Groundwater Recharge in Texas

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ABSTRACT

Groundwater recharge is critical in evaluating water resources. Recharge estimates are required for groundwater models being developed as part of the Groundwater Availability Modeling program at the Texas Water Development Board. The purpose of this study was to assess the status of data on recharge for the major aquifers in Texas, evaluate the reliability of the recharge estimates, develop conceptual models for recharge for each of the aquifers, review techniques for quantifying recharge, and recommend appropriate techniques for quantifying the recharge of each of the major aquifers.

Recharge rates for all major aquifers were compiled from published reports. The Edwards aquifer is the most dynamic, and recharge rates are highly variable spatially and temporally. Recharge is fairly accurately quantified using stream-gauge data. Estimates of recharge rates in the Carrizo-Wilcox aquifer range from 0.1 to 5.8 in/yr. The higher recharge rates occur in the sandy portions of the aquifer (i.e., the Carrizo and Simsboro Formations). Reported recharge rates for the Gulf Coast aquifer (0.0004 to 2 in/yr) are generally lower than those for the Carrizo-Wilcox aquifer. In both the Carrizo-Wilcox and Gulf Coast aquifers, higher recharge rates are estimated in upland areas containing sandy soils. Regional recharge rates in the High Plains aquifer, outside irrigated areas, are generally low (0.004 to 1.7 in/yr), whereas playa-focused recharge rates are much higher (0.5 to 8.6 in/yr). Irrigated areas also have fairly high recharge rates (0.6 to 11 in/yr). Recharge rates in the Trinity and Edwards-Trinity Plateau aquifers generally range from 0.1 to 2 in/yr. The Seymour aquifer has recharge rates that range from 1 to 2.5 in/yr. Recharge rates for the Hueco-Mesilla Bolson and the Cenozoic Pecos Alluvium are represented as total recharge along mountain fronts and valley floors.

The main techniques that have been used for estimating recharge are Darcy's Law, groundwater modeling, base-flow discharge, and stream loss. Darcy's Law is widely applied in the confined sections of the Carrizo-Wilcox and Gulf Coast aquifers; however, these recharge estimates may not be very reliable because of uncertainties in estimates of regional hydraulic conductivity. Groundwater modeling has been used to estimate recharge in most aquifers; however, only hydraulic-head data were available to calibrate the models. Model calibration based on hydraulic head data alone can only be used to estimate the ratio of recharge to hydraulic conductivity. Additional information, such as groundwater-age data or base-flow discharge data, is required for estimating recharge using groundwater models. Base-flow discharge has been used to estimate recharge primarily in the Trinity, Edwards-Trinity, Seymour, and Cenozoic Pecos Alluvium aquifers. Stream loss has been used to quantify recharge to the Edwards aquifer. Environmental tracers have been used only to a limited extent (chloride mass balance, tritium, and carbon-14) in the Ogallala and Carrizo Wilcox aquifers. This review of existing data indicates that additional studies are required to provide more quantitative estimates of recharge to the major aquifers.

Techniques for quantifying groundwater recharge have been subdivided into surface water, unsaturated-, and saturated-zone techniques and include those based on physical, chemical, and modeling data. The range of recharge rates and spatial and temporal scales represented by each technique are described. Determination of appropriate techniques for quantifying recharge to the major aquifers depends in part on the recharge rates; however, recharge is what we are trying to quantify. Therefore, we can suggest only different approaches that are likely to provide the most quantitative estimates of recharge. Results from initial studies should provide additional data for optimizing the techniques and refining the recharge estimates.

A phased approach may be required to quantify recharge rates accurately. A variety of approaches should be applied because of uncertainties in recharge estimates. Results from the various techniques can be compared to determine uncertainty in the recharge rates.

INTRODUCTION

Water resources management is critical in Texas because of diminishing water supplies and projected rapid increases in population growth (19 million in 1997 to 36 million in 2050) (Texas Water Development Board, 2002). Recent droughts in Texas have caused researchers to focus attention on recharge issues. For future water resources to be managed, we must understand how much water is recharging groundwater aquifers and how this recharge varies spatially within and between the nine major aquifers. Development of a database containing a consistent set of recharge values for each of the major aquifers will allow various modeling groups to have access to similar data and will help avoid use of widely ranging recharge rates in different groundwater models. Studies of recharge have been conducted throughout the state; however, no database compiles existing information on recharge. Space and time scales represented by the recharge rates will be important for using the data in numerical modeling. Some recharge rates represent point estimates, whereas others reflect lumped values for large areas. In addition, some recharge rates represent current recharge, whereas others integrate much longer time scales. An analysis of the space and time scales of the recharge data will be included in the summary report.

Evaluation of various techniques for incorporating recharge in groundwater models for the Groundwater Availability Modeling (GAM) process is required for accurate prediction of future groundwater resources of the major aquifers. There are various methods of modeling recharge to the system. Some techniques, such as use of the "general head boundary" in MODFLOW, may overpredict recharge and overestimate future water resources. Uncertainty estimates should be included in all recharge rates, and such uncertainties should be propagated through the models to estimate in uncertainties in water resources.

In addition to a literature review, it is important to develop a conceptual understanding of the recharge processes for each of the major aquifers. Recharge may be classified as diffuse or focused. Diffuse recharge describes areally extensive recharge to an aquifer, whereas focused recharge reflects concentrated recharge in areas such as playas in the Southern High Plains. An understanding of the spatial distribution of recharge may be important for numerically modeling water resources and also protecting such areas from contamination. Different stages of aquifer development may also result in a change in recharge (Sophocleous, 1998). In the early stages of aquifer development, groundwater pumpage is derived primarily from groundwater storage; however, groundwater pumpage may ultimately be derived from induced recharge from streams. A conceptual understanding of the recharge processes is a prerequisite to implementing recharge in models of the aquifers and will help in determining how to represent recharge in these models. An understanding of recharge processes is essential if we are going to control recharge or enhance recharge in the future.

The purpose of this study is to

1. review available techniques for estimating recharge;
2. compile all existing information on recharge rates on the basis of physical, chemical, isotopic, and modeling techniques for the nine major aquifers in the state by examining databases and literature;

3. evaluate the range of recharge rates for each aquifer in light of the techniques used to estimate recharge and assess whether the various techniques are appropriate;
4. develop a conceptual model for the recharge processes in each aquifer and determine local and regional controls on recharge;
5. evaluate techniques for simulating recharge in groundwater models; and
6. determine which aquifers require additional recharge studies in order to recommend appropriate techniques for quantifying recharge in these aquifers.

TECHNIQUES FOR ESTIMATING RECHARGE

For purposes of discussion, techniques for estimating recharge are subdivided into three: those based on data from surface water, unsaturated zones, and saturated zones. This subdivision of techniques is somewhat arbitrary and may not be ideal. The different zones provide recharge estimates over varying space and time scales. Within each zone, techniques can generally be classified into physical, tracer, or numerical-modeling approaches. This overview focuses on aspects of each approach that are important in choosing appropriate techniques, such as the space/time scales, range, and reliability of recharge estimates. The range of recharge rates for different techniques is based on evaluation of the literature and general evaluation of uncertainties and should be considered only approximate. Because many techniques in the different zones are based on the water-budget equation, this topic is described separately.

WATER BUDGET

The water budget for a basin can be stated as

$$P + Q_{in} = ET + Q_{out} + \Delta S \quad (1)$$

where P is precipitation (and may also include irrigation); Q_{in} and Q_{out} are water flow into and out of the basin, respectively; ET is evapotranspiration; and DS is change in water storage. All components are given as rates (e.g., in/d or in/yr). Individual components can be broken down into subcomponents. Water flow into or out of the basin can be written as the sum of surface flow, interflow, and groundwater flow. ET can be distinguished on the basis of the source of evaporated water (surface, unsaturated, or saturated zones). Water storage takes place in snow, surface-water reservoirs, the unsaturated zone, and the saturated zone. Rewriting the water-budget equation to incorporate many of these subcomponents results in

$$P + Q_{in}^{sw} + Q_{in}^{gw} = ET^{sw} + ET^{uz} + ET^{gw} + R_0 + Q_{out}^{gw} + Q^{bf} + \Delta S^{snow} + \Delta S^{sw} + \Delta S^{uz} + \Delta S^{gw} \quad (2)$$

where superscripts refer to the subcomponents described above, R_0 (runoff) is surface-water flow off the basin, and Q^{bf} is base flow (groundwater discharge to streams or springs). Groundwater recharge, R, includes any infiltrating water that reaches the saturated zone and can be written as (Schicht and Walton, 1961)

$$R = Q_{off}^{gw} - Q_{on}^{gw} + Q^{bf} + ET^{gw} + \Delta S^{gw} \quad (3)$$

This equation simply states that all water arriving at the water table flows out of the basin as groundwater flow, is discharged to the surface, is evapotranspired, or is retained in storage. Substituting this equation into equation 2 produces the following version of the water budget:

$$R = P + Q_{on}^{sw} - R_0 - ET^{sw} - ET^{uz} - \Delta S^{snow} - \Delta S^{sw} - \Delta S^{uz} \quad (4)$$

Water-budget methods are those that are based, in one form or another, on a water-budget equation. They include most hydrologic models, such as surface-water and groundwater flow models. For any site, some of the terms in equation 4 are likely to be negligible in magnitude and therefore may be ignored.

The most common way of estimating recharge by the water-budget method is the indirect or "residual" approach, whereby all of the variables in the water-budget equation, except for *R*, are measured or estimated and *R* is set equal to the residual. Recharge may be underestimated using the water budget approach in karst systems because evapotranspiration may be overestimated when recharge is rapid. An advantage of water-budget methods is flexibility. Few assumptions are inherent in equation 1. The methods are not hindered by any presuppositions as to the mechanisms that control the individual components. Hence, they can be applied over a wide range of space and time scales, ranging from lysimeters (cm, s) to global climate models (km, centuries).

The major limitation of the residual approach is that accuracy of the recharge estimate depends on the accuracy with which the other components in the water-budget equation can be measured. This limitation is critical when the magnitude of the recharge rate is small relative to that of the other variables, particularly *ET*. In this case, small inaccuracies in values of those variables commonly result in large uncertainties in the recharge rate. Some authors (e.g., Gee and Hillel, 1988; Lerner et al., 1990; and Hendrickx and Walker, 1997) therefore questioned the usefulness of water-budget methods in arid and semiarid regions. However, if the water budget is calculated on a daily time step, *P* can greatly exceed *ET* on a single day, even in arid settings. Averaging over longer time periods tends to dampen out extreme precipitation events (those most responsible for recharge events). Methods for measuring or estimating various components of the water budget are in Hillel (1980), Rosenberg et al. (1983), and Tindall and Kinkel (1999).

RECHARGE ESTIMATION TECHNIQUES BASED ON SURFACE-WATER STUDIES

The status of recharge related to surface-water bodies depends on the degree of connection between surface water and groundwater systems (Fig. 1) (Sophocleous, 2002). Humid regions are generally characterized by gaining surface-water bodies because groundwater discharges to streams and lakes. In contrast, arid regions are generally characterized by losing surface-water bodies because surface-water and groundwater systems are often separated by thick unsaturated sections. Therefore, surface-water bodies often form localized recharge sources in arid settings. Recharge can be estimated using surface-water data in gaining and losing surface-water bodies.

Physical Techniques

Channel-Water Budget

Surface-water gains or losses can be estimated using channel-water budgets based on stream-gauging data. Lerner (1997), Lerner et al. (1990), and Rushton (1997) provided detailed reviews of this approach. The channel-water budget is described as (Lerner 1997)

$$R = Q_{up} - Q_{down} + \sum Q_{in} - \sum Q_{out} - E_a - \frac{\Delta S}{\Delta t} \tag{5}$$

where R is recharge rate, Q is flow rate, Q_{up} and Q_{down} are flows at the upstream and downstream ends of the reaches, ΣQ_{in} and ΣQ_{out} refer to tributary inflows and outflows along the reach, E_a is the evaporation from surface water or streambed, and ΔS is change in channel and unsaturated-zone storage over change in time (Δt). The term *transmission loss* refers to the loss in stream flow between upstream and downstream gauging stations (Lerner et al., 1990). This loss reflects potential recharge that can result in an overestimate of actual recharge because of bank storage and subsequent evapotranspiration, development of perched aquifers, and inability of the aquifer to accept recharge because of a shallow water table or low transmissivity (Lerner et al., 1990). Recharge values generally reach a constant rate when the water-table depth is greater than twice the stream width because flow is generally controlled by gravity at these depths (Bouwer and Maddock, 1997) (Fig. 1).

The range of recharge rates that can be measured using this technique depends on the magnitude of the transmission losses relative to the uncertainties in the gauging data and tributary flows. Gauging data in the U.S. are generally considered to be accurate to ±5% (Rantz, 1982). Lerner et al. (1990) indicated that measurement errors during high flows are often ±25% and can range from –50% to +100% during flash floods in semiarid regions. The recharge estimates represent average values over the reach between the gauging stations. Temporal scales represented by the recharge values range from event scale (minutes to hours) to much longer time scales that are estimated by summation of individual events.

Seepage Meters

Seepage to or from surface-water bodies can be measured by using seepage meters (Kraatz, 1977; Lee and Cherry, 1978), which consist of a cylinder that is pushed into the bottom of the stream or lake. Attached to the cylinder is a reservoir of water; the rate at which water within the cylinder infiltrates is determined by changes in reservoir volume. This method is inexpensive and easy to apply. An automated seepage meter was described by Taniguchi and Fukuo (1993). Uncertainties in estimated fluxes can be determined from replicate measurements. Seepage fluxes measured in different U.S. studies vary greatly, from approximately 0.04 to 118 in/d at a site in Minnesota (Rosenberry, 2000), 0.4 to 8.8 in/d in Minnesota and Wisconsin (Lee, 1977), and 0.5 to 4.8 in/d in Nevada (Woessner and Sullivan, 1984). Because seepage meters provide point estimates of water fluxes, measurements may be required at many locations for a representative value to be obtained. Time scales range from those based on individual events to days. Recharge over longer times is estimated from a summation of shorter times.

Base-Flow Discharge

In watersheds with gaining streams, groundwater recharge can be estimated from stream hydrograph separation (Meyboom, 1961; Rorabough, 1964; Mau and Winter, 1997; Rutledge, 1997; Halford and Mayer, 2000). Use of base-flow discharge to estimate recharge is based on a water-budget approach (equation 3), in which recharge is equated to discharge. Base-flow discharge, however, is not necessarily directly equated to recharge because pumpage, evapotranspiration, and underflow to deep aquifers may also be significant. These other discharge components should be estimated independently. Bank storage may complicate hydrograph analysis because water discharging from bank storage is generally derived from short-term fluctuations in surface-water flow and not from areal aquifer recharge and could result in overestimation of recharge. Various approaches are used for hydrograph separation, including digital filtering (Nathan and McMahon, 1990; Arnold et al., 1995) and recession-curve displacement methods (Rorabough, 1964). The accuracy of the reported recharge rates depends on the validity of the various assumptions. Recharge estimates based on hydrograph separation range from 6 to 50 in/yr in 89 basins (Rutledge and Mesko, 1996) and from 5 to 25 in/yr in 15 basins (Rutledge and Daniel, 1994) in the eastern U.S. Rutledge (1998) recommended an upper limit on basin size of 500 mi^2 for application of this method because of difficulties in separating surface-water and groundwater flow and bank-storage effects in larger systems and because of the areally uniform recharge assumption. The minimum time scale is a few months. Recharge over longer times can be estimated by summation of estimates over shorter times. Recent progress has been made on the use of chemical and isotopic techniques to infer the sources of stream flow from end members, such as rainfall, soil water, groundwater, and bank storage (Hooper et al., 1990; Christophersen and Hooper, 1992). This approach is data intensive, but it provides information that is useful in conducting hydrograph separation. Suecker (1995) used sodium concentrations in a two-component mixing model to determine the subsurface contribution to three alpine streams in Colorado.

Tracer Techniques

Heat Tracer

Installation and maintenance of stream-gauging stations are expensive and difficult, particularly in ephemeral streams in semiarid regions that are subject to erosion. As an alternative to stream gauging, heat can be used as a tracer to provide information on when surface water is flowing in ephemeral streams and to estimate infiltration from surface-water bodies (Stallman, 1964; Lapham, 1989; Constantz et al., 1994; Ronan et al., 1998). Monitoring depths vary, depending on time scales, sediment types, and anticipated water fluxes beneath the stream. Diurnal temperature fluctuations are generally monitored at depths of ~ 0.16 to 3.3 ft for fine-grained material, and 1 to 10 ft for coarse-grained material. Depths for monitoring annual temperature fluctuations are generally an order of magnitude greater. Measured temperature is used with inverse modeling that uses a nonisothermal variably saturated flow code, such as VS2DH (Healy and Ronan, 1996), to estimate hydraulic conductivity of the sediments. Data analysis is complex, and inverse solutions may not be unique. Percolation rates can be estimated if the hydraulic head is calculated from measured data. Temperature can be monitored accurately and inexpensively using thermistors or thermocouples. Heat dissipation sensors can be used to monitor temperature and matric potential simultaneously in unsaturated media.

The minimum net infiltration rate that can be estimated using heat as a tracer depends on the range of surface-water temperature fluctuations and the time scale considered. Stallman (1964) suggested a minimum recharge rate of ~0.8 in/d in the streambed using diurnal temperature fluctuations and ~0.04 in/d using annual temperature fluctuations in natural media with average heat properties. Reported infiltration rates from various studies include from 0.002 to 0.25 in/d (Maurer and Thodal, 2000), 0.71 to 1.46 in/d (Bartolino and Niswonger, 1999), and 18 in/d (Lapham, 1989). Previous studies generally use a single vertical array of temperature sensors and therefore provide an estimate of one-dimensional flow at a point; however, some ongoing studies include two- and three-dimensional arrays of sensors to provide more realistic, three-dimensional flux estimates beneath streams (R. Niswonger, U.S. Geol. Survey, personal communication, 2001). Recharge can be estimated for time periods ranging from hours to years.

Numerical Modeling

Watershed (rainfall/runoff) modeling is used to estimate recharge rates over large areas. Singh (1995) reviewed many watershed models, which generally provide recharge estimates as a residual term in the water-budget equation (equation 4) (Arnold et al., 1989; Leavesley and Stannard, 1995; Hatton, 1998). The minimum recharge rate that can be estimated is controlled by the accuracy with which the various parameters in the water budget can be measured (~±10%) and the time scale considered. The various watershed models differ in spatial resolution of the recharge estimates. Some models are termed *lumped* and provide a single recharge estimate for the entire catchment (Kite, 1995). Others are spatially disaggregated into hydrologic response units (HRU's) or hydrogeomorphological units (HGU's) (Salama et al., 1993; Leavesley and Stannard, 1995). Watershed models are applied at a variety of scales. Bauer and Mastin (1997) applied the Deep Percolation Model to three small watersheds (average size 0.15 mi^2) in Puget Sound in Washington, USA. Average annual recharge rates are 1.5, 5.4, and 6.8 in for the three basins. Arnold et al. (2000) applied the SWAT model to the upper Mississippi River Basin (189,962 mi^2). The basin was divided into 131 hydrologic-response units with an average area of 1,448 mi^2. Estimated annual recharge ranged from 0.4 to15.7 in. Small-scale applications allow more precise methods to be used to measure or estimate individual parameters of the water-budget equation (Healy et al., 1989). Time scales in models are daily, monthly, or yearly. Daily time steps are desirable for estimation of recharge because recharge is generally a larger component of the water budget at smaller time scales. Other recent applications of watershed models to estimate recharge include Arnold and Allen (1996; recharge rates 3.3 to 7.5 in/yr, Illinois, USA), Sami and Hughes (1996; recharge rate ~ 0.2 in/yr in a fractured system, South Africa), and Flint et al. (2002; recharge rate 0.11 in/yr at Yucca Mountain, Nevada, USA).

RECHARGE ESTIMATION TECHNIQUES BASED ON UNSATURATED-ZONE STUDIES

Unsaturated-zone techniques for estimating recharge are applied mostly in semiarid and arid regions, where the unsaturated zone is generally thick. These techniques are described in detail in Gee and Hillel (1988), Hendrickx and Walker (1997), Scanlon et al. (1997), and Zhang (1998). The recharge estimates generally apply to smaller spatial scales than those calculated from surface-water or groundwater approaches. Unsaturated-zone techniques provide estimates of potential recharge that are based on drainage rates below the root zone; however, in some cases, drainage is diverted laterally and does not reach the water table. In addition, drainage rates in thick, unsaturated zones do not always reflect current recharge rates at the water table.

Physical Techniques

Lysimeters

The various components of the soil-water budget are accurately measured by using lysimeters (Brutsaert, 1982; Allen et al., 1991; Young et al., 1996). Lysimeters consist of containers filled with disturbed or undisturbed soil, with or without vegetation, that are hydrologically isolated from the surrounding soil for purposes of measuring the components of the water balance. All lysimeters are designed to allow collection and measurement of drainage. Precipitation and water storage are measured separately in drainage lysimeters (also termed pan or nonweighing lysimeters). Weighing lysimeters are generally used for accurate measurements of evapotranspiration. They are constructed on delicate balances capable of measuring slight changes in weight that represent precipitation and water-storage changes. Surface areas of lysimeters range from 16 in^2 (Evett et al., 1995) to 46 in^2 for large pan lysimeters (Ward and Gee, 1997); depths range from tens of inches to 33 to 66 ft (Gee et al., 1994). If the base of the lysimeter is not deeper than the root zone, measured drainage fluxes will overestimate aquifer recharge rates. Therefore, lysimeters are generally unsuitable for areas with deep-rooted vegetation. Recharge rates can be estimated at time scales from minutes to years. The minimum water flux that can be measured using a lysimeter depends on the accuracy of the drainage measurements and the surface area of the lysimeter. For large lysimeters (surface area 1,076 ft^2), recharge rates of about 0.04 in/yr can be resolved. The upper flux that can be measured depends on the design of the drainage system but should exceed drainage fluxes in most natural settings. A wide variety of recharge rates have been measured using lysimeters: 13.5 to 18.8 in/yr over a 3-yr period for the Bunter Sandstone, England (surface area, 1076 ft^2; Kitching et al., 1977), 7.9 in/yr for the Chalk aquifer, England (surface area, 269 ft^2; Kitching and Shearer, 1982), and 0.04 to 7.9 in/yr in a 60-ft-deep lysimeter in a semiarid site (Hanford, Washington, USA, Gee et al., 1992). Most lysimeters have a drainage-collection system that is open to the atmosphere, which creates a seepage-face lower-boundary condition. For thick unsaturated zones, this artifact causes different moisture and pressure-head profiles in the lysimeter relative to those in the adjacent undisturbed area (van Bavel, 1961). To minimize the influence of the bottom boundary, some lysimeters have been built with a porous plate on the bottom that is set at a prescribed pressure head. Lysimeters are not routinely used to estimate recharge because they are expensive and difficult to construct and have high maintenance requirements. They are more suitable for evaluation of evapotranspiration than recharge.

Zero Flux Plane

The soil-water budget can be simplified by equating recharge to changes in soil-water storage below the zero flux plane (ZFP), which represents the plane where the vertical hydraulic gradient is zero. The ZFP separates upward (ET) from downward (drainage) water movement. The rate of change in the storage term between successive measurements is assumed to be equal to the drainage rate to the water table or the recharge rate. The ZFP requires soil matric-potential measurements to locate the position of the ZFP and soil-water-content measurements to estimate storage changes. The ZFP method, first described in Richards et al. (1956), has been used in various studies (Royer and Vachaud, 1974; Wellings, 1984; Dreiss and Anderson, 1985; Healy et al., 1989). The minimum recharge rate that can be measured is controlled by the accuracy of the water-content measurements (generally ± 0.01 m^3/m^3). Recharge rates estimated by this method range from 1.3 to 5.9 in/yr (eight sites, semiarid region, W. Australia; Sharma et al., 1991), 3.1 to 11.9 in/yr (chalk and sandstone aquifers, England; Cooper et al., 1990), and 13.6 to 18.5 in/yr (Upper Chalk aquifer, southern England; Wellings, 1984). The ZFP provides a recharge estimate at the measurement point. Time scales range from event scales to years.

The ZFP technique cannot be used when water fluxes are downward throughout the entire profile or when water storage is increasing because downward movement of a wetting front generally masks the zero flux plane. A simplified water-budget approach is generally used for these conditions (Hodnett and Bell, 1990; Roman et al., 1996). The ZFP technique is relatively expensive in terms of the required instruments and amount of data collection. This technique works best in regions where large fluctuations exist in soil-water content throughout the year and where the water table is always deeper than the ZFP.

Darcy's Law

Darcy's Law is used to calculate recharge (R) in the unsaturated zone according to the following equation:

$$R = -K(q)dH/dz = -K(q)\frac{d}{dz}(h+z) = -K(q)\left(\frac{dh}{dz}+1\right) \quad (6)$$

where *K(q)* is the hydraulic conductivity at the ambient water content, *q*; *H* is the total head, *h* is the matric pressure head, and *z* is elevation. Application of Darcy's Law requires measurements or estimates of the vertical total-head gradient and the unsaturated hydraulic conductivity at the ambient soil-water content. The method has been applied in many studies under arid and semiarid conditions (Enfield et al., 1973; Sammis et al., 1982; Stephens and Knowlton, 1986) and also under humid conditions (Ahuja and El-Swaify, 1979; Steenhuis et al., 1985; Kengni et al., 1994; Normand et al., 1997). For thick unsaturated zones, below the zone of fluctuations related to climate, in uniform or thickly layered porous media, the matric pressure gradient is often nearly zero, and water movement is essentially gravity driven. Under these conditions, little error results by assuming that the total head gradient is equal to 1 (unit-gradient assumption) (Gardner, 1964; Childs, 1969; Chong et al., 1981; Sisson, 1987). The unit-gradient assumption removes the need to measure the matric pressure gradient and sets recharge equal to the hydraulic conductivity at the ambient water content. The unit-gradient assumption has been used in many studies (Sammis et al., 1982; Stephens and Knowlton, 1986; Healy and Mills, 1991; Nimmo et al., 1994).

The minimum recharge rate that can be estimated by using Darcy's Law depends on the accuracy of the hydraulic conductivity and head-gradient measurement if the latter is not unity. Accurate measurements of hydraulic conductivity as low as 1×10^{-9} cm/s can be obtained by using the steady-state centrifuge (SSC) method; this value corresponds approximately to 0.01 in/yr (Nimmo et al., 1992). However, problems with sample disturbance and drying during collection and spatial variability in hydraulic conductivity generally result in a measurement limit of about 0.8 in/yr. Recharge rates determined by the Darcy method range from 1.5 in/yr in an arid region (New Mexico, USA; Stephens and Knowlton, 1986) to about 19.7 in/yr for an irrigated site having a thin unsaturated zone (near Grenoble, France; Kengni et al., 1994). If hydraulic conductivity is strongly dependent on water content, uncertainty increases (Nimmo et al., 1994). This method provides a point estimate of recharge over a wide range of time scales; however, if applied at significant depths in thick vadose zones, it may represent a larger area. An attractive feature of the Darcy method is that it can be applied throughout the entire year, whereas the ZFP can be applied only at certain times of the year.

Tracer Techniques

Applied Tracers

Chemical or isotopic tracers are applied as a pulse at the soil surface or at some depth within the soil profile to estimate recharge (Athavale and Rangarajan, 1988; Sharma, 1989). Infiltration of precipitation or irrigation transports the tracer to depth. Applied tracers provide recharge estimates at a point scale that may or may not apply to much larger scales. Commonly used tracers include bromide, ^{3}H, and visible dyes (Athavale and Rangarajan, 1988; Kung, 1990; Flury et al., 1994; Aeby, 1998; Forrer et al., 1999). Organic dyes are generally used to evaluate preferential flow (Flury et al., 1994; Scanlon and Goldsmith, 1997). Although ^3H is the most conservative of all tracers, its use is prohibited in many areas because of environmental-protection laws. Kung (1990) showed that bromide uptake by plants is often significant, and sorption is important for organic dyes. The subsurface distribution of applied tracers is determined some time after the application by digging a trench for visual inspection and sampling or by drilling test holes for sampling. The vertical distribution of tracers is used to estimate the velocity (v) and the recharge rate (R):

$$R = vq = \frac{\Delta z}{\Delta t} q \tag{7}$$

where Dz is depth of the tracer peak, Dt is time between tracer application and sampling, and q is volumetric water content. The minimum water flux that can be measured with applied tracers depends on the time between application and sampling and, in the case of surface-applied tracers, the root-zone depth. If the rooting depth is assumed to be 2 ft and the average water content 0.2 ft^3/ft^3, a recharge rate of 0.4 ft/yr would be required to transport the applied tracer through the root zone in 1 yr. Lower recharge rates can be measured when the tracers are applied below the root zone. The maximum water flux that can be measured depends on the depth to the water table. Recharge rates resulting from excess irrigation were evaluated by Rice et al. (1986), who used surface-applied bromide to a bare field followed by 18 in of irrigation water for 159 d. The resultant recharge rate was 0.13 in/d, which exceeded that estimated using a water balance by a factor of 5. The discrepancy was attributed to preferential flow. The tritium injection technique (at a depth of 2.3 ft) was used to estimate recharge rates that ranged from 0.25 to 4 in/yr in several basins in South India (Athavale and Rangarajan, 1990). Sharma et al. (1985) applied bromide at the soil surface under natural precipitation in a vegetated area and estimated a recharge rate of 8.8 in for a 76-d period. Tracers are generally applied at a point or over small areas (108 to 2,150 ft^2). The calculated recharge rates represent the time between application and sampling, which is generally months to years.

Historical Tracers

Historical tracers result from human activities or events in the past, such as contaminant spills (Nativ et al., 1995) or atmospheric nuclear testing (^3H and ^{36}Cl) (Fig. 2). These historical tracers or event markers are used to estimate recharge rates during the past 50 yr (Phillips et al., 1988; Scanlon, 1992; Cook et al., 1994). Contaminants from industrial and agricultural sources, such as bromide, nitrate, atrazine, and arsenic, can provide qualitative evidence of recent recharge; however, uncertainties with respect to source location, concentration, timing of contamination, as well as possible nonconservative behavior of contaminants, make it difficult to quantify recharge. The presence of an event marker in water suggests that a component of that water recharged in a particular time period. The peak concentration of thermonuclear tracers can also be used to estimate water flux by using equation 7, where z is approximated by the depth of the tracer peak concentration, q is the average water content above the tracer peak, and t is the time period between the peak tracer fallout and the time that the samples were collected. The minimum recharge rate that can be estimated using thermonuclear tracers is about 0.4 in/yr because of the time required for movement through the root zone (equation 7) (Cook and Walker, 1995). In many areas where these tracers have been used, the bomb-pulse peak is still in the root zone (^{36}Cl, 0.07 in/yr, Norris et al., 1987; ^{36}Cl, 0.1 to 0.12 in/yr, ^3H, 0.25 to 0.4 in/yr, Phillips et al., 1988; ^{36}Cl, 0.06 in/yr, ^3H, 0.3 in/yr, Scanlon, 1992), indicating that water fluxes at these sites are extremely low, which is important for waste disposal. Because much of this water in the root zone is later evapotranspired, water fluxes estimated from tracers within the root zone overestimate water fluxes below the root zone by as much as several orders of magnitude (Tyler and Walker, 1994; Cook and Walker, 1995). Deep penetration of thermonuclear tracers has been found in sandy soils in arid settings (^{3}H, 0.9 in/yr, Dincer et al., 1974; ^{3}H, 0.87 to 1 in/yr, Aranyossy and Gaye, 1992). The maximum water flux that can be estimated may be limited by depth to groundwater. For example, if the average water content is 0.1 m^3/m^3 and the time since peak fallout is 40 yr, a recharge rate of 1.97 in/yr would result in a peak at a depth of 66 ft. Therefore, this technique is generally unsuitable where recharge rates are much greater than 1.96 in/yr. Theoretically the technique could be used for higher recharge rates if the water table were deeper; however, the difficulty of soil sampling at these depths and locating the tracer peak may be prohibitive. Historical tracers provide point estimates of water flux over the last 50 yr.

Environmental Tracers - Chloride

Environmental tracers such as chloride (Cl) are produced naturally in the Earth's atmosphere and are used to estimate recharge rates (Allison and Hughes, 1978; Scanlon, 1991; 2000; Phillips, 1994). The mass of Cl into the system (precipitation and dry fallout, P) times the Cl concentration in P (C_p) is balanced by the mass out of the system (drainage, D) times the Cl concentration in drainage water in the unsaturated zone (C_{uz}) if surface runoff is assumed to be zero:

$$PC_p = DC_{uz} \qquad D = \frac{PC_p}{C_{uz}} \tag{8}$$

Chloride concentrations generally increase through the root zone as a result of evapotranspiration and then remain constant below this depth. Bulge-shaped Cl profiles at some sites have been attributed to paleoclimatic variations or to diffusion to a shallow water table (Fig. 3). Drainage is inversely related to Cl concentration in the unsaturated-zone pore water (equation 8). This inverse relationship results in the Cl mass balance (CMB) approach being much more accurate at low drainage rates because Cl concentrations change markedly over small changes in drainage (Fig. 4). The CMB approach has been most widely used for estimating low recharge rates, largely because of the lack of other suitable methods. Water fluxes as low as 0.002 to 0.004 in/yr have been estimated in arid regions in Australia and in the U.S. (Allison and Hughes, 1983; Cook et al., 1994; Prudic, 1994; Prych, 1998). Low recharge rates are found to be consistent with radioactive decay of ^{36}Cl at a site in the U.S. (Scanlon, 2000). Somewhat higher recharge rates have been calculated from Cl concentrations measured in sinkholes in Australia (>2.4 in/yr; Allison et al., 1985), sand dunes cleared of vegetation in Australia (0.16 to 1.1 in/yr; Cook et al., 1994), and sands with sparse vegetation in Cyprus (1.3 to 3.7 in/yr; Edmunds et al., 1988). The maximum water flux that can be estimated is based on uncertainties in measuring low Cl concentrations and potential problems with Cl contributions from other sources and is generally considered to be about 11.8 in/yr. Scanlon and Goldsmith (1997) reported uncertainties of an order of magnitude beneath ephemeral lakes (playas) in the U.S. because of uncertainties in Cl input from run-on to playas. The CMB approach provides point estimates of recharge rates. Temporal scales range from decades to thousands of years (Scanlon, 2000).

Numerical Modeling

Unsaturated-zone modeling is used to estimate deep drainage below the root zone or recharge in response to meteorological forcing. Recent advances in computer technology and in computer codes have made long-term simulations of recharge more feasible. A variety of approaches is used to simulate unsaturated flow, including soil-water storage-routing approaches (bucket model, Flint et al., 2002; Walker et al., 2002), quasi-analytical approaches (Kim et al., 1996; Simmons and Meyer, 2000), and numerical solutions to Richards' equation. Examples of codes that use Richards' equation include BREATH (Stothoff, 1995), HYDRUS-1D, HYDRUS-2D (Simunek et al., 1998), SWIM (Ross 1990), VS2DT (Lappala et al., 1987; Hsieh et al., 2000) and UNSATH (Fayer, 2000). Theoretically the range of recharge rates that can be estimated using numerical modeling is infinite; however, the reliability of these estimates should be checked against field information, such as lysimeter data, tracers, water content, and temperature (Scanlon and Milly, 1994; Andraski and Jacobson, 2000; Simmons and Meyer, 2000, Flint et al., 2002). Bucket-type models can be used over large areas (Flint et al., 2002); however, models based on Richards' equation are often restricted to evaluating small areas ($\leq$ 1075 ft^2) or to 1-D flow in the shallow subsurface ($\leq$ 50 ft depth). Time scales that can be evaluated range from hours to decades; however, many recharge modeling studies evaluate periods up to 30 to 100 yr because of availability of meteorological information (Rockhold et al., 1995; Stothoff, 1997; Kearns and Hendrickx, 1998). Because of uncertainties in hydraulic conductivity and nonlinear relationships between hydraulic conductivity and matric potential or water content, recharge estimates based on unsaturated-zone modeling that use Richards' equation may be highly uncertain. Numerical modeling is generally used as a tool to evaluate flow processes and to assess sensitivity of model output to various parameters. Stothoff (1997) evaluated the impact of alluvial cover thickness overlying fractured bedrock on recharge and found high recharge rates ($\leq$50% of precipitation) if alluvial-cover thicknesses were less than 9.8 to 20 in and little or no recharge for alluvial cover thicknesses between 20 and 197 in. The effect of soil texture and vegetation was evaluated by Rockhold et al. (1995) for a 30-yr period (1963 to 1992). Recharge rates range from 0.02 in/yr (sagebrush on sand) to 0.87 in/yr (bare sand). Recharge rates also varied with soil texture (0.3 in/yr, bare silt loam, to 0.87 in/yr, bare sand). A similar study was conducted for a 100-yr period by Kearns and Hendrickx (1998), who also demonstrated the effects of vegetation and soil texture on recharge rates.

RECHARGE ESTIMATION TECHNIQUES BASED ON SATURATED-ZONE STUDIES

Most unsaturated-zone techniques provide point estimates of recharge; saturated-zone techniques commonly integrate over much larger areas. Whereas surface-water and unsaturated-zone approaches provide estimates of drainage or potential recharge, saturated-zone approaches provide evidence of actual recharge because water reaches the water table.

Physical Techniques

Water Table Fluctuation Method

The Water Table Fluctuation (WTF) method is based on the premise that rises in groundwater levels in unconfined aquifers are due to recharge water arriving at the water table. Recharge is calculated as

$$R = S_y \, dh/dt = S_y \, \Delta h/\Delta t \qquad (9)$$

where S_y is specific yield, h is water-table height, and t is time. The effect of regional groundwater discharge is taken into account by extrapolating the antecedent water level recession beneath the peak water level. The WTF method has been used in various studies (Meinzer and Stearns 1929; Rasmussen and Andreasen, 1959; Gerhart, 1986; Hall and Risser, 1993) and was described in detail by Healy and Cook (2002). The method is best applied over short time periods in regions having shallow water tables that display sharp rises and declines in water levels. Analysis of water-level fluctuations can, however, also be useful for determining the magnitude of long-term changes in recharge caused by climate or land-use change. Difficulties in applying the method are related to determining a representative value for specific yield and ensuring that fluctuations in water levels are due to recharge and are not the result of changes in atmospheric pressure, the presence of entrapped air, or other phenomena, such as pumping. The method has been applied over a wide variety of climatic conditions. Recharge rates estimated by this technique range from 0.2 in/yr in the Tabalah Basin of Saudi Arabia (Abdulrazzak et al., 1989) to 9.7 in/yr in a small basin in a humid region of the eastern U.S. (Rasmussen and Andreasen, 1959). Water-level fluctuations occur in response to spatially averaged recharge. The area represented by the recharge rates ranges from tens of square meters to several hundred or thousand square meters. Time periods represented by the recharge estimates range from event scale to the length of the hydrographic record.

Darcy's Law is used to estimate flow through a cross section of an unconfined or confined aquifer. This method assumes steady flow and no water extraction. The subsurface-water flux (q) is calculated by multiplying the hydraulic conductivity by the hydraulic gradient. The hydraulic gradient should be estimated along a flow path at right angles to potentiometric contours. The volumetric flux through a vertical cross section of an aquifer (A) is equated to the recharge rate (R) times the surface area that contributes to flow (S):

$$qA = RS \tag{10}$$

The cross section should be aligned with an equipotential line. The Darcy method, which has been used by Theis (1937) and Belan and Matlock (1973), is easy to apply if information on large-scale, effective hydraulic conductivity and the hydraulic gradient is available. The area should reflect a natural system with minimal pumpage. Recharge estimates based on Darcy's Law are highly uncertain because of the high variability of hydraulic conductivity (several orders of magnitude). The applicability of laboratory-measured hydraulic conductivities at the field scale is also questionable. This technique can be applied to large regions (~0.4 to ≥4,000 mi^2). The time periods represented by the recharge estimates range from years to hundreds of years.

Tracer Techniques

Groundwater Dating

Historical tracers or event markers such as bomb-pulse tritium (^3H) are used in both unsaturated and saturated zones to estimate recharge. Tritium has been used widely in the past (Egboka et al., 1983; Robertson and Cherry, 1989); however, bomb-pulse ^3H concentrations have been greatly reduced as a result of radioactive decay. In the southern hemisphere, where ^3H concentrations in precipitation were an order of magnitude lower than in the northern hemisphere (Allison and Hughes, 1977), it is often difficult to distinguish bomb-pulse ^3H from current ^3H concentrations in precipitation. The use of ^3H to date groundwater is generally being replaced by the use of tracers such as chlorofluorocarbons (CFCs) and tritium/helium-3 ($^3H/^3He$). These gas tracers can be used only as water tracers in the saturated zone, where they can no longer exchange with the atmosphere. The first appearance of tracers such as CFCs or $^3H/^3He$ can be used to estimate recharge rates where flow is primarily vertical, as in recharge areas near groundwater divides. Recharge rates can also be determined by estimating ages of groundwater, age being defined as the time since water entered the saturated zone. Groundwater ages are readily estimated from CFCs by comparing CFC concentrations in groundwater with those in precipitation (Fig. 2). The age of the groundwater, *t*, is calculated from $^3H/^3He$ data using the following equation:

$$t = -\frac{1}{l}\ln\left[1 + \frac{^3He_{trit}}{^3H}\right] \quad (11)$$

where l is the decay constant ($\ln 2/t^{1/2}$), $t^{1/2}$ is the ^3H half life (12.43 yr), and $^3He_{trit}$ is tritiogenic ^3He. Use of this equation assumes that the system is closed (does not allow ^3He to escape) and is characterized by piston flow (no hydrodynamic dispersion).

In unconfined porous-media aquifers, groundwater ages increase with depth, the rate of which depends on aquifer geometry, porosity, and recharge rate (Cook and Bohlke, 2000). The vertical groundwater velocity decreases with depth to zero at the lower boundary of the aquifer. The age increases linearly with depth near the water table and nonlinearly at greater depths. Near the water table, the influence of the aquifer geometry is greatly reduced. The recharge rate can be determined by dating water at several points in a vertical profile, calculating the groundwater velocity by inverting the age gradient, extrapolating the velocity to the water table if it is not measured near the water table (Cook and Solomon, 1997), and multiplying the velocity by the porosity for the depth interval (which is similar to equation 7).

The range of recharge rates that can be estimated by using groundwater dating depends on the ranges of ages that can be determined. CFCs and $^3H/^3He$ are used to determine groundwater ages up to approximately 50 yr, with a precision of 2 to 3 yr (Cook and Solomon 1997). Radioactive decay of ^{14}C can be used to estimate groundwater ages between 200 and 20,000 yr. The estimated recharge rates are average rates over the time period represented by the groundwater age. Groundwater recharge rates between 3.9 and 39 in/yr have been determined using $^3H/^3He$ (Schlosser et al. 1989; Solomon et al. 1995) and CFCs (Dunkle et al., 1993; Cook et al. 1998). Recharge rates of much less than 1.2 in/yr are difficult to determine accurately using these tracers because of problems associated with diffusion of ^3He into the unsaturated zone (Cook and Solomon, 1997) and the difficulty of obtaining discrete samples near the water table. Dispersive mixing can result in ± 50% uncertainty in $^3H/^3He$ ages prior to 1970, when ^3H input varied markedly during the bomb pulse (Solomon and Sudicky, 1991). Recharge rates between 0.004 and 3.9 inches/yr have been determined using ^{14}C (Vogel 1967; Leaney and Allison, 1986; Verhagen 1992), although diffusional transport at very low recharge rates probably means a lower limit to the method of about 0.04 in/yr (Walker and Cook, 1991). The method is most accurate where piezometers have been completed with relatively short well screens. Recharge rates calculated using groundwater dating spatially integrate recharge over an area upgradient from the measurement point. Therefore, spatial scales can range from local (decameter scale) if samples are collected near a groundwater divide (Szabo et al., 1996) to regional (kilometer scale; Pearson and White, 1967).

Horizontal flow velocities can be estimated from radioactive decay of ^{14}C or ^{36}Cl in a confined aquifer. These data can be used to estimate recharge rates (R):

$$R = vnA/S \quad (12)$$

where v is velocity, n is porosity, A is the cross-sectional area of the confined aquifer where the velocity is determined, and S is the surface area of the recharge zone. If necessary, corrections should be made for any leakage to or from the confined aquifer. Using this method, recharge rates of ~ 2 in/yr were estimated for the Carrizo aquifer in Texas on the basis of ^{14}C data from Pearson and White (1967), assuming no vertical leakage to the confined aquifer.

There are various restrictions to the use of these tracers to date groundwater. CFCs can be used only in rural areas not affected by septic tanks because of contamination associated with industrial and residential areas, whereas $^{3}H/^{3}He$ can be used in contaminated and uncontaminated areas. The concentrations of $^{3}H/^{3}He$ and CFCs at the water table are assumed to be equal to those in the atmosphere. Any difference in concentrations would result in errors in the estimated ages. Cook and Solomon (1995) concluded that errors are negligible if the unsaturated-zone thickness is ≤ 33 ft (≤ 10 m). Other issues that need to be considered when using these tracers include the effect of excess air for both $^{3}H/^{3}He$ and CFCs and the effect of recharge temperature, sorption, and degradation on CFCs. Sampling for $^{3}H/^{3}He$ and CFCs is complex, and analysis is relatively expensive.

Environmental Tracers - Chloride

The chloride mass balance (CMB) approach can be used in the unsaturated and saturated zones to estimate groundwater recharge. If sources, in addition to precipitation, contribute chloride to the system, the chloride input from these sources needs to be quantified to use the CMB approach. The CMB approach was originally applied in the saturated zone by Eriksson and Khunakasem (1969) to estimate recharge rates (1.2 to 12.8 in/yr) on the Coastal Plain of Israel. Recharge rates estimated from groundwater Cl concentrations range from 0 to 0.31 in/yr in South Africa (Sami and Hughes 1996), 0.43 in/yr in the Southern High Plains, US (Wood and Sanford, 1995), 0.5 to 3.9 in/yr in southwestern Australia (Johnston 1987), and 5.9 to 26 in/yr in northeast Australia (Cook et al., 2001). Recharge rates based on groundwater Cl by Johnston (1987) are up to two orders of magnitude greater than those based on Cl in the unsaturated-zone pore water. The discrepancy between the two rates is attributed to preferential flow. Slightly higher recharge rates can be estimated using Cl in groundwater than they can in soil water because extraction of water from the soil generally requires additional dilution. The CMB approach spatially integrates recharge over areas upgradient from the measurement point. Spatial scales range from ~ 655 ft (Harrington et al., personal communication) to several km (Wood and Sanford, 1995). The time scales range from years to thousands of years.

Numerical Modeling

Groundwater recharge was estimated previously by graphical analysis of flow nets for both unconfined and confined aquifers (Cedegren, 1989); however, this approach has largely been replaced by groundwater flow models. Groundwater model calibration or inversion is used to predict recharge rates from information on hydraulic heads, hydraulic conductivity, and other parameters (Sanford, 2002). Because recharge and hydraulic conductivity are often highly correlated, model inversion using hydraulic-head data only is limited to estimating the ratio of recharge to hydraulic conductivity (Fig. 5). The reliability of the recharge estimates depends on the accuracy of the hydraulic-conductivity data. Because hydraulic conductivity ranges over several orders of magnitude, estimation of recharge rates using model calibration may not be very accurate. The estimated recharge may be nonunique because the same distribution of hydraulic heads can be produced with a range of recharge rates, as long as the ratio of recharge to hydraulic conductivity remains the same (Fig. 5). Recharge and hydraulic conductivity are fixed for steady-state simulations, whereas transient simulations reproduce temporal variations in recharge that further constrain the recharge estimates.

Recent studies have used joint inversions that combine hydraulic heads and groundwater ages to further constrain inverse modeling of recharge (Reilly et al., 1994; Szabo et al., 1996; Portniaguine and Solomon, 1998). Manual trial-and-error procedures or automated procedures, which use nonlinear regression between measured and simulated data, are used. Whereas hydraulic heads are sensitive to the ratio of recharge to hydraulic conductivity, groundwater ages are sensitive to the ratio of recharge to porosity (Portniaguine and Solomon, 1998). Use of both head and age data provides constraints on recharge, hydraulic conductivity, and porosity. Because these three parameters are highly correlated, a unique solution requires information on one of these parameters. Porosity generally varies much less than recharge or hydraulic conductivity; therefore, porosity can be estimated for the system (Portniaguine and Solomon, 1998). Automated inversions provide information on the nonuniqueness of the solutions. Joint inversions were used to estimate zonal recharge rates that range from 0.39 to 79 in/yr at a site in the U.S. (Portniaguine and Solomon, 1998). Spatial scales are generally much greater than those for unsaturated-zone modeling and range from several ft^2 to 386,100 mi^2 or greater. Time scales generally range from days to 100 yr because of the availability of hydrologic data.

Mixing-cell models (compartment models, lumped models, and black-box models) have been used to delineate sources of recharge and estimate recharge rates on the basis of chemical and isotopic data. The hydrologic system is treated as a series of interconnected cells or compartments, which are fully mixed internally. Each cell can have more than one input and output. Fluxes between cells are varied iteratively until a good fit between measured and simulated hydrologic, chemical, and/or isotopic data is obtained. An estimate of the mean recharge to the system is calculated by dividing the volume of the system by the mean residence time. Allison and Hughes (1975) used a mixing cell model based on conservation of ^3H to assess the relative contribution of lateral inflow from a mountain range and recharge through the unsaturated zone. Yurtsever and Payne (1986) used a nine-compartment mixing-cell model having shallow, intermediate, and deep reservoirs to reproduce discharge and ^3H concentrations in a large karst system in southern Turkey. Hydrochemical and isotopic data were used by Adar et al. (1992) to define a mixing-cell model in the Arava Valley, Israel. Multivariate cluster analysis was used to define recharge sources and delineate mixing cells. Mass-balance equations were developed for each cell on the basis of conservation of water, dissolved chemical species, and isotopes. These equations were solved simultaneously for unknown recharge rates into the various cells.

COMPARISON OF RANGE OF RECHARGE RATES AND SPATIAL AND TEMPORAL SCALES OF THE VARIOUS TECHNIQUES

The various techniques for quantifying recharge differ in the range of recharge rates that they estimate (Fig. 6) and the space and time scales they represent (Figs. 7, 8). The range of recharge rates estimated with a particular technique should be evaluated on a site-specific basis by conducting detailed uncertainty analyses that include uncertainties in the conceptual model and in the input and output parameters. The ranges shown in Fig. 6 are based primarily on measured ranges from the literature discussed previously and provide some indication of possible ranges for each technique. Numerical-modeling approaches can generally be used to estimate any range in recharge rates; however, the reliability of these recharge estimates should be evaluated in terms of the uncertainties in the model parameters. Some techniques have definite restrictions on the recharge rates that they can estimate. Surface-applied and historical tracers in the unsaturated zone require a minimum recharge rate to transport the tracers through the root zone. In addition, historical tracers in the saturated zone, such as $^{3}H/^{3}He$, require a minimum recharge rate of ~1.2 in/yr to confine the ^{3}He. Use of environmental tracers, such as Cl, is one of the few techniques that can estimate very low recharge rates and is generally more accurate in this range. The upper range of recharge rates shown for the various techniques (Fig. 6) generally reflects the measured rates in the literature and may not reflect a true upper limit for the technique. Upper recharge limits for applied and historical tracers in the unsaturated zone may reflect limitations of the thickness of the unsaturated zone or the ability to locate these tracers at depth. In many cases, where recharge rates are high, the unsaturated zone is not very thick. Analytical uncertainties in Cl measurements and uncertainties in Cl inputs restrict the upper range of recharge rates that can be estimated with the CMB technique.

The surface areas represented by the recharge estimates vary markedly among the different techniques (Fig. 7). In general, many techniques based on unsaturated-zone data provide point estimates or represent relatively small areas, whereas some of the surface-water techniques and many of the groundwater approaches represent much larger areas. Surface-water techniques such as seepage meters and heat tracers provide point estimates of recharge. Watershed modeling can be used to estimate recharge rates over a large range of scales, as shown by previous studies (up to 193,050 mi^2 [500,000 km^2]; Arnold et al., 2000). Although many of the unsaturated-zone techniques provide point estimates of recharge, such estimates may represent much larger areas, as shown by comparisons of point estimates from several basins (Phillips 1994) and by relating point data to geomorphic settings (Scanlon et al. 1999a) in the southwestern US. In addition, electromagnetic induction has proved to be a useful tool to regionalize point estimates (Cook et al., 1992; Scanlon et al., 1999b). Saturated-zone studies spatially integrate recharge fluxes over large areas. This spatial integration is important for water-resource assessments where large-scale estimates of recharge are often required. Using tracers to date water near groundwater divides may provide estimates of local recharge rates.

The time scales represented by recharge rates are variable (Fig. 8). Many surface-water approaches provide recharge estimates on short time scales (event scales), and estimates over longer time scales are obtained by summing those from individual events. Unsaturated-zone techniques, such as lysimeters, zero flux plane, and applied tracers, and saturated-zone techniques, such as water-table fluctuations, provide recharge estimates on event time scales also. These techniques are restricted to providing recharge estimates for the length of the monitoring record. Numerical-modeling approaches can be used to predict recharge over any time scale; however, recharge estimation based on climatic data is generally restricted to about 100 yr. The only techniques that can provide integrated, long-term estimates of recharge are tracers such as ^{36}Cl, ^{3}H, $^{3}H/^{3}He$, CFCs, ^{14}C, and Cl. Tracers are very useful for estimating net recharge over long time periods but generally do not provide detailed time series information on variations in recharge.

APPLICATION OF MULTIPLE TECHNIQUES

Because of uncertainties associated with each approach for estimating recharge, the use of many different approaches is recommended to constrain the recharge estimates. In many cases, different approaches complement each other and help refine the conceptual model of recharge processes. Examples of multiple approaches include the use of various tracers in unsaturated zones (e.g., Cl, ^{36}Cl, and ^{3}H; Scanlon, 1992; Cook et al. 1994; Nativ et al., 1995; Tyler et al., 1996; Prych, 1998) and saturated zones (CFC-11, CFC-12, $^{3}H/^{3}He$; Ekwurzel et al., 1994; Szabo et al., 1996). Other studies have combined soil physics and environmental tracers (Scanlon et al., 1999a) and also numerical modeling (Scanlon and Milly, 1994; Fayer et al., 1996).

Ideally as many different approaches as possible should be used to estimate recharge. Techniques based on data from surface water and unsaturated and saturated zones can also be combined. Sophocleous (1991) showed how unsaturated-zone water-balance monitoring could be combined with water-table fluctuations to increase the reliability of recharge estimates. Some studies have used catchment-scale surface-water models to provide estimates of recharge to groundwater models (Davies-Smith et al., 1988; Handman et al., 1990); however, such an approach assumes that no time lag occurs between infiltration and groundwater recharge. More recently, surface-water and groundwater models have been integrated, such as the SWAT and MODFLOW codes by Sophocleous and Perkins (2000). This integrated model provides a framework for the total system that can be used to check continuity and better constrain model parameters. Parameter optimization is conducted by calibrating against multiple targets, such as groundwater levels, stream-flow data, and other data, that should result in more reliable results than obtained when using watershed or groundwater models separately. In this integrated model, recharge is constrained by an overall water budget for the surface-water system, and stream-aquifer interactions are constrained by the watershed model.

RECHARGE RATES FOR THE MAJOR AQUIFERS BASED ON REVIEW OF EXISTING DATA

A database was developed that compiles existing information on recharge rates in the State. Table 1 contains recharge rates for eight major aquifers, including the Carrizo-Wilcox, Cenozoic Pecos Alluvium, Edwards-Trinity, Gulf Coast, Hueco-Mesilla Bolson, Ogallala, Seymour, and Trinity. Recharge to the Edwards aquifer is much more dynamic than recharge to the other major aquifers and cannot readily be represented as a single value in Table 1. Recharge rates for the Edwards aquifer can be found in Slattery et al. (1998) and in annual reports published by the Edwards Aquifer Authority (e.g., 2000). Recharge data were compiled from reports published by the Texas Water Development Board, U.S. Geological Survey, and other publications. The table lists the study areas (counties or general area), underlying aquifers, recharge rates (units of mm/yr, in/yr, or total recharge in acre-feet/yr), data sources, and techniques used to estimate recharge. Additional notes are provided in some cases. The full reference citations are listed separately.

Estimates of recharge rates in the Carrizo-Wilcox aquifer range from 0.1 to 5.8 in/yr. The higher recharge rates occur in the sandy portions of the aquifer (i.e., Carrizo and Simsboro Formations). Recharge rates are generally lower in the Gulf Coast aquifer, ranging from 0.0004 to 2 in/yr. In both the Carrizo-Wilcox and Gulf Coast aquifers, higher recharge rates are in upland areas with sandy soils. Regional recharge rates in the High Plains aquifer, outside irrigated areas, are generally low (0.004 to 1.7 in/yr), whereas playa-focused recharge rates are much higher (0.5 to 8.6 in/yr). Irrigated areas also have fairly high recharge rates (0.6 to 11 in/yr). Recharge rates in the Trinity and Edwards-Trinity aquifers generally range from 0.1 to 2 in/yr. The Seymour aquifer has recharge rates that range from 1 to 2.5 in/yr. Recharge rates for the Hueco-Mesilla Bolson and the Cenozoic Pecos Alluvium are represented as total recharge along mountain fronts and valley floors.

EVALUATION OF TECHNIQUES USED TO QUANTIFY RECHARGE IN THE MAJOR AQUIFERS

The main techniques for estimating recharge are Darcy's Law, groundwater modeling, and base-flow discharge. Darcy's Law is widely applied in the confined sections of the Carrizo-Wilcox and Gulf Coast aquifers. Groundwater modeling is used in most aquifers. Base-flow discharge is used primarily in the Cenozoic Pecos Alluvium, Edwards-Trinity Plateau, Seymour, and Trinity aquifers. Base-flow discharge, however, is not necessarily directly equated to recharge because pumpage, evapotranspiration, and underflow to deep aquifers may also be significant. In some cases base-flow discharge is calculated for the winter period to minimize the effect of evapotranspiration (Price, 1978; Preston 1978). Rutledge (1998) recommended an upper limit on basin size of 500 mi^2 for application of this method because of difficulties in separating surface-water and groundwater flow and bank-storage effects in larger systems and because of the areally uniform recharge assumption. Some estimates are from contributing areas that far exceed the recommended 500-mi^2 area (e.g., Iglehart 1967; 3,800-mi^2 area). Bank storage effects may also complicate recharge estimates using base-flow discharge, e.g., base-flow discharge estimates along the Brazos River adjacent to the Seymour aquifer. The time period over which base-flow discharges are calculated also affects the recharge estimate. Higher recharge estimates for the 1974 through 1977 period by Kuniansky (1989) are attributed to higher precipitation during that time. Extending the time period from 1940 through 1960 (Ashworth 1983) to 1997 (Mace et al., 2000) increased the recharge estimate from 1.3 to 2.2 in/yr. Some recharge estimates are based on very short flow records, e.g., several days (Price et al. 1978; Preston 1978). In many areas, the density of gauging stations is insufficient to reliably estimate recharge using base-flow discharge. Environmental tracers have only been used to a limited extent (chloride mass balance, tritium, and carbon-14) to estimate recharge. The chloride mass balance approach has been used to provide a regional estimate of recharge to the northern segment of the Ogallala in Texas (Wood and Sanford, 1995). The recharge estimate for this region may be affected by irrigation return flow; however, the estimate is considered fairly reliable. Recharge estimates beneath playas using chloride data are highly uncertain because of uncertainties in the chloride input to the system (Scanlon and Goldsmith, 1997). Tritium data beneath a playa and in groundwater provide a fairly accurate estimate of recharge in the southern Ogallala aquifer (Nativ, 1988; Wood and Sanford, 1995). Carbon-14 data from the Carrizo Wilcox aquifer provide a reliable estimate of groundwater velocity in the confined section of the aquifer. These data can provide an upper bound on the recharge for the outcrop region by assuming no leakage through the confining layer. Neutron-probe logging has been used in the Ogallala aquifer to quantify recharge in irrigated and nonirrigated regions (Klemt, 1981). Water-content changes below 10-ft depth were used to estimate recharge; however, water-content changes at these depths were too low to be accurately monitored with a neutron probe. Comparison of water-content data between nearby irrigated and nonirrigated regions to estimate the depth of wetting fronts in irrigated regions assumes that the soil texture in the profiles is uniform. This is unlikely to be the case in these systems, making it difficult to compare water-content data because water content will vary with texture. Water-budget approaches have been used in some aquifers to quantify recharge; however, recharge is generally the smallest term in the water-budget equation; therefore, 5% to 10% uncertainties in the various terms of the water-budget equation can result in errors in the recharge estimate of more than 100%.

CONCEPTUAL MODELS OF RECHARGE PROCESSES FOR THE MAJOR AQUIFERS

Development of a conceptual model for recharge for the major aquifers is critical to understanding recharge processes and rates, although understanding the sources of recharge and the spatial and temporal variability in recharge is basic to developing a conceptual model of recharge. Distribution of the major aquifers is shown in Fig. 9. Potential sources of recharge include

- precipitation, including rainfall and snowmelt;
- return flow from irrigation, where more water is applied than consumed by evapotranspiration (ET);
- surface water (rivers, lakes, floods); and
- cross formational flow.

Climate ranges from humid to arid in Texas (Larkin and Bomar 1983). Mean annual precipitation decreases from a maximum of about 56 in/yr in East Texas to a minimum of about 8 in/yr in West Texas (Fig. 10). The seasonal distribution of precipitation is also an important factor controlling recharge. Winter precipitation is generally much more effective than that of other seasons for recharging underlying aquifers because vegetation is normally dormant during this time (Nativ and Riggio 1989).

In the humid regions of Texas, precipitation is the dominant source of recharge, which is fairly uniform spatially in interstream areas. Irrigation is generally negligible in humid regions. Also, much of the recharge across the outcrop of an unconfined aquifer ends up discharging (providing base flow) to rivers and streams that cross the outcrop. Only a fraction of recharge reaches the deeper, confined part of the aquifer.

Precipitation, of course, is much lower in semiarid and arid regions than in humid regions, and much of the infiltrated water in the soil or unsaturated zone is evapotranspired before it can recharge groundwater at the water table. The thickness of the unsaturated zone is generally much greater than in humid regions. Recharge from precipitation across interstream areas (diffuse recharge) is generally much lower than it is from other sources. Recharge in semiarid and arid regions is primarily from surface water, such as streams, lakes, and playas, irrigation return flow, and other local sources of perennial wetness.

Evaporation is also important in controlling recharge (Fig. 11). Pan evaporation in Texas ranges from a maximum of 45 in in East Texas to a maximum of 81 in/yr in West Texas. These are the maximum amounts of evaporation that would occur if water were available—that is, exposed at or above ground surface such as at surface-water reservoirs. Evaporation from bare or vegetated soils and transpiration by plants may be less than pan evaporation. The greatest evaporation rates occur during summer months. Although annual pan evaporation cannot be subtracted directly from precipitation to estimate recharge, pan-evaporation values indicate that there is a higher potential for evaporation in the semiarid and arid regions of the state. In areas where annual precipitation is much less than pan evaporation, it is easy to see that reservoirs must rely on inflows of surface-water runoff or groundwater discharge.

The primary controls on recharge include climate (precipitation, evaporation), vegetation (plant transpiration), land use (impervious cover and irrigation), soil thickness and physical properties, and bedrock type (porous or fractured). Recharge generally increases with increased precipitation. Seasonal distribution in precipitation may be more important than the average annual precipitation because winter precipitation is much more effective in recharging groundwater than summer precipitation. Many think that if average annual potential evaporation is much greater than precipitation, there should be no groundwater recharge. However, the time scale of the calculations is important. Use of long time scales, such as yearly or monthly, can lead to an underestimation of recharge. Water-budget estimates should be conducted using daily or hourly data because precipitation can greatly exceed evapotranspiration on these short time scales and result in episodic recharge. Recharge is generally much greater in nonvegetated than in vegetated regions (Gee et al. 1994) and greater in areas of annual crops and grasses than in areas of trees and shrubs (Prych 1998). The impact of vegetation was clearly demonstrate in Australia, where replacement of deep-rooted native Eucalyptus trees with shallow-rooted crops resulted in recharge increases of about two orders of magnitude (<0.004 in/yr) for native mallee vegetation to 0.2 to 1.2 in/yr for crop/pasture rotations) (Allison et al. 1990). Brush-control projects are being conducted in Texas to increase recharge in the Concho valley near San Angelo, Texas (Dugas et al., 1998). The effectiveness of brush control in increasing recharge is questionable. Results from previous studies demonstrated that ET was only reduced during the first 2 years after brush was removed (Dugas et al., 1998). Therefore, information on land use/land cover is also important for evaluating recharge. Irrigated areas should be identified as well because irrigation return flow can contribute significant amounts of recharge. Impervious cover may increase recharge because runoff from such covers can focus flow and more effectively recharge the underlying aquifer. Soil texture and permeability, too, are important because coarse-grained soils generally result in higher recharge rates than do fine-grained soils. Cook et al. (1992) noted an apparent negative correlation between clay content in the upper 6.6 ft and the recharge rate. Thick soils generally provide large storage capacity for infiltrated water and allow the water to remain near the soil surface where it can readily be evapotranspired. In contrast, bare rock, particularly fractured rock, allows water to move rapidly through fractures and minimizes evapotranspiration.

Recharge is a critical part of the water budget of an aquifer. As previously stated, part of the water recharged to an unconfined aquifer is discharged to surface water, transpired by plants, or withdrawn by wells across the outcrop of the aquifer, and only a fraction of recharge ends up moving downdip to the confined part of the aquifer. The water that moves downdip eventually discharges by seeping into overlying aquifers. With increased groundwater development in unconfined and confined sections of aquifers and lowering of water tables and piezometric heads, groundwater discharge through base flow to streams and evapotranspiration should decrease. As groundwater development continues, streams may change from gaining to losing and provide a source of recharge to the groundwater. Computer models of future conditions should account for spatially variable groundwater recharge, evapotranspiration that will vary with water-table elevation, surface water–groundwater interactions that may vary as water tables are lowered, and water withdrawal by wells in both the unconfined and confined parts of aquifers. Accurate simulation of the impact of increased groundwater withdrawals through pumpage on surface-water–groundwater interactions (i.e., gaining to losing streams) is critical to water-resource prediction in aquifer systems.

CARRIZO-WILCOX AQUIFER

The Carrizo-Wilcox aquifer consists of sand interbedded with gravel, silt, clay, and lignite deposited during the Tertiary period. The formations of the Carrizo-Wilcox aquifer mainly crop out in a band that is 10 to more than 20 mi wide and parallel to the Gulf Coast; another extensive outcrop straddles the Texas-Louisiana border (Fig. 9). The formations and the aquifers they contain dip toward the coast or into the East Texas Basin beneath younger sediments. In the central part of the Texas coastal plain between the Trinity and Colorado Rivers, the Simsboro Sand is a distinct, mappable formation as much as 400 ft thick within the Wilcox Group. Here the Simsboro and Carrizo are the main units of the Carrizo-Wilcox aquifer, although groundwater is also withdrawn from other parts of the aquifer system. To the north of the Trinity River and south of the Colorado River, the thickness of the Simsboro Sand is much less, and that unit of the aquifer is neither recognized nor mapped separately from the rest of the Carrizo-Wilcox aquifer system. Following this geological pattern, the aquifer has been divided into northern, central, and southern zones for the purposes of the Groundwater Availability Modeling program.

Precipitation in the north and central parts of the Carrizo-Wilcox aquifer ranges from approximately 30 to 56 in/yr. Diffuse recharge occurs primarily from precipitation across interstream areas, and soil type probably also affects the spatial distribution of recharge. Soil type varies with the underlying geologic formations; i.e., sandy soils overlie the Carrizo and Simsboro Formations, and more clay-rich soils are developed on the predominantly clayey Hooper and Calvert Bluff Formations. Also, clay-rich "hardpan" soil horizons occur in parts of the outcrop, developed naturally by weathering of surficial sandy soils. Much of the groundwater in the outcropping, unconfined part of the aquifer, recharged at the water table, is discharged as base flow to streams and rivers and through evapotranspiration by phreatophytic vegetation over shallow water tables adjacent to streams.

The south part of the Carrizo-Wilcox differs from the central and north parts in a number of ways. Precipitation is lower in the south part and ranges from about 30 in/yr between the Colorado and San Antonio Rivers to about 20 in/yr on the south edge of the aquifer (Maverick and Zavala Counties). Diffuse areal recharge results from precipitation and irrigation return flow. The streams are losing rather than gaining in this part of the aquifer. Focused recharge from stream-channel and flood flows also constitutes a significant component of the total recharge (L.B.G. Guyton and Associates and H.D.R. Engineering, Inc., 1998).

Most municipal and industrial pumpage has occurred in the northeast (e.g., near Tyler, Lufkin–Nacogdoches and Bryan–College Station [≥ 400 ft of drawdown]) (Ashworth and Hopkins 1995). A good deal of pumping for agricultural purposes has also occurred in the south part of the aquifer. In the future, large decreases in water levels where rivers cross the aquifer outcrop may change the classification of those river segments from gaining to losing. In these cases, surface water would become a source of aquifer recharge.

CENOZOIC PECOS ALLUVIUM AQUIFER

The Cenozoic Pecos Alluvium aquifer is located in the upper part of the Pecos

River valley where ground surface slopes gently toward the Pecos River. The aquifer consists of as much as 1,500 ft (457 m) of alluvial fill in two separate basins: the Pecos Trough to the west and the Monument Draw Trough to the east (Ashworth and Hopkins 1995). Although groundwater is generally unconfined to semiconfined, it may exhibit confined conditions locally. Depth to water early in the 20th century averaged 50 ft (15 m) below ground surface; locally there has been as much as 250 ft (76 m) of drawdown owing to well withdrawal in areas where irrigation has been intensive. Groundwater flow is generally toward the Pecos River, although locally it may move toward areas of high irrigation pumpage. Precipitation ranges from 10 to 14 in/yr in the region (Larkin and Bomar, 1983). Recharge occurs by infiltration of precipitation (primarily in sand dunes in Winkler and northeastern Ward Counties), seepage from ephemeral stream channels with headwaters in the Rustler Hills and Davis Mountains, irrigation return flow, and subsurface cross-formational flow from adjacent systems (Edwards-Trinity aquifer, Rustler Formation, and volcanic aquifers [Ashworth, 1990])). Intensive groundwater development has reduced base-flow discharge to the Pecos River.

EDWARDS AQUIFER

The Edwards (Balcones Fault Zone, BFZ) aquifer lies in a narrow belt that stretches from Bell County southwestward to Kinney County. The aquifer consists of Cretaceous-age marine carbonates (limestone and dolomite). The aquifer is unconfined in the outcrop area and under artesian conditions, where it is confined by the Del Rio Clay (Ashworth an Hopkins 1995). Recharge occurs primarily from losing streams in the outcrop area and to a lesser extent directly from precipitation in interstream areas. For the Central Texas area, Slade et al. (1985) estimated 85% of recharge is in streambeds, and 15% is across uplands. The contributing area of the streams extends to the west and north from the outcrop area. As much as 60,000 acre-feet/yr may enter the Edwards in the subsurface from the underlying Trinity aquifer in the Hill Country (Mace et al., 2000). Recharge to the Edwards from the losing rivers has been extensively monitored and studied providing recharge estimates as early as 1934 (Edwards Aquifer Authority annual hydrologic report). The methodology for estimating recharge is currently being reevaluated and revised (Ozuna, 2001, personal communication). Groundwater discharges to major springs and to wells through pumping. The aquifer is extremely dynamic, and groundwater generally has a short residence time that can range from days to several hundred years (Campana and Mahin 1985). Artificial recharge is also being used to increase groundwater recharge. Recharge structures are being installed to divert surface water to recharge features during high flow conditions.

EDWARDS-TRINITY PLATEAU AQUIFER

The Edwards-Trinity (Plateau) aquifer extends from the Hill Country in Central Texas to the Trans-Pecos region in West Texas. The aquifer consists of the Glen Rose Limestone (south part of the plateau) and the Antlers Sand (north part of the plateau) of the Lower Cretaceous-age Trinity Group overlain by limestones and dolomites of the Comanche Peak, Edwards, and Georgetown Formations (Ashworth and Hopkins 1995). The aquifer is generally unconfined, although the aquifer may be confined in the Trinity when overlain by low-permeability formations at the base of the Edwards. Regionally, the base of the Cretaceous slopes to the south and southeast; however, groundwater locally moves toward the major streams. Precipitation ranges from 10 in/yr in Culberson County in the west to 32 in/yr in Blanco County in the east. Lake evaporation decreases from 90 in/yr in the west to approximately 50 in/yr in the east. Recharge occurs primarily from precipitation and irrigation return flow. Soils are thin on the plateau, and numerous features on the carbonate terrain may focus recharge, such as collapse features (a few feet wide and a few tens of feet long to 1,500 ft wide and 3 mi long; Freeman, 1968), faults, and lineaments. Groundwater discharges to streams. Numerous discharge zones, mainly along bedding planes between crystalline beds and underlying marly units, are exposed in canyons on the periphery of the plateau.

GULF COAST AQUIFER

The Gulf Coast aquifer consists of interbedded clays, silts, sands, and gravels of Cenozoic age, which are hydraulically connected to form a leaky confined aquifer system. The formations (from lower to upper) are the Catahoula (which as a whole is a confining system, although the very sandy parts are generally included in the overlying Jasper aquifer), Jasper, Burkeville (confining system), Fleming and Goliad sands (which make up the Evangeline aquifer), and the Lissie, Willis, Bentley, Montgomery, and Beaumont Formations (which usually are grouped as the Chicot aquifer) (Baker 1986; Ashworth and Hopkins 1995). The Lissie and Willis make up the most transmissive part of the Chicot, and the Montgomery can act as a local confining layer within the Chicot. Likewise, although it can contain significant sand deposits (Kreitler et al. 1977), the Beaumont can also act as a confining layer at the top of the Chicot. Vertical profiles of water levels and water chemical composition in the Evangeline and Chicot aquifers in Matagorda and Wharton Counties suggest that the Beaumont should be modeled as a hydrological layer separate from the rest of the Chicot (Dutton 1994).

Sources of recharge include the Rio Grande and precipitation. The aquifer extends over a wide range in precipitation regimes, from 56 in/yr in the northeast to 22 to 26 in/yr in the southwest (Larkin and Bomar 1983). As with the Carrizo-Wilcox aquifer, diffuse recharge occurs primarily from precipitation across interstream areas. In addition, soil type influences recharge rates because soil type varies with the underlying geologic formations. There is probably more recharge across sandy soils of the Lissie and Willis Formations than across the more clay-rich, finer grain sandy soils in the Montgomery and Beaumont Formations. Much of the groundwater in the outcropping, unconfined part of the aquifer, recharged at the water table, is discharged as base flow to streams and rivers and through evapotranspiration by phreatophytic vegetation over shallow water tables adjacent to streams.

Although groundwater in the outcrop area is generally less than 100 ft below ground surface, heavy municipal and industrial pumpage has resulted in significant water-level declines in areas such as near Houston and Galveston. In the future, large decreases in water levels that reach the outcropping, unconfined aquifer may change river segments from gaining to losing where the rivers cross the aquifer outcrop. In these cases, surface water would become a source of aquifer recharge.

HUECO-MESILLA BOLSON AQUIFER

The Hueco-Mesilla Bolson aquifers are composed of Tertiary and Quaternary basin-fill (bolson) deposits. The Mesilla Bolson aquifer, west of the Franklin Mountains, extends into New Mexico to the north. The bolson is as much as 2,000 ft thick, and there are three water-bearing zones. Most of the recharge to the Hueco-Mesilla Bolson aquifers occurs from mountain-front recharge. Recharge is primarily from the alluvial fans along the Franklin Mountains. There is also inflow from New Mexico to Texas within the Mesilla Bolson, seepage from the Rio Grande, discharge from groundwater in the Franklin Mountain bedrock recharged within the mountains, and return flow from irrigation. The Hueco Bolson, east of the Franklin Mountains, can be locally as much as 9,000 ft thick, with a thick unsaturated zone. Recharge is primarily from the upland areas, discharge from groundwater beneath the Diablo Plateau (Mullican and Senger, 1992), and seepage from the Rio Grande. Fresh to brackish (TDS <10,000 mg/L) water lies only within the upper several hundred feet beneath the water table. The use of artificial recharge is currently being evaluated in El Paso—wastewater that has been treated to drinking-water standards is being recharged in a dry well and in a 0.5-acre recharge basin. Preliminary results indicate that this system is very successful, with recharge rates of 13 ft/d in the recharge basin and 800 L/min in the dry well (M. Ankeny and D. B. Stephens, personal communication, 2001).

OGALLALA AQUIFER

The Ogallala aquifer consists of sand, gravel, clay, and silt deposited during the Tertiary period (Ashworth and Hopkins 1995). Approximately 95% of the water pumped from the Ogallala aquifer is used for irrigation. It is the main aquifer beneath the High Plains of Texas; other local aquifers are used only where the saturated part of the Ogallala is thin. The Ogallala is generally unconfined; however, locally it may be partially confined. Precipitation ranges from 14 to 16 in/yr in the west to 20 in/yr in the east. Sources of possible recharge include playas, streams, and other locations of perennial wetness; irrigation return flow; and precipitation across interplaya areas. Most recharge is focused beneath playas (Scanlon et al. 1997). Low precipitation in the High Plains and high potential evaporation indicate that precipitation recharge in interplaya and interstream areas may be negligible (Aronovici and Schneider 1972; Scanlon et al. 1997). The spatial focusing of recharge beneath playas may not be critical for water-resource modeling, as shown by Mullican et al. (1997). The spatial distribution of recharge may be affected by the distribution of precipitation and soil types. The soil types generally reflect the underlying distribution of geologic materials. Recharge is generally greater in the sand-rich sediments than in more clay-rich sediments. Tributary streams may provide local recharge, but much of such recharge may discharge farther downgradient, lower in the tributary valley. Major streams, such as the Canadian River, generally have received discharge from Ogallala groundwater.

Many areas of the aquifer have been irrigated since the 1940's. Average annual withdrawal for irrigation was greatest during the 1980's, but during the 1990's, the total rate of irrigation withdrawal decreased. Irrigation inefficiency probably was high during the 1940's and 1950's but decreased during the past few decades. Luckey and Becker (1999) estimated that irrigation inefficiency decreased from 24% during the 1940's and

1950's to less than 4% by the 1980's. Irrigation return flow may contribute a good deal of recharge to the aquifer. The amount of return flow depends on irrigation rate, irrigation inefficiency, soil type, depth to water, and velocity or rate of downward movement of water from the root zone to the water table. Return flow may reach the water table later than the year or even the decade in which irrigation was applied, and the delay or lag may increase through time as depth to water increases. Velocity of water moving downward through the unsaturated zone is also an important, although poorly constrained, variable. If the velocity is much greater than the rate of water-level decline, return flow quickly reaches the water table. If the downward velocity is similar to the rate of water-level decline, much of the return flow may be significantly delayed in reaching the water table, leaving more water in storage in the unsaturated zone. The magnitude and effect of return flow in different parts of the High Plains aquifer remain poorly understood.

SEYMOUR AQUIFER

The Seymour Formation lies within the Rolling Plains and consists of isolated areas of unconsolidated Quaternary sand and gravel, silty clay, and caliche eroded from the High Plains. Although the thickness of the Seymour unit varies greatly, it is generally less than 100 ft, and the upper portion is generally finer grained than the base. Precipitation ranges from 24 to 28 in/yr in the east to 20 in/yr in the west (Larkin and Bomar 1983). Nearly all recharge occurs by direct infiltration of precipitation. Groundwater is unconfined. Recharge is higher in the sand hills area and in other areas where the sandier materials crop out. Groundwater follows arcuate flow paths toward the east-southeast, heading to the perimeter of the Seymour deposits. Streams are gaining because stream stage is at a lower elevation than groundwater in the Seymour aquifer. Natural discharge from the aquifer is through seeps, springs, evapotranspiration, and seepage to the underlying Permian units. Approximately 90% of the water pumped from the aquifer is used for irrigation (Ashworth and Hopkins 1995).

TRINITY AQUIFER

The Trinity aquifer consists predominantly of sandstones in the north-central section of the aquifer and of flat-lying limestones in the central section (Hill Country area). In the north-central section, the Trinity aquifer includes groundwater in Lower Cretaceous-age Twin Mountains and Paluxy Formations. In the subsurface, these formations are separated by the Glen Rose Formation. Beyond the updip limit of the Glen Rose to the north and west, the Twin Mountains and Paluxy Formations are mapped together in the outcrop as the Antlers Formation. The Trinity aquifer in the Hill Country consists of lower, middle, and upper aquifers based on hydraulic characteristics of the sediments. The aquifer-bearing formations crop out along the dissected margin of the Edwards Plateau.

Precipitation is the dominant source of recharge on the outcrop in both the north central and central regions; however, seepage losses occur in some locations from lakes and streams (at headwaters). Groundwater flow generally follows topography in the outcrop area. Cross-formational flow from the underlying Antlers Formation may also provide recharge to the Trinity aquifer in the north central section. Groundwater discharges to streams and to phreatophytic vegetation. Water levels have declined significantly near Dallas-Fort Worth and near Waco (down to 900 ft, Mace et al., 2000).

TECHNIQUES FOR MODELING RECHARGE

The purpose of this section is to discuss how recharge can be simulated in the major aquifers in Texas using MODFLOW. Groundwater recharge can be considered in a general sense as the *net groundwater inflow,* which refers to the right-hand side of the groundwater balance equation solved by MODFLOW:

$$(dS/dt)_{gw} = Q_{gw} + Q_{rech} + Q_{rseep} - Q_{gdiv} - Q_{et\text{-}gw} - Q_{base} \tag{13}$$

On the left-hand side of equation 13 is the rate of change in groundwater storage, $(dS/dt)_{gw}$. On the right, Q_{gw} = net lateral inflow, Q_{rech} = recharge, Q_{gdiv} = diversions (for example, irrigation pumping), $Q_{et\text{-}gw}$ = evapotranspiration from shallow ground water; Q_{rseep} = stream seepage to underlying aquifer, and Q_{base} = base-flow discharge to streams. Some of the methods used to represent the net groundwater inflow for a groundwater simulation model are summarized here. All of the components of the net groundwater inflow can be spatially distributed over the domain of the model simulated by MODFLOW. Temporal variability in each parameter, including recharge can also be incorporated into the model. The various approaches to simulating recharge in MODFLOW include (McDonald and Harbaugh, 1988; Harbaugh and McDonald, 1996):

1. Specified head
2. Specified flow (using the Recharge package or Well package)
3. Head-dependent flow
   General Head Boundary
      River package
      Stream package

The ET and Drain packages simulate discharge in the outcrop areas, which is important in calculating recharge from outcrop areas to confined sections of aquifers, such as the Carrizo Wilcox and Gulf Coast aquifers. Accurate simulation of groundwater discharge by baseflow to streams and by evapotranspiration is important because increased groundwater development in the future may capture some of this water. A schematic of the water fluxes associated with the different MODFLOW packages is shown in Fig. 12. The following descriptions of the different approaches to simulating recharge are adapted from Anderson and Woessner (1992).

Simulating Recharge Fluxes Using Various Approaches in MODFLOW

Specified Head

Specified-head or constant-head values can be assigned to model grid nodes when the value of head at those nodes remains fixed at the specified level, such as may be the case with lakes, the sea coast, impoundments, rivers, and canals that are hydraulically connected to the aquifer. (In the case of rivers and canals, specified head will vary spatially.) The elevation of specified head for any surface water body at the boundary is normally assumed to equal the elevation of the water table at the specified location. When constant-head boundaries are used, the MODFLOW code determines the fluxes entering or leaving the constant-head grid cells; therefore, this approach can be used in certain circumstances to delineate recharge and discharge zones.

In order to maintain the value of head constant, the constant-head approach provides an infinite source of water to the constant-head cells. This approach may be unrealistic for simulations of transient or future conditions, where head changes are expected. Thus, recharge and water availability may be greatly overestimated for simulations of future conditions using specified head boundaries. The constant-head approach could be used in preliminary steady-state simulations of predevelopment conditions to delineate recharge and discharge zones and to provide preliminary estimates of recharge rates. These steady-state simulations can then be rerun using the calculated recharge rates from the constant-head approach and incorporating discharge explicitly using the River, Stream, ET, or Drain packages. Once constant head is prescribed, the model will keep it at the prescribed level irrespective of whatever pumping or recharge is taking place at the constant-head cell(s) or anywhere else in the model grid. In other words, the specified-head condition behaves as if it were an infinite source or sink of water in order to maintain the head level constant. Therefore, caution should be exercised in using this type of boundary condition, especially in transient simulations, where head values are likely to change with time.

Specified Flow

Specified flow or prescribed flux boundaries are specified where the groundwater flux is known. Such conditions can be simulated by the recharge package or well package. The **Recharge package** simulates the water fluxes across the water table from an outside source. Recharge is specified as a rate (L/T) either to the top grid layer or to the highest active cell in each vertical column, or the vertical distribution of recharge is specified. MODFLOW computes the volume of water added to the model by multiplying nodal recharge rates by the area of the top of the cell per unit of time. Recharge can either be uniform over the whole modeled area or varied among cells, and can vary in time.

The **Well package** uses injection (or pumping) wells to inject (or extract) water at the specified rate. The user specifies the injection (or pumping) rate and location of the well screen. Inflows are treated as volumes of water "placed" into the model grid cell. Conceptually, water may enter the top of the block as groundwater recharge or side of the block as underflow. The flux is assumed to be uniformly distributed over the face of the cell. The well package allows great flexibility and parsimony in assigning fluxes to specific cells in each layer. If pumping and injection wells are also used in the model, use of the well package to specify recharge could be confusing.

A special type of specified flux boundary is the no-flow or zero flux boundary (also known as barrier or impermeable boundary). Such boundaries are normally specified at the limits of the geologic units that form the aquifer. Flow lines are parallel to no-flow boundaries. Groundwater divides are also represented as no-flow boundaries.

Very little information is available on recharge rates for the various aquifers (Scanlon, 2000). Most of the estimates are based on previous groundwater modeling studies. Preliminary estimates of recharge rates could be obtained from running steady-state simulations with a constant-head boundary condition for predevelopment conditions.

Head-Dependent Flow

Head-dependent flow is flow that is dependent on the difference between a user-supplied head on one side of the boundary and the model-calculated head on the other side. Leakage to or from a surface water body such as a river, lake, or reservoir can be simulated using head-dependent conditions. The flux or leakage rate is calculated according to the following equation:

$$Q = C\,(h_s - h), \qquad (14)$$

where the conductance term, $C = (k_s A / b_s)$, k_s is the hydraulic conductivity of the interface in the flux direction (e.g., riverbed sediments), A is the area of the cell through which leakage occurs, b_s is the thickness of the interface, h_s is the head in the source reservoir (e.g., a lake or river), and h is the simulated head in the model layer representing the aquifer immediately below or adjacent to the source. Examples of head-dependent flow include the General Head Boundary, River, Stream, ET, and Drain packages.

The **General Head Boundary (GHB)** package simulates recharge according to equation 14, where the conductance refers to that of the boundary and h_s refers to the head in the boundary. The GHB package assumes continuous linear leakage. Because there are no limits to recharge when the GHB package is used, recharge can be overestimated during simulations of transient or future conditions (Anderson and Woessner, 1992). The River and Stream packages differ from the GHB package in imposing limits to recharge.

The **River package** is used to simulate the water flow between a surface water body (e.g., river or lake) and the underlying aquifer. The River package can be used to simulate recharge from or discharge to streams, lakes, or reservoirs. Some studies are using the River package to simulate recharge in outcrops, which does not require explicitly representing recharge and discharge separately (E. Strom, USGS, personal communication, 2001). The River package assumes that the recharge from losing rivers or streams is independent of the river or stream discharge (which is not the case with the Stream package, to be described later.) Thus a losing reach of stream could recharge the aquifer with more water than is being carried in the stream. No adjustment is made in the stream stage.

The **Stream package** differs from the river package in that it considers the volume of stream flow in each stream segment; therefore, it will increase stream flow in areas of gaining reaches and reduce stream flow by water lost through streambed seepage in losing reaches. The reach will go dry if leakage or surface-water diversions for a given reach exceed stream flow. In this case, leakage is set to zero, and downstream reaches are prevented from leaking until additional water is added by tributaries or groundwater seepage. Because the river package may overestimate recharge from losing streams, the GAM specifications require the use of the Stream package rather than the River package.

Evapotranspiration (ET) across the water table also can be represented as a head-dependent boundary, where the flux across the boundary is proportional to the depth of the water table below the land surface. Groundwater ET may occur when the water table is close to the land surface or when phreatophytes draw water from below the water table. Input information required by the ET package includes the maximum ET rate for each cell and the extinction depth at which ET is assumed to be zero. ET is assumed to decrease linearly with depth from the maximum rate when the water table is at the land surface to zero at the extinction depth. The extinction depth can be assigned the value of the rooting depth.

Discharge through springs and seeps can be simulated using the **Drain package**. The elevation of the spring or seep as it emerges at the land surface is the elevation of the drain. Diffuse flows, such as seepage to wetlands, can be simulated by specifying drain nodes in the general area where seepage is likely to occur. Leakage to the drain is simulated whenever h in equation 14 is greater than h_s (the elevation of the drain). Leakage rate equals zero if h is less than h_s. The drain nodes will be activated only when the head in the aquifer equals or exceeds the land-surface elevation. The River package also can simulate a spring or a drain by setting the bottom elevation of the streambed equal to the head in the source reservoir (i.e., the river stage).

Inverse Modeling

Inverse modeling can be used to estimate recharge or to refine initial recharge estimates. In conventional groundwater models, recharge is postulated as known and hydraulic heads computed, whereas in inverse groundwater models, recharge is computed from field measurements of hydraulic head. There are a variety of codes available to do inverse modeling, including PEST (Doherty 1994) and UCODE (Poeter and Hill 1998), among others. However, it is important to remember that if only head data are used for model inversions, only the ratio of recharge to hydraulic conductivity can be estimated. Additional information, such as base-flow discharge or groundwater dating, is required in order to estimate recharge rates using inverse methods.

Spatial Variability in Recharge

Recharge may vary spatially as a result of variations in precipitation, vegetation and land use, soil type, geology, and depth to water table. Ideally, site-specific information on recharge should be used to distribute recharge to the aquifer; however, most of the recharge estimates for the various aquifers are based on previous model calibrations. Various techniques need to be used to spatially distribute recharge as a function of the different controls on recharge. Initial estimates of recharge can be obtained from the literature or from running a simulation with constant-head conditions for predevelopment. These estimates of recharge can then be varied spatially on the basis of the distribution of controlling factors. Dutton et al. (2001) demonstrated how recharge was varied in the north part of the Ogallala aquifer on the basis of spatial variations in precipitation, geology (Ogallala and Blackwater Draw Formations), and irrigation. A segmented linear relationship between recharge and precipitation was assumed. Minimum and maximum values of recharge were estimated for minimum and maximum values of precipitation. These initial estimates of recharge were optimized through model calibration by minimizing the root mean square error (RMSE) between measured and simulated heads. A multiplication factor was then applied to this relationship depending on the soil types. This multiplication factor was also optimized by minimizing the RMSE. The potential impact of irrigation return flow was also considered for this aquifer by considering irrigation amounts, irrigation efficiency, time of travel through the unsaturated zone, and declining water tables. Some scenarios suggest that water from irrigation return flow may still reside in the unsaturated zone, whereas other scenarios indicate that irrigation return flow may already have reached the water table.

The spatial variability in recharge can be predicted by various techniques such as the soil water balance (Sophocleous 1991), combining field-based water-balance methods with GIS and statistical analyses (Sophocleous 1992), watershed models such as SWAT (Soil Water Assessment Tool; Arnold et al., 1994), and combined watershed and groundwater models (integrated SWAT and MODFLOW; Sophocleous et al., 1999; Sophocleous and Perkins, 2000); however, these various techniques will be described in a separate paper on techniques for estimating recharge.

Groundwater Discharge

Groundwater discharge may be a critical issue for assessing groundwater resources under future development conditions. Although much of the recharge in humid settings discharges to nearby streams, it is important to include this recharge and discharge in the GAM models because increased groundwater development in the future may capture some of this discharge. Increased groundwater pumpage would reduce groundwater discharge to streams and could ultimately result in changing streams from gaining to losing.

Specific Comments Related to Modeling Recharge in the Major Aquifers

The constant-head approach can be used in the water table portion of any of the aquifers to obtain a preliminary estimate of recharge to the aquifer. The constant-head approach should preferably be used in the steady-state simulations for predevelopment conditions because this boundary condition would provide an inexhaustible source of water for simulations of transient or future conditions that may be unrealistic. This approach would delineate recharge and discharge areas in the aquifer. The steady-state simulations can then be rerun using the calculated water fluxes from the constant-head approach.

Carrizo-Wilcox Aquifer

The main source of recharge is precipitation. Losing streams and flood water provide an additional source of recharge in the southern section of the Carrizo-Wilcox aquifer (L.B.G. Guyton & Associates and HDR Engineering, Inc., 1998). Recharge can be simulated using the Recharge package. Specified recharge fluxes can be varied regionally depending on long-term average precipitation amount, vegetation and land use, irrigation, and soil type. The Carrizo Wilcox aquifer spans the entire upper coastal plain, and precipitation varies from 20 to 56 in/yr. Soil types from the STATSGO database (USDA, 1994) can be grouped on the basis of correlations with underlying geology. The geologic atlas developed by Henry and others (1979) may also be helpful in grouping soils by geology across part of the aquifer. Because the recharge package only simulates the input to the system, the outputs have to be simulated explicitly using the ET and Stream packages. The ET package can be used to simulate discharge from areas of shallow water tables (< ~10 ft) adjacent to stream beds, and the distribution of phreatophytes can be used to determine where such ET is occurring. Information on vegetative cover can be obtained from the National Land Cover Characterization Dataset (NLCD). The ET output may be captured later as the water table is lowered during simulations of future conditions. The Stream package should be used to simulate the larger streams in the study area. Streams are generally gaining throughout most of the aquifer, and base-flow discharge to streams can be simulated using the Stream package. In the southwest portion of the Carrizo, streams such as the Nueces and Frio are losing streams and constitute a source of recharge (L.B.G. Guyton & Associates and HDR Engineering, Inc., 1998). Recharge can be simulated in these areas using the Stream package, and flood conditions can be represented by recharge in cells adjacent to the streams.

Cenozoic Pecos Alluvium Aquifer

The primary sources of recharge are precipitation, irrigation return flow, focused recharge from playas, and cross-formational flow. The main attributes of the aquifer, including important factors for assessing recharge, are described in Jones (2001). Although the playas constitute point sources of recharge, it may not be necessary to simulate these explicitly because a previous modeling study (Mullican et al. 1997) indicates that similar results are obtained whether focused or diffuse recharge is used as input. However, this modeling study was conducted for the Ogallala aquifer and the results may not apply to the Cenozoic Pecos Alluvium aquifer. Recharge can be simulated using the Recharge package, and initial estimates of recharge can be obtained from the specified-head approach discussed previously. Specified recharge fluxes can be varied regionally, depending on long-term average precipitation amount, vegetation and land use, and soil type. Soil types from the STATSGO database can be grouped on the basis of their hydrologic properties. The alluvial soils are probably more closely related to the geology of the respective sediment source areas (Davis, Delaware, and Apache Mountains and Ogallala aquifer) than with the underlying geology (Dockum, Edwards, etc.) (Jones, personal communication, 2002). Higher recharge rates may be specified for the thick sand dunes in Monument Draw. The ET package can be used to simulate discharge from areas of shallow water tables (< ~10 ft) adjacent to the Pecos River. The distribution of phreatophytes can also be used to determine where ET is occurring. The Stream package should be used to simulate the Pecos River in the study area. Increased groundwater development in the future may change the Pecos River from gaining to losing status.

Edwards Aquifer

The primary source of recharge to the Edwards aquifer is focused in the streams, although precipitation-based recharge also occurs in the upland areas, and cross-formational flow from the Trinity aquifer in the Hill Country may recharge the Edwards aquifer. Stream recharge has generally been simulated with the Recharge package (Barrett and Charbeneau 1997; Scanlon et al., 2001). Recharge can be uniformly distributed along the stream sections in the outcrop zone or varied spatially along the streams if information on spatial variability is available. Data from the Balcones Fault Zone section of the aquifer suggest that stream recharge is greatest in the upstream section and much less toward the downstream region. Interstream recharge can also be simulated using the recharge package. Previous studies indicate that interstream recharge is about 15% of the total recharge (Barrett and Charbeneau 1997). Temporal variability in recharge can be simulated using monthly or daily inputs.

Edwards-Trinity Plateau Aquifer

The primary sources of recharge include precipitation and irrigation return flow that can be simulated using the Recharge package. Recharge rates can be varied spatially depending on precipitation, irrigation, vegetation, soil types, and topographic slope. Mean annual precipitation varies from 10 in/yr in Culberson County to 32 in/yr in Blanco County. Higher recharge rates can be applied to represent focused recharge along collapse features, faults, and lineaments. Steeper slopes may result in greater runoff and lower recharge. Groundwater discharge includes well pumping, base-flow discharge to streams (simulated using the Stream package), and evapotranspiration by phreatophytes along streams (simulated using the ET package). Discharge along bedding planes between crystalline beds and underlying marly units exposed on canyons on the periphery of the plateau can be represented by the Drain package.

Gulf Coast Aquifer

The Gulf Coast is recharged primarily by precipitation. Losing streams and irrigation canals provide additional sources of recharge in the southwest portion of the aquifer along the Rio Grande. Spatial variability in recharge can be represented on the basis of relationships between recharge and precipitation amount, vegetation and land use, irrigation, and soil type. Precipitation ranges from 56 in/yr in the northeast to 20 to 26 in/yr in the southwest portions of the aquifer. Similar soils may be grouped according to variations in the underlying geology. Soils in the Jasper and Goliad units of the Evangeline aquifer and in the Lissie and Willis units of the overlying Chicot aquifer may have higher permeabilities than those in overlying confining units. Areally distributed recharge can be simulated using the Recharge package. The stream package can be used to simulate recharge in the southwest portion of the aquifer, where streams are losing, such as the Rio Grande and base flow in the central and northeast portions of the aquifer, where streams are gaining. Evapotranspiration adjacent to streams can be simulated using the ET package, particularly in the southwest in areas of shallow water tables, where phreatophytes occur.

Hueco-Mesilla Bolson Aquifer

The Hueco-Mesilla Bolson is recharged primarily through the alluvial fans along the Franklin Mountains and by cross-formational flow. Mountain front recharge can be represented by specifying a flux in the Recharge package in a band along the mountain front. Additional recharge is provided by seepage along the Rio Grande and irrigation canals, which can be simulated using the Stream package.

Ogallala Aquifer

The primary source of recharge to the Ogallala aquifer is the playas. In addition, irrigation return flow may contribute a significant amount of recharge. Recharge from precipitation may be important in areas of coarse-grained soils having a relatively shallow water table. Previous modeling studies by Mullican et al. (1997) demonstrate that focused recharge through playas can be represented as distributed recharge, provided an overall estimate of playa recharge can be made. Recharge to the Ogallala aquifer can be simulated using the Recharge package. Dutton et al. (2001) described how recharge can be varied spatially on the basis of variations in precipitation and soil type for part of the central segment of the Ogallala aquifer. Incorporating irrigation return flow into the simulations and lagging return flow on the basis of the depth to the groundwater are also described. The Stream package should be used to simulate possible recharge from streams along tributaries and base-flow discharge to the Canadian River. A similar approach to simulating recharge can be used for the south segment of the Ogallala aquifer.

Seymour Aquifer

Most recharge occurs by direct infiltration of precipitation. Groundwater is unconfined. Recharge can be simulated using the Recharge package. Recharge can be varied spatially depending on precipitation, soil type, and geology. Higher recharge rates can be assigned to the sandier materials where they crop out. Base-flow discharge to streams can be simulated using the Stream package. Discharge caused by phreatophytes can be simulated using the ET package. Discharge through seeps and springs can be simulated using the Drain package.

Trinity Aquifer

The Trinity aquifer can be subdivided into the north-central section between the Colorado and Red Rivers that consists predominantly of sandstones that dip to the east and a central section south of the Colorado River that consists of mostly flat-lying limestones in the Hill Country area. The main source of recharge in the north-central section is precipitation, which can be simulated using the Recharge package. The strike of the outcrop area parallels the isopleth contours of precipitation; therefore, spatial variability in precipitation may not be significant, although the Paluxy Formation may have less than the Twin Mountains Formation. Information on soil types, underlying geology, and vegetation type can be used to spatially distribute recharge. Groundwater discharge resulting from base-flow discharge to streams can be simulated using the Stream package and that from evapotranspiration can be simulated using the ET package.

Recharge in the Hill Country can also be simulated using the Recharge package. Recharge estimates for this aquifer have generally been determined from base-flow discharge studies and therefore represent an estimate of net recharge after evapotranspiration has occurred (Mace et al, 2000). Recharge can be varied spatially according to precipitation amount, geology, soil type, and vegetation. Groundwater discharge resulting from base-flow discharge to streams can be simulated using the Stream package.

RECOMMENDED APPROACHES FOR QUANTIFYING RECHARGE FOR THE MAJOR AQUIFERS

Detailed descriptions of the various techniques for quantifying recharge are provided at the beginning of this report. It is apparent from the review of existing recharge estimates that all of the major aquifers require additional recharge studies to better quantify recharge. One of the difficulties of determining appropriate techniques for quantifying recharge is that many techniques are restricted to measuring recharge rates within a certain range; however, we do not know what the recharge rate is. Therefore, in many cases, an iterative approach should be adopted in which a particular technique is initially applied. Preliminary recharge estimates from this technique can then be used to determine a more appropriate technique for quantifying recharge. In this process, the recharge estimates are continually improved and refined. Proposed techniques for quantifying groundwater recharge are shown in Table 2.

CARRIZO WILCOX AND GULF COAST AQUIFERS

Recharge rates may vary markedly throughout the Carrizo Wilcox and Gulf Coast aquifers. The climatic conditions range from humid in the northeast to semiarid in the southwest. Recharge rates may be much greater in high permeability units (e.g., Carrizo, Simsboro, Jasper, Goliad, and Lissie Formations) and much lower in low permeability units (e.g., Calvert Bluff and Hooper, Catahoula, and Burkeville Formations). In addition, recharge in the outcrop area should be distinguished from recharge to the confined section of the aquifer (net recharge) because much of the groundwater in the outcrop is discharged through evapotranspiration and base flow to streams and never reaches the confined aquifer. Most techniques described in this section apply to recharge in the outcrop area, where the aquifers are unconfined. It is much more difficult to quantify net recharge to the confined section of the aquifers.

Surface-water techniques can be used to quantify recharge in different areas of the aquifers. The channel-water-budget approach may be appropriate for the southwest part of the Carrizo-Wilcox aquifer, where there are losing streams such as the Nueces and Frio Rivers. Stream-flow-gauging data are required upstream and downstream of the outcrop area. Additional gauging stations would have to be installed in many areas to meet this requirement (Fig. 13). For example, L.B.G. Guyton and Associates and HDR Engineering, Inc. (1998), noted that there are no upstream gauging stations for San Miguel and Atascosa watersheds. In addition, information on tributary inflows and outflows should be quantified. These tributary flows could be estimated using periodic seepage runs with pygmy meters. Watershed modeling could also be used to quantify recharge, particularly in the more humid regions of the Carrizo-Wilcox and Gulf Coast aquifers in the northeast, where recharge should be a greater portion of the total water budget. Stream flow is greater in these areas also, providing data to calibrate the watershed models. Monitoring soil-water content to compare with simulation results would also increase confidence in simulations results.

Unsaturated-zone techniques should be used in areas where the water table is relatively deep (≥ 50 ft [15 m]). The zero flux plane requires monitoring of pressure heads to establish the zero flux plane and water-content monitoring below this zone in order to estimate recharge. This technique would require long-term monitoring of water content and cannot be used for time periods when water movement is downward from the land surface. The chloride mass balance approach is the simplest technique to apply in the unsaturated zone and should be appropriate in the central and south sections of the aquifer. The use of this technique should be restricted to the high-permeability units, such as the Simsboro and Carrizo units, because connate water may not be flushed out of the low-permeability units, such as the Wilcox, Hooper, and Calvert Bluff units. Estimates of chloride input in precipitation are available from the National Atmospheric Deposition Program (NADP, Fig. 14) These data represent wet fallout only. Recent studies conducted by Izbicki (USGS, personal communication, 2002) in California indicated that wet fallout underpredicts the total input of chloride to the system by a factor of 2 to 3. Therefore, the NADP values should be increased by at least a factor of 2. More quantitative estimates of recharge may be obtained by locating the position of the bomb-pulse tracer peaks in the unsaturated zone, such as tritium and chlorine-36. Use of these tracers should be restricted to areas where the bomb peak has not reached the water table, such as the central and south sections of the Carrizo-Wilcox aquifer. One-dimensional unsaturated-zone modeling can be used to estimate recharge in different parts of the aquifer using climate data from the National Oceanic and Atmospheric Administration (NOAA) stations (www.noaa.gov). Long-term climate records (~ 30 yr) should be used for the simulations in order to evaluate the potential range of recharge rates. Time resolution of meteorologic inputs for the upper boundary condition should be at least daily. Various codes are available to simulate unsaturated flow, such as HYDRUS-1D (Simunek et al. 1998) and UNSATH (Fayer, 2000). The simulations can be conducted for a range of sediment types to obtain estimates of the possible range of recharge rates on the basis of climatic variations. Additional information on vertical hydraulic conductivity can be obtained from the STATSGO database. These simulations may not provide accurate estimates of recharge in areas of low-permeability materials, where runoff and preferential flow may be important.

Saturated-zone techniques may be appropriate in areas of high recharge, most likely the humid northeast section of the aquifers. Water-table fluctuations may be used in areas of shallow water tables to quantify recharge pulses but cannot be used to quantify recharge that does not change over time. The chloride mass balance approach can also be applied to the saturated zone in the high-permeability units. Tracers such as $^3H/^3He$ and CFCs provide the most accurate estimates of recharge under ideal conditions. These techniques may be appropriate in the northeast section of the aquifers. Recharge rates need to be high enough so that the bomb peak has moved into the saturated zone. The $^3H/^3He$ technique requires a recharge rate of at least 1.2 in/yr to confine the ^3He in the groundwater (Cook and Solomon, 1997). CFCs can only be used in rural settings outside the zone of influence of septic tanks. This technique requires a fairly shallow water table (≤ 30 ft). The ^{14}C tracer has been used to date groundwater in the confined section of the aquifer in Atascosa County (Pearson and White 1967). These data can be used to estimate recharge rates (R):

$$R = vnA/S \tag{12}$$

where v is velocity estimated from the ^{14}C age and the distance from the recharge zone, n is porosity, A is the cross-sectional area of the confined aquifer where the velocity is determined, and S is the surface area of the recharge zone. This method provides an upper bound on the recharge rate of the confined aquifer because it assumes no leakage through the confining unit. Inverse modeling of groundwater flow can be used to estimate recharge if there is information in addition to hydraulic heads, such as groundwater dating, base-flow discharge, or other flux data. Information on hydraulic heads alone can only provide estimates of the ratio of recharge to hydraulic conductivity. Inverse modeling based on hydraulic head data and ^{14}C dating may be used to estimate net recharge to the confined section of the Carrizo-Wilcox aquifer.

OGALLALA AQUIFER

Many of the previous studies emphasize the importance of playa recharge and indicate that recharge in interplaya settings is negligible (Scanlon et al. 1997). Although the major streams, such as the Canadian River, are gaining, tributaries to these streams may be losing, particularly in the upstream areas. Recharge is also expected to occur in sandy sediments and in irrigated regions. Because of the thick unsaturated zone, many of the techniques for estimating recharge to the Ogallala aquifer should be based on the unsaturated zone. The absence of calcic soils or caliche may be used as a qualitative indicator of recharge. The absence of calcic soils or low levels of calcium carbonate suggest high recharge rates, such as beneath playas (Scanlon et al., 1997). Surface-water techniques may be appropriate for quantifying recharge from upstream sections of tributaries using heat tracers, seepage meters, or channel-water budgets. The heat-tracer approach has been used to evaluate flow in ephemeral streams and to estimate infiltration from such streams (Constantz et al. 1999).

Appropriate unsaturated-zone techniques include the use of chloride concentrations in soil water. Chloride data can be used to provide a qualitative indicator of recharge and may also provide quantitative estimates of recharge. Low chloride concentrations beneath playas suggest high recharge rates, whereas high chloride concentrations in interplaya settings suggest low recharge rates. The chloride mass balance approach may also be used in sandy areas to quantify recharge rates; however, the accuracy of this approach decreases as recharge rates increase. It would be difficult to use chloride to quantify recharge rates in irrigated regions because of uncertainties in the chloride input to the system. Bomb-pulse tritium may be appropriate for quantifying recharge in sandy areas where the bomb peak is expected to have moved beneath the root zone. The presence or absence of bomb-pulse tritium may also be used in irrigated regions to provide estimates of recharge; however, use of this technique is complicated because the irrigation water probably does not contain bomb tritium. Bomb-pulse $^{36}Cl/Cl$ ratios could also be used to quantify recharge in sandy areas where recharge is expected to be higher than in finer grained sediments. The $^{36}Cl/Cl$ bomb peak may be much more obvious than the ^{3}H peak because of the long half-life of ^{36}Cl (301,000 yr) relative to that of ^{3}H (12.43 yr). Unsaturated-zone modeling could be used to estimate recharge rates in irrigated and nonirrigated regions. Long-term climate records (~ 30 yr) should be used for the upper boundary condition in nonirrigated settings. Simulations of irrigated regions could use a percentage of the applied water as the upper boundary condition. These simulations may provide bounding estimates of recharge rates for different types of sediments and for different estimates of return flow.

Saturated-zone techniques for estimating recharge may be more appropriate in areas of playa recharge because the saturated-zone methods provide a more spatially averaged recharge rate than the point estimates provided by unsaturated-zone techniques. Previous studies have used chloride concentrations in groundwater to estimate recharge in the north half of the Southern Ogallala (Wood and Sanford 1995). Anthropogenic substances such as arsenic used for cotton production and the herbicide atrazine may provide qualitative indicators of high recharge rates as shown by Nativ (1988). Tracers such as ^{3}H, $^{3}H/^{3}He$ can be used to quantify recharge in sandy areas (e.g., southeast section of the Ogallala) and irrigated areas. Inverse groundwater modeling may be combined with groundwater-age data on the basis of ^{3}H, $^{3}H/^{3}He$, and ^{14}C to provide regional estimates of groundwater recharge.

CENOZOIC PECOS ALLUVIUM AQUIFER

This aquifer is in a semiarid region. Calcic-soil development could be used as a qualitative tool to estimate where recharge occurs. The absence of calcic soils would indicate high recharge rates; however, the presence of calcic soils may not accurately reflect current recharge conditions because of the long time required to dissolve these calcic soils. Approaches for quantifying recharge should focus on unsaturated- and saturated-zone techniques rather than surface-water techniques because an integrated surface-water drainage system is not developed. Most surface water drains internally to playas.

Appropriate unsaturated-zone techniques may include the zero flux plane technique in sandy areas where relatively higher recharge rates would be expected. However, monitoring matric potentials and water content should be conducted over a long time period because precipitation is much more variable in semiarid regions than in humid regions. The chloride mass balance approach may be appropriate in nonirrigated regions. Bomb-pulse tracers such as ^{3}H and $^{36}Cl/Cl$ may be used to provide quantitative estimates of recharge in areas where these tracers have moved below the root zone. Unsaturated-zone modeling can be used to estimate ranges of recharge rates for different precipitation, soil, and vegetation conditions. Sensitivity analyses may be used to evaluate important controls on recharge.

Saturated-zone techniques may also be used, particularly in areas where recharge is expected to be higher, such as the Monument Draw Trough north of the Pecos River. Groundwater chemistry may provide qualitative indicators of areas of low and high recharge to the aquifer. Jones (2001) noted that the total dissolved solids content of groundwater in the Monument Draw Trough (<1,000 mg/L TDS) was much lower than in Pecos Trough (<1,000 to >3,000 mg/L TDS), which he attributed to higher recharge in Monument Draw Trough. Water-table fluctuations may be used in areas of relatively high recharge where water tables are shallow (≤30 ft [≤10 m]). Bomb-pulse tracers, such as ^{3}H and $^{3}H/^{3}He$ may be used to quantify recharge in some sandy areas, where high recharge is expected. Inverse groundwater modeling using groundwater-age data, hydraulic-head data, or other flux data may be used to estimate the regional distribution of recharge.

EDWARDS AQUIFER

Quantification of stream recharge is fairly accurate for the Edwards aquifer, as a result of long-term gauging records upstream and downstream of the recharge zone. Future studies should evaluate the spatial variability of recharge along the stream reaches by manual monitoring of stream flow. Studies should also be conducted to quantify interstream recharge. Current estimates of interstream recharge are based on regional watershed studies. Unsaturated-zone studies should be conducted, including the monitoring of water content or water pressure near the contact between soil and underlying fractured rock to determine what threshold rainfall events are required to reach bedrock. Once water has reached bedrock, it should readily recharge the underlying aquifer. The effect of shallow depressions and other features on interstream recharge could also be investigated through monitoring of the unsaturated zone. Data from unsaturated zone monitoring can be used to calibrate the watershed models.

EDWARDS-TRINITY PLATEAU AQUIFER

Previous studies of the Edwards Trinity Plateau have used base-flow discharge to quantify recharge; however the density of stream-flow-gauging stations is very low (Fig. 13). Rutledge (1998) suggested that drainage-basin areas should be less than about 500 mi^2 for this technique because assumptions of the technique, such as uniform recharge over the catchment area, may not be valid for larger basin areas. Additional gauging stations should be installed in the Edwards Trinity Plateau to estimate recharge using base-flow discharge. The heat-tracer approach may be used to estimate infiltration in ephemeral streams. Watershed modeling may be used to provide preliminary estimates of recharge. The unsaturated zone is relatively thick in the Edwards Trinity Plateau; however, stony soils and fractured rock make it difficult to use unsaturated-zone techniques for quantifying recharge. Monitoring matric potential or water content near the soil-bedrock interface may be used to determine threshold rainfall events required to initiate recharge. Techniques for estimating recharge that integrate over large areas are particularly appropriate for fractured systems. Therefore, saturated-zone techniques should be more appropriate than unsaturated-zone techniques. Preliminary studies by H.S. Nance (BEG, personal communication, 2002) suggest that chloride concentrations in groundwater may be suitable for quantifying recharge in the Edwards region. Tritium may also be useful for distinguishing zones of low and high recharge.

SEYMOUR AQUIFER

Qualitative indicators of recharge to the Seymour aquifer may be provided by groundwater-level fluctuations in response to wet periods, total dissolved solids content, and anthropogenic contamination, such as nitrate and atrazine. The highest recharge should occur in the sand hills area southwest of Knox City. Recorded water-level rises after wet periods were greatest in this region (Harden and Associates, 1978). Total dissolved solids are also lowest in this region (Harden and Associates, 1978). Nitrate contamination from cultivation can be used to provide an estimate of recharge on the basis of the time period of cultivation.

Previous studies have used base-flow discharge to the Brazos River to estimate recharge (Table 1); however, bank storage in river alluvium could greatly affect these recharge estimates. If recharge rates are sufficiently high, particularly in the sand hills area, saturated-zone techniques may be used to quantify recharge. Water-table fluctuations may be used in some areas to quantify recharge pulses; however, previous studies indicate that estimates of specific yield may vary markedly, resulting in large uncertainties in recharge estimates (Healy and Cook, 2002). Tritium data can be used to estimate areas of low and high recharge rates. Dating with $^3H/^3He$ could be used to provide accurate estimates of recharge. The CFC tracer may be unsuitable for estimating recharge in the Seymour aquifer because of widespread contamination from septic tanks (Harden and Associates, 1978). Use of chloride to estimate recharge may also be unsuitable because of brine contamination from oil-field activities and possible chloride input from the underlying Permian system. Inverse groundwater modeling could be used with age data or flux data, in addition to hydraulic-head data, to provide regional estimates of recharge.

Outside the sand hills area, lower recharge rates may require the use of unsaturated-zone techniques. The distribution of bomb-pulse tritium below the root zone may be used to estimate recharge in these areas. Unsaturated-zone modeling may also be used to estimate recharge in these areas.

TRINITY AQUIFER

Many of the techniques discussed for quantifying recharge in the Carrizo-Wilcox and Gulf Coast aquifers may also be applicable to the north-central section of the Trinity aquifer, which consists of unconfined and confined sections. Base-flow discharge could be used to estimate recharge as all streams are gaining. Watershed modeling may also provide estimates of recharge by calibrating the models using stream-flow and base-flow discharge data. In areas of thick unsaturated zones, the chloride mass balance approach and bomb-pulse tracers may be used to estimate recharge. Unsaturated-zone modeling may also be used to estimate recharge rates. In areas of higher water flux, chloride, ^{3}H, $^{3}H/^{3}He$, and CFCs in the saturated zone in the outcrop area can be used to quantify recharge. ^{14}C could be used to date water in the confined section of the aquifer. Inverse groundwater modeling can be used with head data, age data, and/or flux data (e.g., base-flow discharge) to estimate the regional distribution of recharge.

Recharge in the Hill Country section of the Trinity aquifer has been quantified using base-flow discharge. Watershed modeling could also be used to simulate recharge. The watershed models can be calibrated using base-flow discharge data. Because much of the Hill Country section of the aquifer has thin soil development and recharge probably occurs primarily through fractured rock, unsaturated zone techniques may not be appropriate for quantifying recharge. Saturated-zone techniques may be suitable in areas of high recharge, such as chloride, ^{3}H, $^{3}H/^{3}He$, and CFCs. Inverse groundwater modeling based on hydraulic head data, age data, and/or flux data could be used to constrain recharge estimates.

HUECO-MESILLA BOLSON AQUIFER

Recharge to the Hueco-Mesilla Bolson occurs primarily through alluvial fans along the Franklin Mountains. Recharge is expected to occur beneath ephemeral streams in the alluvial fans. Calcic-soil development may provide a qualitative indicator of higher recharge beneath the ephemeral streams. Flow in ephemeral streams may be difficult to gauge because of erosion and ephemeral nature of flow. Constantz et al. (1999) used heat as a tracer to determine when ephemeral streams are flowing and to quantify infiltration from these ephemeral streams. Similar approaches may be suitable for ephemeral streams on the alluvial fans near the Franklin Mountains. Although information on the infiltration rate provides an estimate of potential recharge to the groundwater, much of the infiltrated water may later evapotranspire. The thick unsaturated zone may require the use of unsaturated-zone techniques to estimate recharge. Tracers such as chloride and bomb-pulse tritium and ^{36}Cl may be used to estimate recharge beneath the ephemeral streams. Groundwater tracers may also be used in areas where recharge is expected to be high and water tables are shallow.

CONCLUSIONS

A wide variety of techniques are available for quantifying groundwater recharge. These techniques include those based on surface-water, unsaturated-zone, and saturated-zone data and can be based on physical, chemical, or numerical modeling approaches. These techniques vary in the space and time scales that they represent and in the range of recharge rates that they can estimate. Descriptions of the various techniques for quantifying recharge are provided, and estimates of the spatial and temporal scales and range in recharge rates that can be estimated using the various techniques are provided. Groundwater-tracer techniques are considered the most reliable estimators of recharge because the distribution of the tracers indicates that the water has reached the groundwater, and the accuracy of the groundwater dates provided by these techniques ensures that the recharge estimates are highly accurate. Techniques based on surface-water and unsaturated-zone data provide estimates of potential recharge.

Compilation of recharge rates for all major aquifers provides information on the state of knowledge of recharge to the aquifers. The Edwards aquifer is the most dynamic of all the major aquifers, and recharge rates are highly variable spatially and temporally. Stream-channel recharge is fairly accurately quantified using stream-gauge data. Interstream recharge rates are less well known. Estimated recharge rates in the Carrizo-Wilcox aquifer range from 0.1 to 5.8 in/yr. The higher recharge rates occur in the sandy portions of the aquifer (i.e., Carrizo and Simsboro Formations). Reported recharge rates for the Gulf Coast aquifer (0.0004 to 2 in/yr) are generally lower than those in the Carrizo-Wilcox aquifer. In both the Carrizo-Wilcox and Gulf Coast aquifers, higher recharge rates are estimated in upland areas containing sandy soils. Regional recharge rates in the High Plains aquifer, outside irrigated areas, are generally low (0.004 to 1.7 in/yr), whereas playa-focused recharge rates are much higher (0.5 to 8.6 in/yr). Irrigated areas also have fairly high recharge rates (0.6 to 11 in/yr). Recharge rates in the Trinity and Edwards-Trinity Plateau aquifers generally range from 0.1 to 2 in/yr. The Seymour aquifer has recharge rates that range from 1 to 2.5 in/yr. Recharge rates for the Hueco-Mesilla Bolson and the Cenozoic Pecos Alluvium are represented as total recharge along mountain fronts and valley floors.

The main techniques that have been used for estimating recharge in the major aquifers in Texas are Darcy's Law, groundwater modeling, and base-flow discharge. Darcy's Law is widely applied in the confined sections of the Carrizo-Wilcox and Gulf Coast aquifers; however, these recharge estimates may not be very reliable because of uncertainties in estimates of the regional hydraulic conductivity. Groundwater modeling has been used to estimate recharge in most aquifers; however, only hydraulic-head data were available to calibrate the models. Model calibration based on hydraulic-head data alone can only be used to estimate the ratio of recharge to hydraulic conductivity. Additional information, such as groundwater-age data or base-flow discharge data, is required to estimate recharge using groundwater models. Base-flow discharge has been used to estimate recharge primarily in the Cenozoic Pecos Alluvium, Edwards-Trinity, Seymour, and Trinity aquifers. A recent study by Jennings et al. (2001) used water level fluctuations to estimate recharge to the Trinity aquifer in the Hill Country region. Environmental tracers have only been used to a limited extent (chloride mass balance, tritium, and carbon-14) in the Ogallala and Carrizo-Wilcox aquifers. This review of existing data indicates that additional studies are required to provide more quantitative estimates of recharge for the major aquifers.

Determination of appropriate techniques for quantifying recharge to the major aquifers depends in part on the recharge rates; however, recharge is what we are trying to quantify. Therefore, we can only suggest different approaches that are likely to provide the most quantitative estimates of recharge. Results provided by initial studies should provide additional data to optimize the techniques and refine the recharge estimates. A phased approach may be required to accurately quantify recharge rates and a variety of approaches should be applied because of uncertainties in recharge estimates. Results from the various techniques can be compared to determine uncertainties in recharge estimates.

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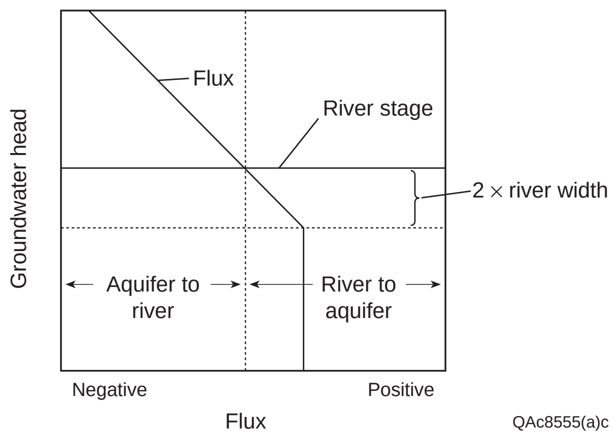

Figure 1. Water fluxes related to the degree of connection between rivers and aquifers. The aquifer discharges to the river when the groundwater head is greater than the river stage, whereas the river recharges the aquifer when the river stage is greater than the groundwater head. Recharge values generally reach a constant rate when the water-table depth is greater than twice the river width (Bouwer and Maddock 1997).

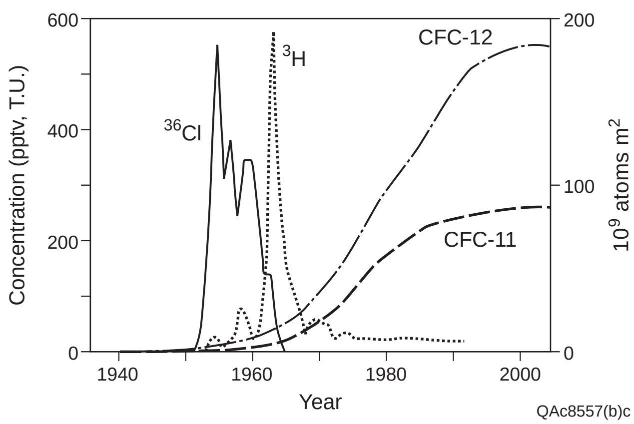

Figure 2. Input functions for historical tracers, including ^3H, ^{36}Cl, CFC-11 and CFC-12 (^3H data from Ottawa, Canada; ^{36}Cl data, Phillips 2000; CFC data, http://water.usgs.gov/lab/cfc/background/air_curve.html).

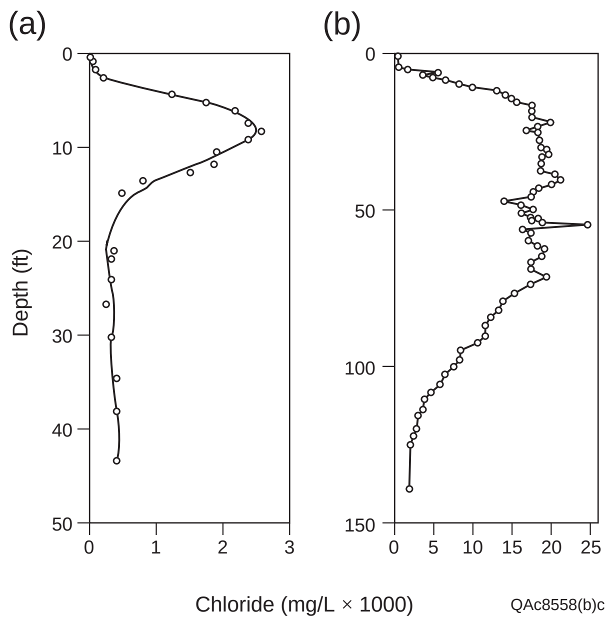

Figure 3. Typical chloride profiles in unsaturated systems: (a) bulge-shaped chloride profile attributed to paleoclimatic variation (Scanlon 1991); (b) bulge-shaped chloride profile attributed to diffusion to a shallow water table (Cook et al. 1989).

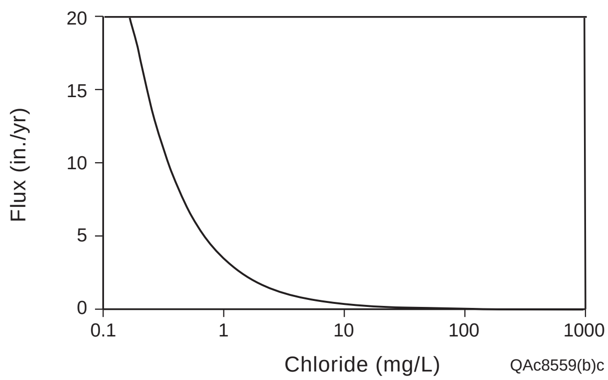

Figure 4. Sensitivity of drainage calculated using the chloride mass-balance approach to chloride concentrations, based on data from boreholes at a site in the Chihuahuan Desert, Texas, USA (Scanlon 2000).

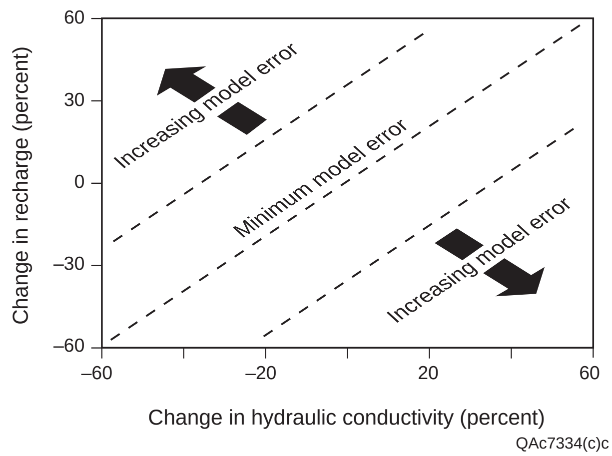

Figure 5. Diagram showing relation between change in recharge and change in hydraulic conductivity. Groundwater-model calibration using hydraulic heads only provides information on the ratio of recharge to hydraulic conductivity. Model error can be minimized using a wide range of recharge and hydraulic-conductivity values if the ratio between recharge and hydraulic conductivity is constant (modified from Luckey et al. 1986).

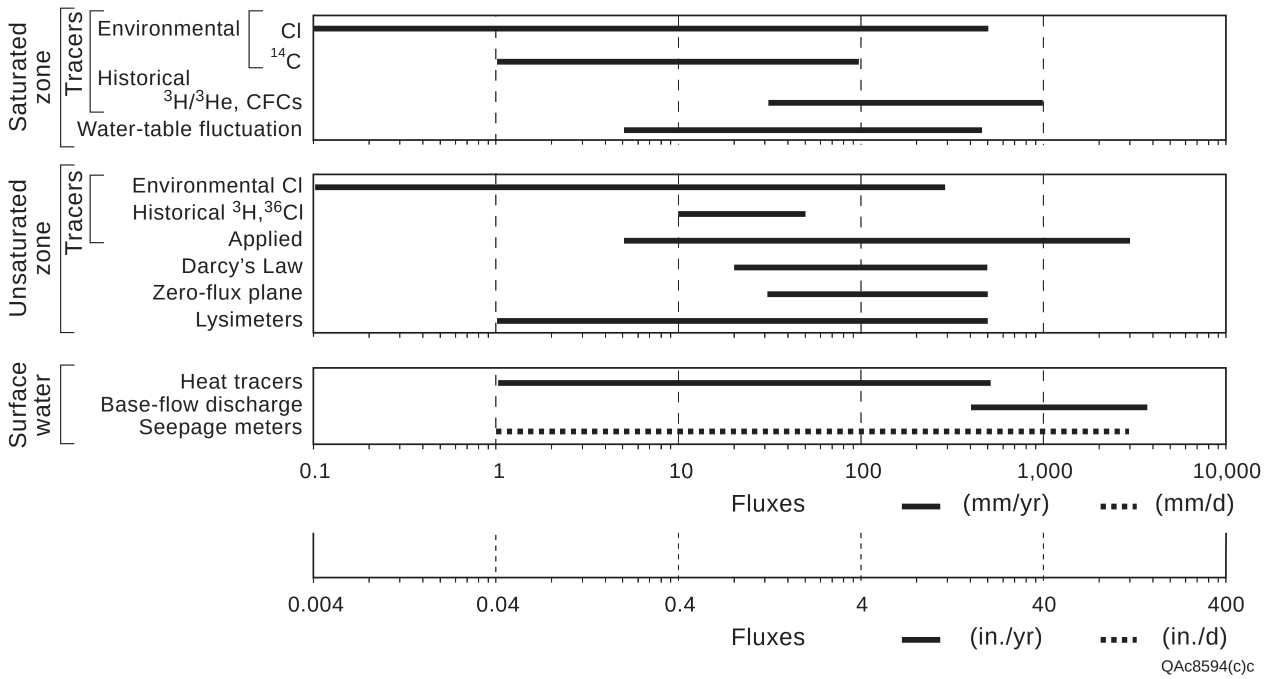

Figure 6. Range of fluxes that can be estimated using various techniques.

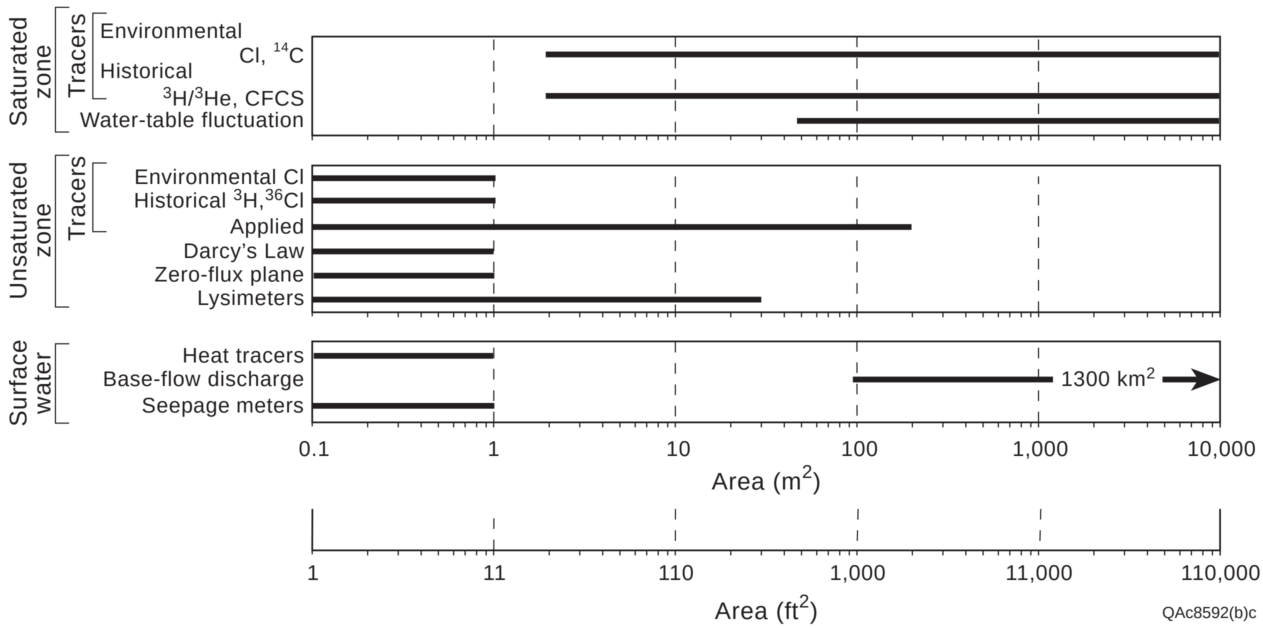

Figure 7. Spatial scales represented by various techniques for estimating recharge. Point-scale estimates are represented by the range of 0 to 1 m.

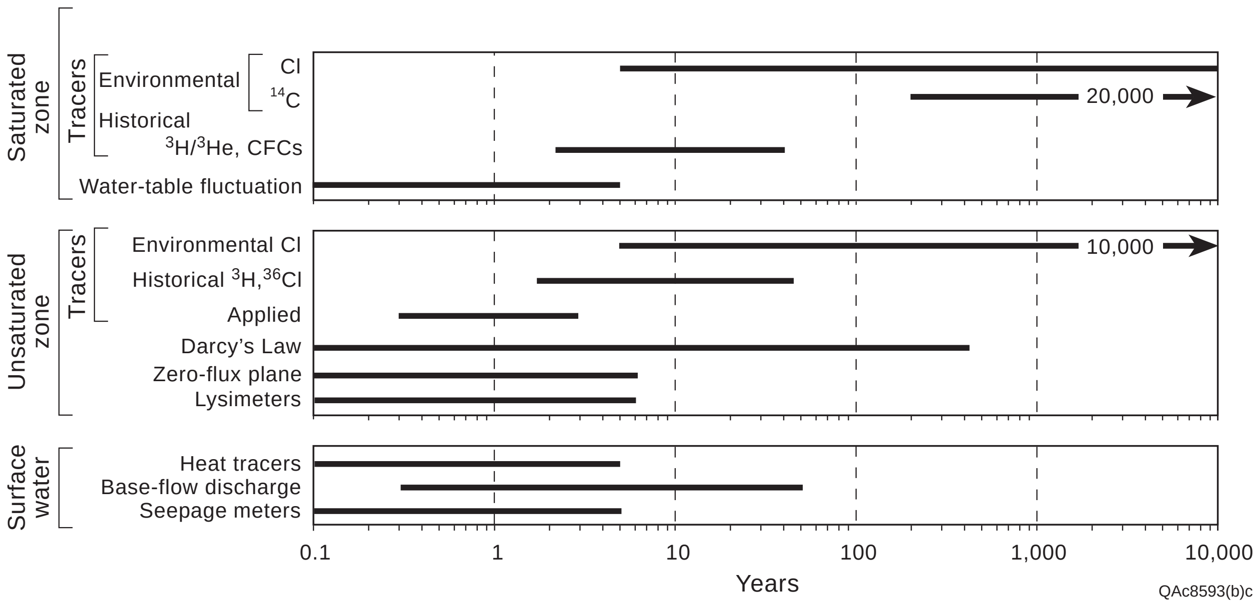

Figure 8. Time periods represented by recharge rates estimated using various techniques. Time periods for unsaturated- and saturated-zone tracers may extend beyond the range shown.

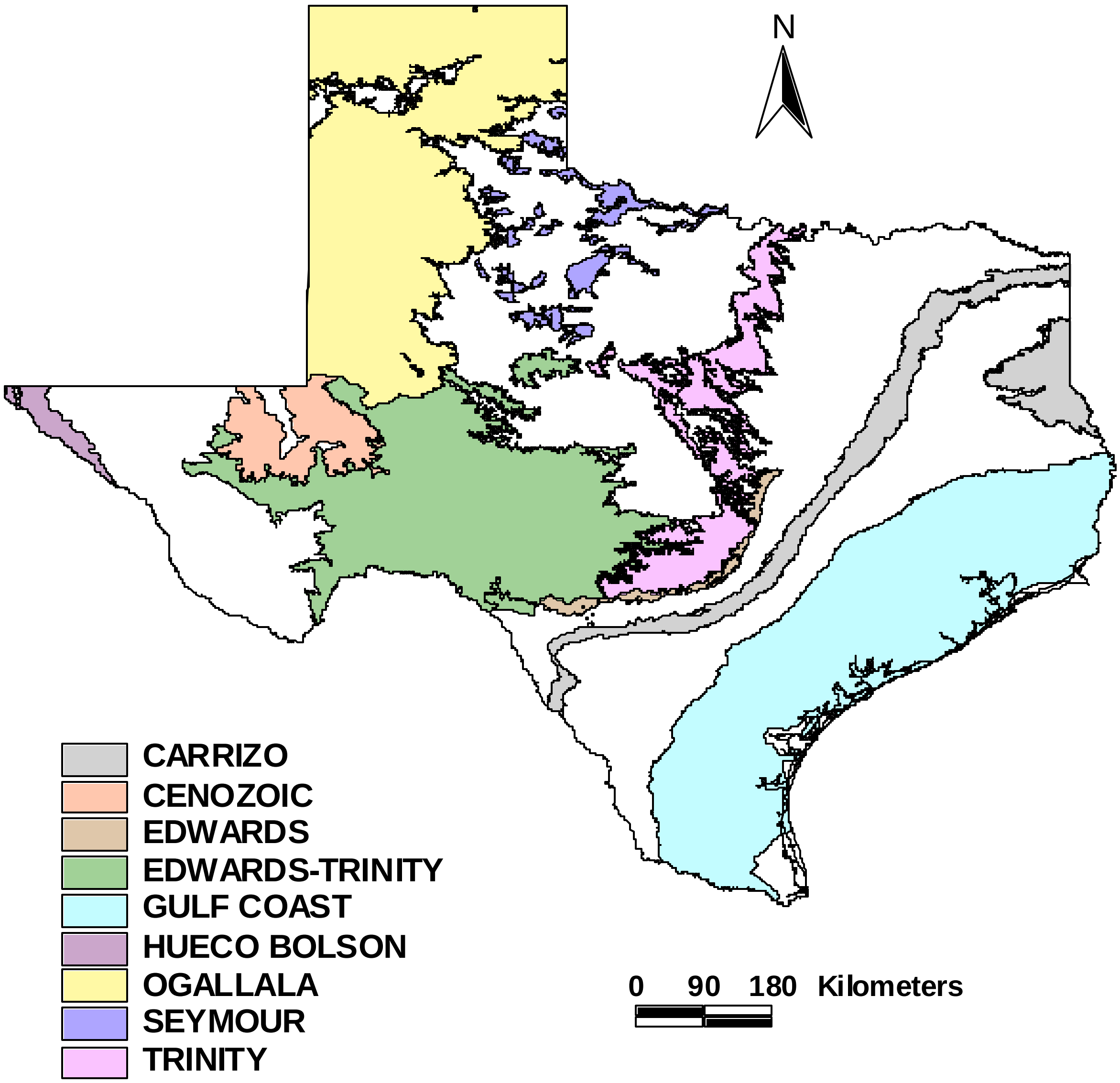

Figure 9. Major aquifers of Texas (www.tnris.state.tx.us).

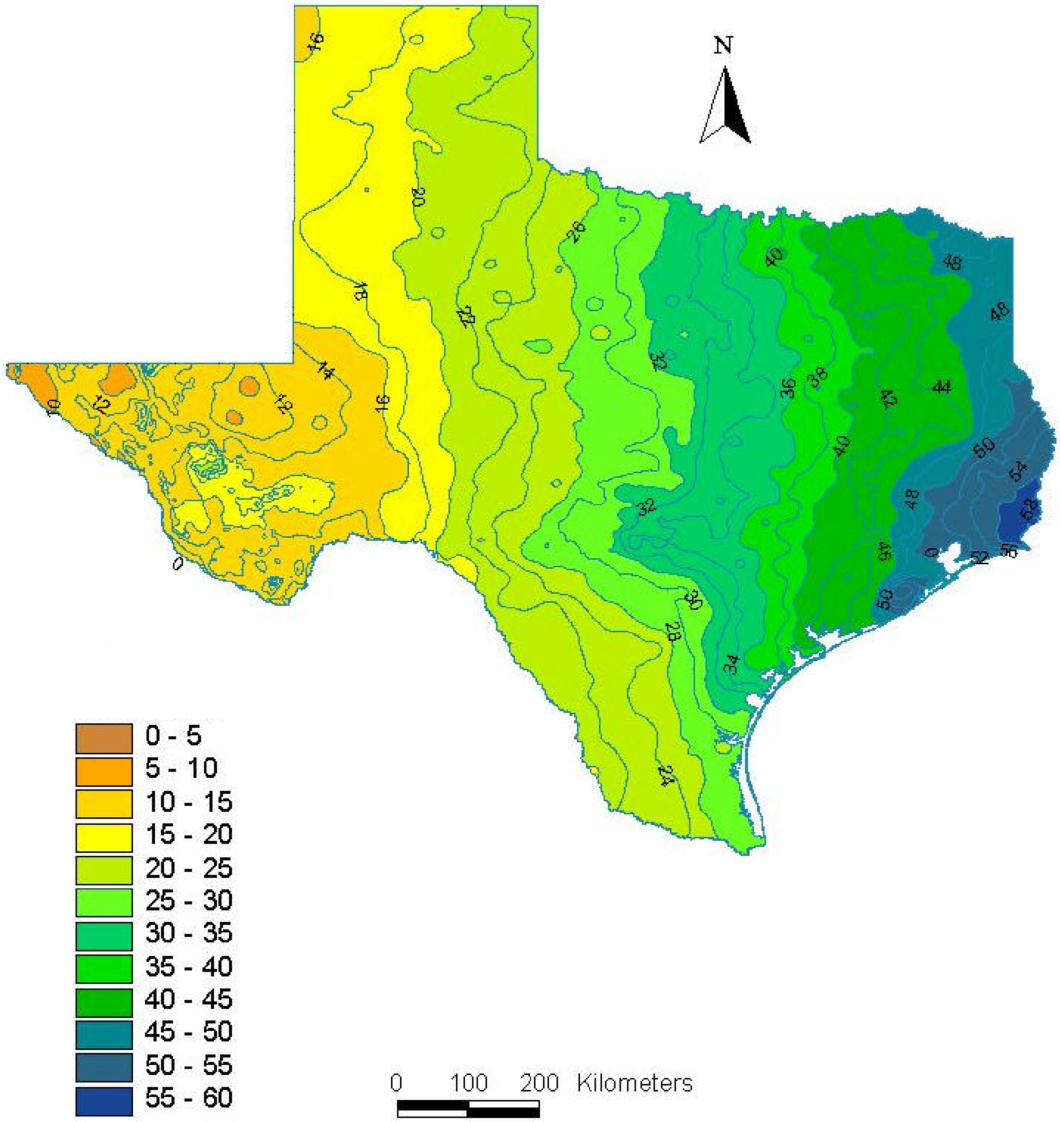

Figure 10. Distribution of mean annual precipitation in Texas (inches) (www.tnris.state.tx.us).

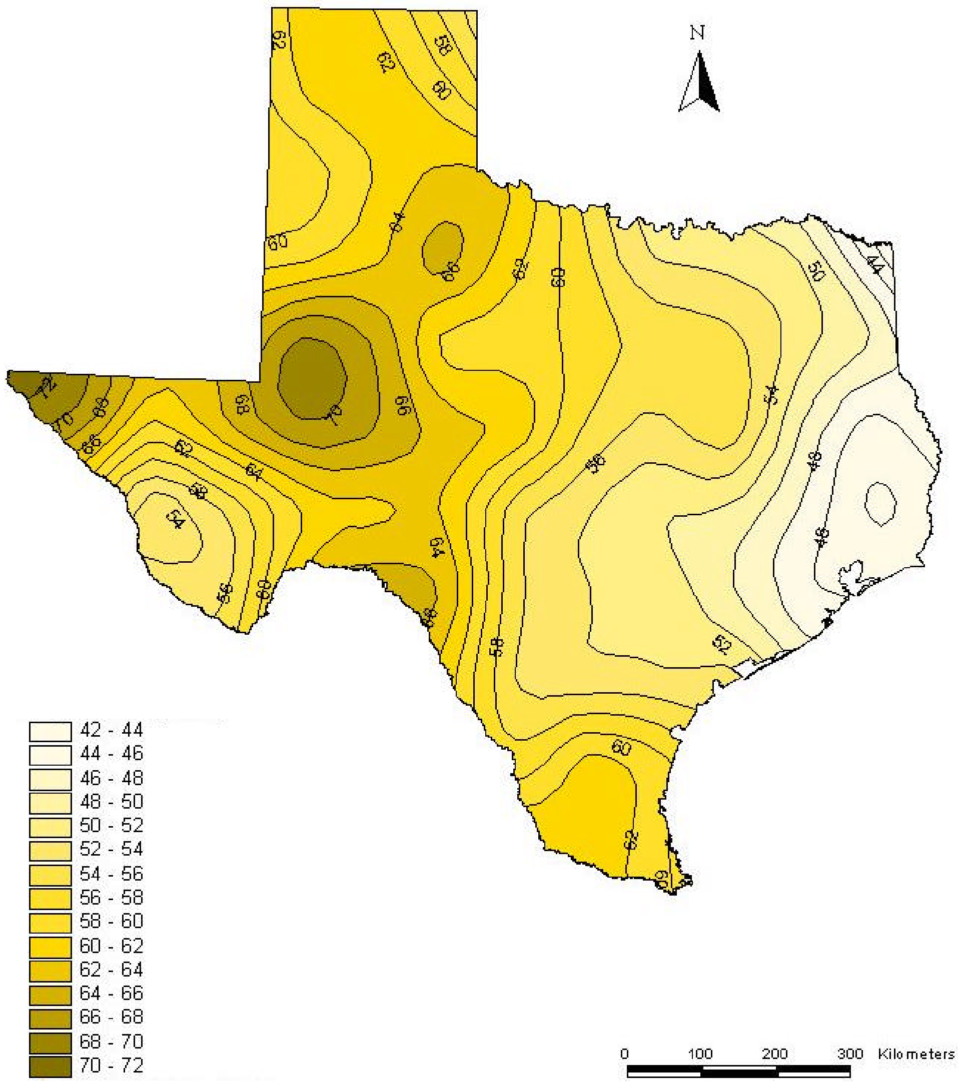

Figure 11. Distribution of mean annual lake evaporation in Texas (inches) (www.tnris.state.tx.us).

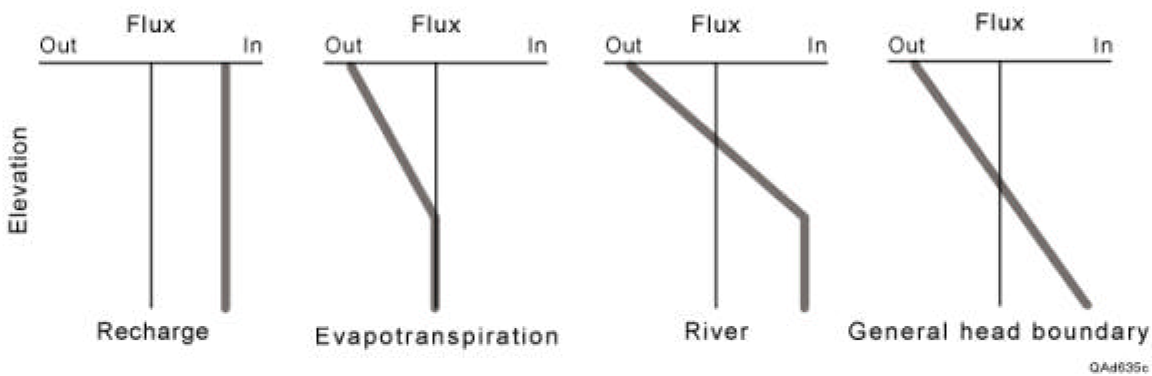

Figure 12. Schematic of various MODFLOW packages that can be used to simulate recharge and discharge.

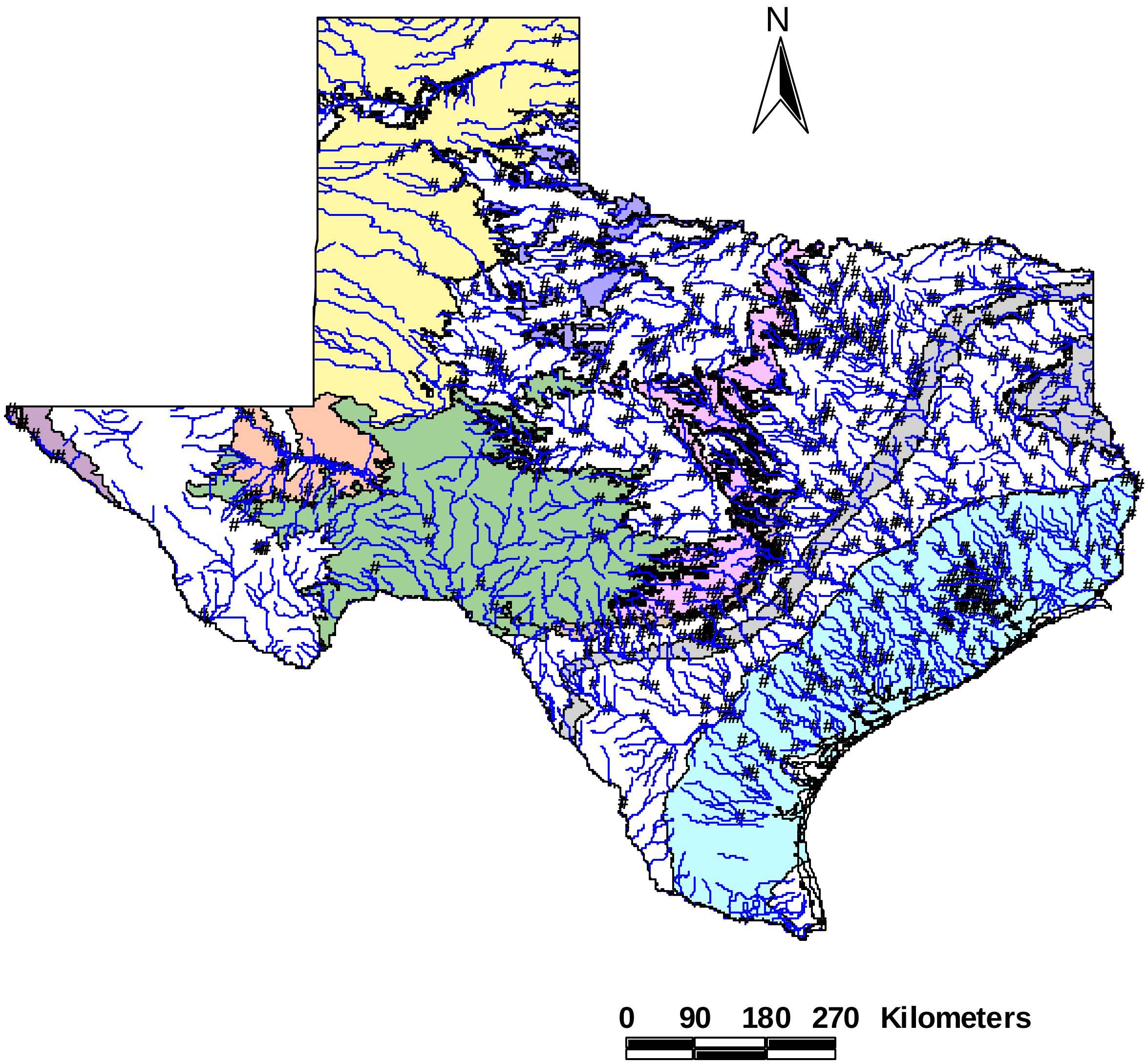

Figure 13. Distribution of gauged streams in Texas and relation to major aquifers.

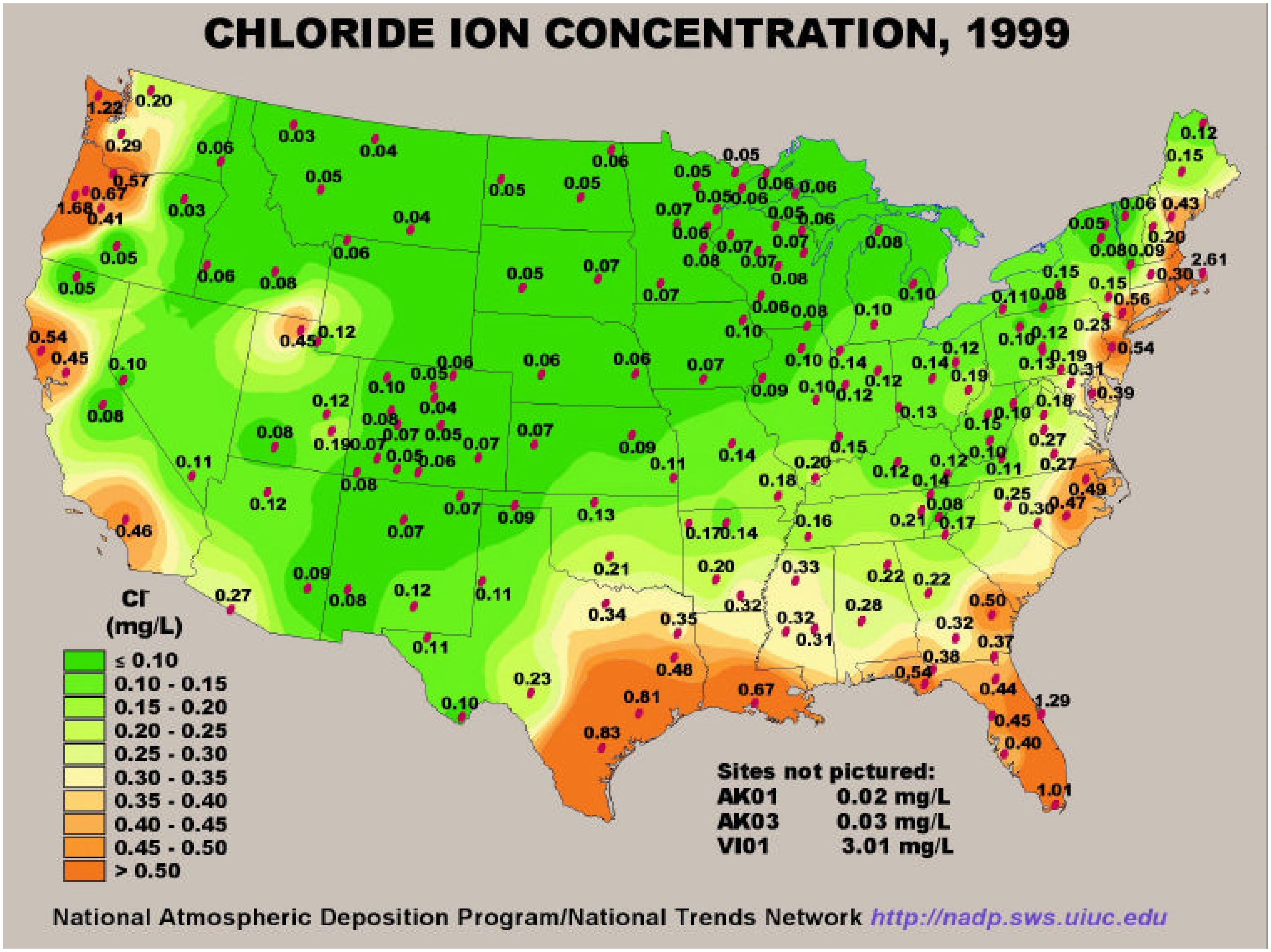

Figure 14. Wet chloride fallout in precipitation from the National Atmospheric Deposition Program.

Table 1. Review of recharge rates for the major aquifers of Texas.

Major Aquifer	Location (County/Area)	Aquifer	Recharge rate (mm/yr)	Recharge rate (in/yr)	Total recharge (af/yr)	Reference	Technique	Notes
Carrizo Wilcox	Atascosa, Frio	Carrizo sand	45.7	1.8		Alexander and White, 1966	^{14}C, Darcy's Law	
	Sabine, San Augustine	undifferentiated	50.8	2.0		Anders, 1967	Darcy's Law	
	Sabine, San Augustine	undifferentiated	25.4	1.0		Anders, 1967	baseflow discharge	
	Camp, Franklin, Morris, Titus	Carrizo Wilcox			12,000	Broom et al., 1965	baseflow discharge	
	Harrison	Cypress	7.9	0.3	15,000	Broom and Meyers, 1966	Darcy's Law	
	Harrison	Cypress	7.9	0.3	40,000	Broom and Meyers, 1966	baseflow discharge	
	Wood	Carrizo	12.7	0.5	3,000	Broom, 1968	Darcy's Law	
	Freestone	Calvert Bluff sands	100	3.9		Dutton, 1990	soil water budget	
	Bastrop, Lee, Milam	Simsboro, Carrizo	51 - 102	2.0 - 4.0		Dutton, 1999	groundwater modeling	
	Bastrop	Carrizo, Wilcox sand	38	1.5		Follett, 1981	Darcy's Law	
	Winter Garden area	undifferentiated	5 - 127	0.2 - 5.0		Guyton & Assoc. and HDR, 1998	modeling, water budget	
	Bastrop, Lee, Milam, Robertson, Falls, Limestone, Freestone, Navarro	Carrizo, Simsboro	76 - 127	3.0 - 5.0		Harden, 2000	groundwater modeling	
	Bastrop, Lee, Milam, Robertson, Falls, Limestone, Freestone, Navarro	Calvert Bluff, Hooper,	12.7	0.5		Harden, 2000	groundwater modeling	
	Bexar	Hooper, Simsboro, Calvert Bluff	45.7	1.8		HDR Engineering Inc., 2000	groundwater modeling	
	Winter Garden area	undifferentiated			100,000	Klemt et al., 1976	groundwater modeling	
	Atascosa	Carrizo	147	5.8		Opfel and Elder, 1978	neutron probe logging	
	Rusk	Carrizo	<25.4	< 1.0		Sandeen, 1987	Darcy's Law	
	Navarro	Carrizo Wilcox	12.7	0.5		Thompson, 1972	estimate	
	Caldwell, Bastrop, Lee, Milam, Robertson, Limestone, Freestone	undifferentiated	25.4	1.0		Thorkildsen and Price, 1991	groundwater modeling	
	Bastrop, Lee, Fayette	undifferentiated	25.4	1.0		Thorkildsen et al., 1989	groundwater modeling	
	Atascosa, Bexar, Dimmit, Frio, Gonzales, Guadalupe, Medina, Uvalde, Wilson, Zavala	undifferentiated			25,000	Turner et al., 1960	Darcy's Law	
	Rains, Van Zandt	Carrizo Wilcox	3	0.1	5,000	White, 1973	Darcy's Law	
Gulf Coast	Matagorda, Wharton	Beaumont, Chicot, Evangeline	0 - 10	0.0 - 0.4		Dutton and Richter, 1990	groundwater modeling	
	Duval, Jim Wells	Evangeline	1.5	0.1		Groschen, 1985	groundwater modeling	
	Aransas, Bee, Brooks, Calhoun, De Witt, Duval, Goliad, Hidalgo, Jackson, Jim Hogg, Jim Wells, Karnes, Kenedy, Kleberg, Lavaca, Live Oak, McMullen, Nueces, Refugio, San Patricio, Starr, Victoria, Webb, Willacy	Chicot, Evangeline, Jasper	0.01 - 3.0	0.0004 - 0.12		Hay, 1999	groundwater modeling	
	Colorado, Lavaca, Wharton	Chicot, Evangeline	30 - 34	1.2 - 1.3		Loskot et al., 1982	Darcy's Law	
	Jim Wells	Evangeline	<2.5	< 0.1		Mason, 1963	Darcy's Law	
	Gulf Coast					Muller and Price, 1979	groundwater modeling	
	Brooks	Evangeline			5,600	Myers and Dale, 1967	Darcy's Law	
	Harris, Montgomery, Walker	Chicot, Evangeline	< 152.4	< 6.0		Noble et al., 1996	tritium	
	Montgomery	Chicot, Evangeline, Jasper	43.2	1.7		Popkin, 1971	Transmission capacity, artificial gradient	
	Gulf Coast		18.8 (0 - 152)	0.7 (0.0 - 6.0)		Ryder, 1988	groundwater modeling	
	Washington	Jasper, Evangeline, Catahoula,	17.8	0.7		Sandeen, 1972	base flow discharge	
	Nueces, San Patricio				5,400	Shafer, 1968	Darcy's Law	
	Polk	Jasper and Evangeline	50.8	2.0		Tarver, 1968	Darcy's Law	
	Gulf Coast		0 - 16.8	0.0 - 0.7		Williamson et al., 1990	groundwater modeling	

Table 1. Review of recharge rates for the major aquifers of Texas.

Major Aquifer	Location (County/Area)	Aquifer	Recharge rate (mm/yr)	Recharge rate (in/yr)	Total recharge (af/yr)	Reference	Technique	Notes
Ogallala		Southern	0.6 - 2	0.02 - 0.08		Brown and Signor, 1973	Darcy's Law	regional
		Southern	13	0.5		Cronin, 1961	Darcy's Law	regional
			13 - 38	0.5 - 1.5		Dugan et al., 1994	water budget	regional
		Central	3.6 - 42.7	0.1 - 1.7		Dutton et al., 2000	groundwater modeling	regional
	Armstrong, Carson, Collingsworth, Donley, Gray, Hansford, Hemphill, Hutchinson, Lipscomb, Ochiltree, Potter, Roberts, Wheeler		152	6.0		Gould, 1906	observation	regional
	Lea Co., New Mexico		20.6	0.8		Havens, 1966	water budget	regional
			76 - 102	3.0 - 4.0		Johnson, 1901	observation	regional
		Central, Southern	4.8	0.2		Klemt, 1981	neutron probe logging	regional
		Central, Southern	2.8 - 5.1	0.1 - 0.2		Klemt, 1981	neutron probe logging	nonirrigated
		Central, Southern	15 - 279 (irrigated)	0.6 - 11.0		Klemt, 1981	neutron probe logging	irrigated
		Central, Southern	5.1 (1.5 - 20)	0.2 (0.06 - 0.8)		Knowles et al., 1984	groundwater modeling	regional
	Dallam, Hartley		16 - 24	0.6 - 0.9		Luckey and Becker, 1999	groundwater modeling	sand dunes
	Sherman, Moore, Hansford, Hutchinson, Ochiltree, Lipscomb		1.6 - 2.1	0.06 - 0.08		Luckey and Becker, 1999	groundwater modeling	low permeability soils
		Southern	3.3 (2.5- 25.4)	0.13 (0.1 - 1.0)		Luckey et al., 1986	groundwater modeling	regional
		Central	3.8 (1.5-	0.14 (0.06 -		Luckey et al., 1986	groundwater modeling,	regional
	Carson, Potter		205.5	8.1		Mullican et al., 1994	groundwater modeling	playa
	Carson, Potter		6.00	0.2		Mullican et al., 1994	groundwater modeling	Blackwater Draw
	Armstrong, Carson, Donley, Gray, Hemphill, Hutchinson, Potter, Randall, Roberts, Wheeler		9.00	0.4		Mullican et al., 1997	groundwater modeling	Ogallala outcrop area
	Armstrong, Carson, Donley, Gray, Hemphill, Hutchinson, Potter, Randall, Roberts, Wheeler		219.00	8.6		Mullican et al., 1997	groundwater modeling	playas Blackwater Draw
	Lubbock		41 (13 - 82)	1.6 (0.5 - 3.2)		Nativ, 1988	tritium	playa
	Carson, Potter		60 - 100	2.4 - 3.9		Scanlon and Goldsmith, 1997	chloride mass balance	playa
	Carson, Potter		0.1 - 4	0.004 - 0.16		Scanlon and Goldsmith, 1997	chloride mass balance	interplaya
		Southern	71.1 (15.2- 139.7)	2.8 (0.6 - 5.5)		Stovall et al., 2000	groundwater modeling	regional
		Central, Southern	3.2 - 17.0	0.1 - 0.7		Theis, 1937	Darcy's law	regional
			24	0.9		U.S. Bur. Reclamation, 1982		regional
		Central, Southern	2.5	0.1		Wood and Osterkamp, 1984	literature	regional
		Central, Southern	40	1.6		Wood and Osterkamp, 1984	literature	playa annulus
		N. half of Southern	11	0.4		Wood and Sanford, 1995	chloride mass balance	regional
	Lynn		77	3.0		Wood et al., 1997	tritium	playa

Table 1. Review of recharge rates for the major aquifers of Texas.

Major Aquifer	Location (County/Area)	Aquifer	Recharge rate (mm/yr)	Recharge rate (in/yr)	Total recharge (af/yr)	Reference	Technique	Notes
Trinity	Kendall		33	1.3		Ashworth, 1983	baseflow discharge	Comfort and Spring Branch on Guadalupe 1940 - 1960
	Bandera, Blanco, Comal, Gillespie, Hays, Kendall, Kerr, Medina, and Travis		38.1 (1.8 in 1956; - 116.8 in 1975)	1.5 (0.07 - 4.6)		Bluntzer, 1992	baseflow discharge	
	Dallas, Kaufman, Parker, Tarrant		111.8	4.4		Dutton et al., 1996	Cross section groundwater model	
	Bell, Brown, Callahan, Comanche, Cook, Coryell, Eastland, Erath, Hamilton, Hood, Mills, Montague, Parker, Somervell, Tarrant, Wise		1.0 to 7.6	0.04 - 0.3		Dutton et al., 1996	groundwater modeling	
	Bandera, Comal, Gillespie, Hays, Kendall, Kerr, Travis				14,000 - 53,000	Jennings et al., 2001	water-level fluctuations	
	Bell, Bosque, Brown, Burnet, Callahan, Comanche, Coryell, Eastland, Erath, Falls, Hamilton, Hill, Lampasas, Limestone, McLennan, Milam, Mills, Somervell, Williamson		30.5	1.2		Klemt et al., 1975	assumed	
	Bandera, Bexar, Comal, Gillespie, Hays, Kendall, Kerr, Travis		55.9	2.2		Kuniansky and Holligan, 1994	groundwater modeling	
	Bandera, Blanco, Gillespie, Kendall, Kerr		53 - 152	2.1 - 6.0		Kuniansky, 1989	baseflow discharge	1974 - 1977
	Kendall		55.9	2.2		Mace et al., 2000	baseflow discharge	Comfort and Spring Branch on Guadalupe, 1940 - 1997
	Bandera, Bexar, Comal, Gillespie, Hays, Kendall, Kerr, Travis		34.5	1.4		Mace et al., 2000	groundwater modeling	
	Kendall		38.1	1.5		Reeves, 1967	baseflow discharge	
	Kerr		25.4	1.0		Reeves, 1969	baseflow discharge	
Seymour	Haskell, Knox		55.9	2.2		Harden and Associates, 1978	water budget	
	Hardeman		25.4	1.0		Maderak, 1972	Darcy's Law	
	Baylor		66	2.6		Preston, 1978	baseflow discharge	
	Jones		45.7	1.8		Price, 1978	baseflow discharge	
	Wilbarger		63.5	2.5		Willis and Knowles, 1953	baseflow discharge	
Hueco-		Hueco Bolson			5,640	Meyer, 1976	groundwater modeling	mountaln-front
Mesilla		Mesilla Valley			18,000	Leggat et al., 1962	Darcy's Law	
Bolson		Mesilla Valley			3,547	Frenzel et al., 1992	empirical	mountaln-front
Cenozoic Pecos	Reeves				10,000 - 50,000	Ogilbee and Osborne, 1962	baseflow discharge	
Alluvium								
Edwards -	Kinney		35.6	1.4		Bennett and Sayre, 1962	baseflow discharge	
Trinity	Crockett		7.6	0.3		Iglehart, 1967	baseflow discharge	
	Real		50.8	2.0		Long, 1958	baseflow discharge	
	Kerr		25.4	1.0		Reeves, 1969	baseflow discharge	

Table 2. Suggested techniques for quantifying recharge to the major aquifers in Texas; E, Edwards aquifer; E-T, Edwards Trinity aquifer; CPA, Cenozoic Pecos Alluvium aquifer; GC, Gulf Coast aquifer; HMB, Hueco Mesilla Bolson; O, Ogallala aquifer; S, Seymour aquifer; and T, Trinity aquifer.

	E	E-T	CW	CPA	GC	HMB	O	S	T
Surface Water									
Channel water budget	✓		✓				✓		
Baseflow discharge		✓	✓						✓
Seepage meters							✓		
Heat tracers						✓	✓		
Watershed modeling	✓				✓				✓
Unsaturated Zone									
Zero Flux Plane	✓	✓	✓	✓	✓				
Tracers									
Cl			✓	✓	✓		✓		✓
$^{36}Cl/Cl$			✓	✓	✓	✓	✓		✓
^{3}H			✓	✓	✓	✓	✓	✓	✓
Modeling			✓	✓	✓	✓	✓	✓	✓
Saturated Zone									
Water table fluctuations			✓		✓			✓	✓
Tracers									
Cl			✓		✓	✓			✓
^{3}H		✓	✓	✓	✓	✓	✓	✓	✓
$^{3}H/^{3}He$			✓	✓	✓	✓	✓	✓	✓
CFCs		✓	✓		✓		✓		✓
^{14}C			✓			✓			✓
Modeling	✓		✓	✓	✓	✓	✓	✓	✓