variable. Alternatively, perhaps the magnitude of the groundwater component is large because much of the recharge over the basin as a whole leaves via regional flow pathways rather than surface runoff from the Colorado River. This assertion is in support of the hypothesis that basins above the regional hinge-line should contribute surface waters to the groundwater flow system, but requires a closer look.

An initial reason for examining the Colorado River basin in detail was the presence of several first and second-order stream channels with relatively even distributions with respect to internal basin area and elevation. The delineations of stream-order per basin are expressed in Figure 5-6, where orange and pink are small and large first-order channels, respectively, and the green denotes basins with second-order channels. Considering the relationship of Qg:R across stream order, we see that groundwater discharge is smaller with respect to recharge in lower elevation first and second-order streams. Inversely, groundwater discharge is larger with respect to recharge in high elevation streams, suggesting that these basins lose more recharge to groundwater flow than those in lower elevation first-order stream basins. This apparent relationship is in support of the postulates of Toth (1963), as well as the initial hypothesis of this study, that high elevation basins should be larger contributors to regional groundwater flow than their lower elevation counterparts within the macro-scale basin.

Thus, the increasing groundwater discharge relative to recharge with increasing area (Figure 5-6) suggests that perhaps groundwater is not exported to a location within the extent of the Colorado River basin. The overall decreasing surface water discharge with basin area while the groundwater counterpart is increasing suggests that the basin is comprised of a network of completely losing streams. Thus, the Colorado River basin is a net-exporter of groundwater, and can be considered an overall “losing basin” sitting somewhere above the regional hinge-line.

The Colorado River basin sits in part within the Edwards-Trinity aquifer system of southwest to central Texas. The topography of the southwestern portion of the basin, within the Edwards plateau, consists of relatively flat-lying rocky plains dissected by steep canyons cut by streams and springs. The general slope over the upper portion of the basin is toward the southeast, with areas of high relief superimposed. Mountains lie to the west and southwest of the Edwards Plateau region (USGS Groundwater Atlas).

The Edwards plateau consists of carbonate rocks with high permeability, while the southeast portion terminates just above the Balcones Fault Zone (Figures 5-10 and 5-11). The carbonate rocks composing the aquifer system are generally southeastward dipping beds that outcrop in the southwestern section of the Colorado River basin (Figure 5-11), and are progressively overlain by onlapping siliciclastic rocks to the southeast. The water table throughout much of the aquifer is deep and disconnected from the land surface, except in the upper regions of high relief, where it may briefly intersect the land surface and form springs (Barker and Ardis, 1996).

Further examination of the USGS RASA file on the Edwards-Trinity aquifer system shows that regional groundwater flow is generally parallel to the Colorado River through much of the basin, until the lower reaches, where the potentiometric contours bend towards the river (Figure 5-10). The upper reaches of the smaller tributary channels to the Colorado river are largely sustained by spring flow, which may account for the slightly higher Qg:R ratios in these basins. Further down gradient, into the Nueces and Guadalupe River basins, the regional flow system encounters the Balcones fault zone, where the regional groundwater flow becomes scattered, often along strike to the northeast, discordant with the potentiometric surface (Barker and Ardis, 1996) (Figure 5-10).

Numerous springs exist along the Balcones heavily faulted zone, often where faulting has placed highly permeable units adjacent to impermeable units, and the general topographic relief is quite steep (figure 5-11). The resulting springs are likely the cause of the high elevation gaining basins within the Nueces river sub-basins, which flow quickly to the south out of the small basins. However, the relief is such that spring waters often re-enter the groundwater flow system along fault planes further downslope. As surface water exits these small basins, it enters channels flowing through the Texas coastal uplands aquifer system, which is characterized by southeastward dipping and progressively thickening sand and silt beds roughly corollary to similar units in the Mississippi Embayment aquifer system to the northeast (Grubb, 1998) (Figure 5-12). The dip of the sedimentary units below the Balcones Fault Zone increases relative to the dip of the Edwards plateau, largely controlled by the steeply sloping base of the sedimentary basin to the southeast.

The horizontal and vertical hydraulic conductivity of the coastal uplands aquifer system, as well as the coastal lowlands aquifer system, tend to increase down-dip toward the coast (Grubb, 1998). This increase in horizontal and vertical conductivity may account for the progressively losing nature of the lowlands basins, where the high elevation spring fed basins feed the lower lying basins, which are characterized by the primarily losing channels of the Nueces and Guadalupe river systems (Ryder and Ardis, 2002). The topographic gradient – controlled by the onlap of the Texas coastal uplands units onto the Balcones Fault Zone – decreases southeastward from ~800 m to ~150 m above sea level, while the water table through the gradient is roughly 600m in the uplands to 50 meters above sea level at the lowest extent of the basin (Figures 5-11 and 5-12).

A regional groundwater flow system has been described and modeled over the extent of the two basins, which suggest that the direction of regional flow is generally towards the coast with increasing volume (Grubb, 1998). In this way, the Nueces and

Guadalupe River basins of the Texas coastal uplands can be considered overall losing basins, or net groundwater exporters, because an overall smaller proportion of the water that is recharged over the basin flows from these rivers. Along the same lines, only a small portion of the water within the regional flow system discharges to the surface within the high elevation faulted zone.

The base of the coastal uplands aquifer system yields to the south-southeast dipping beds of the highly permeable Texas Coastal Plain aquifer system (Ryder and Ardis, 2002). The data from this study do not adequately represent the gulf coast of Texas (many were excluded from the original HCDN data set due to pumpage). However, the coastal basins which do lie within the coastal plain aquifer sit in the upper portion of the latter and generally have Qr:R ratios below Qr:R = 1, suggesting they are mostly groundwater exporting basins.

Unfortunately, insufficient numbers of coastal basins within a close proximity to the shoreline prevent quantification as to whether they may be regions of groundwater discharge. On the other hand, the reader is referred to Figure 5-7**,** where a noticeable zone of abrupt increase in drainage network density along the gulf coast is readily evident (particularly along the southernmost coast of Texas). Slightly up-dip is likely the hinge-line of transition between groundwater recharge and discharge zones. Indeed, a flow regime similar to that suggested by Toth (1963) has been reported here (Williamson and Grubb, 2001), where the upper coastal basins recharge the underlying aquifer and those close to the coast receive discharge from this regional flow system.

Williamson and Grub (2001) also stress the effects of the salt water wedge on the position of the groundwater discharge zone; when recharge is less, and regional flow diminished, the salt wedge moves landward, and thus the zones of regional discharge likewise move landward. When recharge is greater and the flow system stronger, the salt wedge is pushed offshore, such that regional groundwater discharge moves closer to the coast (or, as in predevelopment conditions, potentially offshore). Pumping rates for agriculture and municipal use also greatly effect the position of the hinge line, and hence the encroachment of the saltwater wedge.

Thus, it can be surmised that a hinge line of the regional flow system exists somewhere in the Coastal Plain aquifer system, above which a recharge region exists, extending upland through the Balcones Fault Zone to the Colorado River basin. Topography controls the surface water expression of the groundwater system only to a certain degree; springs in the Colorado River basin occur because of intersects between the water table and the high relief of the topography. However, the springs along the Balcones Fault Zone are controlled by structural features providing conduits to the surface along which groundwater may flow. In this case, the recharge is sufficiently low, and the permeability of the subsurface geology sufficiently high that a network of completely losing basins, and thus a regional groundwater flow system have developed. The negative correlation between Qr:R and basin drainage area over the Colorado River basin further suggests that, for a given recharge value, the geology is disproportionately more efficient at transporting surface input to the water table with increasing area. For this reason, larger basins are greater net-exporters of groundwater. These examples support the hypothesis that in arid climates, the size and extent of losing basins is larger, and therefore the discharge to recharge ratio deviates farther from 1.

5.2.4. Basin 02358000 – Flint and Chattahoochee River Basin, Georgia, Alabama, Florida

The Chattahoochee and Flint River basin sits in the Gulf of Mexico coastal plain and spans over half of the Georgia – Alabama border, with its mouth in northwestern Florida (Figure 5-13) and a total area of 44,548 km^2. The basin received between 530 – 760 mm/yr of recharge over the analysis period, with a mean of 615 mm/yr. The Chattahoochee River basin was selected for the nested relationship of its constituent basins and the good distribution of high and low elevation first-order stream channel basins. This basin may be used to test the hypothesis that, for a given climate (humid), the geology may effect the location of the hinge line; and that the deeper and more permeable the sediment, the further the Qr:R ratio deviates from 1.

Discharge comprised between 46 – 112% of the total recharge over the basin, with mean of 70%, where the calculated Qr:R ratio was 0.46 – 1.12 (mean of .70), with only one data point above 1. A comparison between the discharge to recharge ratio (Qr:R) and basin area suggests a weak linear trend of decreasing Qr:R with increasing basin area ($r = -0.17$) (Figure 5-14). Further comparison of the Qr:R ratio to internal basin elevation shows a weaker linear trend ($r = 0.08$) of increasing discharge per-unit recharge with elevation (Figure 5-14).

The comparison of Qr:R to basin area in the Flint and Chattahoochee River basins would likely be much better if not for the anomalously high value of basin 02349000 in the Whitewater Crk., GA. The basin is small (area = 242 km^2) (Figure 5-13), and therefore risks having a significantly underestimated value for surface recharge since the VIC P and ET output data were split into 0.0625º longitude/latitude grid cells. Since this is the only gaining basin within the Flint and Chattahoochee River basin group, and the data has a high probability of being erroneous, it is regarded with skepticism.

The overall decreasing Qr:R ratio with increasing drainage area and distribution below the Qr:R = 1 line suggests that not only is this basin losing as a whole, but the internal basins become increasingly greater net-exporters of groundwater with increasing stream order. This is not consistent with the hypothesis that lower-order streams ought to be greater groundwater exporters than higher order streams, unless the basins exist entirely above the hinge-line. This may indeed be the case, and the lack of any significant gaining internal basins fits this hypothesis as well.

This case study is similar to the last example from Texas, where the basins examined were overall above the hinge line, and thus net groundwater exporters. However, to say this is an adequate preliminary conclusion is not completely sufficient. The climatologic and geologic conditions through which the Chattahoochee and Flint Rivers traverse is indeed very different than those of the southern Texas basins from the last example. To more completely understand this situation, we must examine these factors more closely, and perhaps investigate the potential location of groundwater outflow from this basin.

The headwaters of the Flint and Chattahoochee Rivers are just south of the valley and ridge aquifer system of the southern Appalachian mountain range and span across the Southern Coastal Plain aquifer system, terminating ~30 km downstream of the confluence of the two rivers forming Appalachicola River in western Florida (Figure 5-15). The fall line transition between the coastal plain sediments and continental bedrock runs through the upper half of the basin (Figure 5-15 and 5-13), and incidentally coincides with the small gaining basin 02349000 (Whitewater Creek, which may help explain its name), as well as basin 02341800, which has the next highest Qr:R value. The significance or implications of this observation is currently unclear.

Further examination of Figure 5-15 shows that the Flint-Chattahoochee River basin terminates within the highly permeable rocks of Florida carbonate platform and its aquifer system, the surface outcrop of which appears as a “U” shape in the southernmost portion of the basin (Figure 5-15). In addition to the already highly permeable limestone of the Floridian Aquifer, extensive karsting has occurred in the system, giving groundwater a matrix of pipelines through the subsurface. This carbonate aquifer system has been identified as a major conduit of regional groundwater flow from the interior coastal plain (as well as Florida) to the Gulf of Mexico (Miller, 1986; Barker and Pernik, 1994). Further, because of the high permeability, the groundwater flow system is able to reach volumes high enough to force the saltwater-freshwater interface into the Gulf, facilitating submarine groundwater discharge off the coast (Renken, 1996). Groundwater flow lines are denoted in Figure 5-15, and represented schematically in cross-section in figure 5-16, after Miller (1986).

Indeed, Radon isotope tracer studies have confirmed significant submarine groundwater discharge in the Gulf Coast region, facilitated by both the highly permeable Floridian Aquifer units, and the thick, unconsolidated sands of the Coastal Plains Aquifer system (Cable, et al., 1996). Submarine groundwater discharge has even been attributed to anomalously high fluxes of phosphorus and nitrogen nutrients to Gulf waters, increasing productivity in areas of locally high influx (Slomp and Cappellen, 2004).

With this evidence in mind, it seems likely that the Flint-Chattahoochee River basin sits above the regional hinge line, exporting much of the incident recharge over its expanse toward the Gulf of Mexico. This case supports the hypothesis that the hinge-line, and thus the position of gaining and losing basins may be controlled by the underlying geology, despite the humid climate and ample recharge received by the basin. The efficiency of aquifer drainage in this case is of seemingly greater influence than the climate regime of the region, although the net effect is somewhat the same (losing basins within close proximity of the coast) as those of the Texas Lowlands and coastal basins.

Further, the climate in this case is most likely responsible for the position of the hinge line such that regional groundwater flow discharge is much further offshore than in the Texas case, and does not discharge (for the most part) terrestrially. In this way, more permeable, drainage-efficient geology, and to a lesser degree, increased recharge, has shifted the hinge line away from the continent and effectively pushed the salt-wedge seaward.

5.2.5. Basin 01646502 – Potomac River Basin, Pennsylvania, Maryland, West Virginia, Virginia

Basin 01646502 houses the Potomac River and sits just above the Appalachian - Atlantic Coastal Plain fall line, which runs through the easternmost corner of the basin (Figure 5-17) and borders the Mississippi River macro-scale basin to the west-northwest. The basin has a total area of 29,940 km^2, and mean annual recharge over the analysis period of 360 mm/yr, with a range between 268 and 448 mm/yr. The Potomac River basin was selected for further analysis because of the presence of several relatively high elevation gaining basins in a tightly nested relationship. This basin can be used to test the hypotheses that high elevation basins should be net groundwater exporters (losing basins), while low elevation basins should be net groundwater exporters (gaining basins), and that the ratio of Qr:R should approach the Qr:R = 1 line with increasing drainage basin area. Also, this case, as in the last, can test the hypothesis that geology controls the position and development of a hinge line.

River discharge accounted for 60% to 150% of the total recharge over the basin interior, with a mean of 102% overall; the calculated Qr:R values show a range between 0.60 and 1.50, with a mean of 1.02. Evaluation of the Qr:R ratio against basin area produces two populations nearly equally distributed across the Qr:R=1 line, suggesting a bimodal distribution with area. Those basins above Qr:R=1 display a decreasing linear trend of Qr:R towards 1 with increasing basin area ($r = -0.25$); while the basins below the Qr:R=1 line suggest an increasing linear trend of Qr:R with basin area towards 1 ($r = 0.34$) (Figure 5-18). Comparison of the Qr:R ratio with basin elevation yields a moderately strong trend of increasing Qr:R with increasing elevation ($r = 0.71$), with equal distribution across the Qr:R=1 line in a single linear relationship (Figure 5-18).

The relationship of Qr:R with area (Figure 5-18), where convergence toward the Qr:R = 1 line increases with basin size, supports the hypothesis that stream discharge should account for a progressively larger portion of the total recharge as basin size increases. The relative gains and losses of individual internal basins effectively balance over the basin as a whole, so the net effect is a Qr:R value of 0.97 for the largest basin, 01646502. However, the significance of this converging relationship with basin area needs to be examined in more detail since the regional geology is known to be complex.

The correlation between the Qr:R ratio and basin elevation is of particular interest (Figure 5-18); a strong linear relationship of increasing discharge to recharge ratio with increasing elevation, and a nearly equal distribution across the Qr:R = 1 line. This apparent relationship is opposite of that hypothesized, that higher elevation basins should be greater groundwater exporters, having lower Qr:R ratios. In this case, the higher elevation basins appear to be groundwater net-importers and have the highest Qr:R ratios.

Figure 5-17 geographically delineates those basins that are gaining from those that are losing, which makes a second spatial trend apparent: the higher elevation gaining basins are predominantly on the west-northwestern side of the basin, while the losing basins sit closer to the coastal region just above the fall-line to the southeast. The line separating gaining from losing basins runs generally northeast to southwest through the center of the basin (Figure 5-17), along the North Mountain Fault line.

The Potomac river basin spans the physiographic provinces of the Appalachian Plateau on the west-northwestern side of the basin, the Valley and Ridge in the west-central region, and Appalachian Piedmont to the southeast (Figure 5-19). The main portion of the basin sits within the Appalachian fold-thrust belt and geologic structure is the primary control over groundwater and surface water movement in the region (Hollyday and Hileman, 1996) (Figure 5-20). Indeed, the morphology of the stream channels in the region suggests strong structural control, where the upper branches of the Potomac River flow at right-angles into the main channel.

Identifying a source of the surplus stream discharge from high elevation basins proves problematic; however, several potential sources may be suggested. Rutledge and Mesko (1996) suggest that the primary groundwater flow systems in the Valley and Ridge region are localized, with some development of intermediate flow systems apparent. These groundwater flow systems generally discharge to streams, which have cut into the bases of syncline features in a northeast to southwest direction, tapping into permeable and often confined units within fold features. They suggest groundwater flows across strike from ridges in fractures and bedding planes, and along strike in stream basin valleys, contributing baseflow to stream discharge.

Rutledge and Mesko (1996) further point to the presence of narrow, but laterally extensive carbonate bands with dissolution features capable of transporting groundwater some distance. These carbonate passages may often be traced into the Appalachian Plateau province, making it a potential source of regional groundwater to the Potomac River basin.

The losing nature of the basins below the North Mountain thrust fault zone (Figure 5-17) is much more readily explained. The presence of extensive limestone units housing the carbonate aquifers of western Maryland and western Virginia (Figure 5-19) have high permeability and are known to be heavily karsted (Hollyday and Hileman, 1996). As in the Flint-Chattahoochee River basin, dissolution channels in the lower carbonate rock aquifer system create subsurface conduits that regionally distribute groundwater flow.

Although the sources of surplus stream discharge, above the level of recharge, cannot be specifically identified in the Valley and Ridge portion of the Potomac River basin, the overall geological control of the groundwater system is evident. Groundwater flow within confined carbonate units along strike, discharging to streams at incision points may be the culprit, but this remains speculation. The complex structure imposed by the Appalachian fold and thrust belt ultimately controls stream location, and areas of discharge and recharge. Groundwater from the portion of the basin coastward of the North Mountain fault flows within and into karsted carbonate systems which ultimately discharge to the coastal plain aquifer system bordering the Chesapeake Bay.

Two broad conclusions may be drawn from this example: Groundwater flow through structurally complex regions may obscure the identification of a hinge line; and the net influx from surface recharge balances with the river discharge over the large area of the Potomac basin. However, the latter point, as evidenced by the convergent relationship between Qr:R and area (Figure 5-18), may agree with the hypothesis that this should occur, but its occurrence may be coincidental. Given the geologic controls over the groundwater system in this case, the distribution of gaining and losing basins happens to balance with respect to area, but the internal zonation of potential flow systems within the basin does not allow the assumption that all the internal basins are hydraulically contiguous. Moreover, it can be said that the surfacial delineation of internal drainage basins – and the direction of stream flow – are not congruent with groundwater flow direction, or the separation of groundwater basins.

The relationship between the Qr:R ratio and basin elevation supports the null hypothesis that higher elevation basins need not be net groundwater exporters, and is consistent with the hypothesis that substrate geology is the primary control over the existence of the hinge-line. Unlike the last two examples, where the location of a hinge line could be readily defined based on rudimentary knowledge of the geology, in this case it can simply be assumed to exist within the Atlantic coastal plain below the fall-line. That is, if it exists at all.

5.2.6. Basin 05465500 – Cedar River Basin, Iowa

The Cedar Rapids River basin sits in east-central Iowa and drains into the upper branch of the Mississippi River (Figure 5-21). The basin has a total drainage area of 32,372.4 km^2, and an average annual recharge of 208 mm/yr with a range of 157 – 255 mm/yr over the analysis period. This set of nested basins was selected to test the hypotheses that high elevation basins should be net groundwater exporters (losing basins), while low elevation basins should be net groundwater exporters (gaining basins), and that the geology for a given climate controls the position of a hinge line.

River discharge from the Cedar River basin accounts for an average of 1.03% of the calculated annual recharge over the record period, with a range between 60% and 138%; the discharge to recharge ratio (Qr:R) values range from 0.60 to 1.38, and average 1.03, while only 3 basins fall below the Qr:R = 1 line. Evaluation of the Qr:R ratio against basin area suggests a linear trend of decreasing Qr:R with increasing area (r = -0.43) (Figure 5.-22a). However, the distribution in Figure 5-22a is suggestive of two populations, as those basins above the Qr:R = 1 line show low variability with basin area, but there are not a sufficient number of basins falling below the Qr:R = 1 line to determine this conclusively.

The Qr:R ratio over the Cedar River basin evaluated against basin average elevation suggests a linear relationship of increasing discharge per-unit recharge (Qr:R) with increasing elevation (Figure 5-22b). Thus, as reinforced by the map in Figure 5-21, the highest discharge per-unit recharge can be found in the small, high elevation basins of the Cedar River basin.

The overall decrease in Qr:R ratio with basin area is the opposite of the hypothesized relationship, where the Qr:R ratio increases with increasing basin area, and the largest basins converge about the Qr:R = 1 line (Figure 5-22a). Keeping their nested relationship in mind, Figure 5-21 reinforces that the largest basins are indeed the greatest groundwater exporters. However, Figure 5.2.6-2a also suggests that a significant population showing little change in the Qr:R ratio exists above the 1-line, distributed between Qr:R = 1 – 1.2 range and spanning linearly 2 orders of magnitude in drainage basin area. The relationship between the discharge to recharge relationship and basin elevation (Figure 5-22b) supports the hypothesis that where basin values straddle the Qr:R = 1 line, some trend with elevation should be apparent. Although, in this case and as in the case of the Potomac River basin, the relationship is the opposite of that expected. We expect to see a decrease in Qr:R with increasing elevation, such that the higher elevation basins are net exporters of groundwater, while the lower elevation basins are net importers.

To understand the relationships represented in figure 5-22, it is necessary to put the Cedar River basin into the context of the Ordovician, Silurian and Devonian aquifer systems of the northern Midwest. A closer look at the geology reveals that the Cedar River headwaters are in the highly permeable Ordovician-Devonian carbonate complex straddling the Minnesota-Iowa state line (Figure 5-23). The upper carbonate extends into Iowa, and pinches out just below the headwater basins of the Cedar River, and is underlain by a thin, but less permeable unit (Young and Siegel, 1992). The upper portion of the Cedar River and its constituents flow within the "U" shaped basin strata shown in line B – B' in Figure 5-23. Groundwater discharge occurs where the down-dip segment of the upper carbonate pinches out, spring-feeding these headwater streams. The streams then flow into the later Devonian strata, where the potentiometric contours shown in Figure 5-24 bend toward the river channels as they do in the headwaters. This Devonian stratum thickens to the south, and dips unidirectionally southwest midway through the basin (Figure 5-23).

The west branch of the river (the Iowa River) originates in the Devonian strata, and flows southeast to meet the Cedar River. Just below the A – A' line in Figure 5-23, the Iowa River encounters the more permeable Mississippian rocks which onlap the Devonian strata. The beds beneath the basin have transitioned from dipping towards the basin center from either side (as in line B – B', Figure 5-23), to an overall southwest dip, towards the Missouri River (A – A', Figure 5-23). The highly permeable and thickening strata of the Mississippian units provides a conduit for surface water in the southern half of the basin, thus, the low elevation losing basins may be a result of this flow.

In this case, the hinge line in the regional flow system can be roughly identified somewhere in northwestern Iowa, near the Dakotas, which appear to be the ultimate source of the regional groundwater flow (Mandle and Kontis, 1992) (Figure 5-24). The lack of the expected decreasing correlation between the Qr:R ratio and elevation within the Cedar River basin supports the hypothesis that the underlying geology is the most influential factor on the development of a hinge line. Further, topography influences the development of local flow systems, which supply the headwaters of the basin, while the lower elevations of the basin as a whole feed into the regional flow system towards the Missouri and Mississippi Rivers.

The regional flow system in place here suggests that water flowing into the Mississippi River and the Missouri River may have originated in South Dakota. It can also be said that a hinge line must exist between the Dakotas (largely losing basins) and the upper reaches of the Upper Mississippi River basin, in Minnesota, which is composed almost entirely of gaining basins. However, the nature of the groundwater potentiometric contours vs. the direction of surface runoff suggests that the surface water and groundwater basins are not synonymous (Figure 5-24); while topography controls surface runoff, it's the underlying geology that dictates the groundwater flow regime.

5.2.7. Basin 13342500 – Clearwater River, Idaho

The Clearwater River basin composes the lower northwestern portion of Idaho, sandwiched between Montana to the east, and Washington and Oregon to the west (Figure 5-25). The total drainage area of the basin is 24,786.3 km^2, and has an average annual recharge between 460 and 5225 mm/yr, with a mean of 475 mm/yr over the analysis period. The Clearwater River basin was initially selected because of its strong internal nesting and intermediate recharge values over mountainous relief, suggesting a relatively sub-humid climate. The basin may be used to test the hypothesis that high elevation basins tend to be groundwater net exporters, while low elevation basins net importers, and that in more humid climates gaining basins tend to be more pronounced.

River discharge from the Clearwater River basin accounts for an annual average of 138% of the basin averaged recharge, with a range between 120% and 154%; Qr:R ratios within the basin range between 1.20 and 1.54, with a mean of 1.38. When compared against basin area (Figure 5-26), the Qr:R ratio shows a strong increasing linear correlation with increasing basin area ($r = 0.50$). Evaluation of the Qr:R ratio in terms of elevation (Figure 5-26) reveals a weaker linear trend of decreasing Qr:R with increasing elevation ($r = -0.31$), suggesting that the lowest discharge per-unit recharge occurs in the high elevation basins.

Although the Qr:R ratio Clearwater River basin does not show a relationship with increasing area that approaches the Qr:R = 1 line (Figure 5-26), the positive correlation suggests that flow accumulation is occurring within the basin. The dashed lines in Figure 5-26 indicate basins that flow into one-another. This increasing Qr:R ratio with increasing area suggests that the basin and its internal constituents are continually receiving groundwater input to stream discharge from up-basin. However, since there are no losing basins within the Clearwater River basin, the groundwater to sustain baseflow to rivers must be sourced elsewhere.

The decreasing trend in river discharge per-unit recharge with elevation suggests that lower elevation streams import more groundwater than do those at higher elevation within the basin. This is in agreement with the hypothesis that, below the hinge line, we expect low elevation basins to have Qr:R values greater than 1. Also, the Qr:R values over all the basins fall above the Qr:R = 1 line, suggesting the Clearwater River basin is an overall groundwater net-importer.

The Clearwater River basin sits in The Northern Rocky Mountains Intermountain aquifer system, where the regional topography is generally high to the east, sloping to the west in Figure 5-25. Since the regional relief is generally high in the higher elevations, groundwater flow occurs through bedrock fractures and along fault scarps into unconsolidated valley alluvium (USGS Groundwater Atlas) (5-27). Springs issuing from fault planes and networks of joints are also important components of surface water flow in the Clearwater River basin (USGS Groundwater Atlas). Likewise, regional groundwater flow occurs along fault plains and through the heavily fractured Miocene to Pliocene basalts.

Because the Clearwater River basin is an overall net importer, the hypothesis of a hinge-line implies that groundwater is probably being exported from an adjacent basin. Although the Clark Fork basin of the Pend Oreille River system (western Montana) has only three internal basins identified in the HCDN data set, it is an overall losing basin of substantial area (27,736 km^2), and has an average elevation of 1,664 m, higher than the Clearwater River basin (1100 m) (Figure 5-25). Thus, these lines of evidence support the hinge-line hypothesis, which we can identify as existing somewhere between the upper Clearwater River basin and the Clark Fork basin of the Pend Oreille River. Again, however, in this situation, it does not appear that the surface water and groundwater basins are congruent – surface flow is roughly southeast to northwest, while regional groundwater flow appears to be roughly northeast to southwest (Figure 5-25)

5.2.8. Basins of the California Central Valley Uplands

The basins of the California Central Valley Uplands are a collection of 45 basins along the highland flanks of California's agricultural district (Figure 5-28). The average recharge over the region is 560 mm/yr, but ranges from 50 to 1160 mm/yr, with higher recharge to the north and less to the south. The mean basin drainage area used in this study was 890 km^2, while basin sizes range from 26 km^2 to 6734 km^2. The basins within the uplands of the California Central Valley were selected for further analysis because of the presence of a well-established and well characterized regional flow system (Williamson, et al., 1989, and others, see discussion above), and to test the hypothesis of a regional hinge line in this context.

Calculation of the Qr:R ratio over the basin reveals that river discharge accounts for between 9% and 158% of the total recharge over the basins, with a mean of 90%; mean Qr:R values of 0.90 and range of 0.09 to 1.58. When evaluated against basin area (Figure 5-29), the Qr:R shows a decreasing trend with increasing area ($r = -0.18$). A comparison with elevation reveals that Qr:R linearly increases with increasing elevation ($r = 0.48$) (figure 5-29b), implying that high elevation basins have higher river discharge per-unit recharge than their lower elevation counterparts.

The poor correlation between the discharge to recharge ratio and drainage basin area is not anomalous considering that few of the basins examined are in a nested relationship with each other. However, the positive correlation between drainage basin elevation and Qr:R does not support the hypothesis that higher elevation basins should be net-exporters of groundwater. The data suggest, in fact, that the high elevation basins in the Sierra Nevada Mountains are importing groundwater, which is becoming a component of the stream discharge in those basins. This is indeed quite anomalous since basins such as these – situated at ridge-tops – have no source from which to import groundwater.

The selected basins sit in the Sierra range, mostly between 1500 and 3000 m elevation, and are generally the highest peaks in the region. These intermountain valleys have very little soil, but are composed of highly fractured basalts and crystalline basement. Winter snow packs in these high basins are crucial in supplying water to the valley floor during the spring melt, particularly through regional groundwater flow pathways originating in the mountain valleys (Williamson, et al., 1989). Groundwater enters the permeable marine and alluvial sediments of the lower Central Valley walls and eventually to the base of the valley, where it supplies river discharge (Figure 5-30). This baseflow discharge sustenance causes spring recharge to be high, but not as high as if spring melt water contributed directly to stream discharge. Thus, spring discharge is buffered from high amplitude events by the groundwater system, while summer dry season discharge is augmented by baseflow from the underlying aquifer (Bertoldi, et al., 1991).

The previous description applies to the ideal, predevelopment system of California's Central Valley. However, substantial pumping for agriculture and municipal use occurs on the valley floor, and since groundwater baseflow discharge is the primary source of streamflow, observations of the latter will reflect some component of pumping. This implies we may need to return to the VIC model results for potential insights on why the high elevation basins may have low recharge values.

A primary reason for using the VIC simulation output (Mauer, et al., 2002) was its high spatial resolution and good match to stream discharge. However, a look at the modeled vs. observed discharge for the Sacramento river (draining California's Central

Valley) shows an overestimation of spring discharge, and underestimation of summer discharge. Overall, the two balance each other so that the net-effect is minimal, however, the VIC simulated river discharge should be the sum of the observed discharge and the pumping discharge (since recharge in VIC is equal to stream discharge). The implication is that the VIC simulation calculated the correct annual mean stream discharge for the Sacramento River, but the value calculated includes pumping from the valley floor. Thus, the discharge values for VIC streamflow should be *higher* than the observed values because they included pumping; the fact that they are roughly equal implies that the simulated river discharge by VIC is too low.

Since stream discharge is equal to recharge in the VIC model, it follows that recharge is duly underestimated as a result. In the calculations conducted here, we've used the ratio of observed stream discharge to VIC calculated recharge. With a calculated recharge that underestimates actual recharge, the ratio of Qr:R is larger than it should be. A further implication of this is that recharge is likely underestimated wherever groundwater pumping is significant and the VIC model matched observed streamflow very well.

Further, the VIC simulation, although in general the best available product for evapotranspiration, has a large grid size (0.125º) relative to the size of many of the California upland basins. The reader will recall that these grid data were split into 0.0625º cells before the recharge data for each basin were extracted, but this does not sharpen the fidelity of the original data. Thus, anything below the grid cell resolution of the VIC model output (less than ~144 km^2) must be taken with caution.

Indeed, 6 of the 31 basins found in Figures 5-29 a and b have areas less than ~144 km^2, all falling above Qr:R = 1 line, implying that these basins may not be accurately simulated. This estimate may even be conservative because model grid cells are equidistant squares of ~12 km by ~12 km, where a ~144 km^2 basin is most probably not a perfect square. The implication is that the VIC simulation may only be useful in this context for basins in which a 12 km by 12 km square will fit.

One last point of interest is the coastal Eel River basin, in NW California, where four basins sit in a nested relationship (Figure 5-30). Visual inspection of the basin, with spatially denoted Qr:R ratio reveals that the high elevation basins are groundwater exporters, having Qr:R ratios between 0.6 and 0.9, while the lower elevation basins are groundwater importers (Qr:R = 1.4-1.1). According to the hinge-line hypothesis, a regional hinge exists somewhere below the high elevation losing basins, such that the lower elevation basins are beneficiaries of exported groundwater.

5.2.9 Summary of the Case Studies

In light of these case studies, a few salient features of the groundwater-surface water partition become apparent. Firstly, that the geology plays an integral role in the partition between groundwater and surface water in a given basin, more so than climate. In every case outlined above, the disagreement between the hypothesized relationships and those observed may be ascribed to heterogeneity in the underlying geologic context of each basin. Groundwater-surface water interactions observed in the Texas basins, the Potomac River basin, and the Clearwater River basin are partially, or dominantly controlled by structural features of the geology. Heterogeneity in the unit thickness and dip of the aquifer units, as well as differential permeability control the groundwater flow regime in all cases. Thus, the substrate conditions exercise the primary control over the development, direction and magnitude of regional groundwater flow; and hence, the distribution of groundwater recharge and discharge zones and the existence of a hinge line.

Secondly, it can be said that a river basin, delineated by surfacial drainage, is in most cases not congruent with that of groundwater flow basins. This is evidenced particularly well by the Cedar River and the Clearwater River, although each of the other cases contains some element of incongruity between the surface water and groundwater basins. The implication of this observation is that surface flow direction may be controlled by the topography and relief of a basin (in a given climate), but these controls have less influence over the groundwater flow regime of that basin. Further, the systematic rejection of the elevation-control hypothesis in favor of the null suggests that elevation may only be influential if the surface water system is congruent with the subsurface system.

Lastly, the climate regime is of overall secondary importance to that of substrate geology in terms of the development of a hinge line or regional flow system. The climate merely affects the variability in flux between the surface water and groundwater reservoirs, but does not alone dictate the development or location of a regional hinge line. This is apparent in the Texas basins, where aridity exacerbates the expression of losing basins at the surface such that the importance of groundwater flow is enhanced relative to surface water discharge. Moreover, the climate affects the distribution of groundwater recharge and discharge areas only by providing surfacial input over areas of appropriate geologic conditions. Thus, the geologic characteristics of a basin effectively create the framework for groundwater flow systems; the nature of the climate regime merely provides the amount and variability (in space and time) of water available to flow.

6. Summary and Conclusions

The separation of groundwater from surface water proposed by this analysis was motivated by the following questions:

5. How much water leaves a given drainage basin without ever passing through the surface outlet of that basin?
6. What is the relative importance of this water in terms of the surface/subsurface reservoirs and the continental water balance?
7. What is the relationship between surface water flow and areas of recharge and discharge to the water table? Are sources and sinks of water to surface basins identifiable in the subsurface flow system?
8. What is the ultimate control over the distribution of gaining and losing basins in terms of their surface water expressions?

With these questions in mind, the hypotheses proposed earlier in this analysis were based on the theoretical relationships outlined by Toth (1963). It was understood that Toth's analysis failed to account for variations in climate and the effects of geologic heterogeneity on the development of groundwater flow regimes. The tenets of Toth's work were an excellent reference point from which to evaluate reality, but the systematic deviations from these postulates were more significant than expected.

The fundamental hypotheses of this study – that distributions in basin area and elevation effect the existence and position of a regional groundwater flow system, and the development of a hinge line delineating groundwater importing and exporting basins – are appropriately derived when considering an aquifer of homogeneous substrate receiving a uniform distribution of surface recharge. Consequently, they do not represent the variability of reality, and are only useful as a fundamental basis of the groundwater-surface water partition and must be expanded upon.

The findings of this study suggest several conclusions:

1. Geology is the dominant, first-order control over the distribution of recharge and discharge (as gaining and losing basins) to the groundwater reservoir, the groundwater flow regime, and the existence and position of a regional hinge line. Fold and thrust, faulting, and dipping and thickening beds are particularly influential, as is the anisotropy in vertical vs. horizontal conductivities and its effects on differential permeability. As such, the postulates of Toth (1963) and Hubbert (1940) – that the groundwater table is a subdued mimic of the topography, and groundwater flow systems a primary function of said topography – are only partially applicable.
2. The delineation and organization of surface water drainage basins is not congruent with the organization of groundwater flow basins, particularly where regional groundwater flow is dominant. As in the case of the Cedar River basin, Iowa, and the Clearwater River basin, Idaho, the direction of groundwater flow is different than surface water flow. The effect is an arrangement of gaining and losing basins within the larger basin that is counterintuitive without consideration of the underlying context. In both cases, regional groundwater flow augments surface water discharge from high elevation basins through either spring discharge or baseflow. In the Clearwater River case, the basin is gaining as a whole (and increasingly gaining downstream); in the Cedar River case, a lithologic transition causes downstream loss of surface water to the groundwater system.

Thus, the distribution of gaining and losing basins is not necessarily dictated by their delineation in terms of surface drainage; as such, the gaining or losing nature of drainage basins is only a surface indicator of deeper hydrologic processes. Groundwater flow basins are defined by their geologic context, and appear to be completely separate entities from surface water basins, in most cases.

3. A hinge line is not apparent if a river network straddles different groundwater basins, or only part of a single groundwater basin. This is no better evidenced than in the Colorado River case, where the Colorado River basin represents a smaller subset of a larger regional flow system, and appears on the surface to be comprised of increasingly losing basins. This is a zone of primary recharge to the groundwater system, and hence, is only a small portion of a larger groundwater basin whose outlet is likely much closer to the coast. Thus, a hinge line in terms of the groundwater system may be identified with little regard for the surface flow system.
4. The groundwater flow system, Qg in this analysis, can be said to be a significant and un-ignorable participant in the terrestrial water cycle. As evidenced by the case studies above, the redistribution of continental waters through subsurface pathways is quite pronounced. This observation has implications for the future of climate modeling – if climate models hope to predict the nature or availability of future water resources, the current methodology is not adequate. To close the continental water budget, it seems apparent that incorporating the position and movement of the groundwater reservoir, and subsurface heterogeneity are necessary.

6.1. Recommendations for Future Work

Ideally, the overall results of the groundwater-surface water partition should be compared to subsurface characteristics throughout the continent. We recognize the complexity of this suggestion, but in light of the emergence of geology as the primary control over the partition, the results cannot be explained fully without such an analysis.

In addition, the precipitation and evaporation/evapotranspiration (and hence river discharge) regimes across the continental U.S. are known to have substantial seasonal variability. A better approach to the separation of groundwater flow from its surface counterpart would be a seasonal evaluation of the stream discharge and basin recharge values across the 39 years of available record. With monthly mean Qr values separated from monthly mean recharge figures, the dynamic characteristics of the partition may become more apparent, particularly their spatial distribution and fluctuation through time.

Lastly, seasonal separation between the continental water reservoirs and flow, combined with a more uniform understanding of subsurface characteristics, suggests that delineation of groundwater basins could be performed. Surface water drainage should only partially influence such a delineation – in that the surface water system becomes a proxy for areas of recharge and discharge to the groundwater system. This follows the conclusion that surface water drainage basins and groundwater flow basins are not apparently congruent. Such a spatial demarcation, with notation of flow direction and flux, may be useful to the climate modeling community for the development of land surface parameterization schemes that incorporate groundwater reservoir dynamics. This may lead to better prediction of future hydrologic conditions.

7. References

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8. Figures

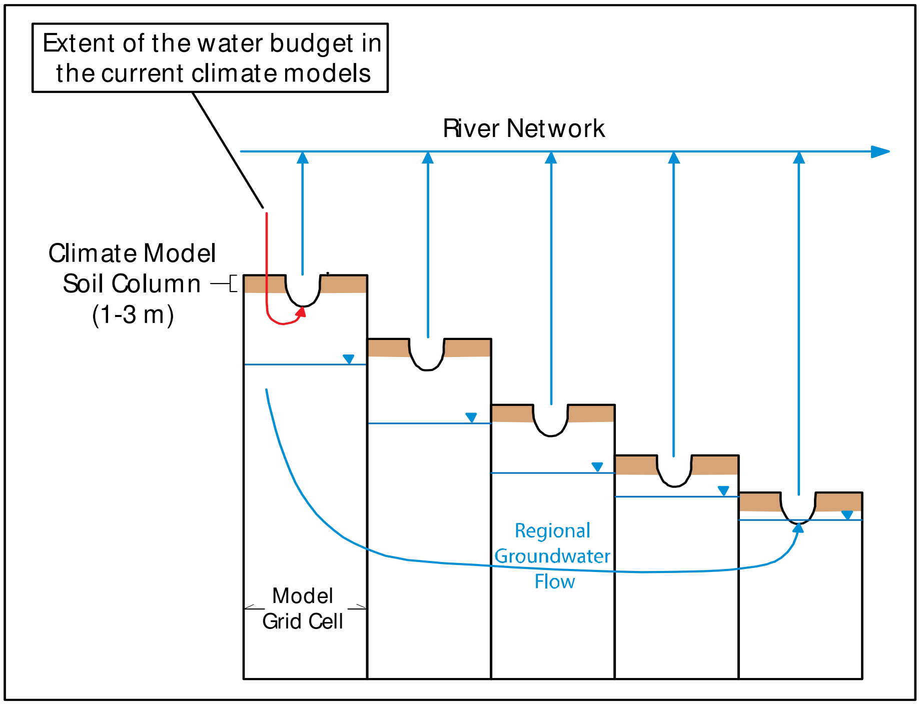

Figure 2-1: A conceptualization of the water budget in a typical General Circulation Climate Model (flow path delineated in red). The current climate models route surface recharge over a cell directly into stream channels within that cell, and neglect the effects of regional flow on importing or exporting water from once cell to an adjacent cell.

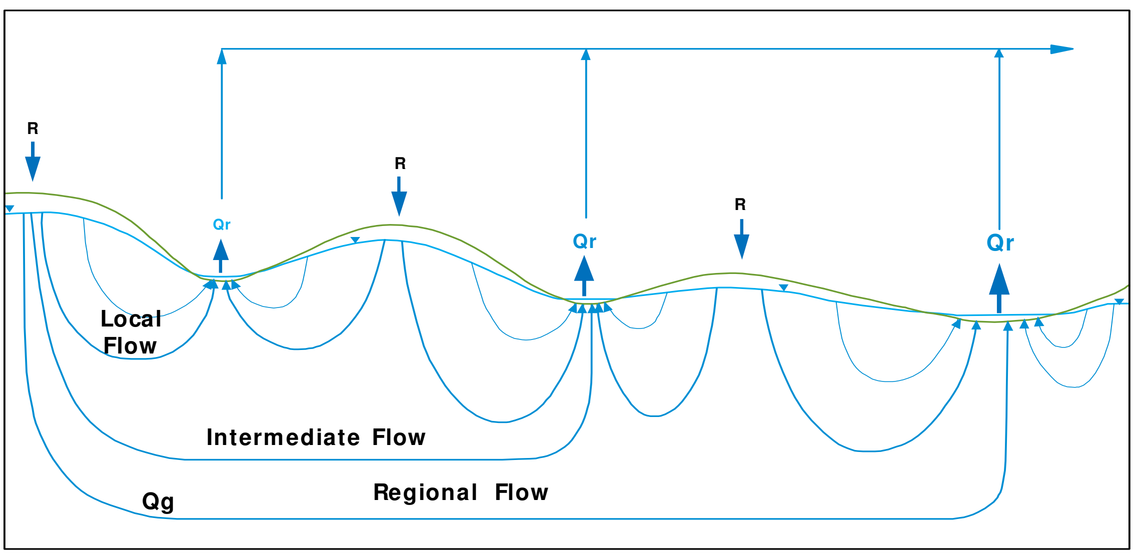

Figure 2-2: Schematic representation of Toth's groundwater flow regimes, local, intermediate and regional, and their relationship with each other. Modified after Toth (1963).

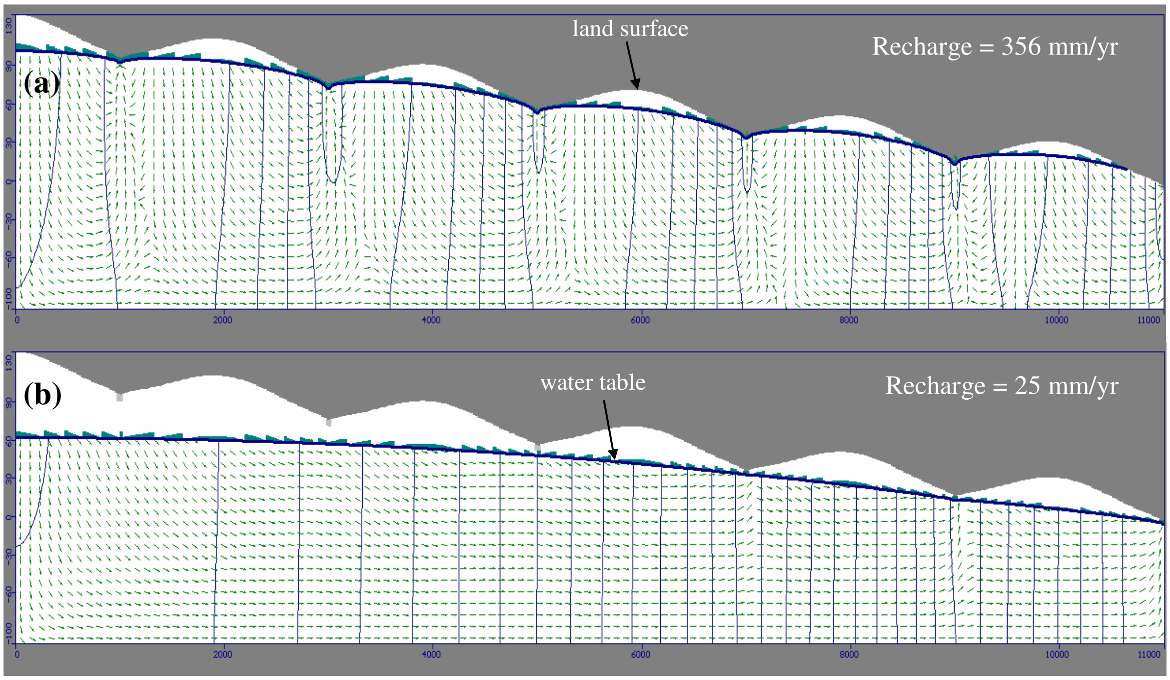

Figure 2-3: A 2-D conceptual model of Toth's (1963) groundwater flow at local (a) and regional (b) scales showing the potential effects of climate. In a humid climate (a) the water table reflects the local topography, local drainage is active with gaining streams, and the regional flow system is diminished. In an arid climate (b) the water table falls below the local relief, streams are primarily losing, and groundwater discharges at regional lows.

Figure 2-4: Map of the High Plains Aquifer with potentiometric contours (blue) and general direction of groundwater flow denoted (blue arrows). Modified from USGS Groundwater Atlas (http://capp.water.usgs.gov/gwa/gwa.html).

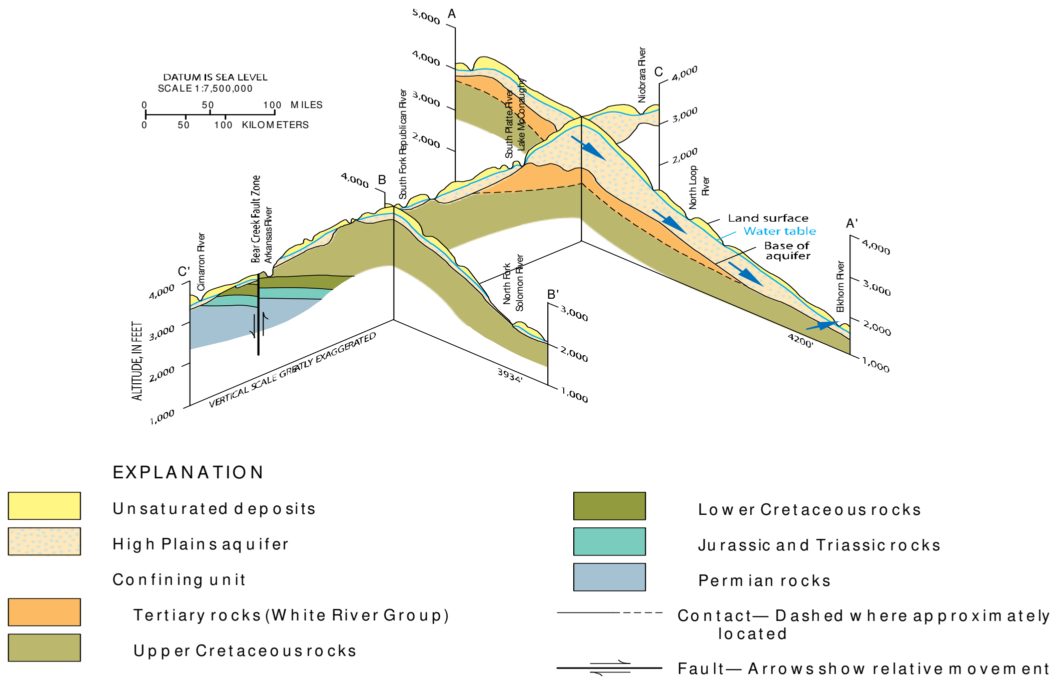

Figure 2-5: A diagrammatic cross section of the High Plains Aquifer system showing the geologic characteristics of the subsurface and direction of groundwater flow. Taken from the USGS Groundwater Atlas (http://capp.water.usgs.gov/gwa/gwa.html).

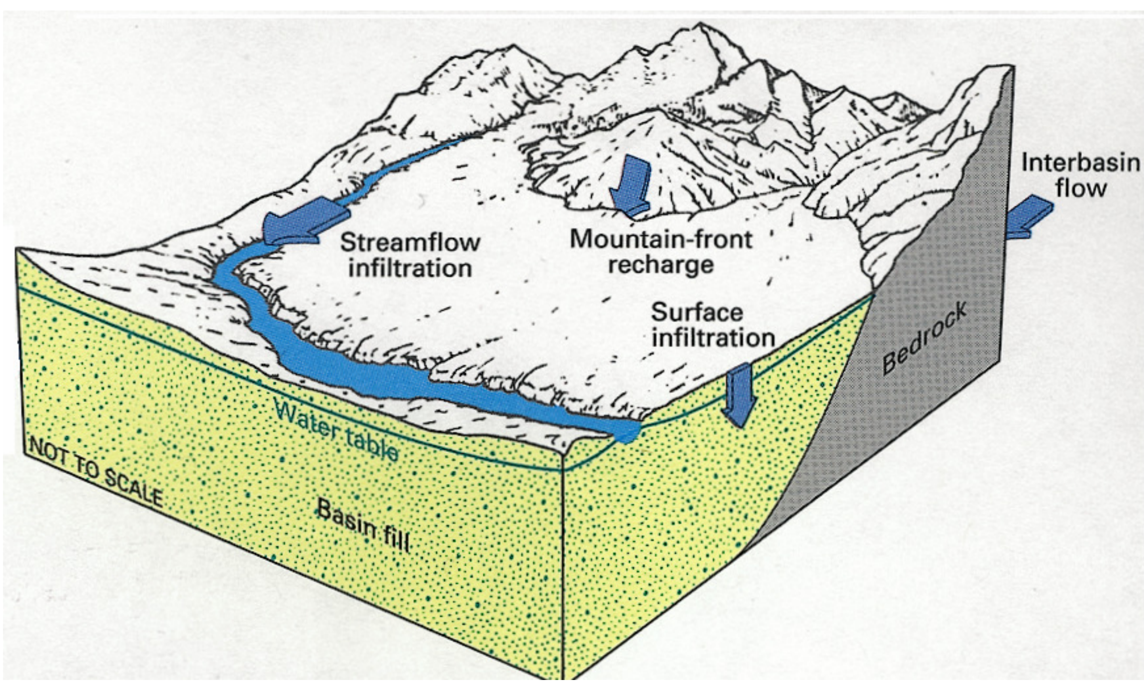

Figure 2-6: A diagrammatic cross section depicting the regional flow system within an intermountain valley, as in the southwestern US. Taken from the USGS Groundwater Atlas (http://capp.water.usgs.gov/gwa/gwa.html).

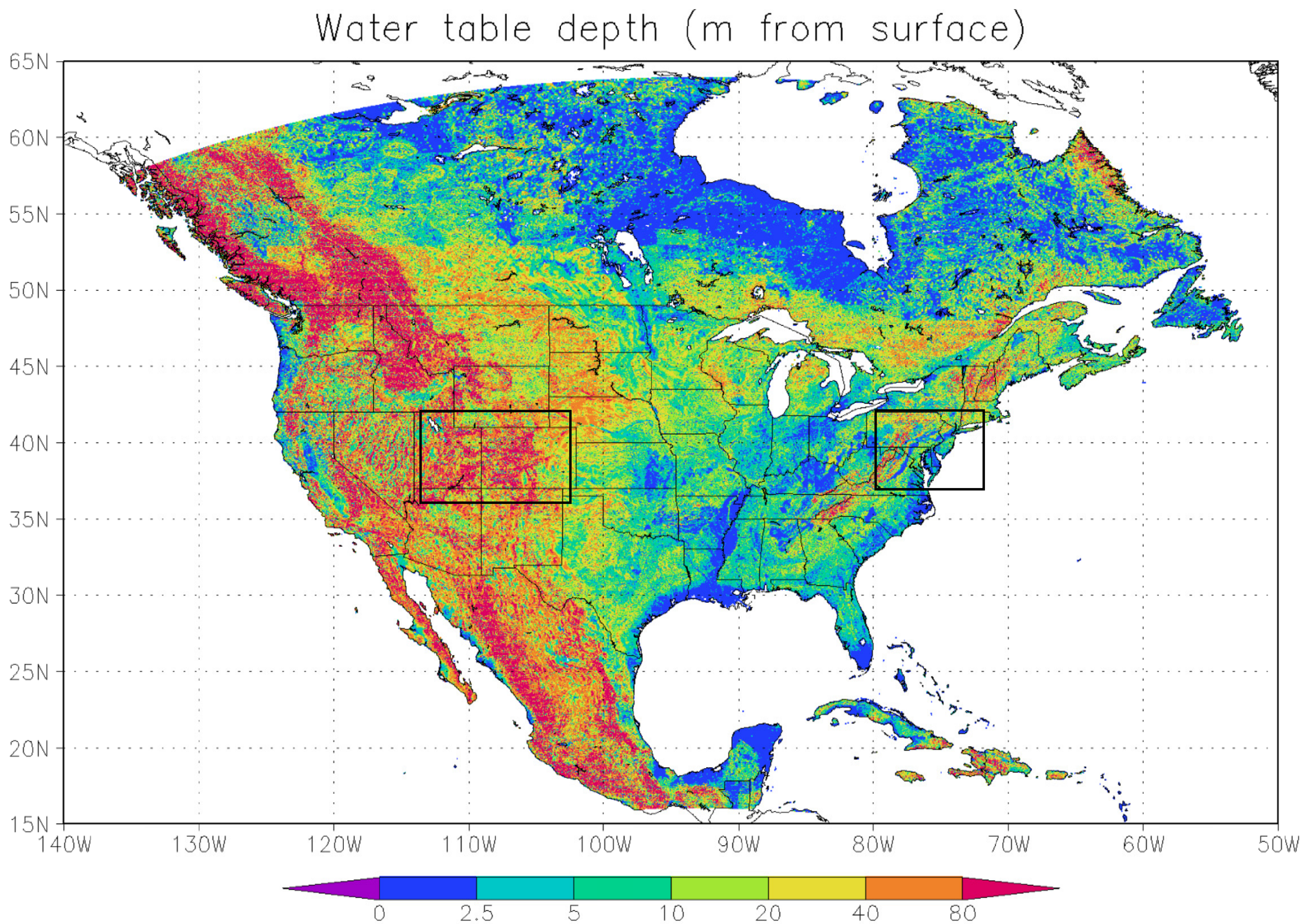

Figure 2-7: Simulated depth to the water table over the contiguous US, modified after Fan, et al., (2007).

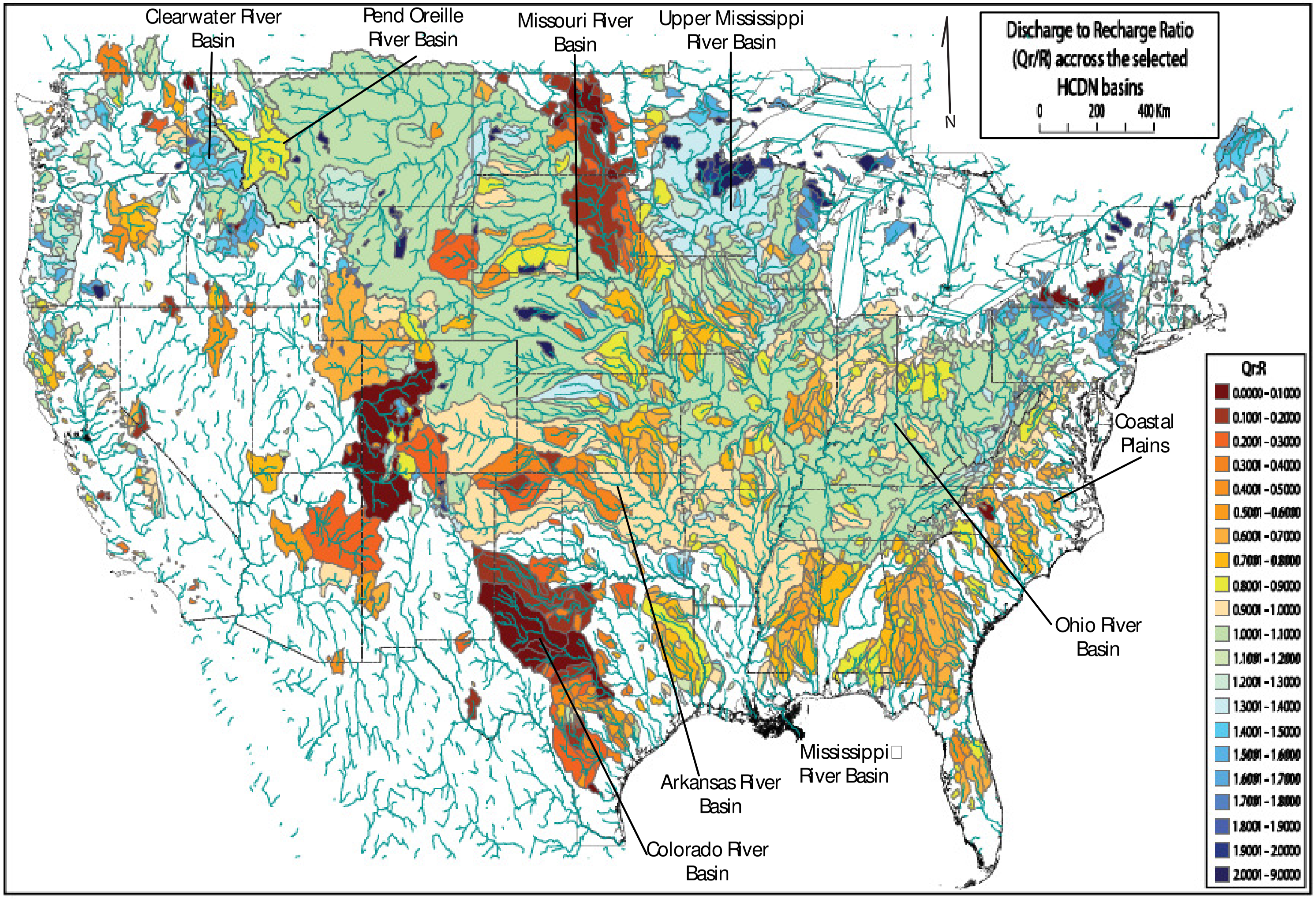

Figure 5 -1: Map depicting the spatial distribution of Qr:R ratio results over the contiguous United States. Basin colors represent the Qr:R ratio, denoted in the color scale to the right. See text for discussion of labeled basins.

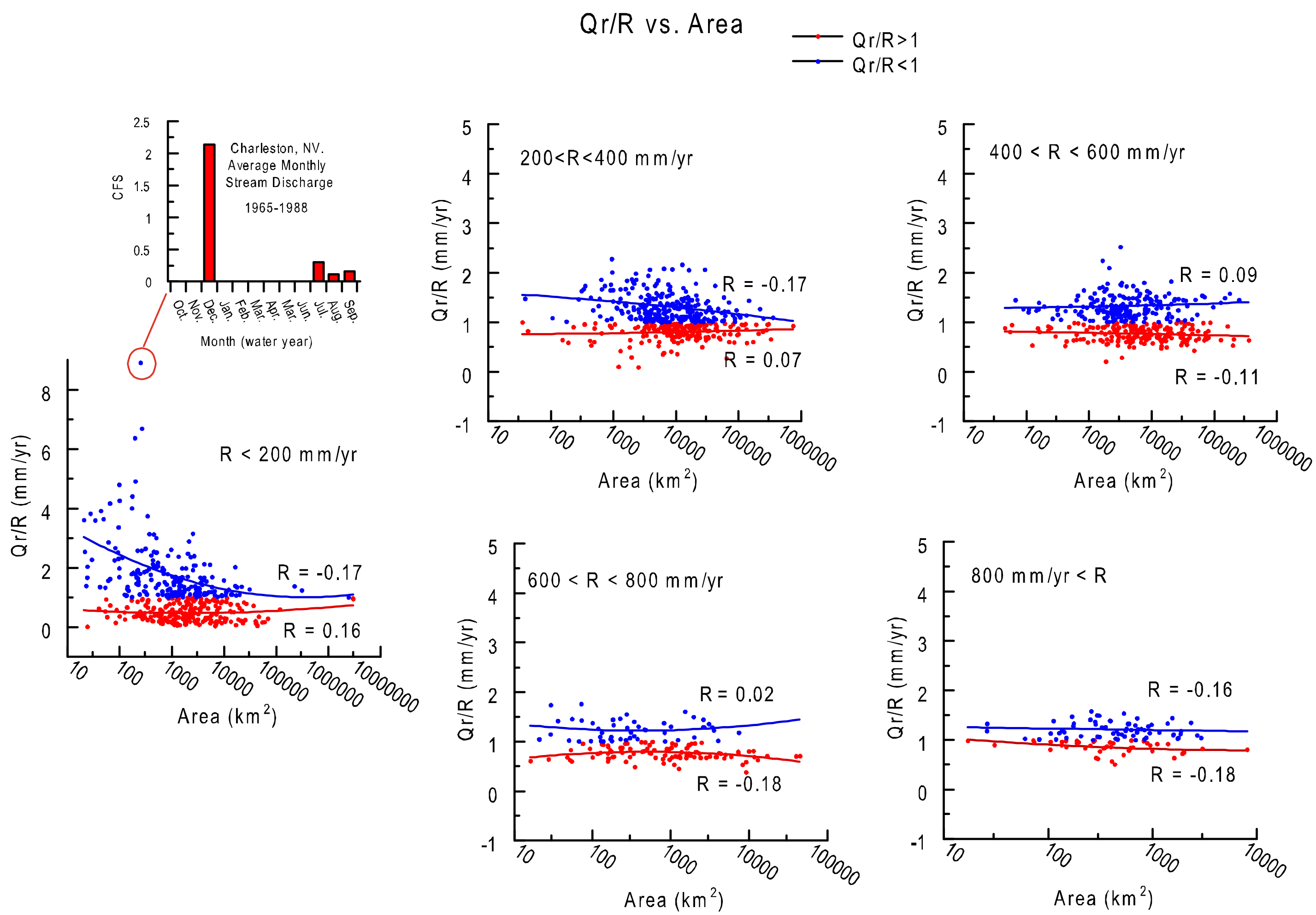

Figure 5-2: Results of comparison between the Qr:R ratio vs. basin drainage area, binned by recharge into 200 mm/yr intervals. Inset plot above the Recharge <200 mm/yr correlation plot corresponds to the highest Qr:R value , the Charleston, NV basin sitting in the foothills of the Sierra Nevada and is receiving substantial groundwater discharge.

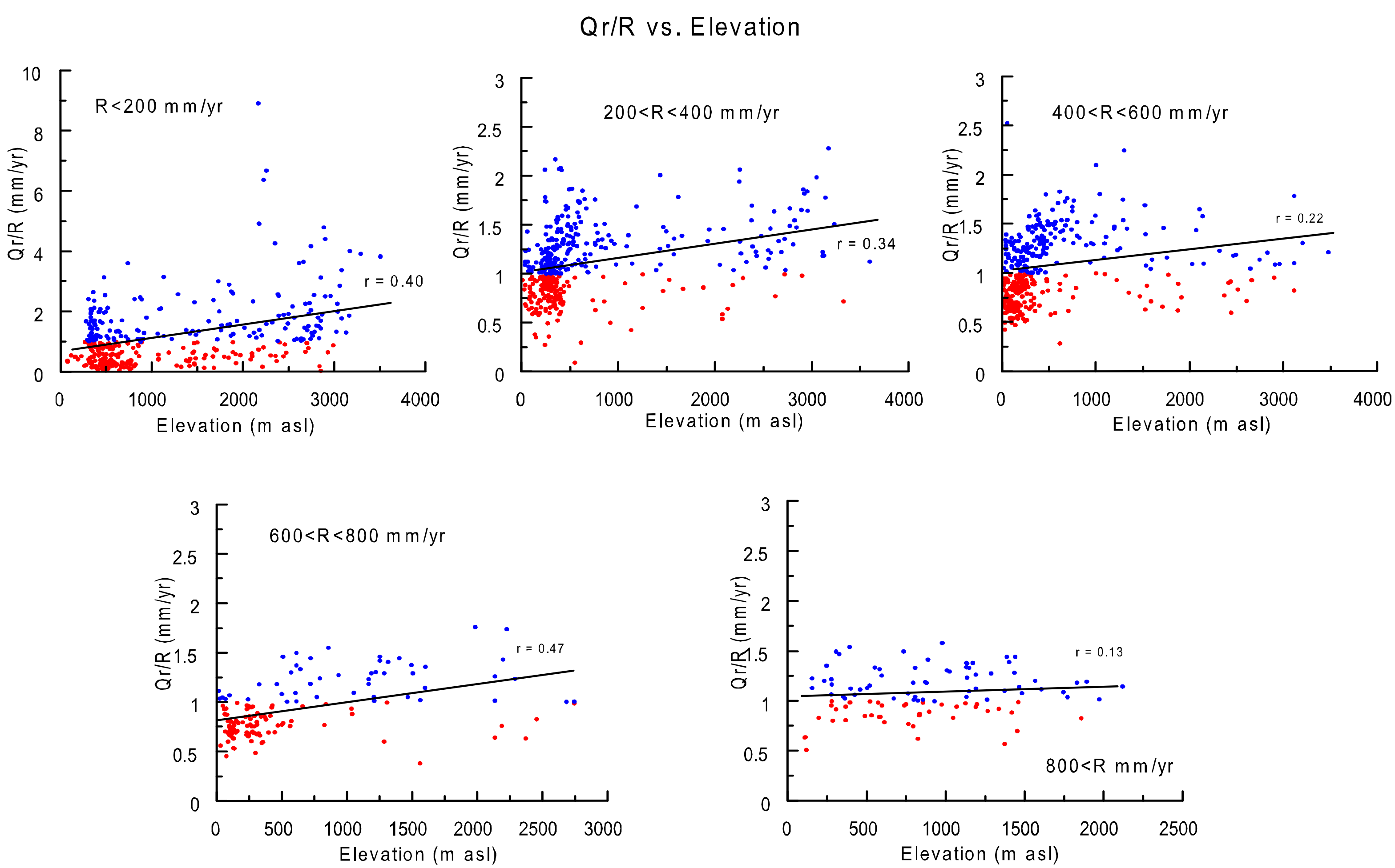

Figure 5-3: Results of comparison between the Qr:R ratio vs. basin mean elevation, binned by recharge into 200 mm/yr intervals.

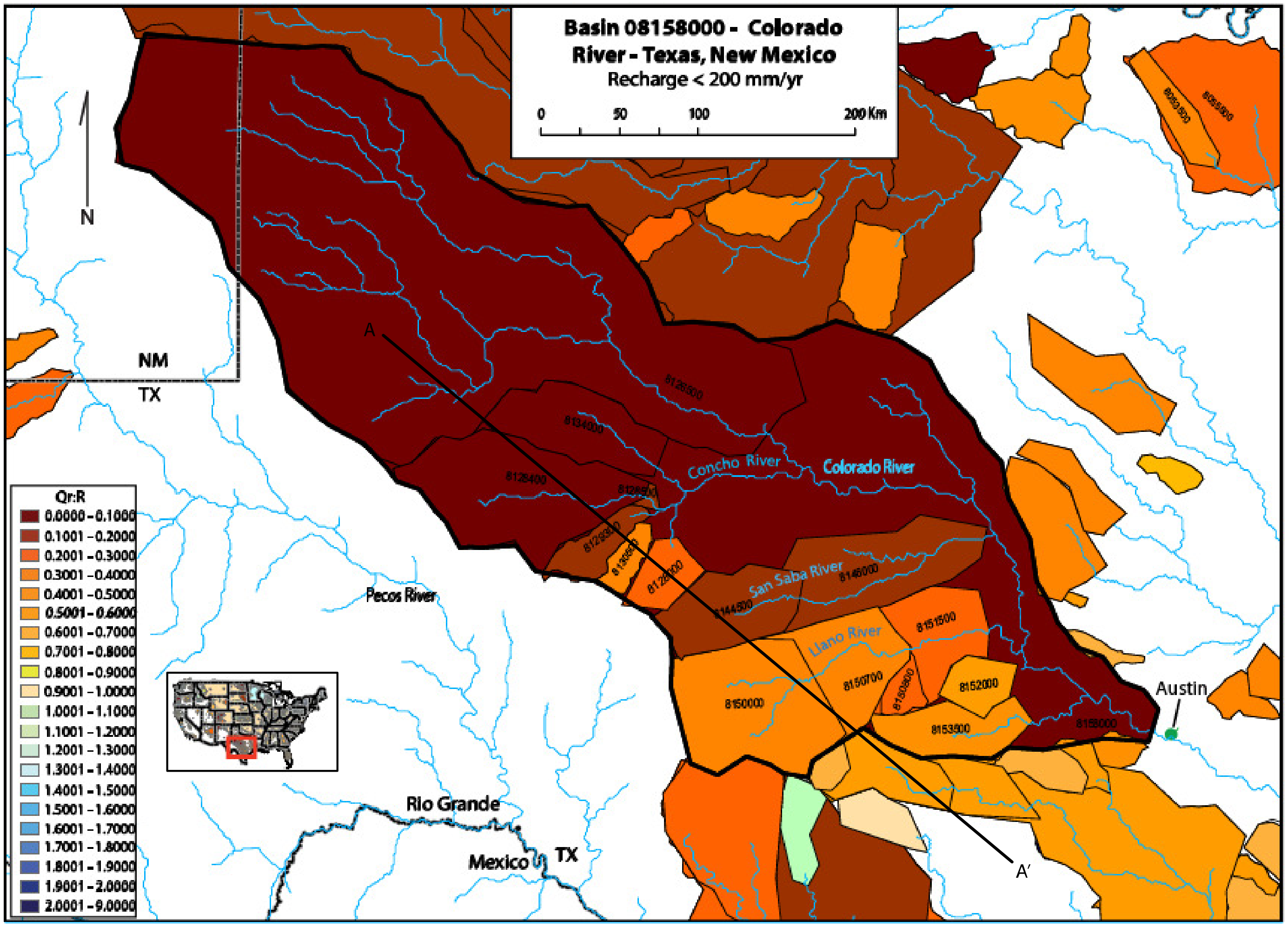

Figure 5-4: Map depicting the extent of the Colorado River basin, Texas, and its internal components. The basin fill colors represent the ratio of surface water discharge to recharge (Qr:R), which are indexed by value in the color-ramp key at the bottom left.

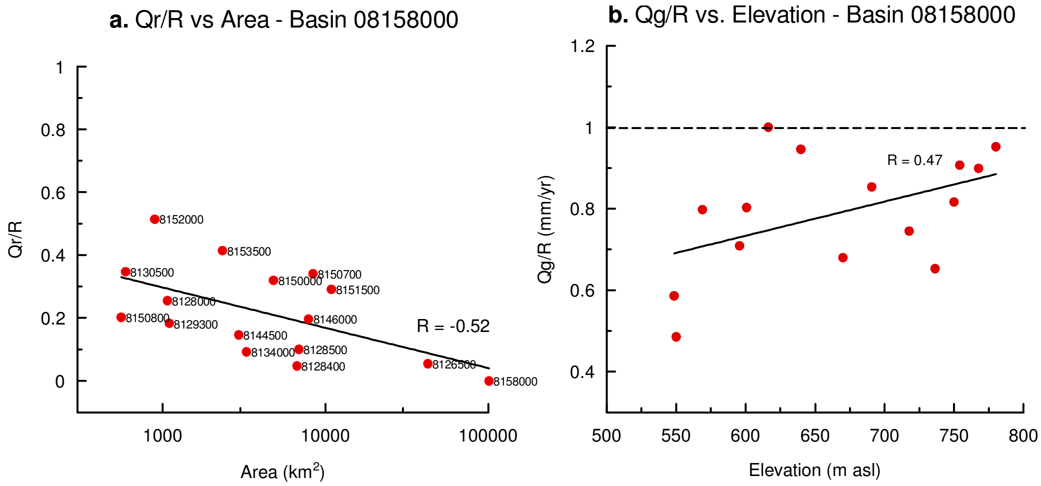

Figure 5-5: (a) Correlation between the Qr:R ratio and basin area; Correlation between the groundwater discharge to recharge ratio (Qg:R) (refer to equations 1, 2, and 3 for discussion) and (b) basin mean elevation.

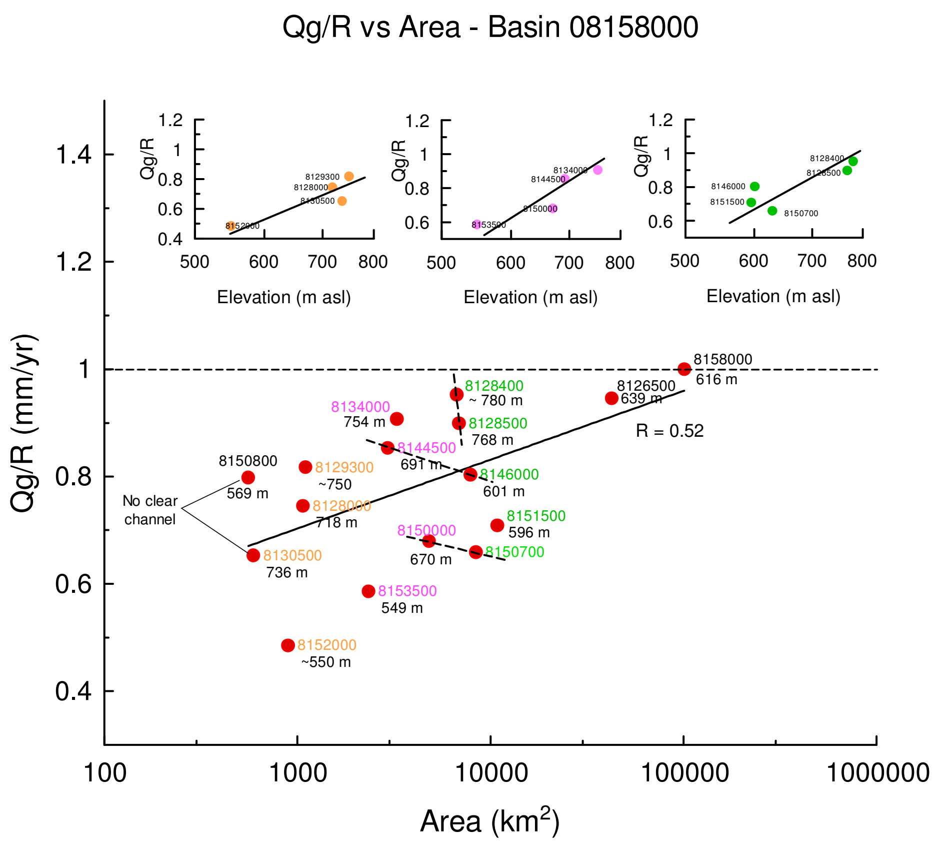

Figure 5-6: Correlation plot between the groundwater discharge to recharge ratio (Qg:R) and basin drainage area. The label colors correspond to inferred stream order, where the orange are first-order ephemeral streams, pink are first-order channels, and the green are second-order channels. The labels correspond to USGS station ID, which are denoted in Figure 5-4. The inset plots show the relationship between the Qg:R ratio and elevation for each of the stream orders delineated, Figure 5-5b shows a correlation between Qg:R over the whole basin.

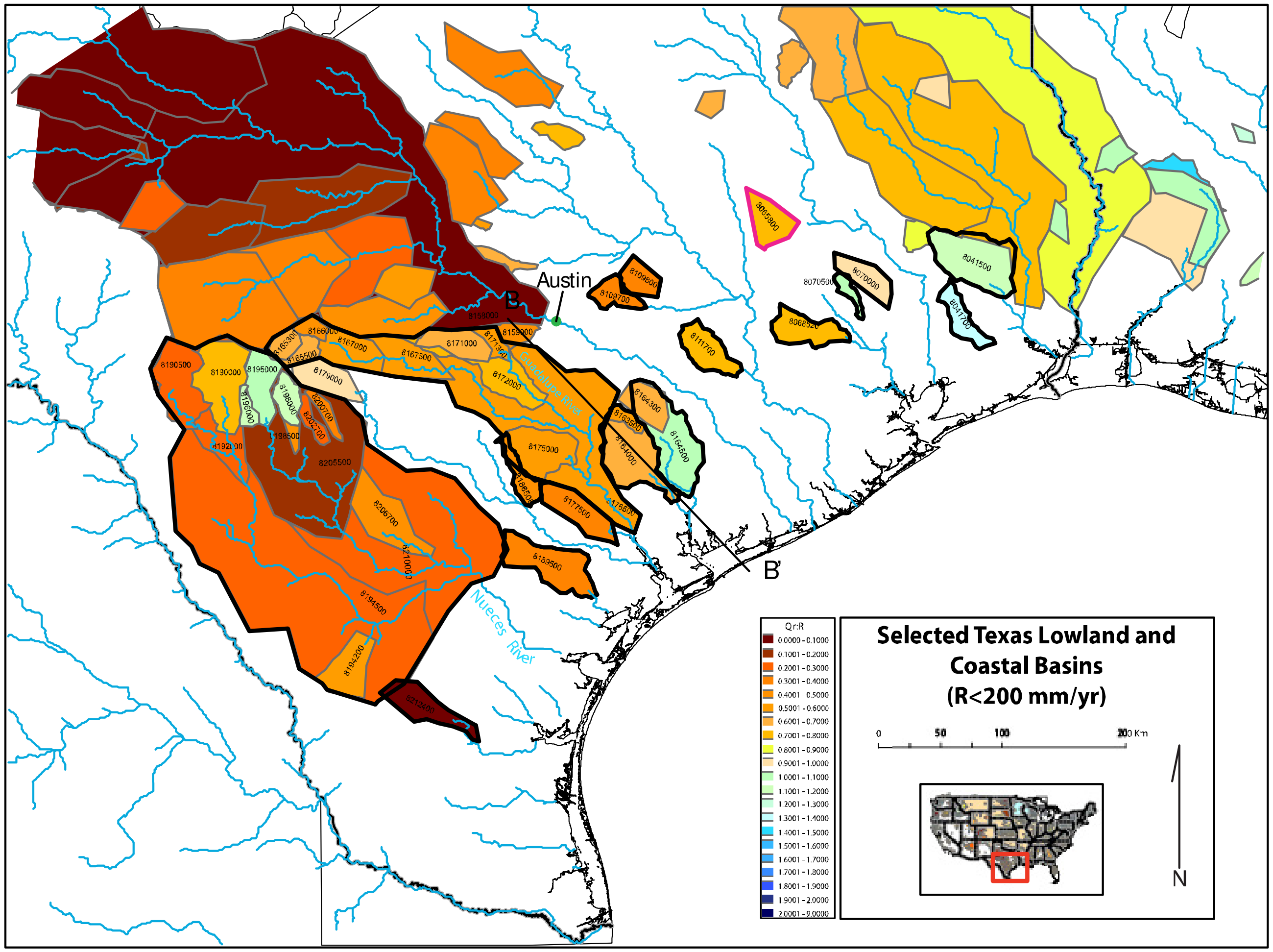

Figure 5-7: Map depicting the Nueces and Guadalupe River basins in southern Texas, southeast of the Colorado River basin. Basin colors correspond to the Qr:R ratio of each basin, with values according to the included color ramp. A subsurface cross section of the line from B – B'

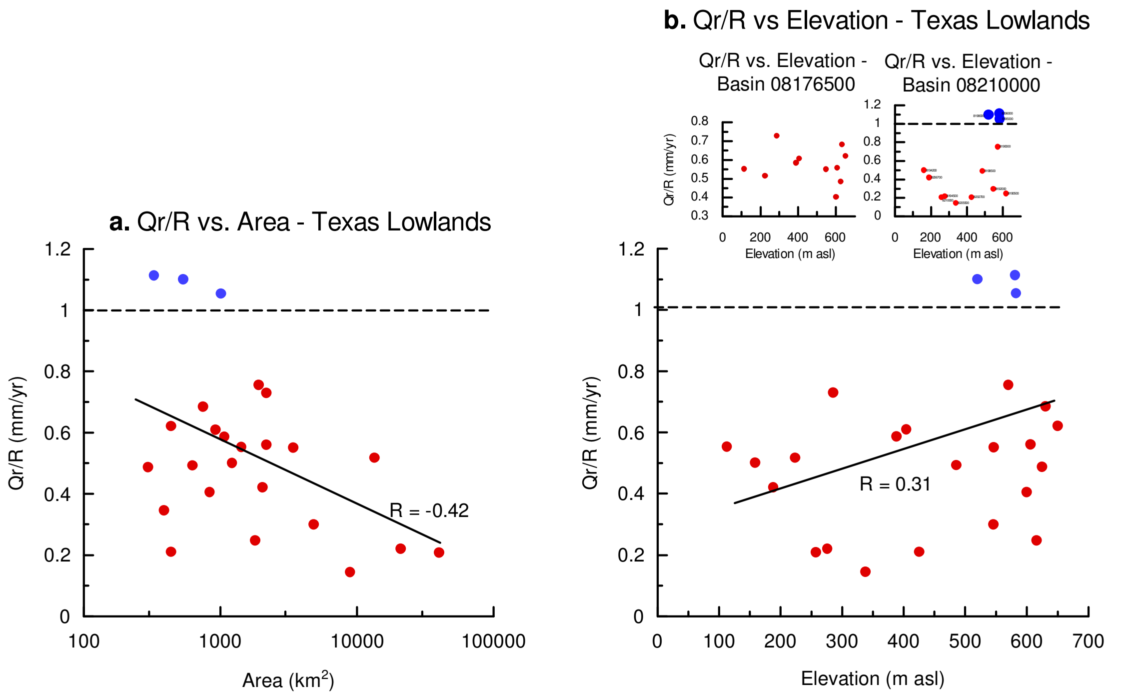

Figure 5-8: (a) Correlation plot between the Qr:R ratio and basin area, and (b) between mean basin elevation for the Guadalupe and Nueces River basins in the Texas Lowlands. The inset plots in (b) display the Qr:R ratio vs. basin elevation for each river basin separately, suggesting the fit is similar regardless of whether or not the two basins are grouped.

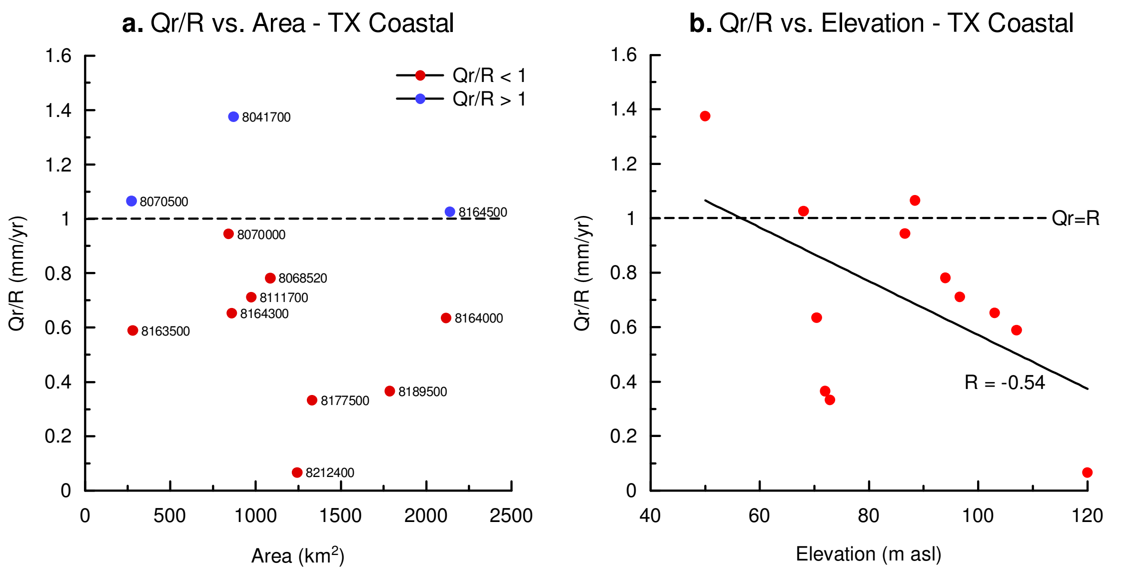

Figure 5-9: (a) Correlation plots between the Qr:R ratio and basin area, and (b) basin mean elevation for the selected Texas coastal basins. No strong correlation with basin area is expected in this case because the basins are not in a nested relationship.

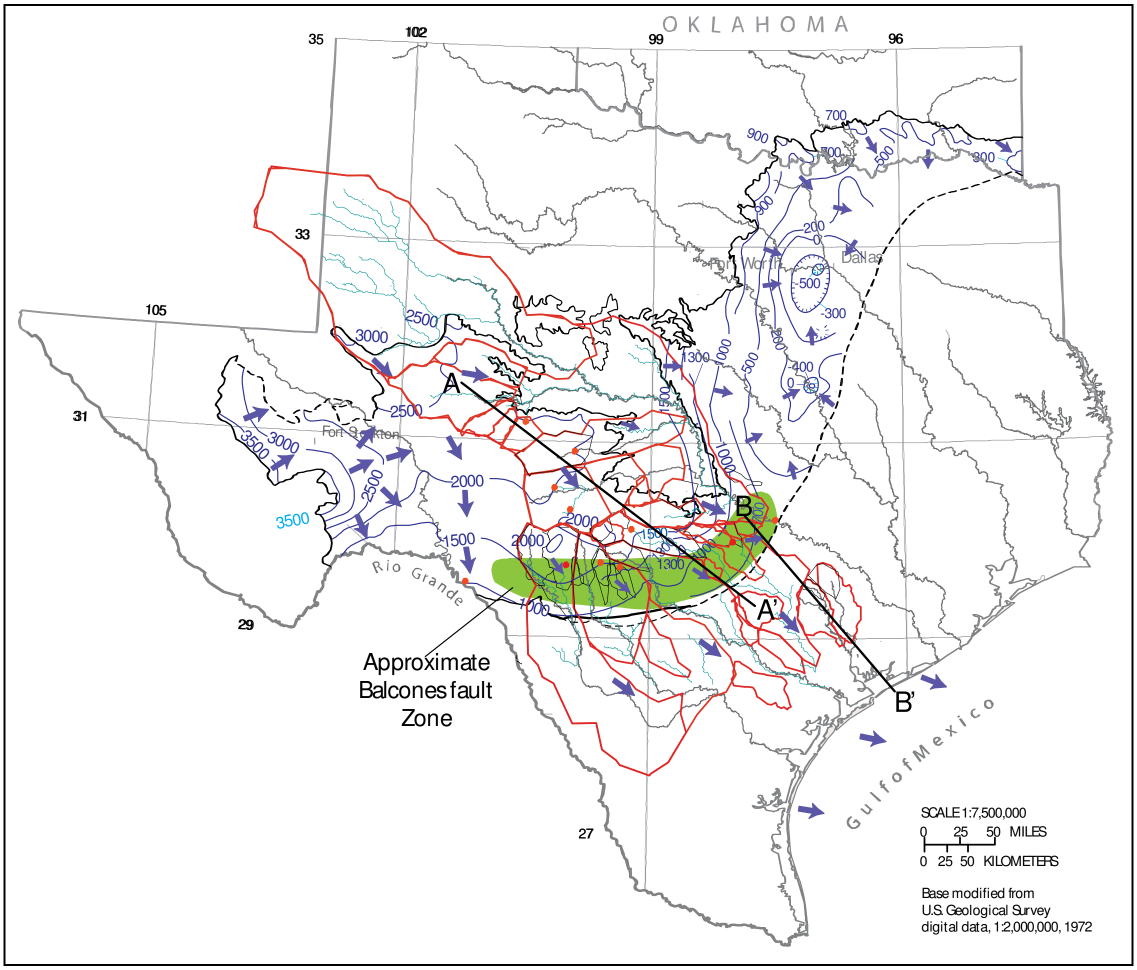

Figure 5-10: Equapotential contours (blue) over the Colorado River basins to the Balcones Fault Zone (green). Basins of interest are outlined in red, and a cross-section corresponding to the A – A' line is found in Figure 5-11. The general direction of groundwater flow (blue arrows) is toward the Gulf of Mexico, where it discharges at, or just offshore of the coast. Map modified from the USGS Groundwater Atlas (http://capp.water.usgs.gov/gwa/gwa.html), flow paths were interpolated based on subsurface topography where not defined.

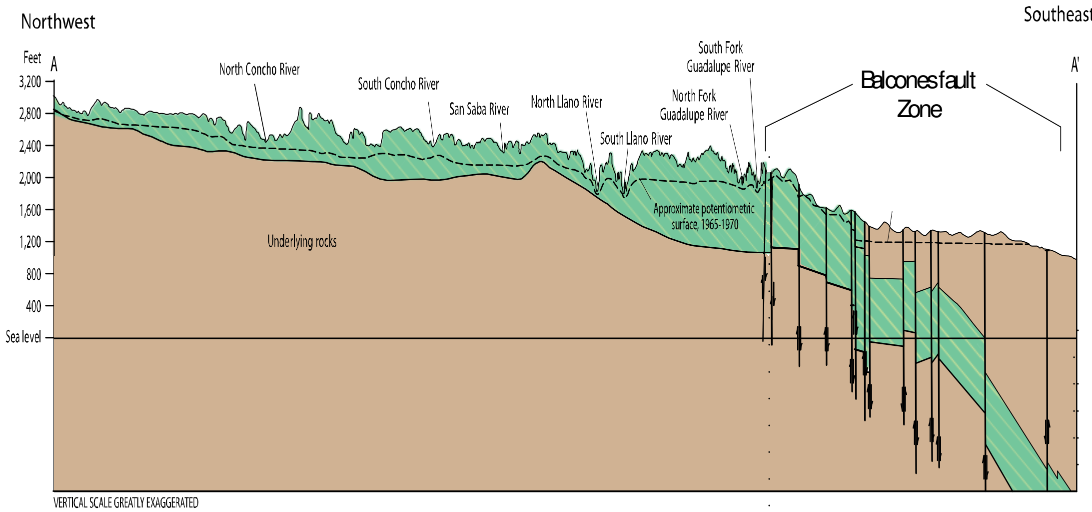

Figure 5-11: A cross section of the western Colorado River basin along the A to A' line found in Figure 5-10. Modified from the USGS Groundwater Atlas (http://capp.water.usgs.gov/gwa/gwa.html).

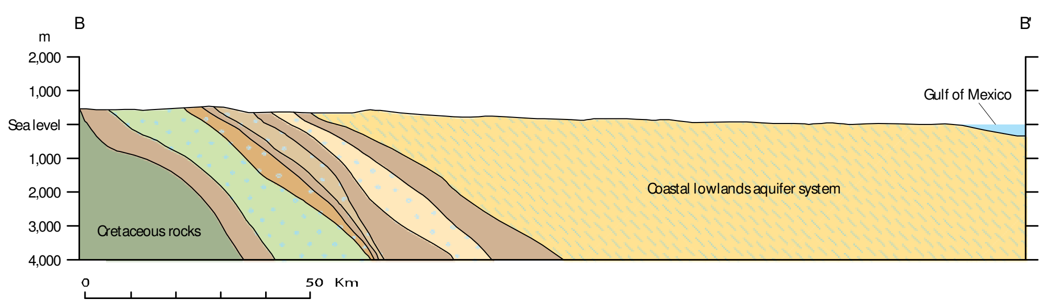

Figure 5-12: A cross-section corresponding to the geology beneath the B to B' line in Figure 5-7 (modified from the USGS Groundwater Atlas (http://capp.water.usgs.gov/gwa/gwa.html).

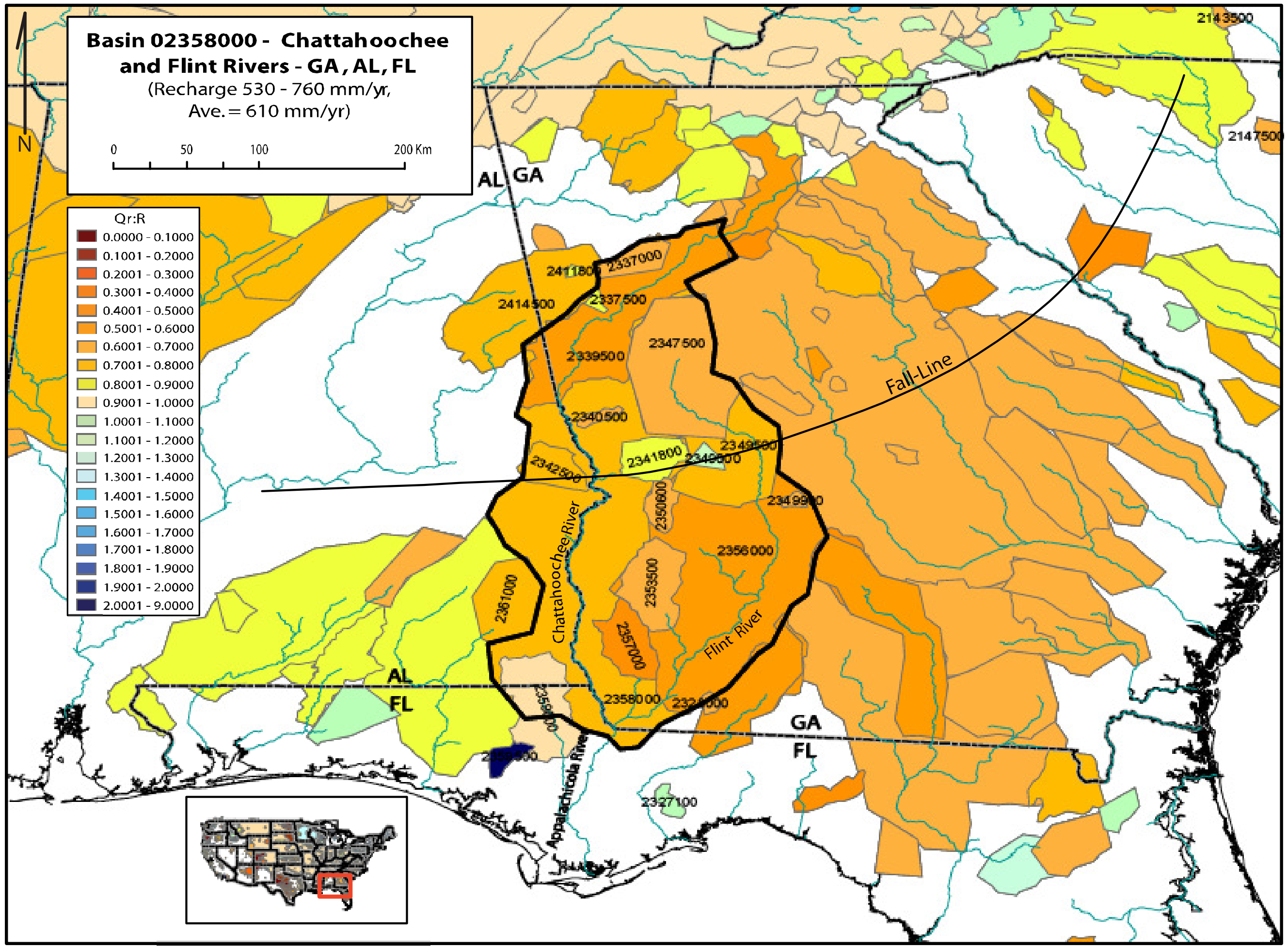

Figure 5-13: Map denoting the Flint and Chattahoochee rivers with their respective basins color coded by Qr:R ratio value (see color scale, left).

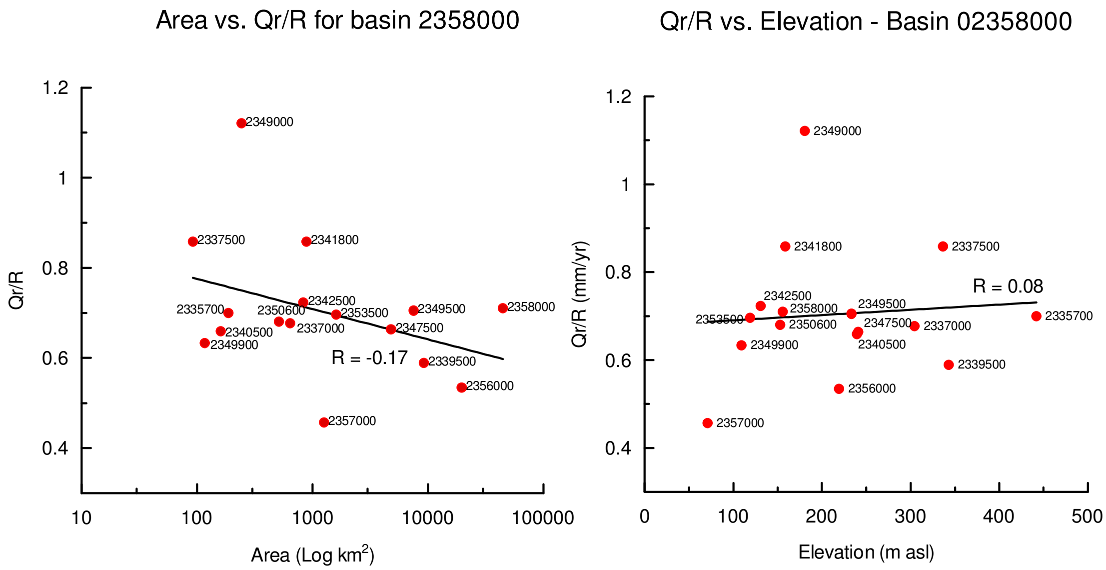

Figure 5-14: (a) Correlation plots between Qr:R and drainage basin area, and (b) mean drainage basin elevation over the Chattahoochee and Flint River basins.

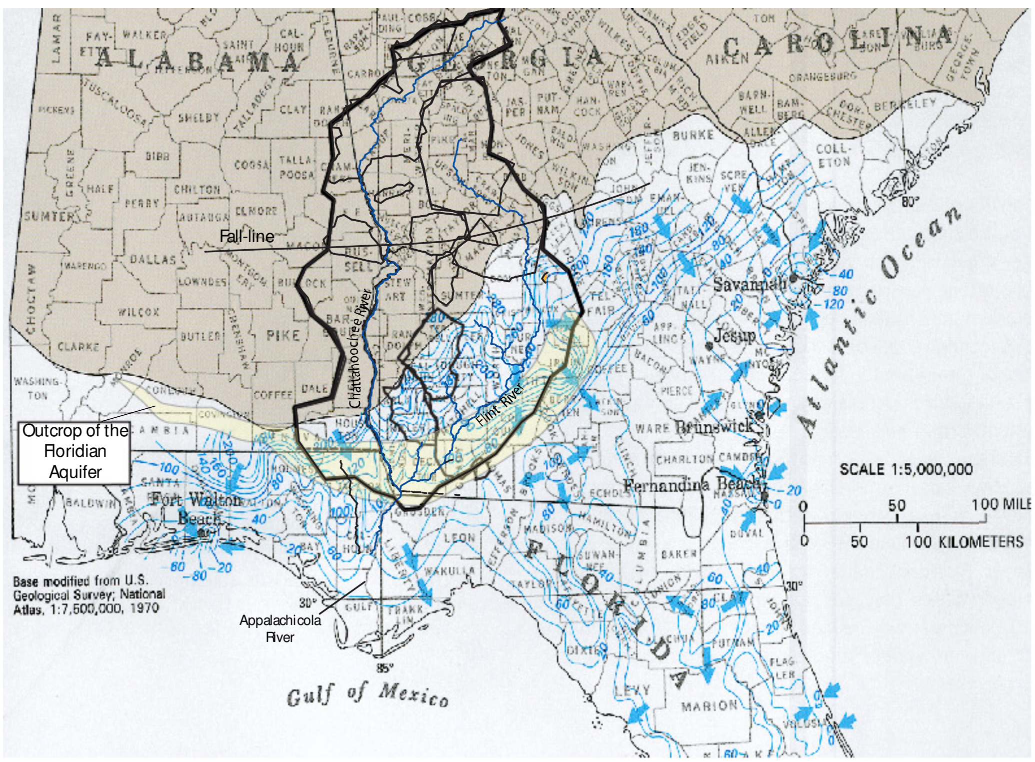

Figure 5-15: Map delineating the Chattahoochee and Flint River basins with flow lines and equapotential contours (modified after USGS Groundwater Atlas (http://capp.water.usgs.gov/gwa/gwa.html). The beige transparency near the confluence of the Flint and Chattahoochee Rivers is the outcropping of the Floridian Aquifer, also denoted schematically in 5-16.

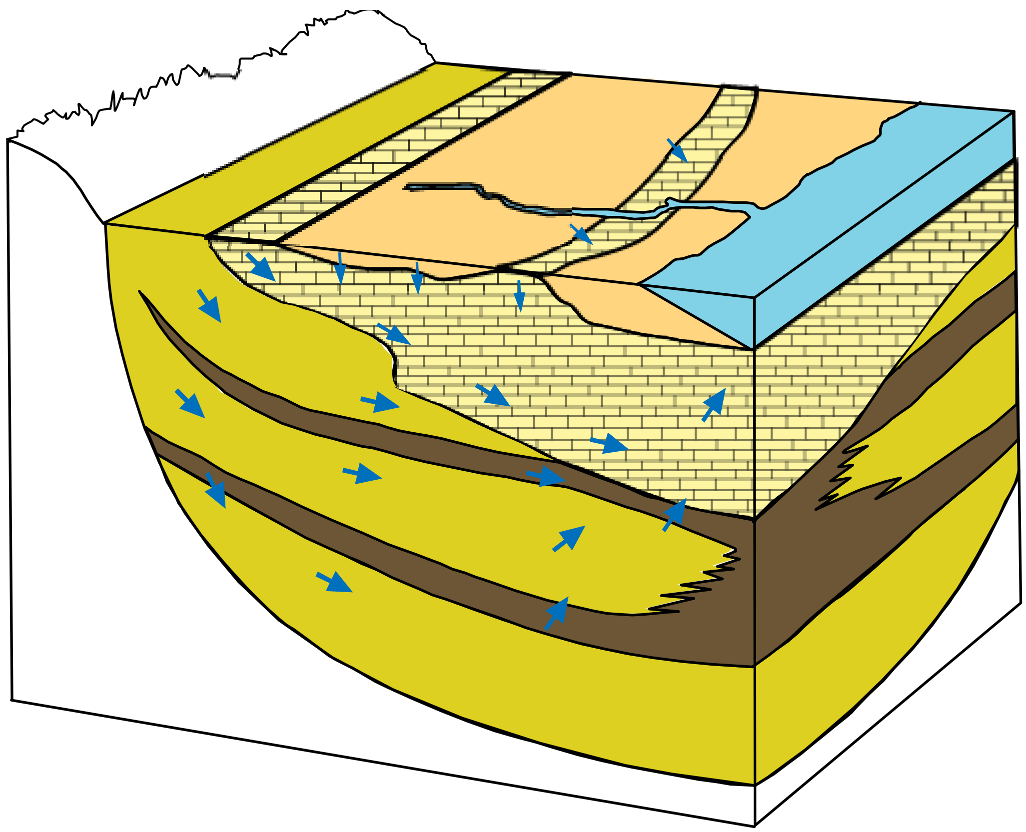

Figure 5-16: An Idealized cross section of the Southeastern Coastal Plain aquifer system showing groundwater flow lines. The surfacial outcrop of the Floridian Aquifer system is shown at top and side (brick pattern).

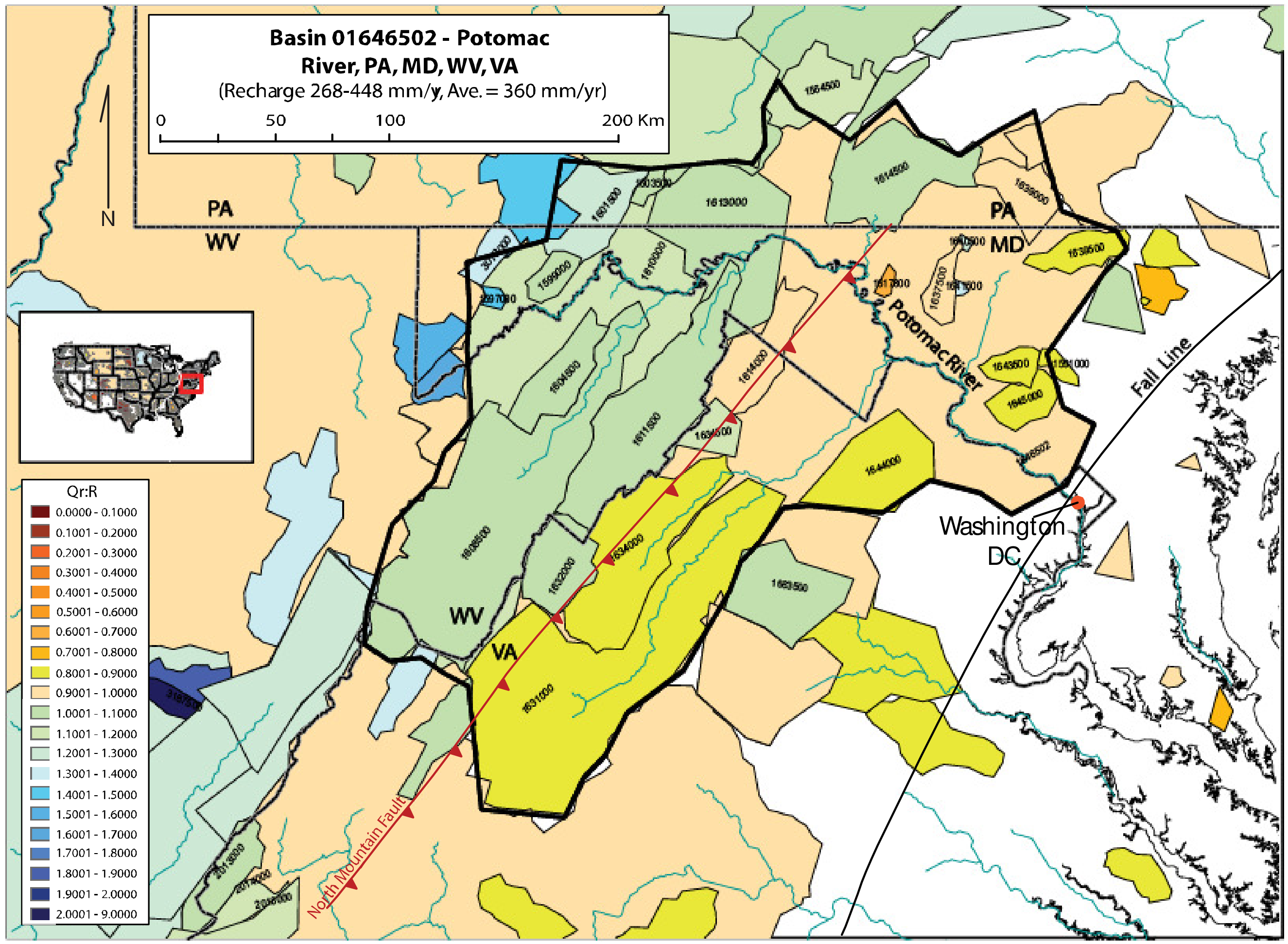

Figure 5-17: Map denoting the Potomac River basin and its internal basins color coded by Qr:R ratio value (colors correspond to scale on left).

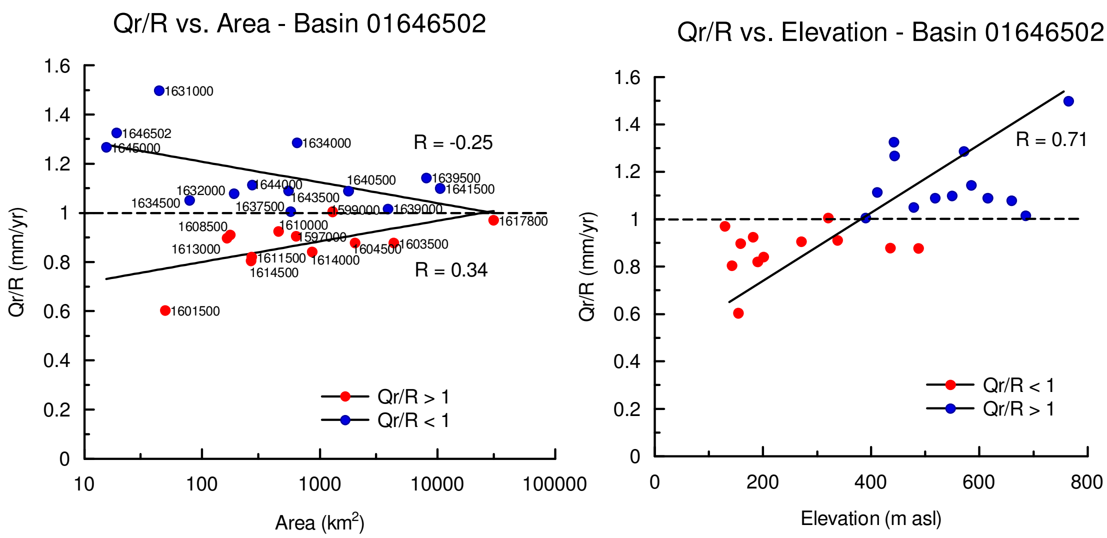

Figure 5-18: (a) Correlation plots between the Qr:R ratio over the Potomac River basin and drainage basin area, and (b) mean basin elevation.

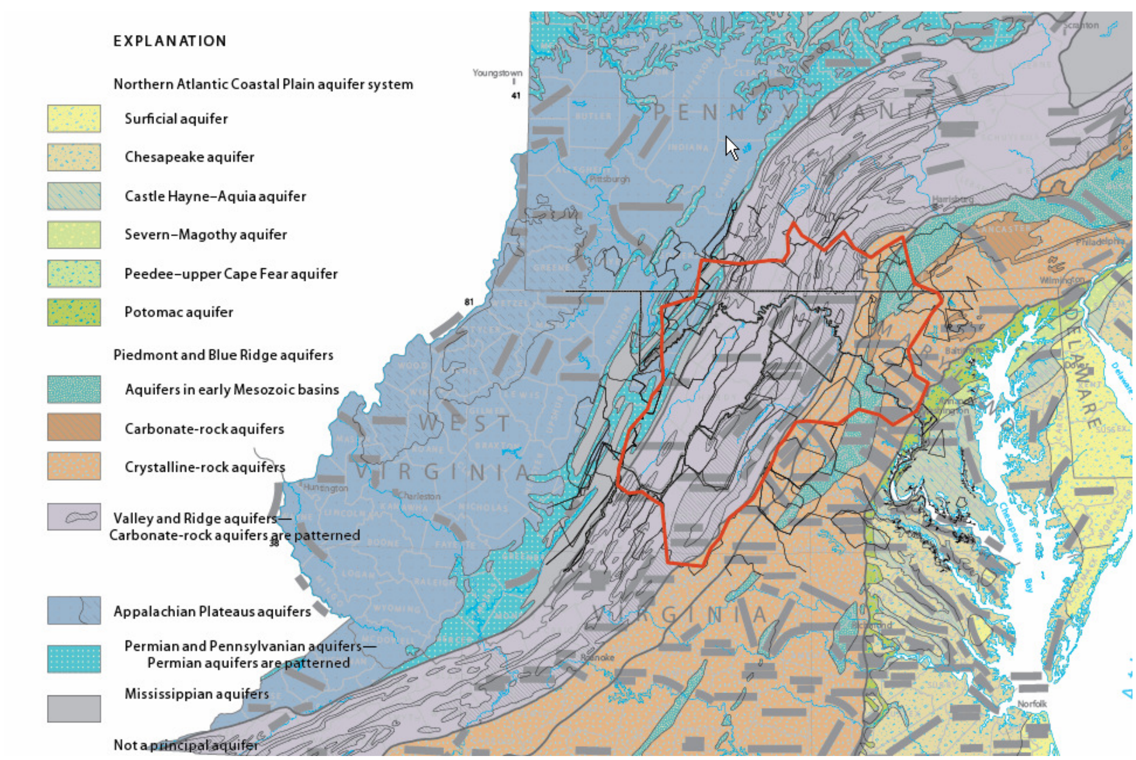

Figure 5-19: Map showing the complexity of the regional geology surrounding the Potomac River basin and the aquifers of the Appalachian Mountain range. The Potomac River exits into the northern west side of Chesapeake Bay, across the bay from Baltimore, Maryland. Figure modified after USGS Groundwater Atlas (http://capp.water.usgs.gov/gwa/gwa.html).

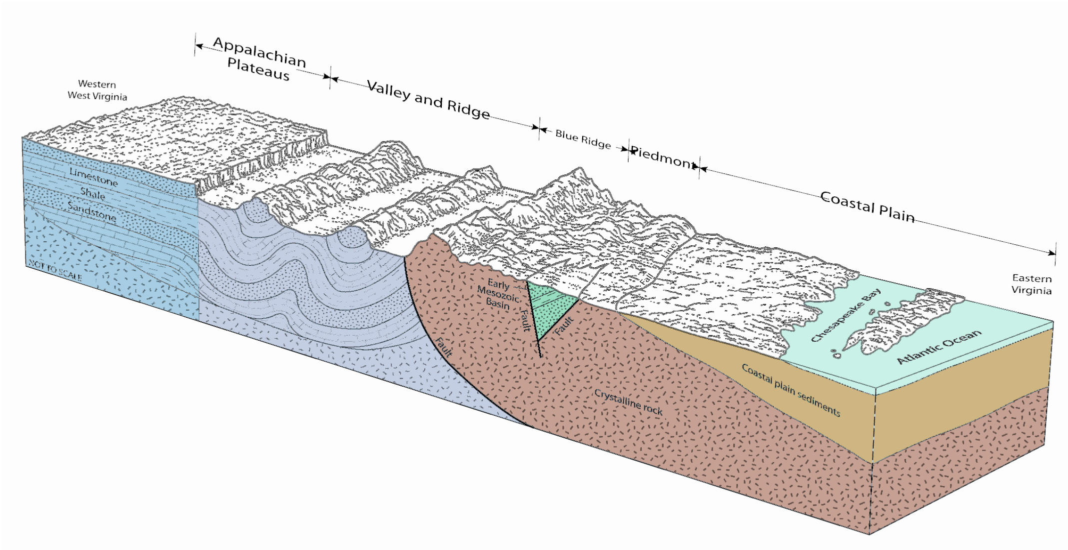

Figure 5-20: Schematic cross section from the Appalachian Plateau to the coastal plain near the Potomac River basin, modified from USGS Groundwater Atlas (http://capp.water.usgs.gov/gwa/gwa.html).

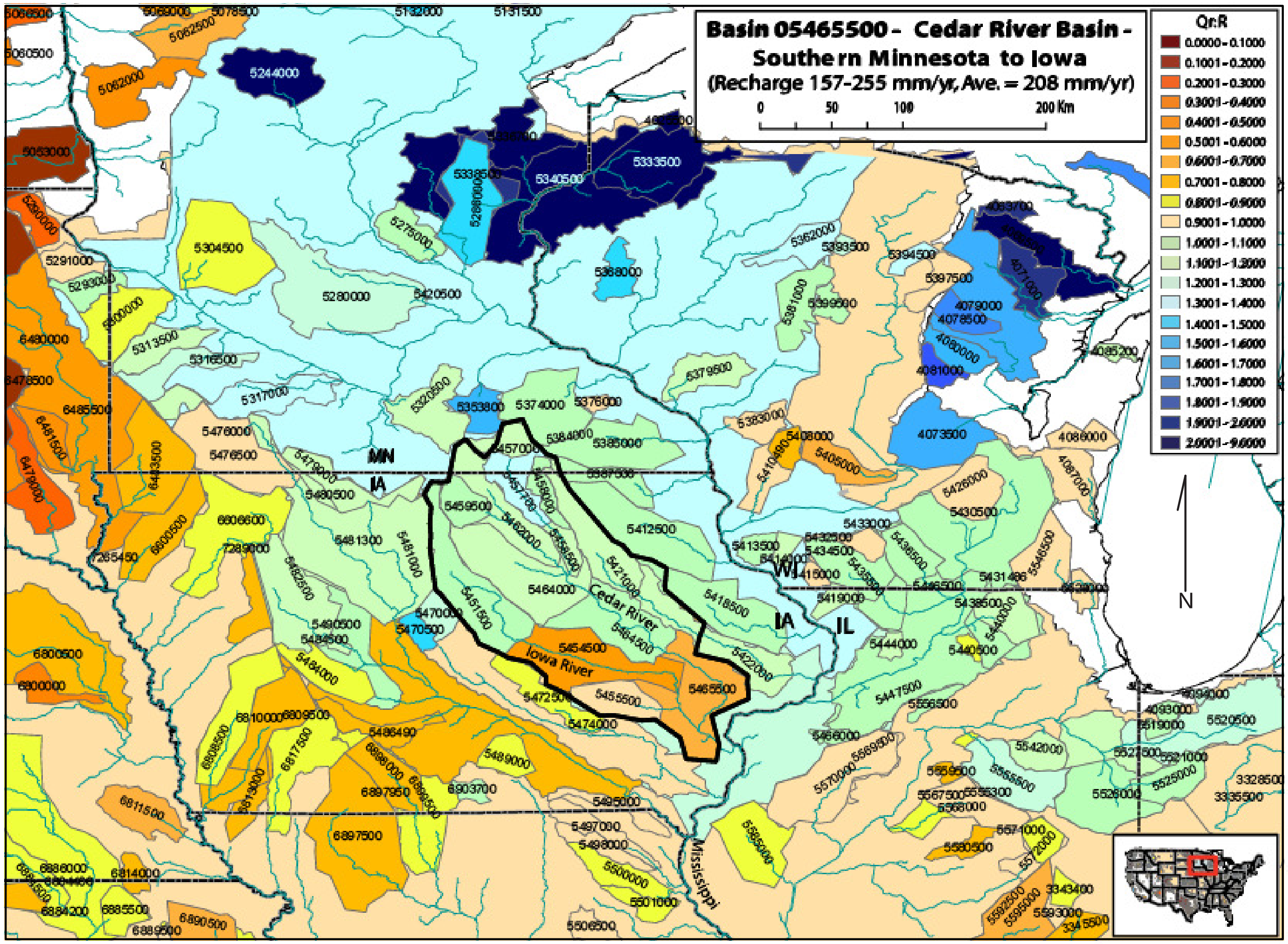

Figure 5-21: Map denoting the extent of the Cedar River basin (black outline) an the surrounding basins, organized by color corresponding to the Qr:R ratio of each basin (see color scale, right).

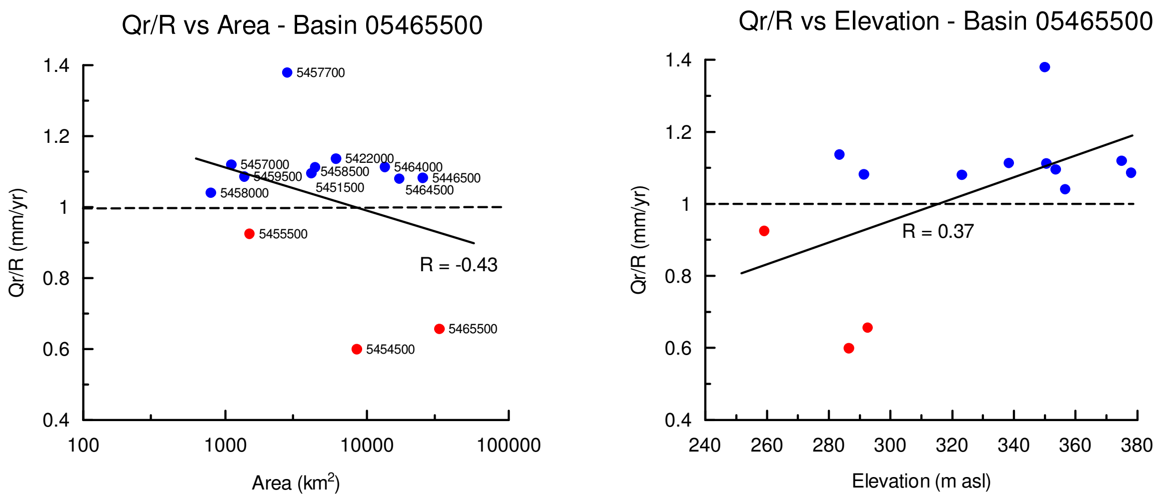

Figure 5-22: (a) Correlation plots between the Qr:R ratio over the Cedar River Basin and basin drainage area, and (b) mean basin elevation.

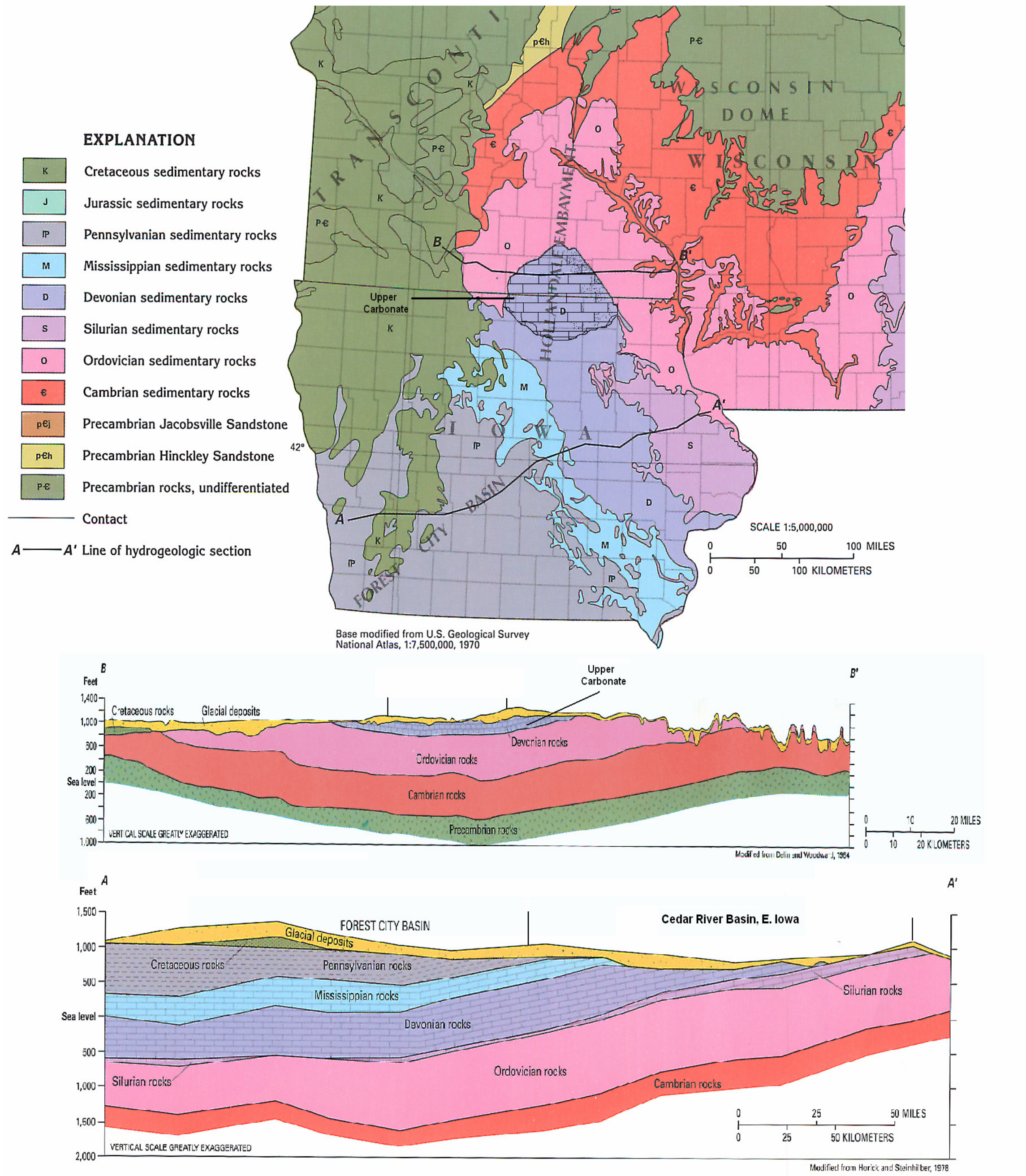

Figure 5-23: The underlying geology beneath the Cedar River Basin, Iowa (modified from USGS Groundwater Atlas (http://capp.water.usgs.gov/gwa/gwa.html)). Cross sections A – A' and B – B' correspond to lines on the map.

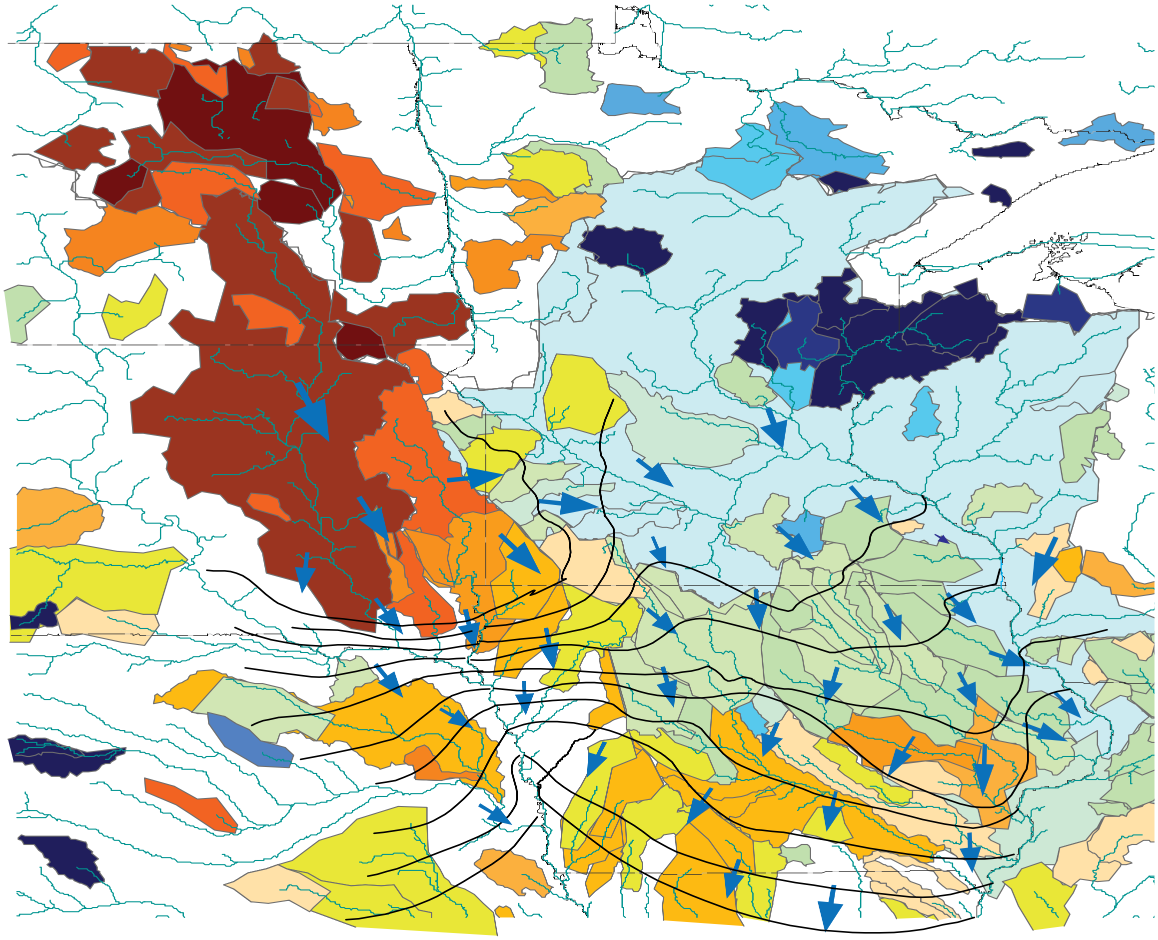

Figure 5-24: Conceptual model of groundwater flow through Iowa in the Devonian Mississippian strata based on USGS potentiometric contours (aggregated from Young (1992), and Mandle and Kontis (1992).

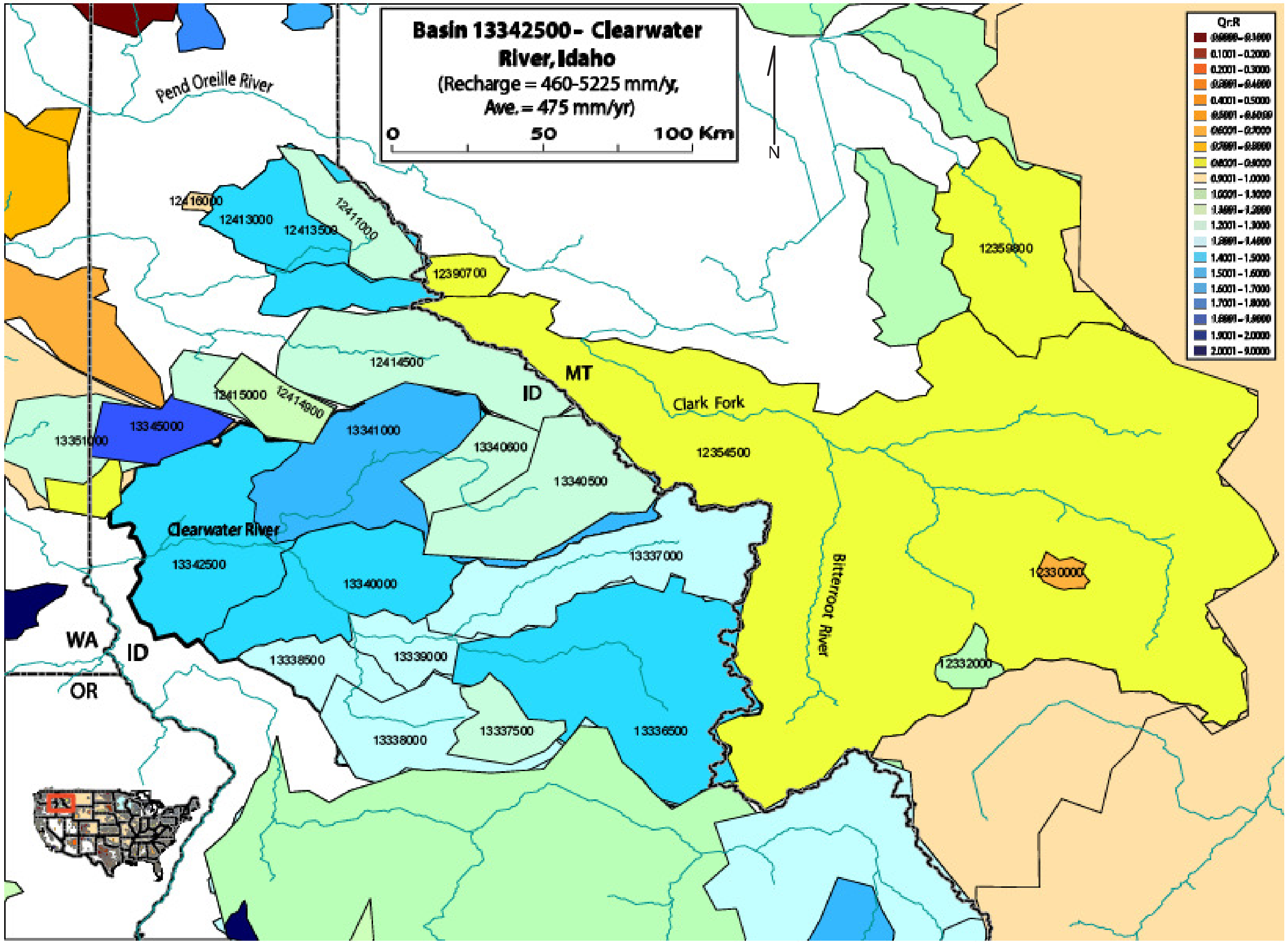

Figure 5-25: Map denoting the extent of the Clearwater River basin (black outline) an the surrounding basins, organized by color corresponding to the Qr:R ratio of each basin (see color scale, right).

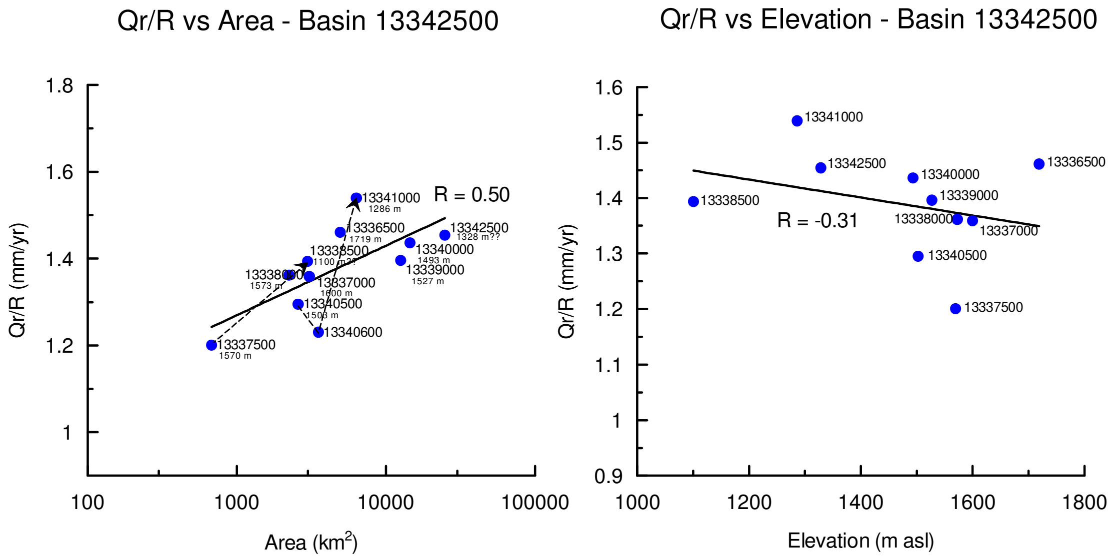

Figure 5-26: (a) Correlation plots between the Qr:R ratio over the Clearwater River basin and internal basin area, and (b) mean basin elevation.

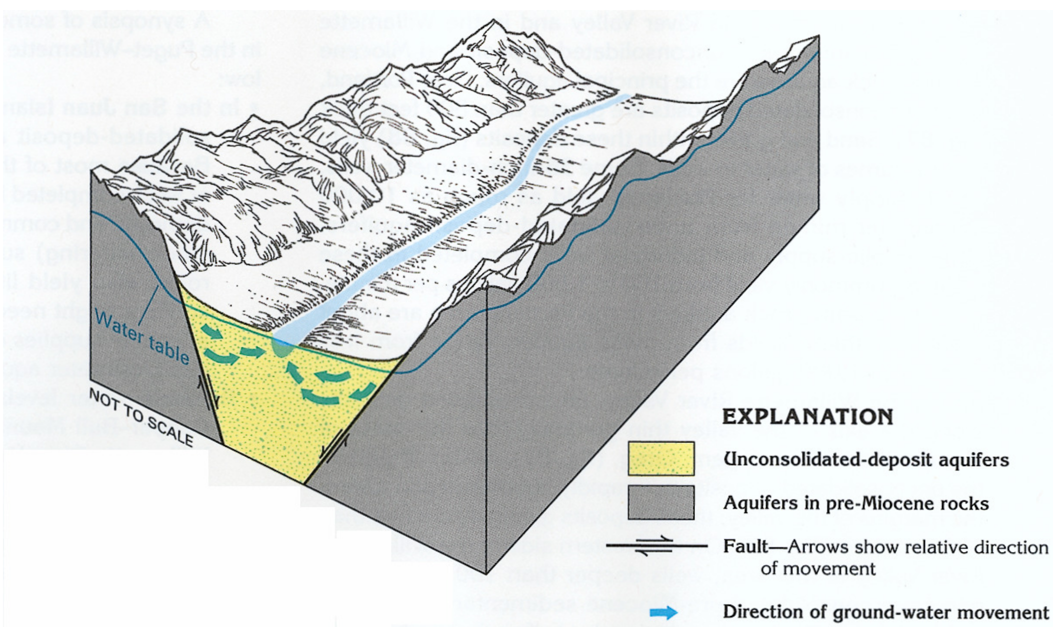

Figure 5-27: Schematic section through the intermountain basins showing groundwater flowing (mostly vertically) along fault lines (modified from USGS Groundwater Atlas: http://capp.water.usgs.gov/gwa/gwa.html). Significant lateral transport of groundwater also occurs along fault scarps toward regional discharge zones at low-lying rivers.

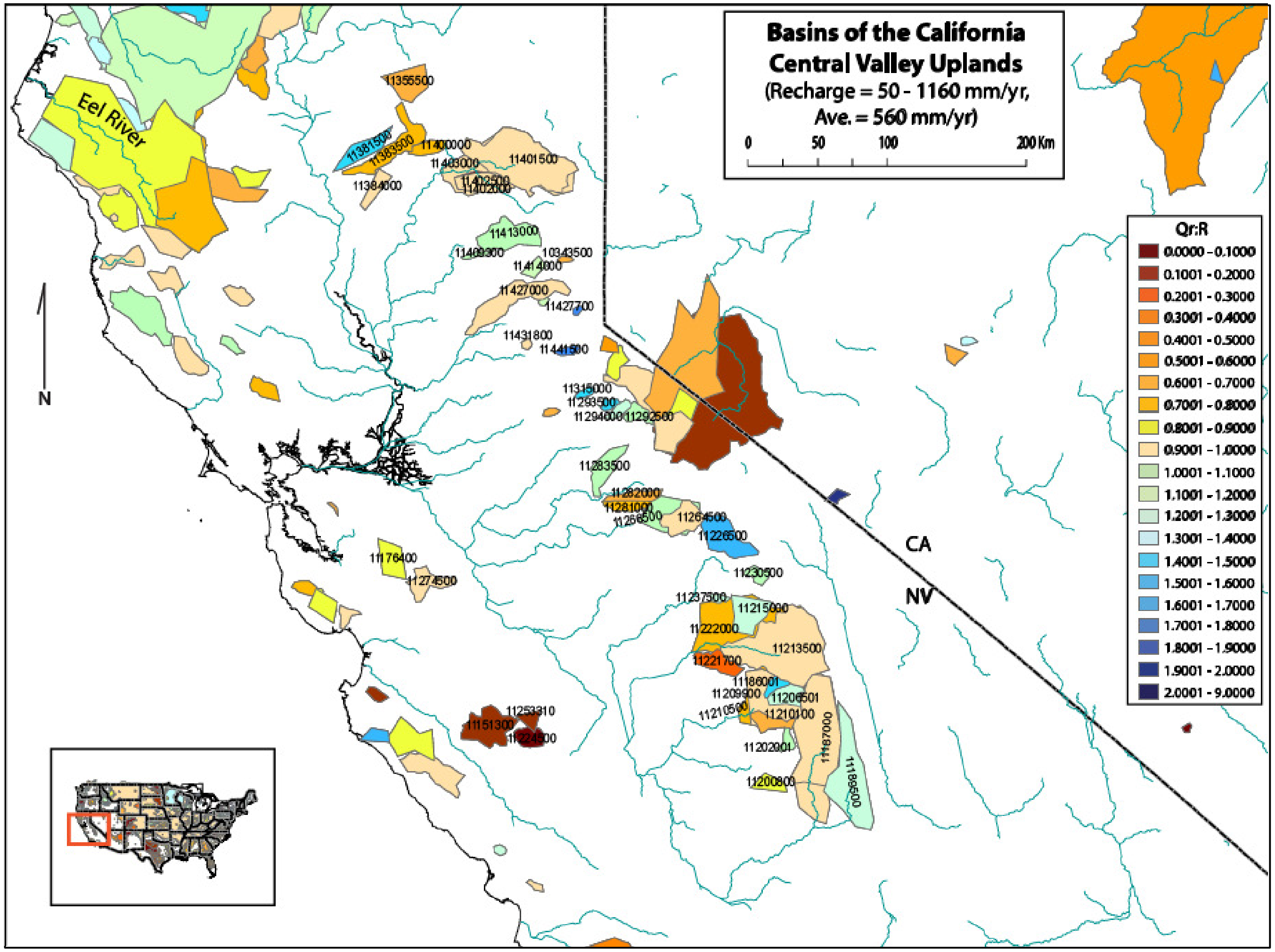

Figure 5-28: Map depicting the California Central Valley Uplands basins color coded by Qr:R ratio, corresponding to the color scale (right).

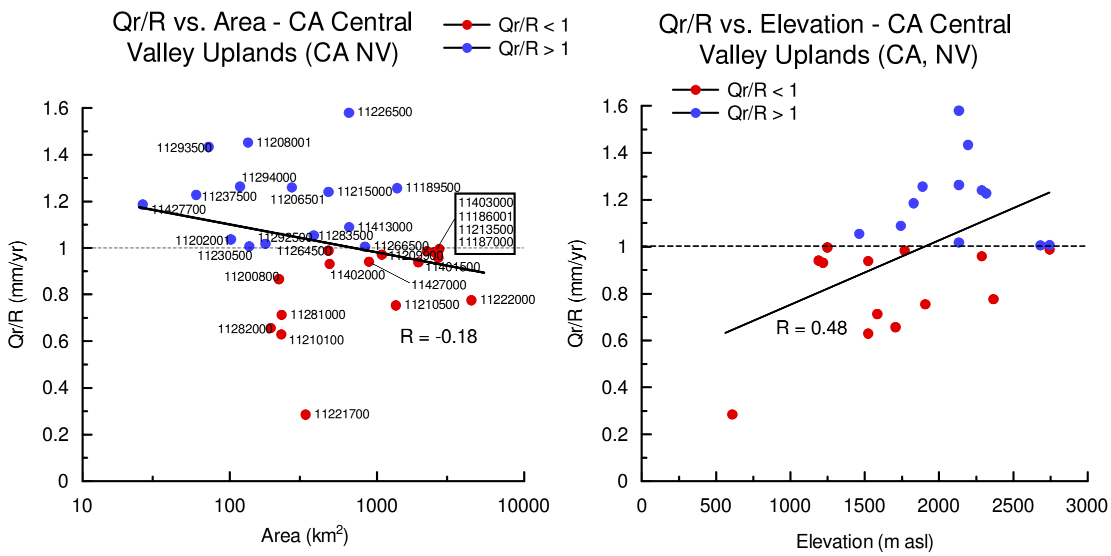

Figure 5-29: (a) Correlation plots between the Qr:R ratio over the California Central Valley Uplands basin and basin drainage area, and (b) basin mean elevation.

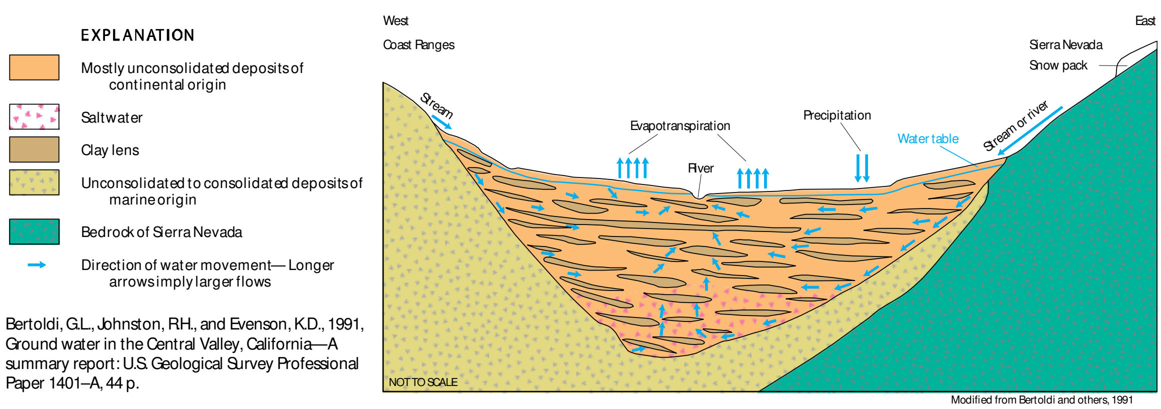

Figure 5-30: A schematic cross section through the California Central Valley showing the pre-development water table and recharge sources. Regional flow recharges the water table at the valley floor sourced primarily from the snow packs in the Sierra Nevada Mountains (figure after USGS Groundwater Atlas http://capp.water.usgs.gov/gwa/gwa.html).