Report for 2002KS6B: Groundwater Recharge in the High Plains Aquifer and Associated Aguifers of Kansas

There are no reported publications resulting from this project.

Report Follows:

GROUNDWATER RECHARGE AND WATER BUDGETS OF THE KANSAS HIGH PLAINS AND RELATED AQUIFERS

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Table of Contents

Figures

Tables

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INTRODUCTION TO GROUNDWATER RECHARGE, DISCHARGE, AND SUSTAINABILITY

As Kansas water resources become fully allocated and demand for groundwater increases, groundwater managers are faced with the difficult task of ensuring the future viability of the resource. With the rise in public environmental awareness, groundwater managers are also concerned with protecting natural environments that are dependent upon the groundwater, such as stream baseflows, riparian vegetation, aquatic ecosystems, and wetlands. Sustainable use of groundwater must ensure not only that the future resource is not threatened by overuse and depletion, but also that natural environments that depend on the resource are protected. There will always be trade-offs between groundwater use and potential environmental impacts, and therefore a balanced approach to water use between development and environmental requirements needs to be advocated. However, to properly manage groundwater resources, managers need accurate information about the inputs (i.e., recharge) and outputs (i.e., pumpage and natural discharge) within each groundwater basin, so that the long-term behavior of the aquifer and its sustainable yield can be estimated or reassessed.

Estimating recharge is critical in any analysis of groundwater systems and the impacts of withdrawing native water from them. In water-resource investigations, groundwater models are often used to simulate the flow of water in aquifers, and, when calibrated, may be used to predict long-term behavior of an aquifer under various management schemes. Without a good estimate of recharge and its spatio-temporal distribution, these models become unreliable. Accurate estimates of recharge and recharge mechanisms are also necessary to assess the risk of groundwater contamination, particularly diffuse agricultural contamination (such as from nitrates and pesticides). Clearly, understanding recharge is critical to managing most groundwater systems.

Under natural or virgin conditions and over long periods of time (before any development), groundwater recharge is balanced by groundwater discharge, i.e., *Recharge* = *Discharge*. As groundwater is nearly always moving, it will naturally flow from the recharge areas to the discharge areas. The discharge from the aquifers may occur in a variety of ways such as flow to streams, lakes, and springs; water use (transpiration) by phreatophytic vegetation that draws its water from the water table or its capillary fringe; evaporation from playas and areas of very shallow water table; leakage to adjacent aquifers; or flow to the sea.

Pumping groundwater constitutes an additional withdrawal from the system that was in a natural state of balance under virgin conditions. In order for the system to reach a new equilibrium (a state of *sustainability*), the pumping must either cause the recharge to increase, and/or it must cause the discharge to decrease. Groundwater pumping usually has little impact on the recharge, especially under arid and semiarid conditions with deep water tables, as recharge is determined mostly by climatic conditions, although in areas of intense irrigation, return flows to the underlying aquifer could be significant. Pumping, however, can decrease groundwater discharge by lowering shallow water tables, thus reducing groundwater evapotranspiration and seepage to streams, springs, lakes, or wetlands. In hydrogeologic terms, pumping can *capture* groundwater discharge. The position of the water table, which normally reflects the distribution of the recharge and discharge areas, as well as the geometry of the aquifer and its hydrogeologic properties, will change as the system adjusts to the change in discharge. Thus, declines in groundwater levels are not necessarily an indication that the sustainable yield of an aquifer is being exceeded, but simply that the water balance has been altered (Cook et al., 2001), and may reflect a temporary decrease in aquifer storage that occurs before a new equilibrium is established.

In order for a groundwater system to be *sustainable,* pumping must be balanced by an equal capture of discharge and/or recharge. If pumping exceeds the total amount of natural recharge or discharge from the system, groundwater *mining* occurs, and the system is no longer sustainable. However, even without regard to the environment, it is not always possible to extract all of the natural aquifer recharge or discharge. In some cases, wells will run dry before natural groundwater discharges are reduced to zero. The fraction of recharge that can theoretically be extracted from an aquifer under steady-state conditions will depend on the geometry of the aquifer system, and, in particular, on the location of the pumping wells relative to the natural recharge and discharge zones (Bredehoeft et al., 1982; Sophocleous, 1998a, 2000a; Bredehoeft, 2002). Therefore, the sustainable yield of aquifers, and thus the environmental impact of groundwater extraction, depends not only upon the volume extracted, but also on the location of pumping wells relative to recharge and discharge areas, and sometimes also on the timing of the extraction. Prediction of environmental impacts of groundwater extraction always requires detailed investigation of natural groundwater recharge and discharge processes.

It is important to note that all levels of groundwater extraction will, in the long run, result in declines of natural discharges, with consequent environmental impacts. Sometimes such impacts will be small and not readily identifiable, while in other cases, they may be much more dramatic, such as in the drying up of springs and streams in western Kansas. However, there will always be a *time lag* between groundwater extraction and reduction in natural discharge, and therefore the current apparent health of an exploited aquifer and the ecosystems that depend upon it does not necessarily indicate that the situation will be sustainable in the longer term (Cook et al., 2001). The task of groundwater managers is to determine what limits of environmental impact are acceptable to the community and to manage extraction to maintain impacts within those limits.

Once the groundwater system is sufficiently perturbed, even cessation of pumping will not stop the adverse impacts. The impact of pumping after it is stopped persists for a time approximately equal to the time of pumping. For example, if one pumps for a year and then stops, the impact of the pumping will persist for approximately another year—that is, it takes a year for the aquifer to recover. The time lag between groundwater extraction and reduction in natural groundwater discharge will depend on the extraction rate of groundwater relative to the natural recharge and discharge rates. For an aquifer discharging to a stream, this time lag is proportional to the square of the distance of groundwater pumping from the stream and inversely proportional to the hydraulic diffusivity of the aquifer (usually expressed as the ratio of aquifer transmissivity to storativity). For relatively large groundwater basins with low recharge fluxes, this time lag can be many hundreds of years (Sophocleous, 1998a, 2000a). In some cases, this allows groundwater extraction at rates well in excess of recharge rates to continue for a number of years before the impact of this policy can be recognized.

Changes in land use, such as intensive irrigation, often result in increased deep drainage, which creates a pressure front that moves down through the soil towards the water table (Jolly et al., 1989). Until the pressure front reaches the water table, aquifer recharge continues at the same rate as it did before irrigation development. When the pressure front reaches the water table, aquifer recharge increases, causing the water table to rise. The time lag between the increase in deep drainage and the increase in aquifer recharge is related to the deep drainage rate, the initial water-table depth, and the soil-water content within the unsaturated zone. This time lag, and thus the manifestation of impacts of land-use changes vis-à-vis groundwater recharge and discharge, can take many years to manifest.

In the following pages, the hydrogeologic framework for understanding natural recharge processes is set out in Part I, together with an outline of recharge estimation methodologies and related uncertainties and challenges facing the field of recharge assessment. A recharge-related glossary is presented as Appendix C of Part I. Part II summarizes most major recharge studies in the Kansas High Plains and associated aquifers as well as their water budgets, with emphasis on assumptions and limitations as well as environmental factors affecting recharge processes. Part III presents a conceptualization of the High Plains aquifer and its recharge characteristics. It also outlines appropriate techniques for quantifying recharge in the High Plains aquifer. Finally, in Part IV, EXCEL spreadsheets with county-by-county and districtwide recharge estimates for the Kansas groundwater management district regions and related statistics are compiled based on Kansas Geological Survey Bulletins and other publications.

PART I. UNDERSTANDING GROUNDWATER RECHARGE

Summary of Part I

This part attempts to establish a hydrogeological framework for the understanding of natural groundwater recharge processes in relation to climate, landform, geology, and biotic factors. It begins with the concepts of groundwater flow systems, which form the basis for comprehending recharge processes. This work then concentrates on the sources and mechanisms of groundwater recharge, and stresses the importance of developing correct conceptualizations of recharge. A variety of recharge estimation methodologies are then outlined, with an emphasis on minimizing uncertainty. This contribution then discusses developing predictive relationships for recharge based on the major recharge-influencing factors, and into regionalizing point recharge data. A discussion of difficulties that face the field of recharge assessment follows with recommendations to minimize these difficulties.

Although there are various well-established methods for the quantitative estimation of recharge, few can be applied successfully in the field. All methods are characterized by large uncertainties. When estimating groundwater recharge it is essential to proceed from a good conceptualization of different recharge mechanisms and their importance in the study area. Besides this conceptualization, the objectives of the study, available data and resources, and possibilities of obtaining supplementary data should guide the choice of recharge-estimation methods. A key to deciding on a recharge estimation methodology is related to the spatial and temporal scale of interest. If the major concern is obtaining good recharge estimates over a limited area, then the need for detailed information is evident. However, for regional studies small-scale variability in local recharge ceases to be a major problem. In addition, the inherent temporal variability of recharge has important implications for the measurement techniques adopted. Different measurement techniques provide recharge estimates with different temporal scales. For example, in arid and semiarid areas where deep drainage fluxes are low and water tables are deep, interpreting groundwater hydrographs and water-table rises may be misleading for estimating rates of groundwater recharge; chemical and isotopic methods are likely to be more successful than physical methods in such cases. A recharge-related glossary is presented as Appendix C.

1. Introduction and Terminology

The endless circulation of water as it moves in its various phases through the atmosphere, to the earth, over and through the land, to the ocean, and back to the atmosphere is known as the *hydrologic cycle*. This cycle is powered by the sun, and, through phase changes of water (i.e., *evaporation* and *condensation*) involving storage and release of *latent heat*, it affects the global circulation of both the atmosphere and oceans, and hence is instrumental in shaping weather and climate. The efficiency of water as a solvent makes geochemistry an intimate part of the hydrologic cycle; all water-soluble elements follow this cycle at least partially. Thus, the hydrologic cycle is the integrating process for the fluxes of water, energy, and the chemical elements. This cycle is the foundation of hydrologic science and occurs over a wide range of space and time scales.

Figure I-1 illustrates different parts of the land-based portion of the hydrologic cycle that affects an individual watershed or catchment (Freeze and Cherry, 1979). Water enters the hydrologic system as *precipitation*, in the form of rainfall or snowmelt. Water leaves the system as streamflow or *runoff*, and as *evapotranspiration*, a combination of evaporation from open bodies of water, evaporation from soil surfaces, and transpiration from the soil by plants. Precipitation is delivered to streams on the land surface as *overland flow* to tributary channels and in the subsurface as *interflow* or *lateral subsurface flow* and *baseflow* following *infiltration* into the soil.

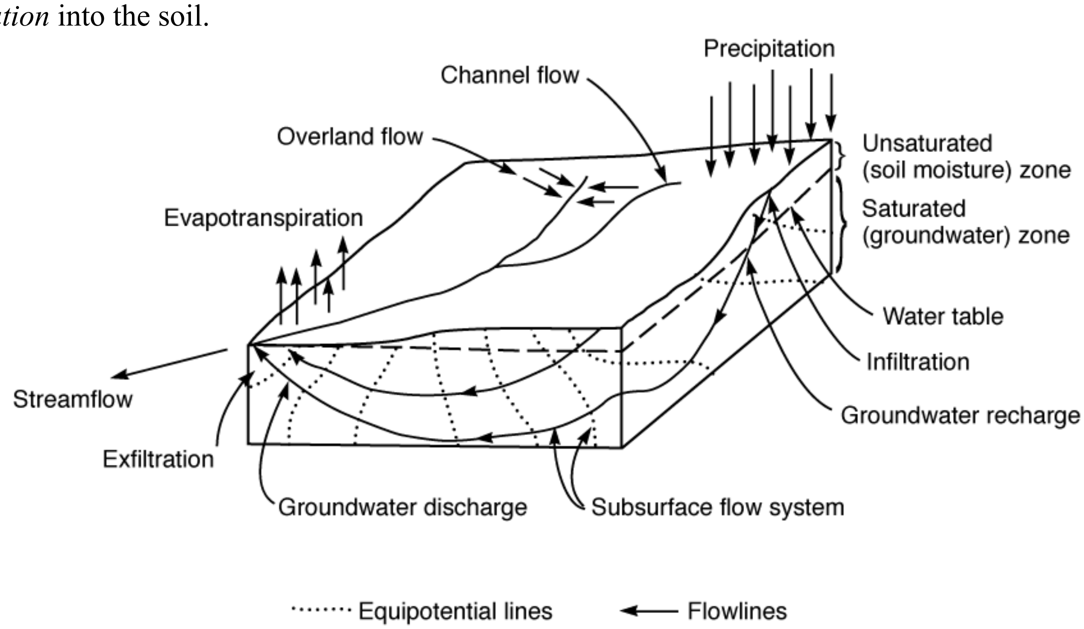

Figure I-1. Schematic representation of the hydrologic cycle (from Freeze, 1974).

A portion of the infiltrated water enters the groundwater or aquifer system by passing through the *vadose* or *unsaturated zone*, and it exits to the atmosphere, surface water, or to plants. As Figure I-1 shows, the flowlines deliver groundwater from the highlands towards the valleys or from the recharge areas to the discharge areas. As Figure I-1 also shows, in a *recharge area* there is a component to the direction of groundwater flow that is downward. *Groundwater recharge* is the entry to the saturated zone of water made available at the water-table surface. Conversely, in a *discharge area* there is a component to the direction of groundwater flow that is upward (Figure I-1). *Groundwater discharge* is the removal of water from the saturated zone across the water-table surface. The patterns of groundwater flow from the recharge to the discharge areas form *groundwater flow systems*, which constitute the framework for understanding recharge processes. Therefore, groundwater flow systems are examined next.

2. Groundwater Flow Systems

The route groundwater takes to a discharge point is known as a flow path. A set of flow paths with common recharge and discharge areas is termed a *groundwater flow system*. The three-dimensional closed system that contains the entire flow paths followed by all water recharging the groundwater system has been termed a *groundwater basin* (Freeze and Witherspoon, 1967). Groundwater possesses energy mainly by virtue of its elevation (elevation or gravitational head) and of its pressure (pressure head). Groundwater can also possess kinetic energy by virtue of its movement, but usually this energy is negligibly small because of groundwater's low velocities. Groundwater moves from regions of higher energy to regions of lower energy. A measure of groundwater's energy is the level at which the water stands in a borehole drilled into an aquifer and measured with reference to an (arbitrary) reference level or datum such as sea level. This height that water stands above a reference datum is called *hydraulic head* or simply *head*. The hydraulic head, for most practical purposes, is composed of the sum of the pressure head and gravitational or elevation head. Both of these component forms of energy (i.e. elevation energy and pressure energy) are known as *potential energy*. The change in hydraulic head over a certain (arbitrary) distance along the groundwater flow path is called hydraulic gradient or head gradient and constitutes the driving force for groundwater movement. According to *Darcy's Law*, which describes the flow of groundwater through an aquifer, the groundwater flow rate is directly proportional to the cross sectional area through which flow is occurring, and directly proportional to the hydraulic gradient. Gravity due to elevation differences is the predominant driving force in groundwater movement. Under natural conditions, groundwater moves “downhill” until it reaches the land surface, such as at a spring, or the root zone, where it is evapotranspired to the atmosphere.

Therefore, groundwater moves from interstream (higher) areas toward streams or the coast (lower areas). Except for minor surface irregularities, the slope of the land surface is also toward streams or the coast. The depth to the water table is greater along the divide between streams than it is beneath the floodplain. In effect, the water table usually is a subdued replica of the land surface.

A groundwater flow pattern is controlled by the configuration of the water table, and by the distribution of hydraulic conductivity in the rocks. The water table, in turn, is affected by the topography, and is controlled by the climate. The flow pattern is therefore a function of topography, geology, and climate. These three parameters have been collectively termed as the *hydrogeologic environment* (Toth, 1970). In addition, biotic influences affect most aspects of the hydrologic cycle, including groundwater. Vegetation, for example, regulates the rate at which a land surface returns water vapor to the atmosphere, and humans alter nearly all aspects of water on land.

Based on their relative position in space, three distinct types of flow systems have been recognized (Toth, 1963, 1999; right-hand side of Figure I-2): (1) a *local system*, which has its recharge area at a topographic high and its discharge area at the immediately adjacent topographic low; (2) an *intermediate system*, which is characterized by one or more topographic highs and lows located between its recharge and discharge areas; and (3) a *regional system*, which has its recharge area at the major topographic high and its discharge area at the bottom of the basin. Regional flow systems are at the top of this hierarchical organization; all other flow systems are nested within them.

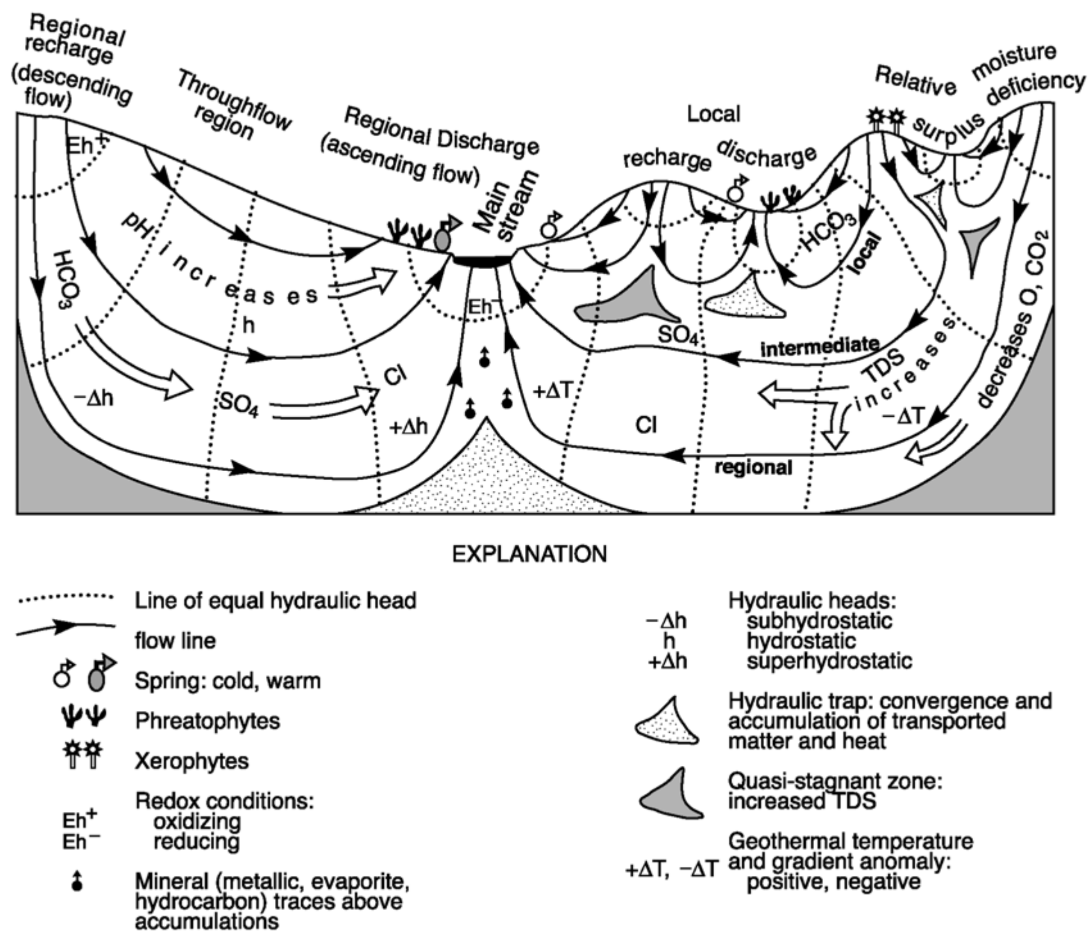

Figure I-2. Effects and manifestations of gravity-driven flow in a regionally unconfined drainage basin (adapted from Toth, 1999).

Based on a comparative study of variations in selected geometric parameters, such as depth to impermeable basement, slope of the valley flanks, and local relief, the conditions under which local, intermediate, and regional systems may develop were elucidated (Toth, 1963). If local relief is negligible, and there is a general slope of topography, only regional systems will develop. Because no extensive unconfined regional system can exist across valleys of large rivers or highly elevated watersheds, pronounced local relief generally is an indicator of a local system. The greater the relief, the deeper the local systems that develop. Under extended flat areas unmarked by local relief, neither regional nor local systems can develop. Waterlogged areas may develop, and the groundwater may be highly mineralized from concentrations of salts.

The recognition that, in topography-controlled flow regimes, groundwater moves in systems of predictable patterns, and that various identifiable natural phenomena are regularly associated with different segments of the flow systems was not made until the 1960's when the system-nature of groundwater flow was first understood (Toth, 1962, 1963; Freeze and Witherspoon, 1967). This recognition of the system-nature of subsurface water flow has provided a unifying theoretical background for the study and understanding of a wide range of natural processes and phenomena and has thus shown flowing groundwater to be a general geologic agent (Toth, 1999).

A schematic overview of groundwater flow distribution and some typical hydrogeologic conditions and natural phenomena associated with it in a gravity-flow environment is presented in Figure I-2 (Toth, 1999). On the left side of the figure, a single flow system is shown in a region with insignificant local relief; on the right side, a hierarchical set of local, intermediate, and regional flow systems is depicted in a region of composite topography. Each flow system has an area of recharge, an area of throughflow, and one of discharge. In the recharge areas, the hydraulic heads, representing the water's potential energy, are relatively high and decrease with increasing depth; water flow is downward and divergent. In discharge areas, the energy and flow conditions are reversed: hydraulic heads are low and increase downward, resulting in ascending and converging water flow. In the areas of throughflow, the water's potential energy is largely invariant with depth (the isolines of hydraulic heads are subvertical) and, consequently, flow is chiefly lateral. The flow systems operate as conveyor belts with the flow serving as the mechanism for mobilization, transport (distribution), and accumulation of mass and energy thus effectively interacting with their ambient environment (Toth, 1999).

3. Flow System Extensions

Studying flow systems in groundwater basins may help gain an understanding of the interrelations between the processes of infiltration and recharge at topographically high parts of the basin and of groundwater discharge through evapotranspiration and baseflow. For example, at least some of the water derived from precipitation that enters the ground in recharge areas will be transmitted to distant discharge points and thus cause a relative moisture deficiency in soils overlying recharge areas. Water that enters the ground in discharge areas may not overcome the upward potential gradient, and therefore becomes subject to evapotranspiration in the vicinity of its point of entry. Water input to saturated discharge areas generates overland flow, but in unsaturated discharge areas infiltrating water and upflowing groundwater are diverted laterally through superficial layers of high hydraulic conductivity. Further, the ramifications of anthropogenic activities in discharge areas are immediately apparent. Some of these include (Domenico, 1972): (1) water-logging problems associated with surface-water irrigation of lowlands; (2) water-logging problems associated with destruction of *phreatophytes*, or plants discharging shallow groundwater; and (3) pollution of shallow groundwaters from gravity-operated sewage and waste-disposal systems located in valley bottoms in semiarid basins where surface water is inadequate for dilution.

The spatial distribution of flow systems will also influence the intensity of natural groundwater discharge. From Figure I-2, the main stream of a basin may receive groundwater from the area immediately within the nearest topographic high and possibly from more distant areas. If baseflow calculations are used as indicators of average recharge, significant error may be introduced in that baseflow may represent only a small part of the total discharge occurring downgradient from the line separating the areas of discharge from the recharge areas.

In groundwater hydrology today, the system concept is fundamental to thinking about a groundwater problem. System thinking is vital to the understanding of practical problems, such as groundwater contamination from point sources, or the impact of a structure such as a dam, waste-disposal facility, or gravel pit. Many such studies suffer irreparably from the failure to place the local site in the context of the larger groundwater system of which the site is only a small part.

4. Sources and Mechanisms of Recharge

The sources of recharge to a groundwater system include both natural and human-induced phenomena. Natural sources include recharge from precipitation, lakes, ponds and rivers (including perennial, seasonal, and ephemeral flows), and from other aquifers. Human-induced sources of recharge include irrigation losses, both from canals and fields, leaking water mains, sewers, septic tanks, and over-irrigation of parks, gardens, and other public amenities. Recharge from these sources has been classified as *direct recharge* from percolation of precipitation and *indirect recharge* from runoff ponding. Other classifications include *localized* or *focused recharge*, *preferential recharge*, *induced recharge*, *mountain front recharge*, and others (Lerner et al., 1990; Simmers, 1997).

Direct or diffuse recharge is defined as water added to the groundwater reservoir in excess of *soil-moisture deficits* and evapotranspiration, by direct vertical percolation of precipitation through the unsaturated zone – that is, recharge below the point of impact of the precipitation. This mode of recharge is spatially distributed (diffuse), and results from widespread percolation through the entire vadose zone. It is typical of humid climates because frequent, regular precipitation maintains a high water content in the soil, so that there is little additional storage capacity in the vadose zone; thus, infiltration can be routed quickly through the vadose zone to the saturated zone. This recharge raises the water table, which leads to increased streamflow. Thus, in humid climates, flowing perennial streams are typically groundwater discharge areas sustained by diffuse recharge in the basin.

Indirect recharge results from percolation to the water table following runoff and localization in joints, as ponding in low-lying areas and lakes, or through the beds of surface water courses. Two distinct categories of indirect recharge are evident: (1) that associated with surface-water courses, and (2) a *localized* or *focused* form resulting from horizontal surface concentration of water in the absence of well-defined channels, such as recharge through *sloughs*, *potholes*, and *playas*. Recharge through such topographic depressions, which are common in the Canadian prairies and Great Plains of the United States, is also known as *depression-focused recharge*, and occurs where surface runoff or lateral flow of subsurface moisture accumulates within or beneath such depressions on the landscape. Thus knowledge of lateral subsurface flow processes becomes important in understanding recharge processes. In arid and semi-arid regions, localized and indirect recharge are often the most important sources of natural recharge.

Percolation to the water table from stream beds takes two forms, depending on whether there is a saturated connection between the stream and the water table. Where no connection exists (Figure I-3a), a situation typical of arid zones where water tables are generally deep, water moves downward from the stream bed to the water table, forming a groundwater mound which then dissipates laterally away from the stream. As long as the mound is recharged by unsaturated flow, there is no hydraulic connection between the groundwater and the streamflow, in the sense that the recharge rate is almost unaffected by the groundwater levels. Yet, even when the unsaturated condition is present, the stream and aquifer may in fact be hydraulically connected in the sense that further lowering of the regional water table could increase channel losses. At some critical depth to the water table, however, further lowering has no influence on channel losses. At this depth, which depends mostly on soil properties and water stage in the channel, the aquifer becomes hydraulically disconnected from the stream. If the distance from the water table to the stream stage is greater than approximately twice the stream width, the seepage begins to rapidly approach the maximum seepage for an infinitely deep water table. The parameters determining the recharge process are the width, depth, and duration of streamflow, and the hydraulic characteristics of the local material in and below the streambed.

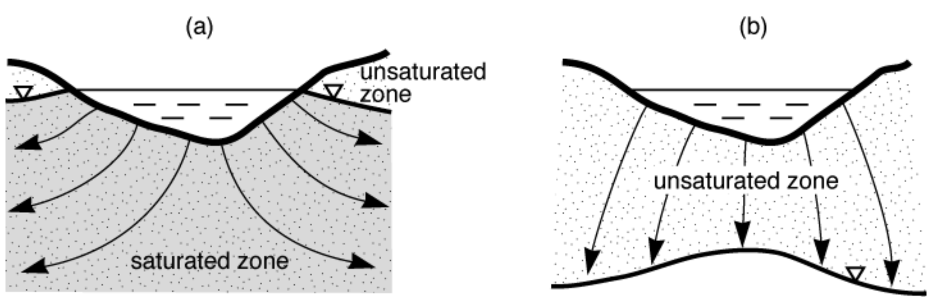

Figure I-3. Recharge from streambeds (a) with hydraulic connection, and (b) with no hydraulic connection.

In less arid areas, water-table levels tend to rise closer to the streambed. In these situations, a hydraulic connection will usually exist between the stream and the groundwater (Figure I-3b), and the recharge rate will decrease as the water table rises. The recharge process will be dominated by horizontal rather than vertical flow, and will have a much shorter *turnover* or *transit time* than when there is no hydraulic connection. In these less arid environments, there is also likely to be recharge from general catchment percolation, and the mix between the two mechanisms may be hard to predict.

Mountain front recharge typically involves complex processes of unsaturated and saturated flow in fractured rocks, as well as infiltration along channels flowing across alluvial fans. On a large scale, mountain front recharge through fractured bedrock is primarily a diffuse recharge process, whereas infiltration from mountain streams is considered a localized recharge process. Vertical leakage across low-permeability strata and underflow from adjacent aquifers (interaquifer flows) can be important sources of recharge but typically they do not involve the vadose zone.

In areas where the potential recharge rate exceeds the rate at which water can flow laterally through the aquifer, the aquifer becomes overfull and available recharge is rejected, a condition known as *rejected recharge*. In this situation, groundwater pumping in recharge areas can increase the rate of underground flow from the area and more water could be drawn into the aquifer as induced recharge.

Two different flow mechanisms, called capillary and viscous flow, drive potential groundwater recharge through the vadose zone (Hendrickx and Walker, 1997). *Capillary flow* takes place in pores with a diameter less than approximately 3 mm in which capillary forces, together with gravity, determine the flow process. A porous medium in which capillary forces are dominant behaves like a sponge; i.e. no free drainage occurs even at high water content, and capillary rise causes water to move upwards against the pull of gravity. The capillary flow process normally leads to stable wetting fronts, but sometimes unstable wetting fronts form that are characterized by *fingered flow*. (Fingered flow is unstable flow whereby the percolating water may concentrate at certain points to break into the sublayer in the form of finger-like or tongue-like protrusions.) Theoretical and experimental research results demonstrate that dry sandy soils are prime sites for the occurrence of unstable wetting fronts (i.e., boundaries between the wetted and dry regions of soil during infiltration), which may be expected in dune fields that often provide a large portion of the recharge under semi-arid conditions. This type of flow also occurs in the transition of percolating water from a fine-textured top layer to a coarser-textured sublayer. Unfortunately, it is not yet possible to quantify the effects of fingered flow on recharge rates (Hendrickx and Walker, 1997).

Macropore flow occurs in pores with a diameter or width larger than 3 mm, such as cracks in clay soils, rock fractures, fissures in sediments, solution channels, worm holes, and old root channels. In macropore flow, the effects of capillarity are no longer felt, and the flow process is dominated by viscous forces and gravity. Flow through macropores is also known as preferential or bypass *flow*, and the resulting recharge is called *preferential recharge,* which preferentially takes place through such macropores, as opposed to diffuse recharge, which takes place through the entire vadose porous medium. The velocity with which water moves from the soil surface to the water table often is several orders of magnitude higher through macropores than through the soil matrix. Saturated flow through macropores can be quantified using *Poiseuille's equation* as opposed to Darcy's equation for diffuse flow. However, capillary and macropore flow frequently occur simultaneously within the same soil mass without the presence of clearly defined macropores. The depth to which preferential flow is effective depends on the nature and connectivity of the macropores or preferred pathways, but rarely are they effective beyond the root zone depth of approximately 2 m (Hendrickx and Walker, 1997).

The process of macropore flow, shown in Figure I-4, is somewhat similar to localized recharge, albeit on a much smaller scale, because horizontal water movement is required. When the overall water input from precipitation or irrigation, *q*(t)*, exceeds the infiltration capacity of the soil, *i(t)*, a horizontal overland flow, *o(t)*, is generated that causes a water flux into the macropores, *q(0,t)*. This flux causes water content inside the macropore, *w(z,t)*, to increase. A fraction of the water, *r*, that occupies a macropore at a given depth will be absorbed by the soil matrix through the macropore walls whereas the remainder will percolate downwards into the macropore, *q(z,t)*. When the infiltration rate, *i(t)*, decreases with time and with increasing antecedent soil water content, the opportunity for overland flow, *o(t)*, and macropore flow, *q(0,t)*, increases.

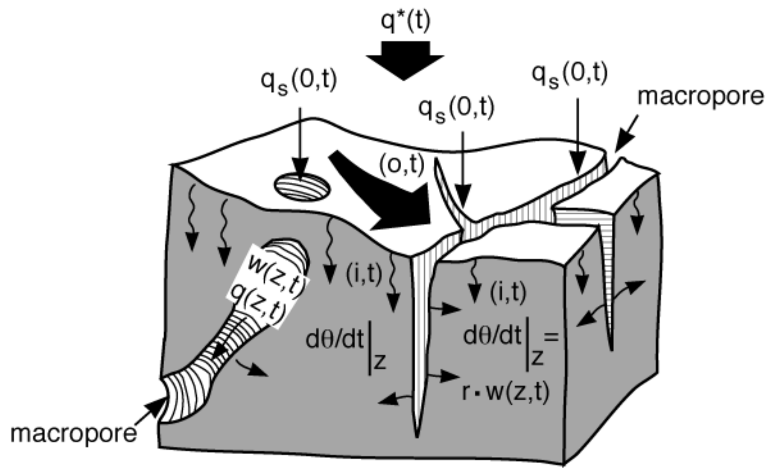

Figure I-4. Schematic representation of the fluxes involved during infiltration into a macroporous soil. See text for explanation of symbols (from Germann and Beven, 1985).

5. **Conceptual Models of Recharge** (Hatton, 1998)

The key to successful hydrological measurement and modeling is the appropriate conceptualization of the system of interest. The *conceptual model* includes the recognition of important hydrological processes, pathways, boundary conditions, spatial and temporal limits, inputs and outputs, and constraints. If the conceptual model is wrong to start with, then recharge estimates based on this model will be unreliable (Hatton, 1998). For example, at the plot scale, important elements of a conceptual water balance model aimed at recharge prediction might be (1) the pattern and amount of evaporation with respect to land cover; (2) the importance of overland flow; (3) the existence of any lateral throughflow; (4) the datum in the profile beyond which drainage will become groundwater recharge; (5) the transience and frequency of recharge events; and (6) the hydraulic pathway(s) that water may take through the profile.

At the catchment scale, the potential complexity of the correct conceptual model increases dramatically, for it includes not only all of the considerations of the plot-scale recharge phenomena, but also the distribution of these phenomena in space, as well as the interaction of the water balance components of adjacent plots (Hatton, 1998). For example, overland flow or shallow throughflow can become groundwater recharge down slope. The complexity of lateral flow systems, and their definition, becomes paramount at the catchment scale. Important considerations include (1) the presence of any *confining bed(s)*, their depth, hydraulic conductivity and distribution across the catchment; (2) the hydraulic head surface of groundwater system(s) and the degree of confinement of aquifers across the catchment; and (3) the geomorphic and geologic features associated with the discharge of groundwater, which define, locate, and control saturated areas within the catchment.

Successful estimation of groundwater recharge depends on first identifying the probable flow mechanisms and important features influencing recharge for a given locality, since it cannot be assumed that a procedure successfully developed for one area will prove equally reliable for another. Thus, in each case involving recharge estimation modeling, conceptual models must be based on local data and experience.

In summary, the vital aspects of a conceptual model of catchment recharge processes must consider (1) what parts of the landscape contribute to groundwater recharge; (2) how these areas change with time; (3) if the topographic catchment is the same as the groundwater catchment; (4) what controls recharge rates from place to place; and (5) the importance of lateral redistribution of runoff and shallow throughflow to recharge downslope (Hatton, 1998).

If appropriate, one can use a mean recharge rate over the entire catchment, or at least over that portion of the catchment subject to recharge. Expected rates or changes in the rate of recharge, however estimated or modeled, can be applied uniformly over this area. In other words, the recharge across the landscape can be treated one-dimensionally. The assumption here is that the lateral redistribution of water in the catchment takes place only after the recharge reaches the groundwater table, and that the subsequent discharge of this groundwater does not in turn change the area subject to recharge (that is, the discharge area does not grow significantly in size). The conditions where such an approach might be appropriate are areas with deep, uniformly permeable soils, deep groundwater, and a very low topographic (hydraulic) gradient (Hatton, 1998).

However, most catchments are heterogeneous in their topography, soil, geology, and land cover. To model catchment recharge in these systems, the spatial pattern of these influences on recharge must be taken into account. There are two basic ways to approach this problem, depending on the nature of the recharge modeling to be undertaken. In the first general approach, the catchment is broken up into land units in which recharge can be expected to respond similarly to climate inferences on recharge and its relation to land use; these units are then distributed spatially on this basis. In the second approach, the individual controls on recharge are distributed independently and serve as input into a spatially explicit water-balance model yielding recharge (Hatton, 1998).

In either approach, an appreciation and understanding of scaling hydrologic parameters is essential (Hatton, 1998). As one looks at larger areas of the landscape and thus incorporates natural heterogeneity into the modeling, the parameter values used to represent hydrologic processes often change. For example, the saturated hydraulic conductivity of a soil profile will be different from the inferred conductivity of a hillslope, which in turn will be different from the inferred conductivity of an entire catchment, even if climate, geology, vegetation, and other variables are held constant. Thus, it is not reasonable to assume scale invariance in model parameters as one moves from point measurements to entire catchments.

This even holds true for the mean of many point estimates of model parameters. The search for scale-invariant model representations of hydrologic phenomena for catchments has yet to yield a solution. Indeed, it is unlikely that any general scaling theory can be developed because of the dependence of hydrological systems on historic and geological perturbations (Beven, 1995). In most watershed models, equations representing hydrologic processes across scales usually involve "effective" parameters— that is, the parameter values change with scale.

Finally, in characterizing groundwater recharge, a distinction between potential and actual recharge needs to be made. *Potential recharge* is soil water that percolates below the root zone, whereas actual recharge is soil water that reaches the aquifer. Most potential recharge water will be stored in the vadose zone at negative pressures (suctions) and is not available for exploitation. In addition, it may still be lost at a later time by an increase in vegetation rooting depth, capillary rise, or upward vapor transport. Conversely, actual recharge is the amount of water that indeed reaches the water table and can be pumped.

6. Methodologies for Recharge Estimation

A number of methodologies are used to estimate recharge. These can be classified as (1) direct or indirect; (2) physical, chemical, or isotopic; (3) methods based on the analysis of inflow, outflow, or aquifer response; (4) methods based on the unsaturated or saturated zones; and (5) methods based on numerical modeling of groundwater flow, soil-water flow, both soil and groundwater flows, or modeling of the hydrologic balance at plot, field, or watershed scales. Additional classifications also exist. Within each methodology, a number of estimation techniques are available (see also Scanlon, Healy, and Cook, 2002 for a recent, comprehensive review of recharge methodologies, as well as other related articles in the special Theme Issue of the Hydrogeology Journal –vol. 10, no 1, February 2002—on groundwater recharge). Here we lump these methodologies into physical methods and tracer methods, and describe them in Appendices A and B, respectively.

7. Accuracy of Recharge Estimates

Because recharge is not easy to measure directly, estimates of it are prone to large errors. Four common types of error are discussed below (Lerner et al., 1990). The most serious and most common type of error is an error in the conceptual model. It arises when the recharge process is not fully understood, or when too many simplifying assumptions are made. For example, in a given study it may be assumed that excess irrigation water applied in parks becomes recharge, whereas in reality, a low-conductivity layer causes perching and horizontal flow to a surface drain. Or a monthly time step might be used for a soil-moisture budgeting model in a semi-arid area, resulting in zero recharge being estimated, whereas occasional short wet spells overcome soil moisture deficits to cause some recharge.

Another common error relates to temporal and spatial variability. Most recharge processes are nonlinear in relation to time. For example, a low intensity rainfall might cause no recharge because of a high rate of evapotranspiration, whereas the same amount over a shorter time period might be sufficient to saturate the soil and cause recharge. Thus, errors will arise if temporal variations are ignored—for example by using monthly, annual, or long-term average data. Recharge is also nonlinear with respect to spatial variations of inputs and physical properties of soils and aquifers.

Measurement error is another type of error and has to do with the equipment used to make measurements. This kind of error is generally not overlooked. The final type of error, calculation errors, can be avoided by care and checking, especially of units. A particularly difficult type of error can occur with numerical computer models unless they are rigorously tested under a wide range of conditions.

Error analysis or *sensitivity analysis* can show which variables in an equation lead to the highest errors, and special effort can be concentrated on obtaining the most accurate estimates for these. However, this approach will not help if the conceptual model is wrong. More than one method of estimation using other data should be used to provide an independent check. Table I-1 summarizes six different methods for estimating natural groundwater recharge from precipitation. These methods were tested and compared in a sandy till area in southeastern Sweden (Johannson, 1988). As mentioned earlier, the desired resolution in time is an important criterion in method selection. The interest may vary from estimation of instantaneous recharge to long-time averages. Table I-2 classifies the methods indicated in Table I-1 according to the time and areal resolution. Clearly, there is a need for comparative studies, in which several methods are applied to minimize the uncertainty in estimations of groundwater recharge.

Table I-1. Comparison of six different methods for estimation for groundwater recharge that were tested in southeastern Sweden[a].

METHOD/MODEL	CATEGORY	NEEDED INPUT DATA		CALIBRATION
		Climatic	*Soil moisture and groundwater*	
One-dimensional soil water flow model (SOIL)	inflow	precipitation, temperature, wind speed, relative humidity	soil water retention properties, hydraulic conductivity, groundwater outflow	measured groundwater levels
Soil moisture budget model	inflow	precipitation, temperature, wind speed, relative humidity	size of soil moisture reservoir, soil moisture-recharge relation	soil water flow model
Groundwater level fluctuations	aquifer response		groundwater levels, specific yield	
Chloride concentration	aquifer response	precipitation, wet and dry deposition of chloride	concentration of chloride in groundwater	
Spring discharge	outflow		spring discharge, size of catchment area	
Catchment area model (PULSE)	outflow	precipitation, temperature, potential evapotranspiration	size of soil moisture reservoir, soil moisture-recharge relation, outflow from the groundwater reservoir	spring discharge

[a]From Johansson, 1988. Methods for estimation of natural groundwater recharge directly from precipitation—Comparative studies in sandy till; in, Estimation of Natural Groundwater Recharge (ed. I. Simmers): Dordrecht: Reidel, 239-270.

Table I-2. Classification of the applied methods for estimation of groundwater recharge (shown in Table I-1) according to the resolution in time of their results. A dashed line indicates point values of groundwater recharge and a solid line indicates an areally integrated value[a].

	TIME SCALE					
METHOD/MODEL	Instantaneous	Events	Monthly	Seasonal	Annual	Long-time average
One-dimensional soil water flow model (SOIL)	– – –	– – –	– – –	– – –	– – –	
Soil moisture budget model	– – –	– – –	– – –	– – –	– – –	
Groundwater level fluctuations	– – –	– – –				
Chloride concentration						– – –
Spring discharge					– – –	– – –
Catchment area model (PULSE)	– – –	– – –	– – –	– – –	– – –	

[a]From Johansson, 1988. Methods for estimation of natural groundwater recharge directly from precipitation—Comparative studies in sandy till; in, Estimation of Natural Groundwater Recharge (ed. I. Simmers): Dordrecht: Reidel, 239-270.

8. Factors Influencing Recharge, Predictive Relationships, and Recharge Regionalization

8.1 Factors Influencing Recharge

The key environmental factors controlling recharge are climate, soils and geology, vegetation and land use, topography, and depth to water table. The water-balance equation is commonly used to quantify the components of the hydrologic cycle:

$$P + I = RO + D + ET + S \tag{I-1}$$

where P is precipitation, I is irrigation, RO is surface runoff, D is deep drainage and recharge, ET is evapotranspiration, and S is water stored in the soil. Under nonirrigated conditions, where $I = 0$, the left-hand side of equation (I-1) is fixed in the sense that it is outside human control. Hence, a decrease in any of the variables on the right-hand side forces an equal increase in the other terms to maintain the equality (i.e., the water balance). For example, a decrease in surface runoff, RO (e.g., as a result of increased infiltration through better tillage practices) may increase the amount stored in the soil profile, S; an increase in S would tend to increase deep drainage (and recharge), D, and evapotranspiration, ET. Clearly, to understand and estimate aquifer recharge, a basic understanding of this water cycling is needed.

Most direct measurements of hydrologic variables related to recharge assessments provide only point measurements or estimates and do not integrate such variables (shown in equation (5)) in relation to space and time. Recharge varies across the landscape because the aforementioned controlling factors vary, but finding ways to estimate and predict this spatial and temporal variability and to regionalize point measurements remains a major problem in recharge assessments (Sophocleous, 1992).

A daily water balance modeling analysis (based on equation (I-1)) for the Rattlesnake Creek basin in south-central Kansas, an approximately 1300-mi^2 semiarid to subhumid agricultural basin, demonstrated that soil factors, plant cover, and land-use practice are important controls on groundwater recharge (Sophocleous and McAllister, 1987, 1990). The importance of each of these factors is detailed below. Although such results are highlighted for the case study of the aforementioned Kansas basin, they are general enough to be valid for any agricultural plain region of similar climate in the world.

Soil factors, such as the *available-water capacity* (AWC) of soil profiles, exert a dominating influence. AWC is the volume of water available to plants if the soil were at *field capacity*, i.e., the moisture content held by soil against the pull of gravity after the excess water has drained out of a saturated or nearly-saturated soil. The AWC of each soil determines the maximum limit of actual evapotranspiration (ET) that can be extracted without additional infiltration and the maximum *soil-moisture deficit* possible. (Soil-moisture deficit is an estimate of the degree to which soil moisture content has dropped below field capacity.) Thus, given the same hydroclimatic conditions and crop cover, a soil with a relatively low AWC will exhibit a relatively small soil-water deficit, and smaller amounts of water will be lost through ET compared to losses from a soil with higher AWC (Figures I-5 and I-6). The AWC also determines the amount of water that can infiltrate into the soil before *deep drainage* occurs. The AWC acts as a buffer for infiltrating water. Given the same initial-moisture conditions, a soil with higher AWC can absorb more infiltrating water than low-AWC soils. Thus, deep drainage decreases with increasing AWC (Figure I-7).

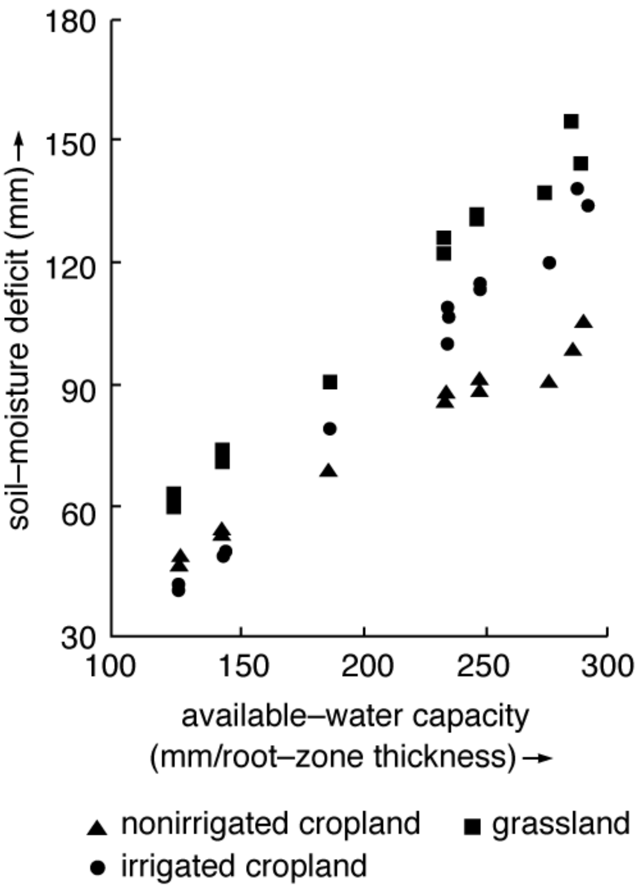

Figure I-5. Soil-moisture deficit versus available water capacity for grassland, dryland, and irrigated cropland for the upper two-thirds of the Rattlesnake Creek basin in south-central Kansas (from Sophocleous and McAllister, 1987).

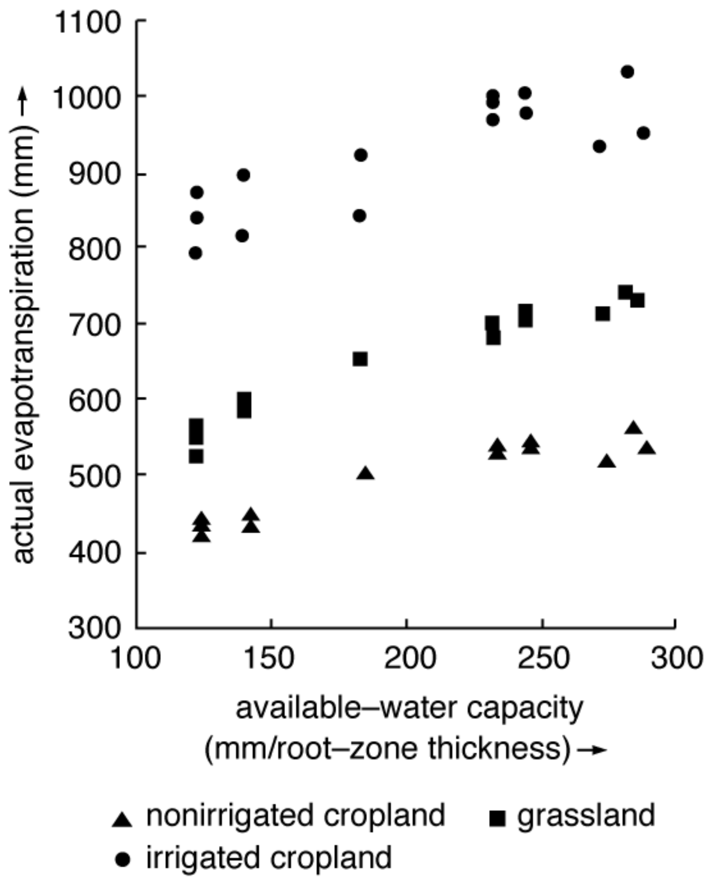

Figure I-6. Actual evapotranspiration versus available-water capacity for grassland, dryland, and irrigated cropland for the upper two-thirds of the Rattlesnake Creek basin in south-central Kansas (from Sophocleous and McAllister, 1987).

The water balance also is greatly influenced by the plant cover and land-use practice. The largest element of the water balance in equation (I-1) in the south-central Kansas basin under study is the ET component, as can be seen for native grassland in Figure I-8. The impact of vegetation on the hydrologic balance is complex and depends on factors such as *crop coefficients* (i.e., empirically determined coefficients relating *potential ET* to crop ET), growth stages, rooting depths, soil, water, and climatic conditions as used in soil-water balance simulation model.

The crop coefficients vary with the stage of crop growth. Mature plants have greater ability to extract soil moisture from all soil horizons and, thus, have larger crop coefficient than young plants. The crop with the largest crop coefficients employed in the soil-water balance model in south-central Kansas is alfalfa. In addition, alfalfa is continuously grown from one year to the next with multiple harvests without replanting or land *fallowing*. Prairie grasses have the next highest overall crop coefficients in the study region with a long growing season. All other crops have lower crop coefficients and are grown only part of the year.

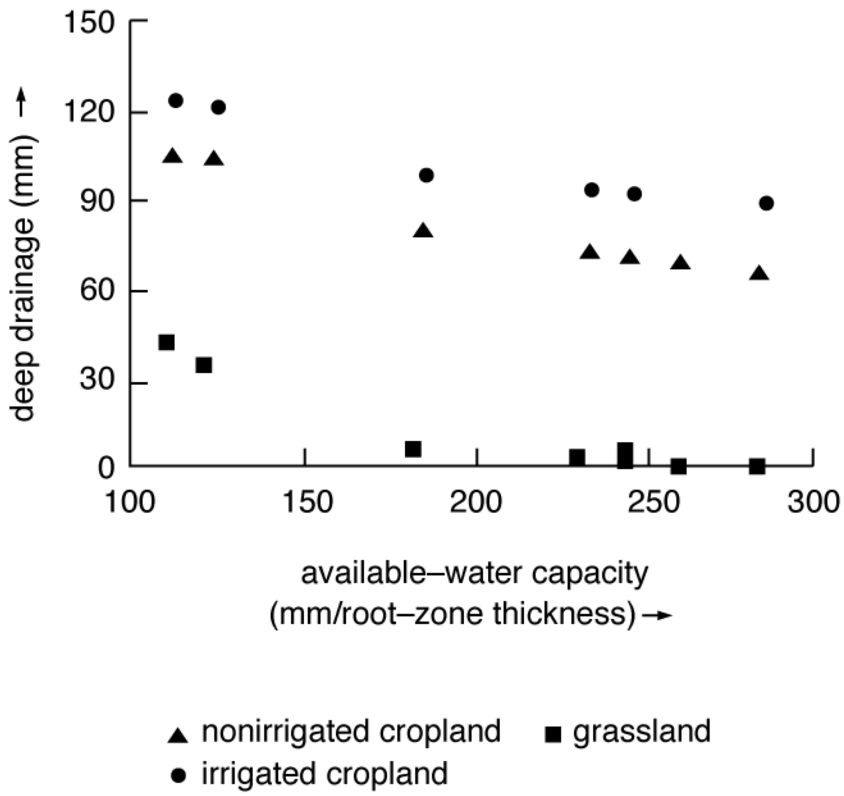

Figure I-7. Deep drainage versus available water capacity for grassland, dryland, and irrigated cropland for the lower one-third of the Rattlesnake Creek basin in south-central Kansas (from Sophocleous and McAllister, 1987).

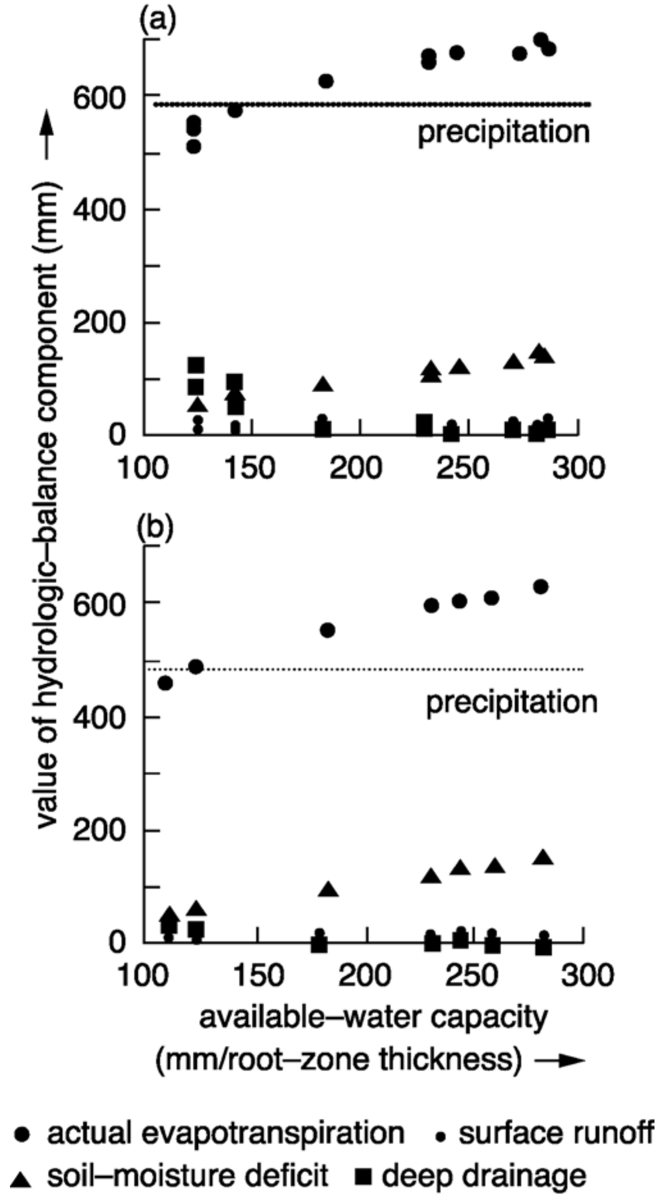

Figure I-8. Grassland water-balance components for (b) the lower one-third and (a) the rest of the Rattlesnake Creek basin in south-central Kansas (from Sophocleous and McAllister, 1987).

The greatest deep drainage occurred in irrigated wheat fields, mainly because of the shallow rooting depth of wheat, while the lowest values occurred in alfalfa and grassland acreages (Figure I-9). Decreased amounts of deep drainage in the northeastern portion of the Kansas study basin (Figure I-8), where precipitation is lower, are from grasslands, indicating the dominant effect precipitation and vegetation exert on deep drainage.

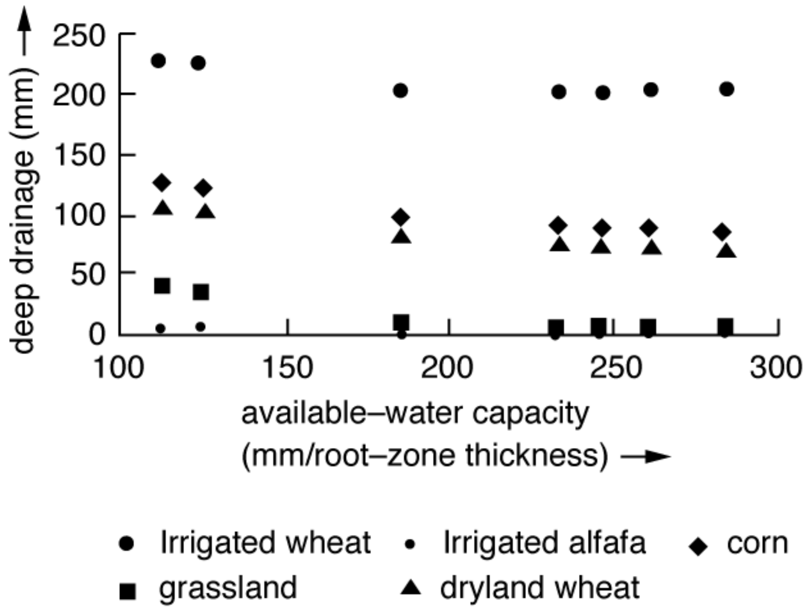

Figure I-9. Effects of vegetation on deep drainage in the lower one-third of the Rattlesnake Creek basin in south-central Kansas (from Sophocleous and McAllister, 1987).

In general, the effect of irrigation is to significantly increase evapotranspiration and also increase deep drainage, as can be seen in Figures I-6 and I-7, respectively. For summer crops, such as sorghum and soybeans, as well as alfalfa, most of the irrigation amounts are spent in evapotranspiration activities, with negligible amounts for deep drainage. The effects of grasslands are reduced deep drainage and runoff, and increased soil-moisture deficits compared to cropland acreages (Figures I-5, I-6, and I-7). From the areal distribution of the various components of the water balance, it was concluded that single average values of hydrologic variables used in management practices are not realistic, and that a spatial-discrimination attempt in managing water resources is needed (Sophocleous and McAllister, 1987, 1990).

A computerized water-balance procedure such as the one used in the Rattlesnake Creek basin study can be used to demonstrate and predict human and natural impacts on the hydrologic cycle (Sophocleous and McAllister, 1987, 1990). The hydrologic effects of vegetation changes, weather modification, extreme weather conditions, and so on can be readily estimated during the planning process using the methodology of combining classification techniques (to identify hydrologically "homogeneous" unit areas within the heterogeneous basin) and water balance modeling employed in the Rattlesnake Creek basin study in Kansas. Thus, had the Rattlesnake Creek basin been entirely covered by prairie grasses, as it probably was during predevelopment time, and had the 1982-83 precipitation pattern and amount prevailed (which is about 10% below average), the overall basin deep drainage is model-predicted to have been 1.13 inches/yr, compared to less than 0.15 inch/yr if alfalfa were planted exclusively in the basin. If the entire basin were planted with dryland wheat under 1982-83 precipitation conditions, the overall basin deep drainage is model-predicted to have been 5.1 inches/yr. Such figures can be arrived at by multiplying the deep-drainage amounts for the corresponding crop and soil complex by the planted area, summing up these figures, and dividing by the area of interest. Similarly, the hydrologic effects of manipulating the proportion of various crops and the amounts of irrigation within any soil-association area can thus be assessed (Sophocleous and McAllister, 1987, 1990).

Provided that future precipitation patterns can be established, then, under known vegetation and land-use practices, various components of the water balance, such as deep drainage and surface runoff within the basin, can be predicted using the presented methodology. An example of the relative effects of an approximately 19 percent precipitation difference on the components of the water balance, keeping the precipitation time-pattern constant, is shown in Figure I-8. This figure represents actual grassland data from the northeastern portion of the Kansas basin study area, which received 18.9 inches of annual precipitation (Figure I-8b) and the rest of the basin, which received an annual precipitation average of 23.3 inches (Figure I-8a). Note the large increase in deep drainage in the higher precipitation region, especially in low-AWC soils, compared to the deep drainage in the lower precipitation region.

8.2 Predictive Relations and Recharge Regionalization

Although there have been numerous studies to estimate recharge in specific areas, there has been no systematic attempt to develop generic, predictive relationships for quantifying recharge based on the aforementioned controlling environmental factors. This is important for groundwater management and protection. Recharge studies in the agricultural plains of central Kansas resulted in the development of such an approach based on classification and statistical analyses and taking advantage of Geographic Information Systems (GIS) capabilities for "mapping" recharge and its controlling factors (Sophocleous, 1992, 2000c). Because of the generality and applicability of the methodology to most semiarid to humid plain regions of the world with relatively shallow water table, the Kansas case study will be outlined below.

Although geostatistical methodologies and multivariate statistical techniques such as cluster analyses are useful tools for regionalizing point measurements, the usually small number of experimental sites for recharge estimation precludes usage of such techniques. To develop practical relationships between annual recharge and easily measured, independent recharge-controlling factors for the south-central Kansas plains (Great Bend Prairie region), advantage was taken of recently completed multi-year (1985-1992) field-based recharge assessment studies at 10 sites in that region. As a result, a number of multiple regression analysis models were developed depending on the number of controlling factors considered. Most of the 10 recharge-assessment field sites were located in grassland and adjacent to irrigated cropland fields. This analysis (Sophocleous, 1992, 2000c) showed that, given the vegetation cover considered, the most influential variables in recharge estimation were, in order of decreasing importance, annual precipitation (PCP; major climatic variable), average maximum soil-profile water storage (AWC; major soil variable), average shallowest depth to water table (DTW; major groundwater condition variable), and average springtime precipitation rate (RATE; secondary climatic variable).

Each of these factors then was used to zone the region for recharge estimation and was mapped as a separate GIS layer or coverage. Thus, four GIS (ARC-INFO) data layers were constructed for the region based on the results of the multiple regression analysis as shown in Figure I-10. Each data layer was classified into the same number of data classes (six in all) and assigned a class rank. Then the overlays were combined to produce a master map of "homogeneous" zones. GIS technology is ideally suited for such overlay analysis.

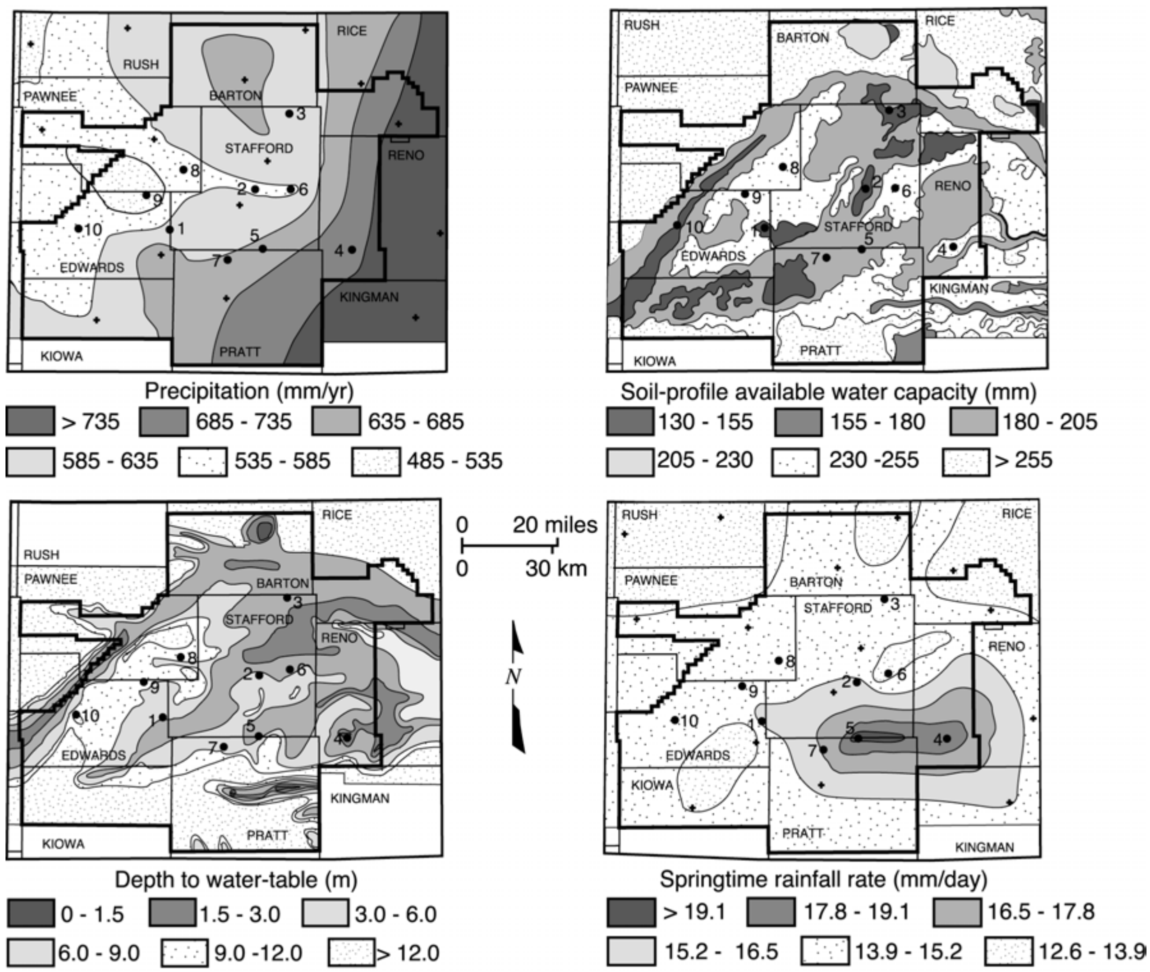

Figure I-10. Four recharge-related GIS coverages for the Great Bend Prairie region of south-central Kansas. Solid circles indicate recharge-assessment sites. Crosses indicate climatic stations (adapted from Sophocleous, 1992).

An ARC-INFO overlay analysis procedure was conducted to identify areas of differing recharge in the south-central Kansas study region. The regression coefficients of the developed multiple regression models, normalized to 1, were used to weigh the class rankings of each recharge-affecting variable. Based on this classification scheme, an area-wide recharge map (Figure I-11) indicating five differing recharge regions was derived. The recharge zonation agreed well with the field-estimated recharge values at the sites (Sophocleous, 1992, 2000c).

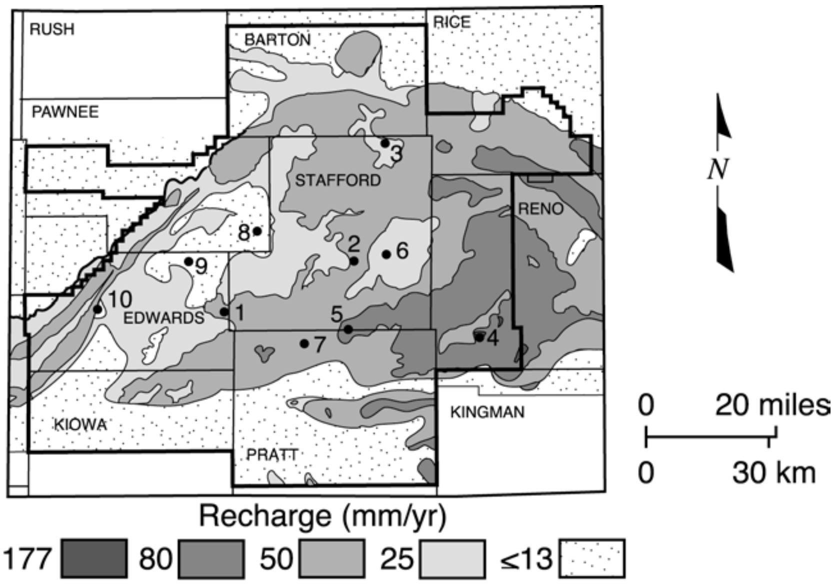

Figure I-11. GIS coverage showing recharge zonations for the Great Bend Prairie region of south-central Kansas. Solid circles indicate recharge-assessment sites (adapted from Sophocleous, 1992).

In addition, the fact that the GIS-based regression estimates for recharge were of the same magnitude as other independent estimates, even during an extreme flooding period during the summer of 1993 (Sophocleous et al., 1996), attest to the robustness of the methodology, although additional tests are desirable. It was concluded that the combination of multiple regression and GIS overlay analyses is a powerful, robust (even under extreme conditions), and practical approach to regionalizing small samples of recharge estimates.

9. Difficulties and Challenges in Recharge Estimation

Quantification of the rate of natural and human-induced groundwater recharge is a basic prerequisite for efficient groundwater resource management, and is particularly vital in arid and semi-arid regions where such resources are often the key to economic development. However, the rate of aquifer replenishment is one of the most difficult factors in the evaluation of groundwater resources to measure.

Although there are various well-established methods for the quantitative estimation of recharge, few can be applied successfully in the field. A 1988 international recharge-estimation workshop (Simmers, 1988) concluded that "no single comprehensive estimation technique can yet be identified from the spectrum of methods available; all are reported to give suspect results."

Difficulties in reliably quantifying groundwater recharge stem from a variety of factors. These include the limited capability to identify and quantify the probable recharge mechanisms and important features influencing recharge for a given locality, the nonlinear recharge response with time, the highly variable areal distribution of groundwater recharge, the scarcity of hydrogeologic data, and the complexities of the hydrologic balance in general.

Because of these uncertainties, project designs and management strategies need to be flexible enough not to require radical change if initial predictions prove wrong, due to incorrect assumptions about recharge rates and other hydrogeological factors (Foster, 1988). Groundwater recharge estimation must be treated as an iterative process that allows progressive collection of aquifer-response data and resource evaluation. In addition, more than one technique needs to be used to verify results.

When estimating groundwater recharge, one must start with a good conceptualization of different recharge mechanisms and their importance in the study area. To identify the probable flow mechanisms, the field evidence must be examined carefully. The recharge mechanism at a particular location can depend on a variety of factors, which may be different from the influencing factors at another location. Therefore, just because a method has successfully estimated the recharge in one locality, one should not assume the same method can be used elsewhere, even if the situation appears to be similar (Rushton, 1988). Once the recharge mechanisms have been defined, calculations can be carried out to estimate the recharge. In addition to being based on a good conceptualization, the choice of methods should be guided by the objectives of the study, available data and resources, and possibilities of obtaining supplementary data.

Consideration of the following questions may facilitate recharge estimation (Lerner et al., 1990). (1) How much recharge can the aquifer accept? A full aquifer will reject further water, which must then find another destination. (2) How much water can the unsaturated zone transmit? High potential recharge rates (for example, from rivers or irrigation canals) may not be able to pass through low conductivity layers. (3) What other destinations are there for potential recharge, and how large are they? (4) How much potential recharge is there? (5) What is the actual recharge? This step considers the balance and destinations of all water from the source, based upon the first four questions. (6) How do other estimates compare? As previously mentioned, whenever possible, more than one method should be used.

The key to deciding on a recharge estimation methodology is the spatial and temporal scale of interest. If the major concern is obtaining good recharge estimates over a limited area (e.g., for waste disposal or local water supply purposes), then the need for detailed information is evident. In this situation multiple site investigations are needed, which also require identification of preferential flow contributions. Conversely, for projects on a regional scale, or those requiring only preliminary recharge estimates, groundwater-based methods (such as those involving interpretation of fluctuations in groundwater levels) are relevant and small-scale variability in local recharge ceases to be a problem (Simmers, 1997).

The inherent temporal variability of recharge has important implications for the measurement techniques adopted (Cook, 1993; Scanlon, Healy, and Cook, 2002). Different measurement techniques provide recharge estimates with different temporal scales. For example, applied tracers and lysimeters are only able to provide information on recharge over the period of measurement, usually no more than a few years. Meteorological water-balance techniques, and those involving interpretation of fluctuations in groundwater levels, likewise can provide information only on recharge over the period of record. Chloride displacement techniques provide a mean recharge rate since the change in land use, while bomb tracers give a mean recharge rate since peak fallout (a period of about 40 years). Chloride mass balance methods have a much longer temporal scale, typically in the order of hundreds to thousands of years (dependent on the recharge rate and the thickness of the unsaturated zone). The techniques adopted will depend upon the purpose of the study. Where interest is in estimating long-term recharge rates, a long temporal scale of measurement is desirable. On the other hand, if interest is on the effect of land management on recharge, those techniques with smaller temporal scales are required.

In areas where the annual variability of recharge is very high, measurement techniques with long time scales will be required to estimate the long-term mean annual recharge rates with any accuracy. Where the annual variability of recharge is lower, measurement techniques with shorter time scales will be suitable.

The temporal variability of the soil water flux should decrease with depth (Cook, 1993). If the recharge rate is sufficiently low, and the water table sufficiently deep, then below some depth, the temporal variability of drainage will approach zero. At these depths, even the measurement of the soil water flux over a short time scale should be sufficient to infer the long term drainage rate. This decrease in temporal variability of soil-water flux with depth may cause problems in estimating recharge from hydrograph records. If much of the temporal variability is lost during passage through the unsaturated zone, then these methods may underestimate the recharge rate. For example, the coefficient of variation (CV) of annual drainage below the root zone of Banksia woodland (33 ft deep) was found to be 64%, whereas the CV for drainage at 66 ft (water table recharge level) was 11%. Thus, in areas where deep drainage fluxes are low and water tables are deep, groundwater-based methods may be inappropriate for estimating rates of groundwater recharge. In particular, methods which involve interpretation of hydrograph records will underestimate recharge. Individual recharge events will not usually be seen as rises in water tables if the time lag is greater than a few days. Seasonal variations in drainage will likewise not be reflected in water table variations if the time lag is much greater than a few months.

Tracer methods seem the most reliable for point recharge measurements in arid areas. The best method will depend on the magnitude of recharge flux. Chloride appears most reliable over drainage rates from less than 0.04 to 4 inches/yr. At deep drainage rates of more than 4 inches/yr, measurement errors and anion exclusion may become important. Bomb tracers ^{3}H and ^{36}Cl are suitable for recharge rates greater than 0.8 inch/yr. The accuracy of drainage estimates obtained with natural tracers should generally not be assumed to be better than ±50%.

Temporal variability of recharge is related to temporal variability of precipitation. The variability of annual recharge increases rapidly as the mean annual recharge decreases. For mean annual recharge of 1.2 inches/yr, measurements over at least 15 to 20 years have been suggested (Cook, 1993).

When recharge rates are only a few millimeters per year or less, chemical and isotopic methods are likely to be more successful than physical methods, such as water balance methods, which rely on measured or estimated values of water flux. Water-flux estimates are often in error by as much as one order of magnitude or more, especially when measuring physical parameters in the drier ranges. An advantage of tracers is that they integrate all the processes that combine to affect water flow in the unsaturated zone. Tracer behavior is generally a much more robust indicator of water movement in a porous medium than is the solution of the equations of water flow, especially when soils are relatively dry (e.g., for arid sites).

In estimating recharge, a method such as ^{3}H-profiling relies on estimating the amount of the tracer beneath the soil surface; thus, the precision of the estimate of recharge will increase with recharge rate. In contrast, for a tracer such as Cl, the concentration of which is inversely proportional to recharge rate, the precision of the estimate will increase with decreasing recharge rate. Combining Cl (tracer) and suction profile (physical) information, at sites where changing land use has significantly altered recharge rates is useful in quantifying both past and present recharge rates at such locations (Allison et al., 1994).

Indirect, physical approaches, such as water balance and Darcy flux measurements, were found the least successful, whereas methods using tracers (e.g., Cl, ^{3}H, and ^{36}Cl) have been found to be the most successful in estimating groundwater recharge in arid regions. Of the tracer techniques available, Cl-balance techniques appear to be the simplest, least expensive, and most universal for recharge estimation.

Nevertheless, advances in recent years show that the value of water-balance and Darcian methods should not be underestimated. Reliability of water-balance methods for recharge estimation depends on the precision with which the water balance components have been determined. In arid and semiarid regions, application of this method is more difficult than in humid regions because precipitation is frequently only slightly different from actual evapotranspiration; small errors in these two components thus cause large errors in recharge estimates. To minimize such errors, one can use a combination methodology such as the hybrid water fluctuation methodology, which uses a storm-by-storm water balance analysis in combination with analyses of vadose zone moisture and water table fluctuations (see section A1.1). However, the hybrid water fluctuation methodology is applicable predominantly to relatively flat, semiarid to humid regions with relatively shallow water table.

Despite their importance for groundwater management and protection, no generic, predictive relationships for quantifying recharge based on the major controlling factors of climate, soils, vegetation, and land use have been usefully developed. In addition, the problem of regionalizing point measurements, given the spatial and temporal variability of recharge and aquifer heterogeneity, remains a serious one. Although a methodology to address such problems has been developed (see Section 8.2), additional studies and approaches are needed to tackle these challenges.

APPENDIX A. Physical Methods for Recharge Estimation

Physical methods rely on direct measurements of hydrological parameters or on estimates of soil and/or aquifer physical parameters. Physical methods are frequently used to estimate precipitation recharge because they are quick, inexpensive, or straightforward. However, these methods are often problematic in arid and semi-arid regions. There are several reasons for this (Hendrickx and Walker, 1997): 1) The low recharge fluxes depend to a large extent on the vadose zone physical parameters, and significant variations in fluxes may occur with small changes in these physical parameters. Unfortunately, it is almost impossible to detect such small changes in physical parameters. 2) The extreme temporal variability of arid climates means that long time series are needed to assess mean annual recharge rate. 3) Spatial variability caused by changes in local topography, soil type, and vegetation requires a large number of measurement sites to assess the spatially averaged recharge rate. Nevertheless, with prudent appreciation for their limitations, physical methods can be a helpful tool for evaluating precipitation recharge.

A1 Indirect Physical Methods

Indirect physical methods for estimating groundwater recharge consist of (1) empirical methods, (2) water balance methods based on estimates of soil physical properties, and (3) numerical modeling methods. In principle, one of the simplest methods used for estimating diffuse recharge, R, is empirical expressions of the type

$$R = k_1(P - k_2), \tag{A-1}$$

where P is precipitation, and k_1 and k_2 are constants for a particular area. Such expressions have been used with varying degrees of success and are probably most useful for making first-guess estimates of recharge where annual recharge is fairly high, >2 inches/yr, and thus should seldom be used in arid or semiarid regions (Allison et al., 1994).

Methods relying on estimates of soil physical parameters generally fall into the following classes: (1) soil water balance, (2) zero-flux plane method, (3) estimation of water fluxes beneath the root zone using unsaturated hydraulic conductivity and the gradient in soil water potential, and (4) estimation of water fluxes in the saturated zone based on Darcy's law and flow-net analysis. Additional methods also exist that may not fit well into one of these classes, such as gravity surveys for measuring changes in aquifer storage resulting from recharge events. Increased accuracy in measuring temporal variations in the Earth's gravity field has recently allowed the use of gravity observations to deduce subsurface water mass changes resulting from precipitation and consequent recharge events.

A1.1 Water Balances

This group of methods estimates recharge as the *residual* of all other fluxes. The principle is that other fluxes can be measured or estimated more easily than recharge. Examples of water balance methods include (1) soil moisture budgets, in which rainfall and potential evapotranspiration are inputs to a soil-moisture accounting procedure, with actual evapotranspiration and recharge as the outputs; (2) river-channel water balances, when upstream and downstream flows are differenced to calculate recharge or—more accurately—*transmission losses* (a related stream hydrograph analysis technique based on baseflow-separation techniques is founded on steady-state water-balance calculations, whereby the estimate of discharge based on baseflow-separation or baseflow-recession analysis of the stream hydrograph, must equal recharge. This technique is considered too empirical and approximate to give reliable quantitative estimates); and (3) water-table rises, when the volume stored beneath a rising water table is equated to recharge, after allowing for other inflows and outflows such as pumping wells and aquifer throughflow. The simplicity of the latter method made it a popular one. However, calculating the volume of water stored between lowest and highest water table positions over a study period interval involves reliable estimates of the aquifer's *specific yield* values, which may be difficult to obtain.

The advantages of water-balance methods (Lerner et al., 1990) are that they use readily available data (rainfall, runoff, water levels), are easy to apply, and they account for all water entering the system. The major disadvantage is that recharge is the residual or remainder of all other hydrologic components in the water balance equation, and constitutes only a small difference between large-number components, such as precipitation and evapotranspiration. Errors can be high, with the errors in all the other fluxes accumulating in the recharge estimate. For example, high river flow can often only be estimated to ±25 percent. If recharge is 25 percent of flow, the error in estimating it is ±100 percent. Other disadvantages include the difficulty of estimating other fluxes in the water balance equation. For example, evapotranspiration cannot be measured easily, yet it is often the largest outward flux. Physical properties like specific yield are central to some water-balance methods, such as the ones based on water-table rises, but are not easily defined or measured.

The natural time scale for water-balance methods is the duration of a recharge event. Recharge processes are often nonlinear, so that estimates based on longer time intervals should be summed over the individual events rather than calculated for the whole interval at once. Long records are available for much of the data used for balances (rainfall, runoff), so that long time series of recharge can often be calculated.

A methodology consisting of a combination of soil moisture budget and water table rise analyses is known as *hybrid water fluctuation method* (Sophocleous, 1991). This combination methodology, designed to minimize water balance errors, was successfully applied in the central Kansas plains, which are characterized by semiarid to subhumid climate, relatively flat terrain, and relatively shallow water table.

The hybrid water fluctuation methodology can be summarized as follows (Sophocleous, 1991). Neutron-moisture profile readings are collected onsite at least once a week during the recharge season (usually spring and fall). The soil-water balance methodology for each recharge-producing storm period is applied, and the resulting water table rise is noted, provided it is confirmed that the water-table rise is due to incoming soil water from above, as checked with tensiometer readings and/or deeper water content measurements. The recharge estimate resulting from the soil-water-balance is then divided by the corresponding water table rise to obtain an estimate of effective storativity or fillable porosity of the region near the water table. Several such estimates are obtained and averaged. This average is, in effect, the site-calibrated effective storativity value, which can be used to translate each water table rise, tied to a specific storm period, into a corresponding amount of groundwater recharge. In the central Kansas prairies, which are characterized by mostly permeable sandy soils and shallow water table, the time lags between the occurrence of a recharge-causing rainstorm and the corresponding water-table rise usually range from less than a day to just a few days.

Recharge estimation errors in the hybrid water fluctuation method are reduced by running a storm period-based soil water balance throughout the year, in combination with the associated water-level rise, thus avoiding masking short periods of recharge by the averaging effect of monthly or larger time-interval data. Furthermore, during the recharge-producing rainstorm periods under consideration, the evapotranspiration (ET) estimates are usually significantly smaller than the precipitation amounts. Therefore, even a large ET error on a relatively small quantity may not significantly affect the recharge estimate. Also, by employing the Complementary Relation Areal Evapotranspiration (CRAE) methodology, which permits areal ET to be estimated from its effects on the routinely measured temperatures and humidities, the soil-plant system complexities can be avoided, as well as the need for locally calibrated coefficients. (The CRAE concept takes into account interactions between the evaporating surfaces and the overpassing air, whereby a decrease in the availability of water for areal ET causes the overpassing air to become hotter and drier, which in turn causes the potential ET to increase.) In addition, the errors inherent in the soil-water balance approach can be appreciably reduced by corroborating the estimation of recharge with increases in soil-water content at depth, and with unequivocal fluctuations of the water table, provided it is relatively shallow. Furthermore, close monitoring of shallow and deeper hydraulic gradients from multi-level piezometers makes it possible to ascertain whether water table rises are due to lateral inflow at the site or to vertical accretion from rainfall percolation. This combined methodology results in better and more reliable recharge estimates than either the soil-water budgeting procedure or the water-table-rise analysis used singly and does not require additional difficult-to-measure variables (Sophocleous, 1991).

A1.2 Zero-Flux Plane Method

The zero-flux plane (ZFP) method relies on locating a plane of zero hydraulic gradient in the soil profile. Recharge during a time interval is obtained by summation of the changes in water content below this plane. Unfortunately, the method breaks down in periods of high infiltration when the hydraulic gradient becomes positive downward throughout the profile. This is when recharge fluxes are likely to be highest. Use of this technique can give good estimates of recharge for periods during the year when the ZFP exists.

A1.3 Estimation of Unsaturated Water Fluxes

Several studies have reported use of unsaturated zone hydraulic conductivity, $K(\theta)$, or, $K(\psi)$, and water retention data, $\psi(\theta)$, to solve either Darcy's Law or Richards' equation in the unsaturated zone and to estimate soil water flux for periods of months to years (Allison et al., 1994). If the water flux is calculated at such a depth in the profile that no further extraction by roots occurs, then the flux will be equal to groundwater recharge:

$$R = K(\theta)\Delta H_T, \tag{A-2}$$

where ΔH_T is the total head gradient. For most soil systems, $H_T = H_g + H_m$, where H_g is the gravity head and H_m is the *matric suction* head.

Both $K(\theta)$ and $K(\psi)$ relationships are difficult and time consuming to determine, both in the field and in the laboratory, with difficulty and uncertainty increasing with soil dryness. Slight differences in measured water content translate into large differences in unsaturated hydraulic conductivity. As a result, the annual recharge flux could vary significantly, depending on how the mean hydraulic conductivity is computed.

A1.4 Estimation of Saturated Water Fluxes

An equivalent method for recharge estimation based on saturated flow governed by Darcy's Law is simpler, especially when assuming steady state conditions and employing flow-net analysis. The only measurements needed are values of hydraulic head and hydraulic conductivity to construct a quantitative flow net. A *flow net* consists of a set of intersecting lines of equal hydraulic head values (known as equipotential lines) and flow lines representing two-dimensional steady flow through a porous medium (see Figures I-1 and I-2). Two-dimensional, vertical flow nets constructed along the general groundwater flow direction from water table and hydraulic head field measurements provide an approximate but straightforward way of identifying areas of recharge and discharge and estimating recharge.

A1.5 Numerical Models for Estimating Recharge

Different types of models are available for estimating groundwater recharge: (1) numerical models that solve one-, two-, or three-dimensional forms of the water flow or Richards equation; (2) parametric hydrologic models that use a numerical or analytical relationship between infiltration or precipitation and recharge; (3) groundwater flow models; and (4) combined or integrated watershed and groundwater models.

Numerical modeling methods take transient flows and storage changes into account and can include spatial variability of physical properties, of which hydraulic conductivity is one of the most important. However, data requirements and computing load are high. Such models are used to estimate model parameters, in this case recharge, based on known values of hydraulic head. Such an approach is known as a solution of an inverse problem. This is in contrast to the forward or direct problem, where model parameters are considered known, and hydraulic head is computed.

Should one possess the analytical expressions for hydraulic head and transmissivity in the groundwater flow equation, determination of recharge would have been a trivial exercise of calculus in computing the derivatives of the groundwater flow equation. However, hydraulic heads are always measured with inaccuracies. Differentiating such noisy data leads to large errors in recharge estimation.

Integrated watershed and groundwater models allow a complete analysis of the land-based hydrologic cycle, thus providing the means for evaluating the impacts of land use, irrigation development, and climate change on both surface-water and groundwater resources (Sophocleous and Perkins, 2000). Such models allow predictions of the impact of management changes on total water supplies, including recharge. The seasonal variation of water-table levels and recharge can be more accurately predicted by the soil-moisture accounting system employed in the integrated model than by using only a groundwater model. This increased flexibility, however, comes at the expense of increased complexity and expertise needed to effectively use integrated watershed modeling. Although integrated models require extensive data, such integrated modeling constrains the adjustment of model parameters during calibration because overall water budgets must be observed. Whereas traditional methods used to calibrate groundwater models may include adjustments to recharge rates, in an integrated model, recharge is completely constrained by the overall water budget for the surface-water system. In addition, stream-aquifer interactions, including stream-derived recharge, are constrained by the generated amount of surface runoff to streams that in turn, impacts the stream stage and thus the driving forces for stream-aquifer interaction.

The principal advantages of the numerical methods are that they attempt to represent the actual physical processes of interest and that they allow predictions of future recharge regimes resulting from different land uses and climatic changes. These advantages are often countered by the need to make simplifying assumptions in order to reduce the computational effort. For example, numerical models of the soil zone usually assume a single porosity medium with no spatial variation in properties. In practice many soils may have dual porosity, with preferred pathways during high saturation—that is, at times of recharge.

The correct time scale for such models depends on the rate of fluctuation of heads, varying from seconds for rainfall into soil to seasonal or longer for seepage between aquifers. Effectively addressing the multiple temporal (as well as spatial) scales involved in recharge estimation constitutes a major problem in modeling recharge processes. In addition to such obstacles and uncertainties, large data requirements often make application of numerical models difficult.

A2. Direct Physical Methods

In contrast to the numerous indirect physical methods, there is only one direct method for estimating diffuse recharge. This involves the construction of a lysimeter. Lysimeters comprise enclosed blocks of disturbed or undisturbed soil with or without vegetation that are hydrologically isolated from the surrounding soil in order to assess or control various terms of the water balance. There is also only one direct method for estimating indirect recharge associated with surface water bodies in direct hydraulic connection with an underlying aquifer. This involves the use of seepage meters that are inserted in the streambed or lakebed that could provide direct point measurements of localized recharge.

Lysimeters are expensive and permanent instruments (Allison et al., 1994). They are typically filled with disturbed soils, which generally have water content profiles that differ in some degree from those found in surrounding soils. Drainage can occur only when a water table develops at the base of the lysimeter, unless solution samplers (e.g., ceramic cup extractors) and a vacuum system are installed at the base of the lysimeter. This last factor, however, is unlikely to be a problem if the lysimeter is relatively deep and the vegetation is shallow rooted. While lysimeters have been useful in quantifying drainage at waste sites under arid conditions, they have limited ability to document the spatial variability produced by natural and human-induced changes in surface and subsurface flow pathways. Construction cost and logistics limit size and depth to generally no more than a few square feet and a 10-ft depth although lysimeters as deep as 60 ft have been constructed (Allen et al., 1991; Gee et al., 1992). Because lysimeters are effective for the study of recharge mechanisms and yield the high-quality data needed in computer model calibration for simulating the water balance, some specialists recommend that more lysimeter-recharge studies be undertaken worldwide in a variety of climatic and soil conditions. However, the initial construction costs and the long-term monitoring requirements demand serious extended commitment.

Seepage meters were originally developed to measure canal seepage losses. They involve a seepage bell or cylinder that is pushed into the canal-bed sediment, and the infiltration rate is measured by constant or falling head techniques. Their advantages include being (1) lightweight and easily transportable, (2) relatively cheap, (3) simple to operate, (4) rapidly measurable, and (5) directly convertible into a seepage value. Difficulties are encountered in gravelly or stony sediments, or in sandy sediments, which may be washed from around the seepage cylinder by eddy currents, sediment disturbance, and ineffective seal of the inserted seepage cylinder. The number of measurements per unit of area needed to arrive at a reasonable average depends on the degree of heterogeneity in the seepage loss at the specific site. In conclusion, the seepage meter gives a rapid and direct measurement at low cost, but the figures obtained are only point measurements (Lerner et al., 1990).

APPENDIX B. Tracers for Recharge Estimation (Allison et al., 1994)

The natural tracers most commonly used in recharge studies are ^{3}H, ^{14}C, ^{36}Cl, ^{15}N, ^{18}O, ^{2}H, ^{13}C, and Cl. Of these, the first three are radioactive, with half-lives of 12.3, 5700, and 301,000 years, respectively. Their concentrations in the hydrologic cycle have been affected greatly by nuclear testing. Both tritium, ^{3}H and chlorine-36, ^{36}Cl from atmospheric testing have been used for soil-water tracing and recharge studies. Chlorine-36 has been used increasingly as more analytical facilities have become available. Input concentrations of the other isotopes mentioned above have also changed in time, but across a much longer time scale, due to changes in temperature and rainfall patterns. However, little is known of the temporal changes in the fallout of Cl.

Of the tracers mentioned above, tritium (^{3}H), deuterium (^{2}H), and oxygen-18 (^{18}O) most accurately simulate the movement of water because they form part of the water molecule. In most soils, chlorine-36 and nitrate (NO_3) move as the water does, but in some soils with heavier textures, anion exclusion may be a problem, and the tracer may move more rapidly than the water being traced.

Most of the recently developed isotope techniques are aimed at determining the age of water, which in turn permits calculation of groundwater travel time. The recharge rate, R, can then be calculated by $R = L\ \phi_e / t_a$, where ϕ_e is the effective porosity, L is the distance along the flow path, and t_a is the travel time or age of the groundwater at the distance L. There are three basic types of groundwater dating methods: (1) those methods which rely on input concentrations that have changed in time and are well known, such as the radioactive noble gas krypton-85, and the synthetic organic compounds chlorofluorocarbons (CFCs), used for dating young waters (less than ~40 years old); (2) tracers for which input concentrations have been constant, and decreases in concentration with time occur due to radioactive decay, such as ^{14}C, used for dating waters over the time scales of 200 to 20,000 years; or (3) methods where the input concentrations may have changed with time but can be determined because both parent and daughter isotopes are measured, such as $^{3}H/^{3}He$ (tritium/tritiogenic helium), which ratio is also used to date young waters (0 to ~50 years).

Mechanisms of tracer infiltration will affect the interpretation of results. Although piston (or plug) flow is often able to explain the behavior of tracers in the field, there is convincing evidence, particularly from humid regions, that water movement along preferred pathways is the rule rather than the exception. Thus, preferential or non-piston-type flow must be dealt with in any comprehensive analysis of recharge. For example, ^{3}H was found much deeper than the recharge rate would imply in native forest, suggesting preferred flow of water along root channels (Allison and Hughes, 1983).

Three techniques have been used for estimating recharge rates from tracer profiles in the unsaturated zone (Allison et al., 1994).

1. *From the position of the tracer peak*. In this method, the water in the profile above the peak in tracer concentration represents the recharge since the time that peak occurred. Any bypass (preferential) flow will result in recharge being underestimated.
2. *From the shape of the tracer profile in the soil*. This is generally more reliable than Method 1 above because information about flow mechanisms can be obtained. In

order to obtain estimates of mean annual recharge, $\overline{R}$, a weighting function that takes into account year-to-year variations of recharge is needed.

3. *From the total amount of tracer, T_t, stored in the profile*. This is given by

$$T_t = \int_0^z T(z)\theta(z)dz\,, \qquad \text{(B-1)}$$

where *T(z)* is the tracer concentration of water in the unsaturated zone at a distance *z* beneath the surface, and *θ(z)* is the volumetric water content. For evaporative tracers, such as ^3H, mean annual recharge can be estimated by

$$\overline{R} = T_t / \sum_{i=1}^{z} w_i T_{pi} \exp(-t\lambda), \qquad \text{(B-2)}$$

where T_{pi} is the tracer concentration of recharge water *i* years before the present; w_i is the annual recharge weighting factors, and λ is the tracer decay constant. In this analysis, non-piston flow can be handled because $\overline{R}$ is independent of the distribution of the tracer in the profile.

Tracer methods have a number of attractive attributes (Hendrickx and Walker, 1997). Their movement is governed mainly by the long-term mean soil water fluxes that lead to recharge. (Many water-balance or soil-water-pressure-based techniques measure fluxes on a much smaller time scale than is needed for recharge estimates.) The use of tracers does not necessitate frequent visits to the field. With tracers, it is possible to estimate smaller fluxes than with other methods. Finally, they are often the only alternative.

The choice of tracer depends on the situation. In most cases, the tracer is used to follow water movement and hence should move with the water. The tracer thus needs to be mobile and soluble; it should not be strongly retarded by the soil or aquifer matrix. Ideally, the tracer should be nonreactive and not transform during transport. Of course, the tracer needs to be easily measured and easily extracted from the soil. If artificial tracers are used, additional constraints need to be satisfied, such as low natural levels of the tracer in the environment, low toxicity, and low radioactivity. For environmental tracers, it is desirable to have large natural variations of tracer concentrations in the landscape. These constraints usually mean that only anions (Cl, Br, ^{36}Cl) or isotopically labeled water molecules (^2H, ^{18}O, and ^3H) can be used.

The choice of tracer is mainly determined by the time scale of the recharge process (Hendrickx and Walker, 1997). Use of artificial tracers requires that the bulk of the applied tracer has passed through the root zone. The time scale associated with leaching through the root zone is $Z_r\theta / R$, where Z_r is the root zone depth, θ the volumetric water content, and R the recharge rate. For example, in a humid climate (with θ = 0.1, Z_r = 3.3 ft, and R = 3.9 inches/yr), the time scale is one year. However, in an arid climate with a recharge rate of 0.4 inch/yr, the time scale is 10 years. While the former time scale is short enough for the tracer to be applied and the soil sampled in succeeding years, the latter is probably not. However, it would be suitable to use a *bomb tracer* (i.e. a tracer resulting from nuclear testing).

Bromide is the most widely used artificial tracer and ^3H and ^{36}Cl are the most suitable bomb tracers. However, ^3H and ^{36}Cl are too expensive for spatial and temporal variability studies that require many samples. The use of tritium as an artificial tracer is not generally recommended because of concerns about radioactivity and difficulty in application. For such investigations, the use of chloride as an environmental tracer is generally recommended. Employment of multiple tracers can often provide corroborative information needed for correct interpretation.

Because tracers do not measure water flow directly, a number of problems can arise, leading to over- or underestimation of recharge. These problems include secondary (unknown) tracer inputs, mixing and dual flow mechanisms; such problems only arise if the sources, sinks, and pathways of tracer are not fully understood. Part of the recharge going through preferred pathways (such as root channels or fissures) may invalidate the results of a tracer study.

APPENDIX C. Recharge-related glossary

Arid: Said of a climate characterized by dryness, variously defined as rainfall insufficient for plant life or less than 10 inches or 250 mm of annual rainfall.

Artificial recharge: Deliberate act of adding water to an aquifer by means of a recharge project, also the water so added. Artificial recharge can be accomplished via injection wells, spreading basins, or in-stream projects.

Available water capacity: The amount of water released from a wet soil between *field capacity* and the *permanent wilting percentage*.

Bank storage: Change in storage in an aquifer resulting from a change in stage of an adjacent surface-water body.

Baseflow: Stream flow derived mainly from groundwater seepage into the stream.

Bomb tracer: A tracer resulting from nuclear testing, such as tritium or chlorine-36.

Boundary condition: A mathematical expression of a state of the physical system that constrains the equations of the mathematical model.

By-pass flow: See *macropore flow*.

Calibration (model application**):** Process of refining the model representation of the hydrogeologic framework, hydraulic properties, and *boundary conditions* to achieve a desirable degree of correspondence between the model simulation and observations of the groundwater system.

Capillary flow: The flow that takes place in pores with a diameter less than approximately 0.1 inch or 3 mm in which capillary forces, together with gravity, determine the flow process.

Capillary fringe: Unsaturated zone immediately above the water table containing water in direct contact with the water table.

Conceptual model: An interpretation or working description of the characteristics and dynamics of a physical system.

Confined aquifer: Aquifer that is bounded above and below by formations of significantly lower *hydraulic conductivity*.

Confining bed: A geological unit of significantly lower hydraulic conductivity than an aquifer stratigraphically adjacent to one or more aquifers.

Consumptive use: Use that makes water unavailable for other uses, usually by permanently removing it from local surface or groundwater storage as a result of *evaporation* and/or *transpiration*. Does not include evaporation losses from bodies of water.

Crop coefficient: Empirically-determined coefficient relating *potential evapotranspiration* to crop evapotranspiration.

Darcy's equation or **law:** A formula stating that the flow rate of water through a porous medium is proportional to the hydraulic gradient. The factor of proportionality is the *hydraulic conductivity*.

Deep drainage: Drainage of water below the root zone.

Depression-focussed recharge: See *localized recharge*.

Diffuse recharge: Water added to the water table by vertical percolation of precipitation through the unsaturated zone. Also known as *direct recharge*.

Direct recharge: See *diffuse recharge*.

Discharge: The volume of water (and suspended sediment if surface water) that passes a given location within a given period of time.

Discharge area: An area in which water is lost naturally from the saturated zone.

Drainage basin: A hydrologic unit consisting of a part of the surface of the earth covered by a drainage system consisting of a surface stream or body of impounded surface water plus all tributaries. The *runoff* in a drainage basin is distinct from that of adjacent areas. A **river basin** is similarly defined.

Evaporation: The process of liquid water becoming water vapor, including vaporization from water surfaces, land surfaces, and snow fields, but not from leaf surfaces. Compare with *transpiration*.

Evapotranspiration: Sum of *evaporation* and *transpiration*.

Fallow: The period during which land is left to recover its productivity after cropping, mainly through accumulation of water and nutrients, attrition of pathogens, or a combination of these factors.

Field capacity: The quantity of water held back by soil or rock against the pull of gravity when excess water has drained out of a saturated or near-saturated soil. It is sometimes limited to a certain drainage period (2 or 3 days). Field capacity is thought to be the soil moisture condition that will promote maximum plant growth, with transpiration occurring at the potential rate (i.e., transpiration is not limited by moisture availability.)

Fingered flow: Unstable flow whereby the percolating water may concentrate at certain points to break into the sublayer in the form of finger-like or tongue-like protrusions.

Finite-difference method: Numerical technique for solving a system of equations using a rectangular mesh representing the aquifer and solving for the dependent variable in a piece-wise manner.

Finite-element method: Numerical technique for solving a system of equations using an irregular triangular or quadrilateral mesh representing the aquifer and solving for the dependent variable in a continuous manner.

Flow path: The route groundwater takes to a distant point.

Flow net: The set of intersecting lines of equal hydraulic head values and flow lines representing two-dimensional steady flow through a porous medium.

Focussed recharge: See *localized recharge*.

Gaining stream: Stream reach in which the water table adjacent to the stream is higher than the water surface in the stream, causing groundwater to seep into the stream, increasing its flow. Also known as *effluent stream*.

Geographic Information Systems (GIS): Computer-based systems for storing and manipulating geographic (spatial) information.

Groundwater basin: A geologically and hydrologically defined area which contains one or more aquifers which store and transmit water and will yield significant quantities of water to wells.

Groundwater flow system: A set of groundwater flow paths with common recharge and discharge areas. Flow systems are dependent on both the hydrogeologic characteristics of the soil/rock material and landscape position. Areas of steep or undulating relief tend to have dominant *local flow systems* (discharging in nearby topographic lows such as ponds or streams). Areas of gently sloping or nearly flat relief tend to have dominant *regional flow systems* (discharging at much greater distances than local systems in major basin topographic lows or oceans).

Hydrogeologic environment: The physical and chemical conditions resulting from the combination of topography, geology, and climate.

Hydrologic budget or **water balance:** An accounting of the inflow to, outflow from, and storage in a hydrologic unit such as a drainage basin, aquifer, soil zone, lake, or reservoir.

Indirect recharge: Recharge that results from percolation to the water table following runoff and localization in joints, as ponding in low-lying areas and lakes, or through the beds of surface water courses.

Induced recharge: Recharge to groundwater by infiltration, either natural or anthropogenic, from a body of surface water as a result of the lowering of the groundwater level below the surface-water level.

Infiltration: Movement of water from the ground surface into the soil.

Interflow: Subsurface lateral flow that can enter streams quickly enough to contribute to the rising streamflow hydrograph response to a storm.

Localized recharge: Recharge that results from horizontal surface concentration of water in the absence of well-defined channels, such as sloughs, potholes, and playas. Also called *focused* or *depression-focused* recharge.

Losing stream: Stream reach in which the water table adjacent to the stream is lower than the water surface in the stream, causing infiltration from the stream channel, recharging the aquifer and decreasing streamflow. Also known as *influent stream*.

Macropore flow: The flow that takes place in a wide range of large pores such as cracks in clay soils, rock fractures, fissures in sediments, worm holes, and old root channels. *Preferential* and *by-pass* flow are alternative names for macropore flow.

Matric suction: Soil-water potential (energy) resulting from the capillary and absorptive forces due to the soil matrix.

Mountain front recharge: Recharge that involves complex processes of unsaturated and saturated flow in fractured rocks, as well as infiltration along channels flowing across alluvial fans.

Natural recharge: Naturally occurring water added to an aquifer. Natural recharge generally results from snowmelt and precipitation or storm runoff.

Perched groundwater; perching: A superficial body of groundwater separated from an underlying main body of groundwater by an unsaturated zone due to a sufficiently low hydraulic conductivity layer that supports this body of perched groundwater; the act of causing a body of groundwater to form above a low- permeability layer in an unsaturated zone.

Perched water table: Water table of a relatively small groundwater body lying above the general groundwater body.

Percolation: laminar-gravity flow through unsaturated and saturated earth material.

Permanent wilting percentage: The water content of soil when indicator plants growing in that soil wilt and fail to recover when placed in a humid chamber.

Phreatophyte: A plant whose roots generally extend downwards to the water table which customarily feeds on the *capillary fringe*. Phreatophytes are common in *riparian habitats*. Term literary means water-loving plant.

Piston flow or **plug flow:** Purely advective flow without dispersion or diffusion of the dissolved components.

Playa: The flat-floored bottom of an undrained desert basin, becoming at times a shallow muddy lake after heavy rainfall; evaporation of the playa lake may leave a deposit of salt or gypsum.

Potential energy: The energy deriving from elevation and/or pressure.

Potential Evapotranspiration (PET): The maximum amount of soil *evaporation* and *transpiration* from a well-irrigated crop for a given set of environmental conditions.

Potential recharge: Soil-water that percolates below the root zone and has the potential of reaching the aquifer, whereas *actual recharge* is soil-water that actually reaches the aquifer.

Preferential flow: See *macropore* flow.

Preferential recharge: Recharge that takes place preferentially through macropores, as opposed to *diffuse recharge*, which takes place through the entire vadose porous medium.

Recharge area: The area that contributes water to an aquifer. Normally considered to be the natural area of recharge, as contrasted with a constructed recharge basin.

Rejected recharge: Potential recharge that exceeds the rate of flow through an aquifer that is already overfull, and as a result is rejected.

Residence time: The length of time between the input of water as infiltration or recharge and its output as runoff or discharge. Also known as *transit time* or *turnover time*.

Residual: In the case of recharge, the remainder of all other hydrologic components in the *water balance* equation.

Riparian: Of, or pertaining, to rivers and their banks.

Riparian habitat: Natural home of plants and animals occurring in a thin strip of land bordering a stream or river. Dominant vegetation often consists of *phreatophytes*.

Semiarid: Said of a type of climate in which there is slightly more precipitation (10 to 20 inches or 250 to 500 mm) than in an *arid* climate, and in which sparse grasses are the characteristic vegetation.

Sensitivity (analysis): In model application, the degree to which the model result is affected by changes in a selected model input representing hydrogeologic framework, hydraulic properties, and boundary conditions.

Soil moisture deficit: An estimate of the degree to which soil moisture content has dropped below *field capacity*.

Specific discharge: For groundwater, the rate of discharge of groundwater per unit area measured at right angles to the direction of flow.

Specific yield: The fraction of a saturated bulk volume consisting of water which will drain by gravity when the water table drops; specific yield is less than *porosity* because some water is too strongly absorbed to the earth material to drain. The ability of an *unconfined or water table aquifer* to store water is measured by its specific yield. Specific yield can be several orders of magnitude larger than the *storage coefficient*, thus producing more water when developed.

Steady-state flow: Characteristic of a flow system where the magnitude and direction of *specific discharge* are constant in time at any point.

Storativity or **storage coefficient:** The volume of water released per unit area of aquifer and per unit drop in head. Storage coefficient is a function of the compressive qualities of water and matrix structures of the porous material. A *confined aquifer's* ability to store water is measured by its storage coefficient. Storativity is a more general term encompassing both or either storage coefficient and/or *specific yield*.

Texture (soil): Relative proportions of sand, silt, and clay particles in a mass of soil.

Transmissivity: Flow capacity of an aquifer measured in volume per unit time per unit width. Equal to the product of hydraulic conductivity times the saturated thickness of the aquifer.

Transmission losses: Streamflow losses through seepage in ephemeral streams.

Transpiration: The vaporization of water given off by plants.

Unconfined (or **water table) aquifer:** An aquifer in which the *water table* is at the upper boundary of the groundwater flow system that is at atmospheric pressure.

Unsaturated or **vadose zone:** The unsaturated (i.e. not completely filled with water) zone lying between the Earth's surface and the top of the groundwater.

Water balance: See *hydrologic budget*.

Water table: The upper boundary of an *unconfined aquifer* at atmospheric pressure.

Watershed: That surface area which drains to a specified point on a watercourse, usually a confluence of streams or rivers.

Wetting front: The boundary between the wetted region and the dry region of soil during infiltration.

PART II. RECHARGE AND WATER BUDGETS OF THE KANSAS HIGH PLAINS AND ASSOCIATED AQUIFERS

Preface to Part II

This part attempts to summarize most major studies that have quantified groundwater recharge in the Kansas High Plains and associated aquifers. (A note on aquifer nomenclature is presented below.) Those studies are divided into 1) regional climatic soil-water balance studies; 2) regional groundwater modeling or analysis studies; 3) Kansas basin- to county-scale groundwater studies; and 4) field-based experimental studies. For each of those studies, the methodology employed was briefly outlined and the water budget of the study region was summarized in a uniform style of inches per year of the hydrologic quantity of interest over the study area. This compilation and synthesis includes some original research as well, in that information not explicitly stated by the authors was derived from their data, and additional information in studies involving this author is presented in this compilation as well. The assumptions and limitations of the model or analyses used are emphasized; whenever estimation errors or uncertainties were quantified, those are explicitly stated in this report. (A note on uncertainty measures is presented below.) Emphasis was also placed on environmental and land use factors affecting recharge estimates. Thus whenever possible, recharge estimates from predevelopment and development conditions were distinguished, and the recharge impacts of irrigation development are presented.

Note on aquifer nomenclature

The High Plains aquifer is a regional aquifer system underlying parts of eight states in the Great Plains from South Dakota to Texas. In Kansas, the High Plains aquifer lies beneath approximately 33,500 square miles of western and central Kansas, and is composed of several units that are geologically similar and hydraulically connected. The most extensive unit of the High Plains aquifer is the Tertiary-age (Miocene-Pliocene) Ogallala Formation, popularly known as the Ogallala aquifer. The eastern extension of the High Plains aquifer in Kansas is composed of younger, generally Pleistocene sediments, similar to the Ogallala, that include the Great Bend Prairie and Equus Beds aquifers. Also lying above the Ogallala Formation are other Pleistocene and Holocene deposits that also fill the valleys of modern streams. Where these fluvial and eolian deposits and stream valleys are hydraulically connected to the High Plains aquifer, these are considered part of the High Plains aquifer.

Note on measures of uncertainty

To measure how accurate the estimate of the mean value of a variable (such as recharge) is, we can compute its standard deviation from the mean, most often referred to as standard error (Glantz, 1981). The term standard error is a statistical term for the degree of uncertainty inherent in estimating a mean value. The standard error quantifies the reliability of the estimate of the population (true) mean from a sample drawn randomly from the population. Because the certainty with which the mean can be estimated increases as the sample size increases, the standard error of the mean decreases as the sample size increases. Conversely, the more variable the original population, the more variable the possible mean values of samples. Because the population of all sample means follows a normal distribution at least approximately, the true mean of the original population lies within two standard errors of the sample mean about 95% of the time. With this information we can construct an interval that represents the range of values over which the mean can be expected to vary.

A. Climatic soil-water balance studies on regional scales

Two major regional climatic soil-water balance studies have been recently conducted, one for the state of Kansas (Hansen, 1991) and the other for the entire High Plains aquifer (Dugan and Zelt, 2000). Because of their importance, and of the fact that the Hansen (1991) study was mostly adopted by the Division of Water Resources of the Kansas Department of Agriculture in the cases where more specific recharge data were not already available, those studies are analyzed in some detail in this report, especially in view of the fact that the Hansen (1991) study was short on details on methodology and data used.

A1. USGS study on natural recharge for principal aquifers in Kansas (Hansen, 1991)

Hansen (1991) estimated "potential natural recharge" for the entire state of Kansas by extending the results of the soil-water budget model and methodology employed by Dugan and Peckenpaugh (1985) for the Central Midwest Regional Aquifer Systems Analysis (CMRASA) in parts of Arkansas, Colorado, Kansas, Missouri, Nebraska, New Mexico, South Dakota, and Texas, to the High Plains aquifer of Kansas. Potential natural recharge refers to the deep percolation rate of soil water (made available from precipitation) below the root zone, where the water is presumed to be below the zone of influence of evapotranspiration processes, and thus potentially available to move downwards towards the water table and thereby eventually recharge the aquifer.

Because, as mentioned previously, the results of that study have been generally adopted by the Division of Water Resources of the Kansas Department of Agriculture, but the data, methodology, and limitations have not been reported in detail in the Hansen (1991) report, a brief summary of these aspects is presented below based on the reports of Dugan and Peckenpaugh (1985) and Dugan and Zelt (2000), and personal communication with Hansen (July 2002).

The soil-water balance simulation procedure employed requires four types of input: 1) monthly precipitation (P), 2) computed monthly potential evapotranspiration (PET) values, 3) hydrologic properties of soils, and 4) vegetation types. The PET calculation was based on the Jensen-Haise method (Jensen, 1974), which requires monthly temperature and solar radiation data. If direct solar radiation measurements are not available, they can be estimated from the following data (Dugan and Peckenpaugh, 1985): percent of possible sunshine or cloud-cover data, angle of sun's inclination at zenith (noon), hours of possible sunshine, and altitude above sea level.

The soil-water balance approach accounts for soil water entering, leaving, and remaining within the root zone, and can be summarized in the following simple equation:

$$R = ASW + P - SRO - AET - SC, \qquad \text{(II-1)}$$

where R = potential recharge (deep percolation), ASW = antecedent soil water within the root zone, P = precipitation, SRO = surface runoff, AET = actual evapotranspiration, and SC = total available soil water storage capacity of the root zone. All above components are expressed in inches per month.

The soil-water balance approach emphasizes the physical factors (climate, soils, and vegetation) that determine the availability of water for recharge and consumptive water use. Observed monthly climatic data from numerous weather stations in Kansas were compiled for the period of study (1951-1980). Land use data, by county, which provided vegetation patterns, were derived from 1978 statistics collected for the Census of Agriculture (U.S. Department of Commerce, 1980). Changes in vegetation over the period of study were considered small and thus neglected. The soils information was derived from a report by Dugan (1985) that consists of quantitative descriptions and areal distributions of the soils in Kansas and surrounding states based on their hydrologic characteristics, and from other generalized soils information in Kansas.

Six general vegetation types were considered in the state, each with distinctive seasonal consumptive water requirements, rooting depths, and infiltration-runoff relationships that create significant different demands on available moisture (Dugan and Peckenpaugh, 1985): 1) row crops, principally corn, soybeans, and grain sorghum; 2) tame hay, principally alfalfa; 3) small grain, principally winter wheat; 4) native grassland or pasture; 5) fallow or idle land; and 6) woodlands (urban area included). Each vegetation type is characterized by its consumptive water requirement (*CWR*) value, which is the quantity of water that vegetation type will consume if the availability of soil water is not a limiting factor. The *CWR* for each vegetation type was derived by multiplying the monthly value of potential evapotranspiration, *PET*, by a monthly crop coefficient (expressed as a simple ratio of *CWR* to *PET*; Dugan and Peckenpaugh, 1985). The difference between the amount of water required to meet the *CWR* and the water available within the plant root zone is the soil-water deficit (*SWD*).

The numerous soil groups within the state were reduced to ten groups for computational purposes within the soil-water balance program (Dugan and Peckenpaugh, 1985). The availability of water for consumptive water use is influenced by three physical characteristics of the soil: hydraulic conductivity, available water capacity, and slope, by regulating both infiltration and the ability of the soil profile to store water. Infiltration is largely a function of hydraulic conductivity and slope, while the water-storage capacity is determined by the product of the available water capacity (*AWC*) of the soil and the root-zone depth. The relationship between precipitation and infiltration was incorporated in four empirical infiltration curves for varying soils, topography (slope), and land use conditions (vegetation types) (Dugan and Peckenpaugh, 1985). Amounts of surface runoff and infiltration are computed from those infiltration curves. Runoff is that part of precipitation that does not enter the soil and is not accounted for in the soil-water balance. All infiltration is accounted for as either evapotranspiration or recharge. *AET* is a function of both the *CWR* of the vegetation type of interest and the soil water available within the root zone (Dugan and Peckenpaugh, 1985; Dugan and Zelt, 2000).

The soil-water balance program calculates, on a monthly interval, a variety of outputs such as infiltration, surface runoff, soil-water stored in the root zone, consumptive water use, soil water deficits, actual evapotranspiration, and deep percolation or potential recharge. That program computes the results for each climatic station for the various possible combinations of the soils and vegetation types in the study area. This output is then areally distributed through a "water-use program" (Dugan and Peckenpaugh, 1985) that weighs the outputs on the basis of percentage of occurrence of the various vegetation types and soils within the grid elements in the study area. The interpolation procedure is based on the distance of the two or three nearest climatic stations to weigh or adjust the soil-water balance program's output to the centerpoint of each grid element (Dugan and Peckenpaugh, 1985). A simple flow chart of the input and output of the various soil water simulation components is shown in Fig. II-1, and the resultant distribution of potential natural recharge is shown in Fig. II-2.

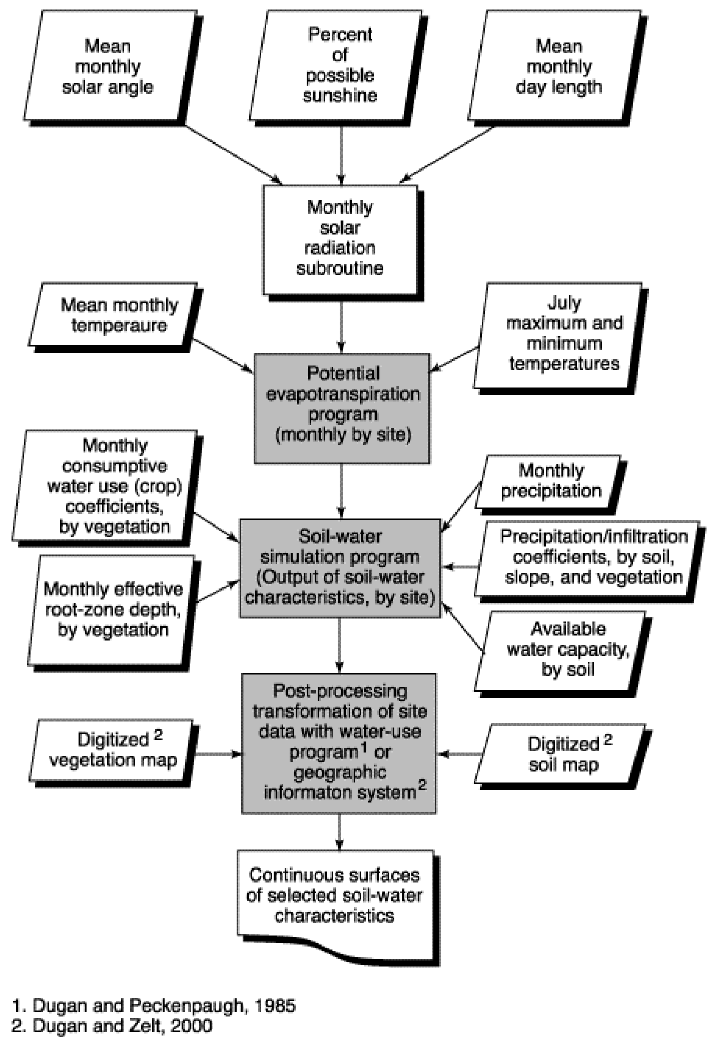

Figure II-1. Flowchart of soil-water simulation (adapted from Dugan and Zelt, 2000).

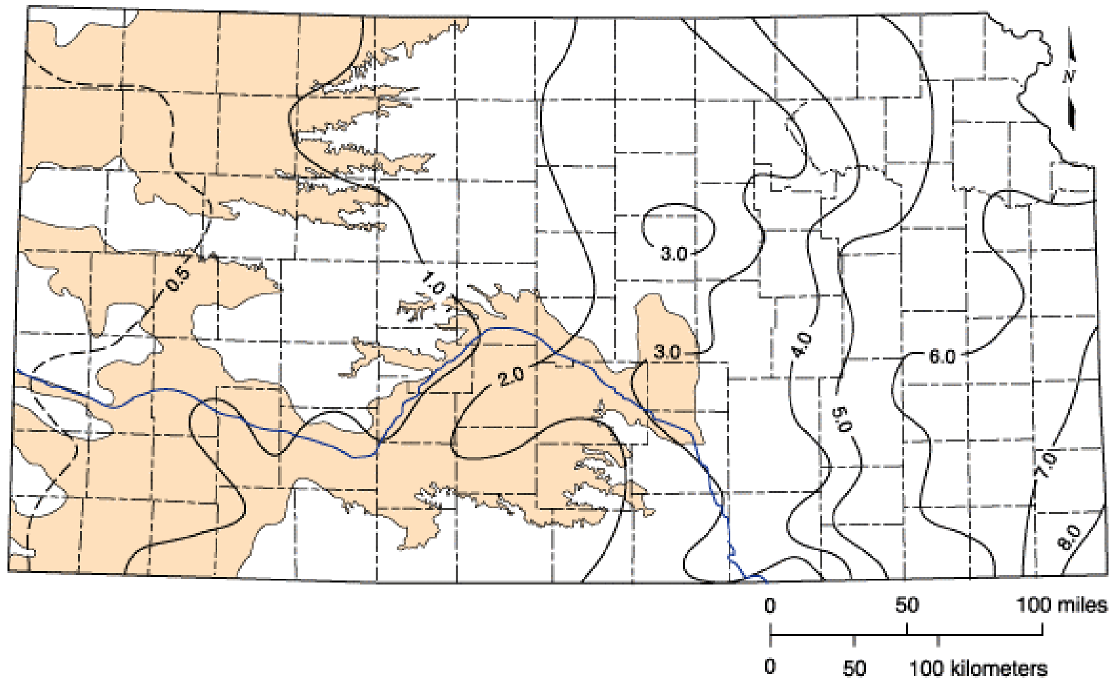

Figure II-2. Mean annual potential natural recharge (in inches per year) and extent of High Plains aquifer in Kansas (adapted from Hansen, 1991).

From the above summary of the methodology and data employed in estimating potential natural recharge, it is obvious that numerous simplifications and approximations were made in such estimation, resulting in relatively large uncertainties in the results. Equating deep percolation to aquifer recharge is premised on the assumptions that the immediate underlying aquifer is unconfined and true water table conditions exist, and that all water that passes through the root zone into the unsaturated zone below ultimately reaches the underlying aquifer. In addition, such potential recharge is determined from factors independent of the properties of the aquifer. It should also be noted that return flow from irrigation and underflow from areas outside Kansas were not considered in estimating potential recharge in the Hansen (1991) study. Also the results of that study have not been calibrated or compared to actual measurements or more detailed estimates. Therefore, caution should be exercised in directly applying the results of that study to specific areas because of the general nature (generalized *CWR/PET* relationships, land use, and soil characteristics) of that study.

However, such analysis provides valuable insights into the hydrologic system in an area. The spatial patterns of potential recharge and consumptive water use, systematically derived from measurable climatic, soil, and vegetation characteristics, are of great use to water resources managers and planners. Such overall patterns of resultant recharge (Fig. II-2) indicate that the controlling elements are the climatic factors themselves, particularly precipitation. The generalization can be made that as precipitation declines, both the magnitude and proportion of precipitation contributed to recharge declines (Dugan and Peckenpaugh, 1985). Also, areas of high cool-season precipitation tend to receive higher amounts of recharge (Dugan and Peckenpaugh, 1985). Smaller variations within local areas, however, are related to differences in soils and vegetation. The effect of soils, for example, is apparent in the westward extension of potential recharge contours in the Equus Beds and Great Bend Prairie aquifers as well as the Arkansas River sand dune areas in southwestern Kansas that coincide with sandy soils (Fig. II-2). The role of vegetation is less apparent because regional vegetation changes are usually gradual (Dugan and Zelt, 2000). However, under similar precipitation and soils, potential recharge tends to be larger for cropland than natural vegetation.

A2. USGS study of soil-water conditions in the Great Plains (Dugan and Zelt, 2000)

Dugan and Zelt (2000) expanded their Central Midwest Regional Aquifer Systems study (CMRASA; Dugan and Peckenpaugh, 1985) to the Great Plains and adjacent areas, including the entire Kansas High Plains aquifer. A major difference of this study to the previous regional climatic soil-water balance studies of Dugan and Peckenpaugh (1985) and Hansen (1991) was that the impacts of irrigation on recharge were considered. Thus, the soil-water balance, computed monthly over the period 1951-1980, is summarized by the following simplified equation (compare to eq. 1):

$$R = ASW + P + I - SRO - AET - SC \qquad \text{(II-2)}$$

where I = irrigation water required in inches per month, and all other terms are identical to the ones shown for eq. 1. The simple flow chart of the input and output of the various soil water simulation components, shown in Fig. II-1, explains the structure of the soil water simulation employed, taking advantage of GIS technology. With the exception of incorporating irrigation return flow to recharge estimation, this study is subject to the same assumptions and limitations as in the previously mentioned USGS soil-water balance studies (Dugan and Peckenpaugh, 1985; Hansen, 1991). In all these studies, the following group of parameters were employed to define the initial physical boundaries of the calculation of the soil-water balance (Zelt and Dugan, 1993): 1) initial soil-water content as a proportion of the available water capacity (*AWC*) of the soil (long-term simulations are usually insensitive to this parameter); 2) infiltration-curve coefficients that define the equations that determine the amount of monthly precipitation that infiltrates the soil and the amount that becomes overland runoff; 3) consumptive water requirement (*CWR*) coefficients that determine the rate at which each vegetation type could potentially consume water, as a proportion of monthly *PET*; 4) monthly effective root-zone depth; 5) *AWC* of the soil. In addition, the following parameters were also specified for that study: 1) the irrigation threshold as a proportion of the *AWC* of the soil below which irrigation is required, and 2) irrigation season, as the months to which irrigation-water application will be restricted. For that study, the irrigation threshold was set at 50 percent of *AWC*, and irrigation season was specified according to the principal irrigated crops in the area of each simulation study (Zelt and Dugan, 1993).

The Dugan and Zelt (2000) study is one of the very few that provides some insight into the impact of irrigation on deep percolation and recharge on a regional scale {see also the Sophocleous and McAllister (1987, 1990) study summarized in Part I, section 8.1}. Figure II-3 shows a comparison of potential recharge for Kansas under non-irrigated conditions (A), irrigated conditions [(B), weighted towards high-water demand row crops—irrigated wheat excluded], and combined nonirrigated and irrigated conditions [(C), a weighted combination of (A) and (B)] extracted from the results of that study. As can be seen in that figure, deep percolation under irrigated conditions is higher than under nonirrigated conditions except where deep percolation under dryland conditions is large as a result of extensive areas of fallow conditions (Dugan and Zelt, 2000). Figure II-3, part (C), probably represents deep percolation more realistically than part (A) because the effects of irrigation are included, although the deep percolation patterns are similar. However, the fact that the closed 2-inches/yr contour near Garden City in southwestern Kansas is missing from the weighted combination part (C) of Fig. II-3, probably represents a display error in the original reference.

The general increase in deep drainage under irrigated conditions does not result from excess irrigation but from an increased available water capacity in the root zone at the end of the irrigation season (Dugan and Zelt, 2000) that maintains the soil profile wetter than otherwise possible, thus making infiltrating precipitation more effective in creating soil-water surpluses available for deep percolation. A comparison of deep percolation under irrigated and nonirrigated conditions for similar soils and crop types at two selected sites is shown in Table II-1 (Dugan and Zelt, 2000). Although the absolute variations between deep percolation under nonirrigated and irrigated conditions are relatively small, the percentage differences can be considerable, particularly when deep percolation is small. The average percentage difference between deep percolation under irrigated and nonirrigated conditions for all soil- and crop-type combinations considered was 13% at Kearney, Nebraska and 24% at Holyoke, northeastern Colorado (Dugan and Zelt, 2000). The generalization can be made that areas where cultivated crops are prevalent have larger potential recharge than areas in grassland with similar climatic and soil conditions. It should be noted, however, that the potential recharge increase under irrigated conditions does not necessarily coincide with a net gain by the underlying aquifer under groundwater-irrigated conditions (Dugan and Zelt, 2000); this gain, derived from an increase of water in the root zone, is at the expense of groundwater in aquifer storage.

Table II-1[1]. Comparison of deep percolation for irrigated conditions (DPI) with deep percolation for nonirrigated conditions (DPD) for selected soils and crop types at Kearney, NE, and Holyoke, CO, 1951-1980 (values in inches/yr).

Crop Type	Clay-Silty Clay Loam soil; flat to undulating topography			Silty Loam to Loam soil; flat to undulating topography		
	DPI	% irr. return	DPD	DPI	% irr. return	DPD
Kearney (mean annual precipitation = 24.52 inches)						
Corn	2.97	7.4	2.75	4.36	4.4	4.17
Sorghum/Soybeans	4.70	6.4	4.40	6.31	4.4	6.03
Alfalfa	0.98	25.5	0.73	1.20	17.5	0.99
Holyoke (mean annual precipitation = 17.63 inches)						
Corn	0.81	12.3	0.71	1.41	8.5	1.29
Sorghum/Soybeans	1.78	9.0	1.62	2.77	5.8	2.61
Alfalfa	0.07	42.9	0.04	0.13	30.8	0.09

[1]Adapted from Dugan and Zelt (2000).

Table II-2 compares deep percolation for actual crops at four sites in North Dakota, Nebraska, Kansas, and Texas that are in areas predominantly in cultivation, with deep percolation for natural grassland only at those sites for Clay-Silty Clay Loam soil. All sites show that substantially greater potential recharge occurs under actual cultivated conditions than under grassland conditions alone (Table II-2). Fallow conditions tend to increase the difference between the two potential recharge conditions (Table II-2; Dugan and Zelt, 2000). The low potential recharge for both conditions at Sharon Springs, Kansas, and Muleshoe, Texas, as compared to those at Grand Forks, North Dakota, and Clay Center, Nebraska, is a result of the larger *PET* and, consequently, *CWR* at those sites (Dugan and Zelt, 2000). It is concluded that areas where cultivated crops are prevalent have greater potential recharge than areas in grassland with similar climatic and soil conditions.

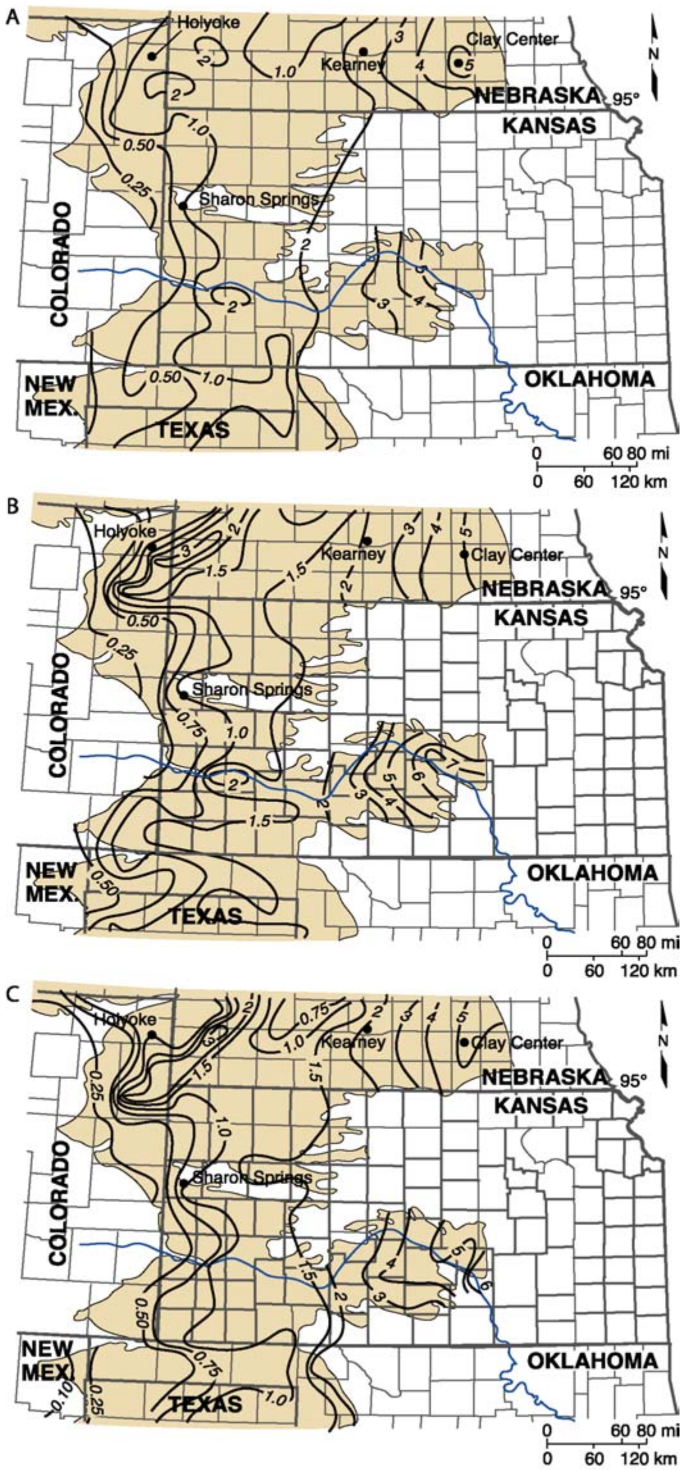

Figure II-3. Potential annual recharge for the High Plains aquifer in Kansas under nonirrigated conditions (A), irrigated conditions weighted towards high-water demand row crops (B), and combined nonirrigated and irrigated conditions (C)—a weighted combination of (A) and (B)— over the period 1951-1980. Contours in inches per year (adapted from Dugan and Zelt, 2000).

Table II-2[1]. Comparison of deep percolation for combined nonirrigated and irrigated conditions for actual vegetation with deep percolation for nonirrigated conditions for grassland for Clay-Silty Clay Loam soil at selected sites, 1951-1980.

[DPW, deep percolation for combined nonirrigated and irrigated conditions; DPD, deep percolation for nonirrigated conditions; PET, potential evapotranspiration; SG, small grain; HRC, high-water-demand row crops (corn, cotton)]

Site	Land use	Predominant vegetation types	Mean annual precipitation (inches)	Mean annual PET (inches)	DPW mean annual potential recharge (inches)	DPD mean annual potential recharge for grassland (inches)
Grand Forks, North Dakota	non-irrigated cropland	Fallow, SG	18.28	28	4.01	1.58
Clay Center, Nebraska	irrigated cropland	Grassland, HRC	26.68	45	4.30	3.61
Sharon Springs, Kansas	non-irrigated cropland	Fallow, SG	18.29	58	0.60	0.33
Muleshoe, Texas	irrigated cropland	Fallow, HRC	16.08	68	0.28	0

[1]Adapted from Dugan and Zelt (2000).

B. Large-area groundwater modeling or analysis

Several regional groundwater modeling and other studies have been conducted in Kansas or in portions of the High Plains aquifer that include parts of Kansas, and these studies are analyzed here.

B1. USGS RASA study of the High Plains aquifer (Luckey et al., 1986)

Luckey et al. (1986) divided the High Plains (HP) aquifer into three segments, the southern HP, the central HP, and the northern HP, and employed a two-dimensional groundwater flow model in each such segment. The HP aquifer of northwest Kansas is included in the northern HP simulation model, whereas the rest of the Kansas HP aquifer (encompassed within GMD1, GMD3, GMD5, and GMD2 districts) is included in the central HP simulation model. The numerical model employed was the USGS Trescott et al. (1976) finite difference model using uniform 10-mile × 10-mile (100-mi^2) grid cells. The model was run and calibrated in two dimensions under both predevelopment, steady state conditions as well as transient conditions up to 1980. The results of that study will be presented for the central HP (which had 513 active grid nodes covering an area of 51,300 square miles) and the northern HP (which had 943 active grid nodes covering an area of 94,300 square miles).

a) Central High Plains aquifer

The recharge distribution that resulted in the predevelopment (pre-1950) calibration is shown in Fig. II-4. The sand dune areas in southwestern Kansas have the maximum recharge rate with a long-term average of 0.84 inch/yr, whereas the rest of south-west and west-central Kansas, consisting of clay-, silt-, and sandy-loam soils, had the minimum recharge value of 0.056 inch/yr. The Great Bend Prairie and southwestern Equus Beds regions, characterized by mostly sandy loam soils, were assigned a recharge rate of 0.28 inch/yr. Overall, the mean, long-term predevelopment recharge rate for the central HP was estimated to be 0.14 inch/yr (Luckey et al., 1986). Another 0.0056 inch/yr flowed into the central HP from the southern and northern HP. Recharge from streamflow losses was difficult to detect because of the coarse grid used in the model; that recharge was included as part of the recharge from precipitation assumed in the sand dune areas (Luckey et al, 1986). The steady-state central HP model calibration resulted in a mean difference between observed and simulated water levels at the 513 active model nodes of – 0.28 ft, with a standard deviation of 38.5 ft. At 98% of the nodes, the simulated water level was within 100 ft of the observed water level. Most of the discrepancies were in areas of sparse water-level data (Luckey et al., 1986).

The long-term groundwater contribution to rivers is generally difficult to evaluate, but estimates are available for some rivers in Kansas. Fader and Stullken (1978) estimated base flow to the North and South Fork Ninnescah River as 38 cfs and 94 cfs, respectively; the model simulated 38 and 41 cfs, respectively. Gutentag et al. (1981) measured the base flow of the Cimarron River as about 60 cfs at the Kansas-Oklahoma state line. The flow computed by the model for the same place was 80 cfs. Winter flow records for 1896-1908 for the Arkansas River indicate a groundwater contribution to the river between Garden City and Hutchinson, KS, of about 80 cfs. The model simulated 71 cfs for the same reach. The simulated outflow to rivers and model boundaries totaled 0.146 inch/yr (see Table 3 of Luckey et al., 1986), which approximately balanced the equivalent total inflow into the model.

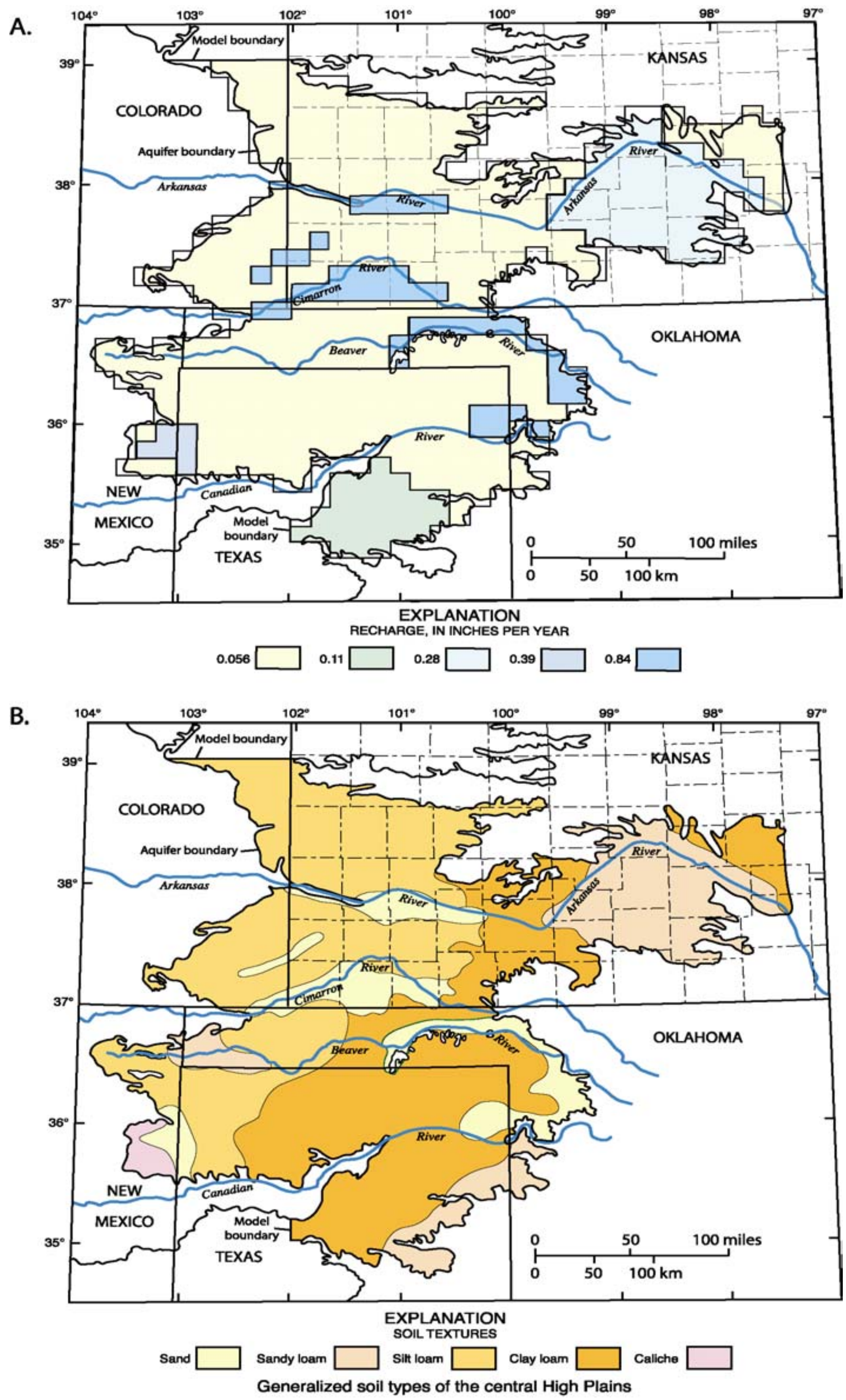

Figure II-4. Estimated predevelopment, long-term average recharge rates (A) and generalized soil types (B) for the central High Plains aquifer (adapted form Luckey et al., 1986).

The development period for the central HP region was considered to be the 1950-1980 period. Additional stresses on the aquifer during the development period consisted of pumpage, return flow to the aquifer from irrigation, and additional recharge caused by human activities.

Pumpage was calculated using the method outlined by Heimes and Luckey (1982) by multiplying the product of irrigated acreage (derived from county-level census data) and composite crop demand for cells of dimension 10 minutes of latitude by 10 minutes of longitude (10-minute cells) by an irrigation efficiency factor (ranging from 45 to 70 percent, with efficiency improving with time) at 5-year intervals from 1949 to 1978 (Luckey et al., 1986). Those pumpage data were redistributed to 100-mi^2 model nodes, thus spreading the pumpage throughout a somewhat broader area than actually occurred (Luckey et al., 1986). Because return flow was assumed to reach the aquifer within the 5-yr pumping period, changing return flow during model calibration was exactly equivalent to changing net withdrawal (total pumpage minus return flow).

For the 1950-1980 simulation, the return flow (to the aquifer from irrigation when net withdrawal was assumed equal to irrigation requirement) ranged from 55% of total pumpage early in the development period to 30% later, and averaged 43%. This return flow appears large, but only the difference between total pumpage and return flow was important and both may be considerably overestimated whereas the difference remains correct (Luckey et al., 1986). The total pumpage for irrigation from the central High Plains during the 30-yr period 1950-1980 was estimated at 18,354,000 ac-ft (refer to Table 4 of Luckey et al., 1986). Assuming that an average 43% of this total pumpage returned to the aquifer, the recharge from irrigation return flow would be 7,892,220 ac-ft over the 1950-1980 period or 263,074 ac-ft/yr or 1.0 inch/yr over the model area. This would be an additional recharge to the estimated predevelopment recharge.

The total volume of aquifer material dewatered was chosen as a calibration target for the development period simulation. Thus, the simulated water-level declines were 9% less than the observed ones on a volumetric basis. On an areal basis, the simulated water-level declines were 6% greater than the observed declines. The simulated change in groundwater storage was 54.9×10^6 ac-ft, whereas the observed change in storage was 50.3×10^6 ac-ft (Luckey et al., 1986). The differences between the observed and simulated water-level changes are believed to be primarily due to errors in the distribution of pumpage (Luckey et al., 1986).

b) Northern High Plains Aquifer

The Northern High Plains was the last area of the High Plains to be developed for irrigation, with development generally starting after 1960. The calibrated predevelopment (pre-1960), long-term average recharge rate for the northwest Kansas High Plains, characterized by silt loam and clay loam soils, was 0.076 inch/yr.

The mean difference between the observed and simulated predevelopment water level at the 943 active model nodes of the northern High Plains was +0.30 ft, with a standard deviation of 55.2 ft. At 92% of the nodes, the simulated water-level altitude was within 100 ft of the observed altitude (Luckey et al., 1986). A comparison of estimated and simulated baseflows show that the simulated baseflow was less than the estimated baseflow for all river systems except the Republican River. (The total simulated baseflow was slightly more than 60% of the estimated baseflow.) According to Luckey et al. (1986), the smaller simulated baseflow probably is because some of the baseflow is contributed from local and intermediate aquifer systems, which were excluded in large-scale regional models such as the northern High Plains model.

The development period for the northern High Plains was simulated to be the 1960-1980 period. In that simulation, the return flow from irrigation ranged from 46% of total pumpage early in the period to 30% later in the period, and averaged 36%. The recharge from precipitation determined during the predevelopment-period calibration was assumed to have continued during the development period. In the development period calibration, the difference between total pumpage and return flow for all areas was close to the estimated irrigation requirement. This was also similar to the central High Plains model simulation.

Recharge to the aquifer in the northern High Plains has been significantly increased by human activities such as leakage from canals and reservoirs, dryland farming, and other factors (Luckey et al., 1986). Cultivation practices associated with dryland farming can increase recharge from precipitation compared to the rate of recharge on rangeland (Ogilbee, 1962). For the northern High Plains, additional recharge due to cultivation was estimated to be 0.5 inch/yr throughout the dryland area.

The development-period composite recharge is shown in Fig. II-5. This recharge is assumed to be the sum of five separate components (Luckey et al., 1986): 1) predevelopment period calibration recharge; 2) canal and reservoir leakage; 3) return flow from surface-water irrigation; 4) increased recharge due to dryland cultivation; and 5) additional recharge east of 100° longitude.

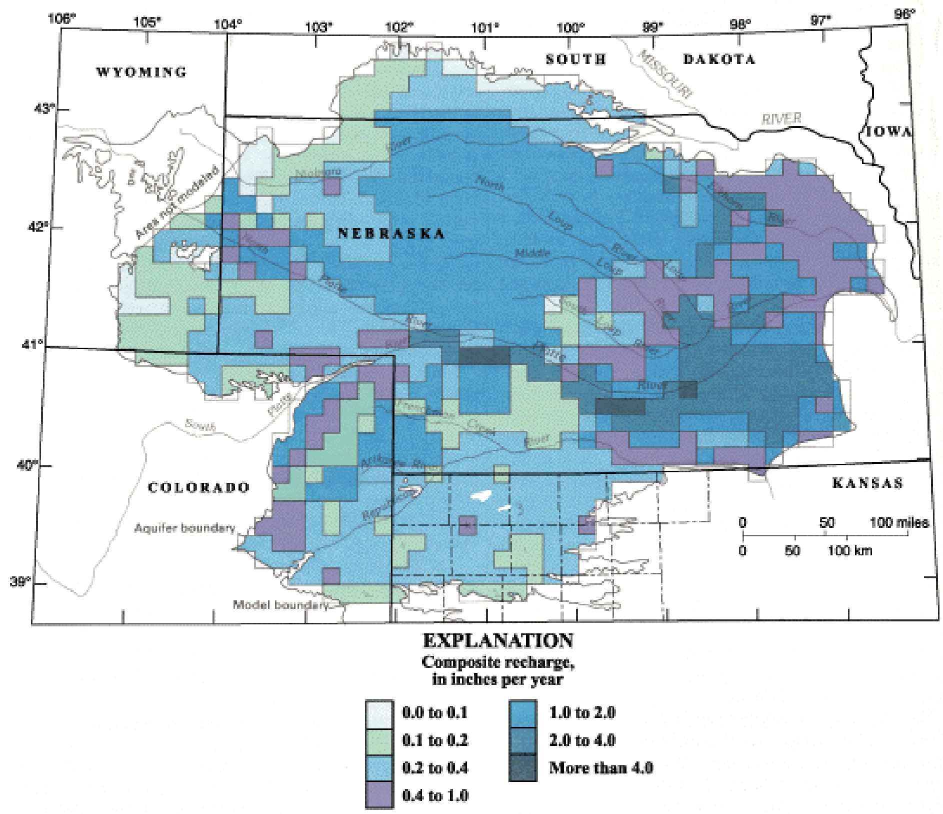

Figure II-5. Composite 1960-1980 recharge for the development-period model of the northern High Plains aquifer (adapted from Luckey et al., 1986).

Luckey et al. (1986) enumerated several additional factors to the above-listed ones that could account for the recharge increase simulated during the development period, such as the following: 1) decrease in runoff to streams due to cultivation; 2) decrease in baseflow to streams due to groundwater withdrawals; 3) change in downward leakage to underlying aquifers; 4) reduction of evapotranspiration due to lower water levels; 5) increase in downward leakage from a saturated zone above the water table; 6) greater specific yield of the aquifer than previously estimated; and 7) smaller total pumpage than estimated. None of these factors individually could account for the apparent increase in recharge, but collectively they might account for such increases.

The simulated net decrease in storage was 15×10^6 ac-ft, whereas the observed decrease in storage was 6×10^6 ac-ft. This simulation was accepted as a reasonable representation of the development-period operation of the aquifer in the northern HP (Luckey et al., 1986).

B2. USGS study of the High Plains aquifer in Oklahoma and adjacent areas, including the High Plains aquifer of southwestern Kansas (Luckey and Becker, 1999).

Luckey and Becker (1999) applied and calibrated a more spatially detailed finite difference numerical model for the central High Plains (exclusive of the High Plains area south of the Canadian River) than in the USGS RASA study (Luckey et al., 1986). They employed a single-layer, two-dimensional (2-D) MODFLOW model (McDonald and Harbaugh, 1988) using a uniform grid cell size of 6000-ft × 6000-ft (1.3-mi^2). The flow model had an active cell area of 27,212 square miles. They calibrated the model for both predevelopment (pre-1946) or steady state, and development (1946-1997) or transient conditions.

Predevelopment period simulation: To estimate recharge from precipitation, the model area was divided into zones of greater or lesser recharge (Fig. II-6). The zones of greater recharge represented either sand dunes or areas of extremely sandy soils, whereas the zones of lesser recharge represented the remainder of the area (Luckey and Becker, 1999). The calibrated mean recharge value for the zones of higher recharge was 0.69 inch/yr, which represents 4% of the mean 1961-1990 precipitation in the area of 16.5 inches/yr. The mean recharge for the lesser recharge areas was 0.068 inch/yr, which represents about 0.37% of mean precipitation in the area. The predevelopment, overall mean simulated recharge from precipitation was 0.155 inch/yr.

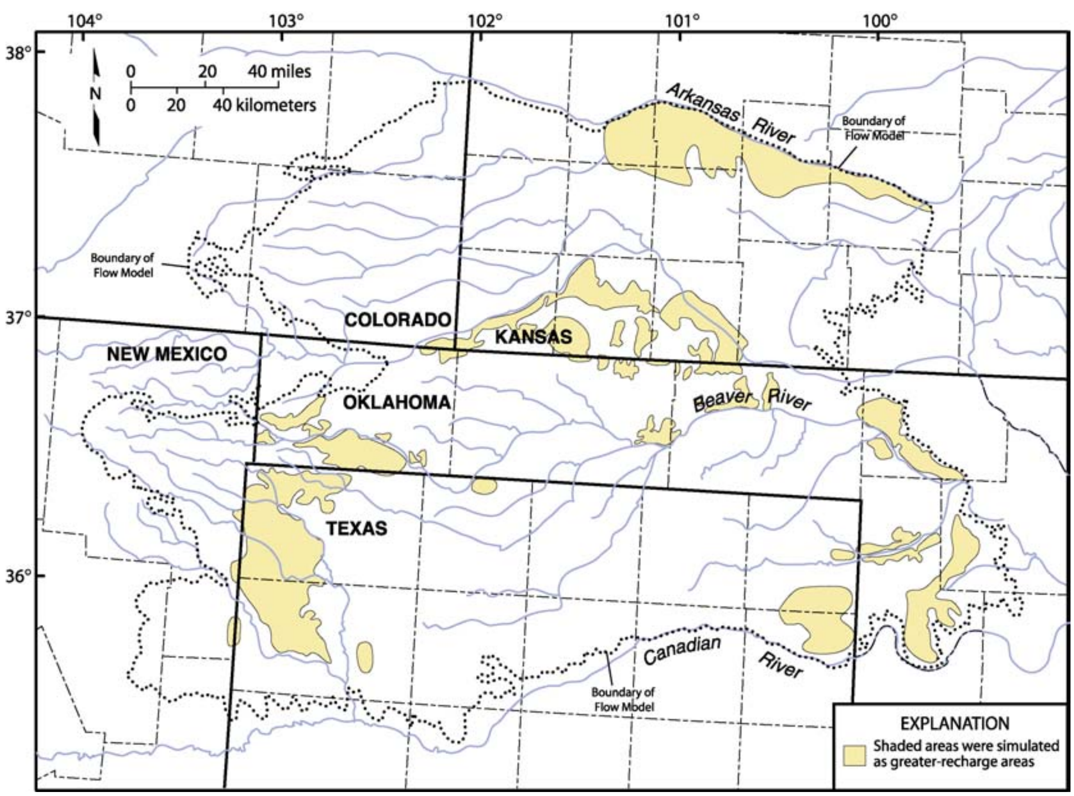

Figure II-6. Simulated greater-recharge areas for the central High Plains aquifer (adapted from Luckey and Becker, 1999).

Simulated predevelopment water levels were compared to observed water levels for the entire model area. The mean residual (i.e. differences between simulated and observed water levels at 21,073 active model nodes) is $-2.9\text{x}10^{-5}$ ft, with a standard deviation of 43.2 ft. At 97.7% of the nodes, the simulated water level was within 100 ft of the observed water level. Simulated predevelopment discharge to the Cimarron River near Liberal, KS, Forgan, OK, and Mocane, OK, was 5 to 10 cfs more than estimated discharge, whereas the estimated and simulated discharge to Crooked Creek were nearly the same (Luckey and Becker, 1999).

<u>Development-period (1946-1997) simulation</u>: The development period was simulated using five stress periods of 10 years each from 1946 to 1995, and one stress period of 2 years (1995-1997). Pumpage was assumed constant within each stress period and was the average of estimated pumpage for all years within the period. Water use for irrigation was estimated based on Census of Agriculture data (U.S. Department of Commerce, 1949-1992) for the period 1946-97 using the method outlined by Heimes and Luckey (1982), that consists of estimating irrigation use as the product of calculated irrigation demand (based on a modified Blaney-Criddle method), reported irrigation acreage, and assumed irrigation efficiency. The estimated pumpage was based on mean 1961-1990 monthly precipitation and temperature. According to Luckey and Becker (1999), the use of mean pumpage, temperature, and precipitation was assumed to have introduced "negligible error" by the end of the 52-year simulation. Time steps used in the simulation were 36.5 days long. Pumpage was also assumed to occur throughout the year, although most pumpage actually occurred during the summer. However, according to Luckey and Becker (1999), that assumption also should have introduced negligible error by the end of the 52-yr simulation.

Simulated recharge due to irrigation averaged 24% of pumpage for the 1940s and 1950s; averaged 14% for the 1960s; averaged 7% for the 1970s; averaged 4% for the 1980s; and averaged 2% for the 1990s (Luckey and Becker, 1999). Recharge due to irrigation was subtracted from total pumpage before the simulation was made, so only net pumpage was input into the model. This operation means that recharge due to irrigation was assumed to occur within the same stress period as the pumpage. According to Luckey and Becker (1999), given the long period simulated and the low recharge near the end of the simulation, that assumption probably did not cause substantial error in the model.

Recharge due to dryland cultivation was estimated to be 3.9% of mean 1961-1990 precipitation (which is 16.5 inches/yr) or 0.64 inch/yr over the area in dryland cultivation (which also included irrigated wheat). Over the entire study area, dryland cultivation recharge amounted to 0.24 inch/yr.

Thus total simulated recharge in the development-period model consists of 0.15 inch/yr background recharge from precipitation, 0.24 inch/yr due to dryland cultivation, and 0.022 inch/yr due to irrigation, totaling 0.41 inch/yr recharge from all sources.

Comparisons of simulated and observed predevelopment to 1998 water-level changes were done only for 162 observation wells in Oklahoma, resulting in a mean residual of -2.9×10^{-3} ft, with root mean square difference of 17.9 ft. 98.1% of simulated values were within 50 ft of observed values.

The simulated discharge to the Cimarron River at the end of 1997 was 51.2 cfs, a decrease of 10.9 cfs from simulated predevelopment discharge. According to Luckey and Becker (1999), both the discharge and simulated change in discharge appear reasonable.

Simulated water-level changes for 1998-2020 using mean 1996-1997 pumpage indicate more than 100-ft declines in Finney and Haskell counties (fig. II-7); an area of simulated decline of 50-100 ft covers most of Grant and Haskell counties, and substantial parts of Stanton, Stevens, Seward, Meade, Gray, Finney, and Kearny counties (Luckey and Becker, 1999). Very little additional decline is simulated to occur in southeastern Gray County, but this is an area of little saturated thickness. A summary of simulated predevelopment, 1998, and 2020 water budgets for the entire model area is shown in Fig. II-8. Groundwater storage over the Central High Plains study area is simulated to decrease by 49×10^{6} ac-ft (or 33.76 inches) from the end of 1997 to the beginning of 2020 (Luckey and Becker, 1999). The increase in stream discharge at the beginning of 2020 compared to the one at the end of 1997, evident in Fig. II-8, occurs mostly in the eastern portion of the simulated area, mainly along Beaver River below Optima Lake and along Canadian River (Fig. II-6). The cause seems to be the extra recharge that was simulated on dryland wheat (R.R. Luckey, written communication, March 2003).

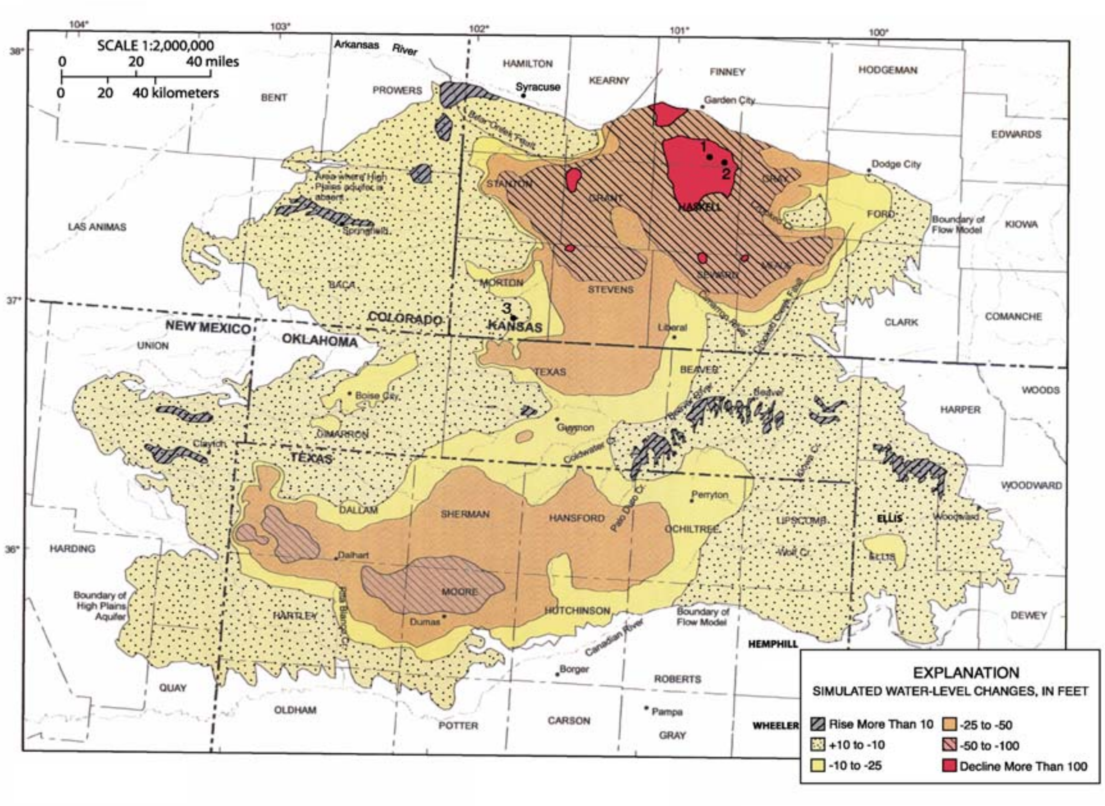

Figure II-7. Simulated water-level changes in the central High Plains aquifer for 1998–2020 using mean 1996–1997 pumpage. Bold dots with numbers 1, 2, and 3 represent experimental recharge sites described in text subsection 4: Field-based experimental recharge studies at the local level (adapted from Luckey and Becker, 1999).

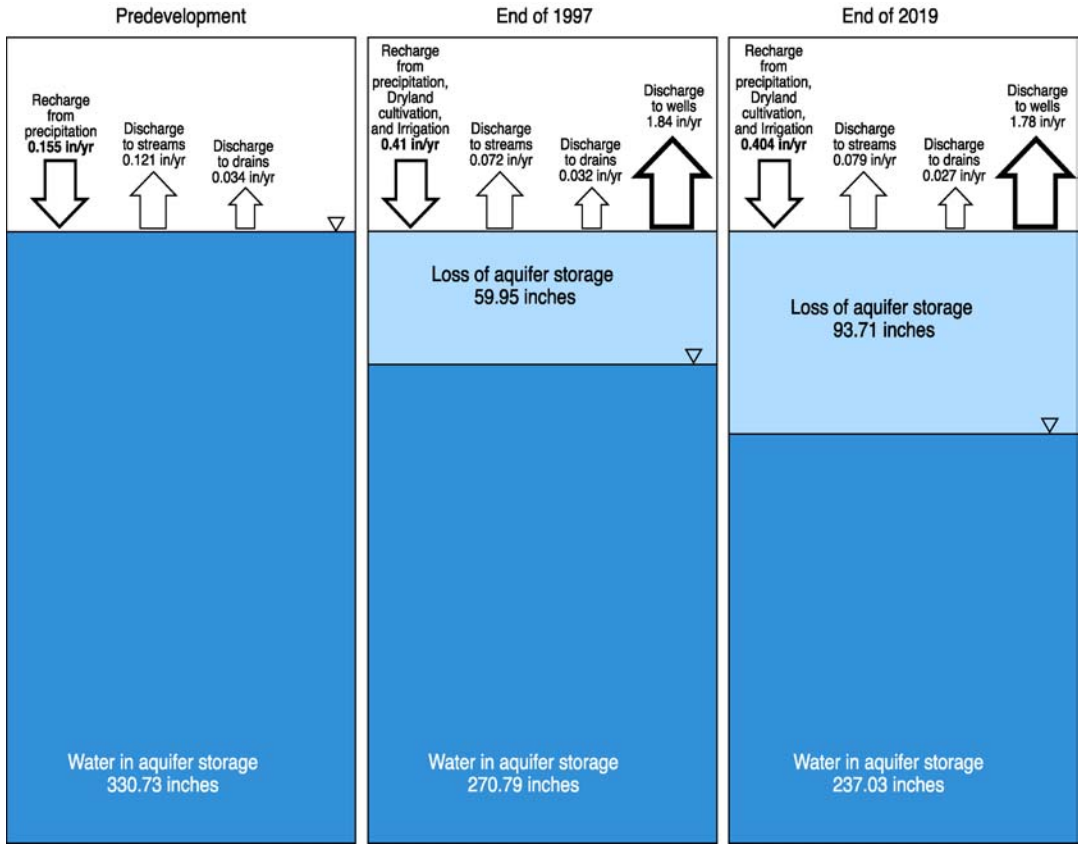

Figure II-8. Simulated predevelopment, end of 1997, and end of 2019 water budgets for the central High Plains studied by Luckey and Becker (adapted from Luckey and Becker, 1999).

B3. USGS study of the High Plains aquifer in western Kansas (Stullken et al., 1985)

Stullken et al. (1985) employed the 2-D finite difference USGS model developed by Trescott et al. (1976) to simulate the steady-state, predevelopment groundwater budgets of northwestern Kansas and southwestern Kansas using uniform 2.84–mi × 2.84–mi (8-mi^2) grid cells.

Steady-state (pre-1950) simulation of the High Plains aquifer in northwestern Kansas:

Following model calibration, the predevelopment recharge varied from 0 to 0.79 inch/yr and averaged 0.20 inch/yr, which is consistent with Jenkins and Pabst's (1975) recharge estimate (see section B10). The mean difference between observed and simulated hydraulic head values (mean residual) was –2.0 ft (buildup), with a standard deviation of 8.12 ft. Simulated hydraulic-head difference in 88% of the nodes was within 10 ft of observed values. According to Stullken et al. (1985), estimated streamflow gains also were simulated "closely" by the model. The predevelopment water budget (Table II-3) indicates that approximately 80% of total recharge came from precipitation, and that approximately 75% of discharge was by leakage to streams.

Table II-3[1]. Water budget from steady-state simulation of High Plains aquifer, northwest Kansas[2] (values in inches/year).

Budget item	Inflow	Outflow
Boundary flow	0.05	0.05
Recharge	0.20	---
Stream leakage	---	0.19
Pumpage	---	0.009
Totals	0.25	0.25

[1.] Adapted from Stullken et al. (1985)
[2.] Active model area approximately 5,220,000 acres

Steady-state (pre-1950) simulation of the High Plains aquifer in southwestern Kansas:

Recharge from precipitation indicated by the calibrated model ranged from 0 to 2.0 inches/yr and averaged 0.24 inch/yr. The greatest recharge occurred in the area south of the Cimarron River and in the area between the Cimarron River and Crooked Creek where a large percentage of the surface is dune sand. The calibrated recharge from precipitation for the Arkansas River valley and dune-sand areas to the south was 0.25 inch/yr (Stullken at al., 1985). The simulated recharge to the aquifer owing to leakage from streams and creeks in the western part of southwestern Kansas (Big Bear, Little Bear, and Sand Arroyo creeks, the North Fork of Cimarron River, and the western reach of the Cimarron River) totaled 0.036 inch/yr, whereas recharge due to stream leakage from the Arkansas River, Crooked Creek, and the eastern reach of the Cimarron River was 0.045 inch/yr, although overall Arkansas River, Cimarron River, and Crooked Creek were gaining streams in the area (Stullken et al., 1985). Estimates of recharge from streams in southwest Kansas using channel geometry methods (Hedman and Osterkamp, 1982) ranged from 0.08 to 0.14 inch/yr in the reaches of the losing streams.

Movement of water from the High Plains aquifer to the underlying Lower Cretaceous (Dakota) sandstone aquifer was simulated in parts of Grant, Haskell, Stevens, Seward, and Meade counties, and totaled 0.032 inch/yr (Table II-4).

The mean residual for all 1,028 active nodes was –1.08 ft (a buildup), with a standard deviation of 10.5 ft.

The simulated steady-state water budget is given in Table II-4 and shows that about 60% of the total recharge came from precipitation, approximately 19% came from boundary inflow, and 21% came from leakage of streams. Also, 38% of the total discharge from the aquifer went to boundary outflow, 8% went to the underlying Lower Cretaceous sandstone aquifer, and 54% went to streams.

Table II-4[1]. Water budget from steady-state simulation of High Plains aquifer, southwest Kansas[2] (values in inches/yr).

Budget item	Inflow	Outflow
Boundary flow	0.07	0.15
Recharge	0.24	---
Stream leakage	0.08	0.21
Loss to Lower Cretaceous sandstone aquifer	---	0.03
Totals	0.39	0.39

1. Adapted from Stullken et al. (1985)
2. Active model area approximately 5,309,917 acres

B4. USGS study of the Dakota and High Plains aquifers of southwestern Kansas (Watts, 1989)

Watts (1989) developed a layered model of the High Plains, Dakota, and Cheyenne Sandstone aquifers separated by the Niobrara-Graneros and Kiowa confining units for southwestern Kansas centered around a study area consisting of Kearny, Finney, Gray, Hodgeman, and Ford counties. The 3-D MODFLOW model was employed for this purpose using a 6-mi × 6-mi square grid for the above-mentioned study area. The model was calibrated to simulate the measured water-level changes in the High Plains and the Dakota aquifers for the period 1975-1982. Each year of the calibration period was divided into three stress periods representing 1) the relatively static water-level period of January to April, 2) the high groundwater-withdrawal period of May through September, and 3) the recovery period of October through December, respectively. Annual irrigation pumpage from the High Plains aquifer was estimated for 1974 and 1978 irrigated acreages multiplied by crop-water demand (Heimes and Luckey, 1982) and divided by an irrigation efficiency of 0.8. Straight-line projection and extrapolation were used to estimate annual pumpage between 1974 and 1978 and 1979-1982, respectively. Estimates of the withdrawal from the Dakota aquifer were based on water-use reports from the Division of Water Resources, Kansas Department of Agriculture.

The model calibration resulted in an average difference between January 1982 measured and simulated water levels in the High Plains aquifer of 1.04 ft, and a standard deviation of about 17 ft. The differences between measured and simulated potentiometric surfaces for the same time period for the Dakota aquifer resulted in an average difference of –0.09 ft and a standard deviation of 55 ft. The simulated 1982 water budget for the High Plains and Dakota aquifers is shown in Table II-5.

Table II-5[1]. Simulated 1982 water budgets for the High Plains and Dakota aquifers[2] (values in inches/yr).

Budget item	High Plains aquifer		Dakota aquifer	
	Inflow	Outflow	Inflow	Outflow
Boundary inflow	0.04	---	0.02	---
Boundary outflow and flow to streams (baseflow) or springs and seeps	---	0.30	---	0.04
Recharge	0.58	---	0.01	---
Streamflow loss	0.05	---	---	---
Well pumpage	---	6.32	---	0.10
Leakage from adjacent aquifers	0.01	---	0.10	---
Leakage to adjacent aquifer	---	0.09	---	0.01
Loss of groundwater storage	6.10		0.01	
Totals	6.78	6.71	0.14	0.15

[1.] Adapted from Watts (1989)

[2.] Active-node model area for the High Plains aquifer: 4,212 square miles (2,695,680 acres); and for the Dakota aquifer: 4,896 square miles (3,133,440 acres)

As can be seen from Table II-5, outflow from and decrease in storage in the High Plains aquifer dominate the water budget of the study area. Average recharge to the High Plains aquifer during 1982 was estimated at 0.6 inch/yr. Although downward leakage was only a minor component of the High Plains aquifer water budget (0.09 inch/yr), it was a major source of inflow to the Dakota aquifer. Leakage across the base of the High Plains aquifer in the model area is simulated as predominantly downward in areas where the Niobrara-Graneros confining unit is present and upward where the Dakota aquifer subcrops below the High Plains aquifer, such as occurs in parts of southern Finney and Kearny counties in the model study area as well as in Stanton, Grant, Haskell, and other counties in the general model area, as shown in Fig. II-9.

B5. USGS study of the High Plains aquifer in Oklahoma, including some Kansas counties (Havens and Christenson, 1984)

Havens and Christenson (1984) simulated the High Plains aquifer in Oklahoma, including the southern tier of Kansas counties (Morton, Stevens, Seward, and Meade) and a portion of Baca County, Colorado, as the northern boundary of the model using the USGS 2-dimensional finite difference model (Trescott et al., 1976). The Canadian River was used as the approximate southern boundary of the study area. The model area was discretized using a regular 5-mi × 5-mi grid network that included 888 active nodes covering an area of 22,200 square miles. The model was calibrated for both predevelopment (1941), steady-state, and transient conditions from 1941 to 1980. Eight five-year pumping periods with one-year time steps were employed in the transient model. Pumpage used in that study was calculated as a percentage of the total crop demand as determined by Heimes and Luckey (1982) and crop distribution data published by the U.S. Department of Commerce as a Census of Agriculture. The predevelopment (1941) and the transient, 1980 simulation water budgets are presented in Table II-6.

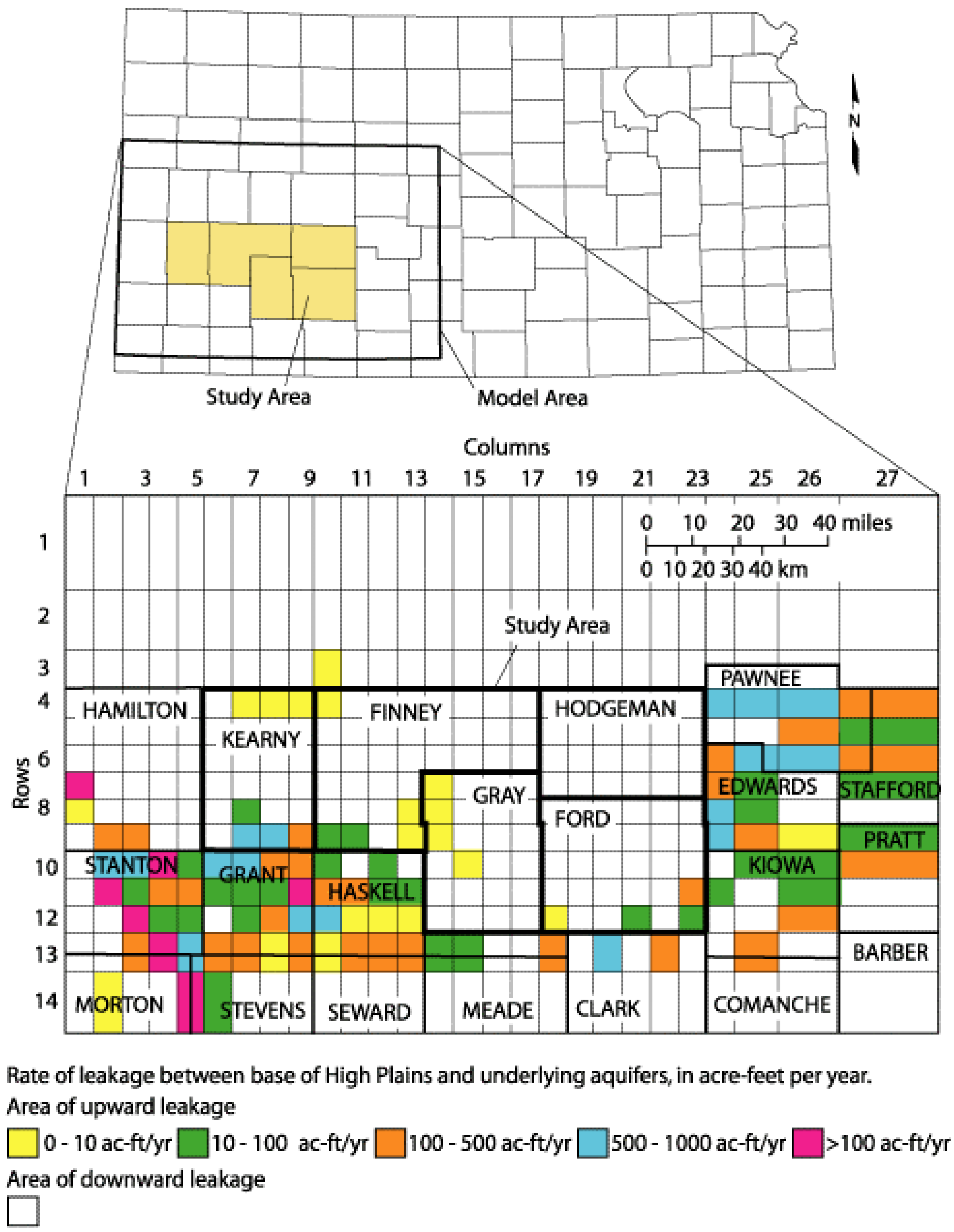

Figure II-9. Finite-difference grid of model and study areas in southwestern Kansas with simulated leakage between High Plains and underlying aquifers, 1982 (adapted from Watts, 1989).

Table II-6[1]. Steady-state (1941) and end of simulation (1980) water budgets for the High Plains regional aquifer in northwestern Oklahoma[2] (values in inches/year).

Budget item	1941 steady-state simulation Input	1941 steady-state simulation Output	1980 transient simulation Input	1980 transient simulation Output
Recharge from precipitation	0.34	---	0.34	---
Discharge to streams	---	0.34	---	0.07
Well pumpage	---	0	---	2.93
Loss of groundwater storage	0	0	2.66	0
Totals	0.34	0.34	3.00	3.00

1. Data from Havens and Christenson (1984)
2. Active-node model area = 22,200 mi^2 = 14,208,000 acres

Recharge from precipitation was estimated as 0.34 inch/yr. However, the eastern half of the model area (that includes Meade, Seward, and the eastern portion of Stevens counties in Kansas), that had also higher precipitation, had a higher value of recharge (0.45 inch/yr), whereas the western half of the model area (including Morton and most of Stevens counties in Kansas) had half of that recharge value (0.23 inch/yr). The steady-state simulation resulted in a mean difference between computed and measured heads of −0.044 ft in the 356 nodes that were in the Oklahoma portion of the model, and a mean absolute value of the differences of 50.1 ft. For the transient simulation, the mean difference between computed and measured heads was −0.011 ft, and the mean of the absolute values of the differences was 48.0 ft. Although no streamflow data were available for predevelopment-conditions calibration, the end-of-simulation (1980) leakage to streams (118.2 cfs) was very close to the total estimate of 117.8 cfs of low-flow discharge of streams draining the High Plains aquifer of northwestern Oklahoma for the period 1970-79.

B6. KGS study of geohydrology of southwestern Kansas (Gutentag et al., 1981)

Gutentag et al. (1981) studied the groundwater resources of southwestern Kansas (11 counties) and compiled a (partial) water budget for the area as of 1975 (Table II-7).

Table II-7[1]. 1975 water budget (partial) for southwestern Kansas[2] (values in inches/yr).

Budget item	Inflow	Outflow
Boundary flow	0.02	0.04
Recharge (from precipitation and streamflow & pond losses)	0.60	---
Streamflow gains	---	0.15
Net well pumpage (total pumpage minus 20% irrigation return flow to aquifer)	---	4.83–6.25

[1.] Adapted from Gutentag et al. (1981)
[2.] Area considered = 6,600 mi^2 = 4,224,000 acres (area that is underlain by sufficient saturated thickness to support irrigation).

Recharge was considered to be 10% of precipitation during the growing season (April-October) on irrigated land, and 1% of precipitation on nonirrigated land. The amount of annual recharge from runoff into surface depressions and streams is assumed to be included with the total estimate of recharge from precipitation. Records for 1975 showed that about 1,400,000 acres were irrigated; the remaining 2,824,000 acres were not irrigated. Annual recharge to the aquifer, based on about 15 inches of precipitation during the growing season, is computed to be 0.6 in over the total area (1.50 inches over the irrigated area and 0.15 inch over the nonirrigated area). Streamflow gains (groundwater contribution to streamflow or baseflow) represent the total average baseflows of Cimarron River at Mocane, Oklahoma, Crooked Creek near Nye, Kansas, and Arkansas River leaving the study area. It was estimated that pumpage for irrigation (of 1,400,000 acres) during 1975 was between 2.1 and 2.8 million acre-feet (6.0–8.0 inches over the total area). Figures derived from Meyer et al. (1970) for irrigated land in Finney County (see section C5) showed about 20% of the water applied to irrigated land was returned to the aquifer by percolation below the root zone. Thus, calculations suggest that about 420,000 to 560,000 ac-ft/yr (1.2 to 1.6 inches/yr) returns to the groundwater reservoir as recharge from irrigation return flow in southwestern Kansas (Gutentag et al., 1981). The number of irrigation wells in January 1975 was estimated at about 7,000 in southwestern Kansas. Groundwater evapotranspiration was considered negligible in southwestern Kansas because the water table in most of the area is too far below the land surface to be affected by evaporation and transpiration (Gutentag et al., 1981). In order to balance this total outflow (well pumpage, baseflow to streams, and boundary outflows–Table II-7), groundwater storage was being depleted by 4.4 to 5.8 inches/yr (as of 1975), and this depletion is reflected in the observed long-term declines in groundwater levels and saturated thickness.

B7. Kansas Governor's Task Force on Water Resources Interim Report (1977)

The Governor's Task Force on Water Resources in its December 1977 Interim Report included a section, entitled "A Case Study of the Ogallala" to provide some basic background information for understanding the physical, economic, legal, and management problems involved on groundwater depletion in western Kansas. The Task Force presented the following information on recharge and withdrawals from the Ogallala aquifer in western Kansas (Table II-8), and summed up the situation in the following way:

"The estimated amount of groundwater withdrawn in Region No. 1 [GMD1] of western Kansas in 1975…was almost *15* times greater than the estimated recharge; in Region No. 3 [GMD3], withdrawals approximated *18* times the recharge; and in Region No. 4 [GMD4] withdrawals were about *seven* times the recharge. For western Kansas as a whole, withdrawals are estimated to average *14* times the recharge rate. With this great a disparity between withdrawals and recharge and with current indications pointing to even greater withdrawals for irrigation in the future, it seems apparent that the water table has nowhere to go but down if current conditions and projections prevail."

Theoretical solutions to increasing recharge were considered by the Task Force, which concluded that "there are no simple physical solutions and we must look beyond the physical to economic, legal and management factors for enlightenment on how to resolve groundwater depletion problems."

B8. KGS study of the Ogallala aquifer in Kansas (O'Connor and McClain, 1982)

O'Connor and McClain (1982) studied the Ogallala aquifer in Kansas (approximately 28,560 square miles in a 32-county area of western Kansas divided into northwest, west-central, and southwest subregions). Based on previous recharge estimates, they approximated the recharge from irrigation return flow and from precipitation on the area overlying the Ogallala and peripheral aquifers as shown in Table II-9. Overall recharge for the Ogallala aquifer in western Kansas was estimated as 0.57 inch/yr. Nonirrigated or dryland average recharge overlying the Ogallala aquifer was estimated as 0.3 inch/yr. This includes recharge from ephemeral streams, depressions, and sand dune tracts that have higher than average recharge. Recharge on irrigated land above the Ogallala (including that from precipitation and irrigation return flow) was estimated to be 10% of an average irrigation application of 18 inches annually, or 1.8 inches/yr. The recharge figures in Table II-9 do not include subsurface inflow or recharge from the Arkansas or Cimarron rivers. Because of declining water levels along those stream valleys, much of the streamflow was lost by influent seepage to the groundwater reservoir and not by flow out the east and south sides of the study area.

Table II-8[1]. Hydrologic budget component estimates for the Ogallala aquifer in western Kansas as of 1975.

Region[2]	1975 Acres		1975 Estimated Groundwater Supply	Avg. Annual Precipitation	Evapotranspiration and Runoff	Recharge		1975 Estimated Groundwater Withdrawn	
	Irrigated	Total	acre-feet	inches	inches	ac-ft	inches	ac-ft	inches
1	411,000	2,400,000	10,000,000	18.5	18.3	40,000	0.20	589,000	2.95
3	1,600,000	6,900,000	223,000,000	18.6	18.4	165,000	0.29	2,900,000	5.04
4	434,000	5,800,000	63,000,000	19.3	19.1	111,000	0.23	763,000	1.58
Total	2,445,000	15,100,000	296,000,000			316,000	0.25	4,252,000	3.38

1. Adapted from Governor's Task Force on Water Resources (1977)
2. These regions include the counties wholly or partially in the corresponding Groundwater Management District

Table II-9[1]. Estimated recharge from precipitation and irrigation return flow to the Ogallala aquifer in western Kansas, as of 1977.

Area	Acreage of Ogallala Aquifer	Irrigated Land: Acres	Irrigated Land: Irrigation Pumpage (acre-feet) @ 18 inches/acre	Irrigated Land: Recharge on Irrigated Land (acre-feet) @ 1.8 inches/acre	Nonirrigated Land: Acres	Nonirrigated Land: Recharge (acre-feet) 0.3 inch/acre	Total Recharge (acre-feet)	Total Recharge (inches)
Northwest Subregion	4,472,000	306,000	459,000	46,000	4,166,000	104,000	150,000	0.40
West-central Subregion	2,478,000	328,000	492,000	49,000	2,150,000	54,000	103,000	0.50
Southwest Subregion	5,119,000	1,548,000	2,322,000	232,000	3,571,000	89,000	321,000	0.75
Western Kansas Total	12,069,000	2,182,000	3,273,000	327,000	9,887,000	247,000	574,000	0.57

[1]Adapted from O'Connor and McClain (1982)

B9. Great Bend Prairie aquifer regional recharge estimates (Fader and Stullken, 1978; Cobb et al., 1983)

Fader and Stullken (1978) evaluated the groundwater resources of the Great Bend Prairie in south-central Kansas. They estimated groundwater recharge for the combined drainage area above the stream-gaging stations of Raymond (Rattlesnake Creek), Arlington (North Fork Ninnescah River), and Murdock (South Fork Ninnescah River), where average annual precipitation was estimated to be 25 inches/yr. The groundwater drainage area above the three stations was estimated to be 2,280 square miles based on a December 1973 potentiometric surface map of the region. Fader and Stullken's recharge by precipitation estimate to the above groundwater drainage area was 240,000 ac-ft/yr or 2 inches/yr. Based on streamflow duration curve analysis, the combined groundwater contribution to streamflow at these stations was estimated as about 110,000 acre-ft/yr or 0.9 inch/yr (Fader and Stullken, 1978).

Recharge to the groundwater reservoir is principally by direct infiltration of precipitation and irrigation on the land surface throughout the area plus underflow laterally from the west, and leakage upward from the bedrock. Recharge to the area by underflow occurs only across the western Kiowa County line and was estimated to be 500 to 1,000 acre-ft/yr. The inflow from the bedrock was estimated to be 5,000 to 10,000 acre-ft/yr (based on the assumptions that the Cedar Hills Sandstone is the major contributor, the hydraulic gradient in that formation is virtually equal to and in the same direction as in the overlying unconsolidated deposits, and the hydraulic conductivity of the Cedar Hills Sandstone is about 25 ft/day).

Fader and Stullken (1978) estimated that 900,000 acre-ft of water was withdrawn by wells through the Great Bend Prairie during 1952-1971, of which 680,000 acre-ft was for irrigation and 220,000 acre-ft was for municipal and industrial use. Sixty-two percent of the wells recorded in May 1974 were within the groundwater drainage area above the stream-gaging stations near Raymond, Arlington, and Murdock.

Cobb et al. (1983), previously of the Kansas Geological Survey, calibrated (by trial and error) the Trescott et al. (1976) USGS two-dimensional finite-difference flow model using a grid spacing of 15,000-ft × 15,000-ft throughout the Great Bend Prairie region. The resulting average recharge was 0.75 inch/yr.

As previously mentioned, Luckey et al. (1986), employed the Trescott et al. (1976) USGS two-dimensional finite-difference flow model using a grid spacing of 10-mi × 10-mi and calibrated it for the central High Plains, which incorporates the Great Bend Prairie region (please also refer to sections C10-C12 and D2-D3 below). The estimated predevelopment long-term average recharge rate for the Great Bend Prairie was 0.28 inch/yr.

B10. Other regional studies involving Kansas High Plains recharge estimates (Jenkins and Pabst, 1975; Landon, 2001, 2002)

Jenkins and Pabst (1975) in a study of Northwest Kansas (nine northwest counties covering an area of 8,050 square miles) estimated annual recharge from precipitation to be 0.25 inch.

Landon (2001, 2002) in a preliminary MODFLOW modeling study of the High Plains aquifer in the Republican River basin in Nebraska, Kansas, and Colorado, estimated average predevelopment recharge rate of 0.26 inch/yr across the entire 30,224-mi^2 active grid-node Republican River basin model area.

C. Basin-scale to county-scale groundwater studies

The basin- to county-scale groundwater studies highlighted here consist of studies that addressed groundwater recharge in some quantitative fashion either through numerical modeling or hydrologic budget analysis. Additional studies in this category are summarized in Part IV.

C1. Wichita and Scott counties.

Dunlap et al. (1980) applied the USGS 2-D finite difference groundwater flow model (Trescott et al., 1976) to a 480-square-mile area centered northeast of the town of Leoti, in Wichita County. The model was calibrated for both predevelopment (1950-51) or steady-state, and transient (1951-1977) conditions. The calibrated, uniform, steady-state recharge was 0.28 inch/yr. A soil-zone model (Lappala, 1978) was used to estimate crop-water demand and irrigation needed to maintain the available soil moisture at 50 percent. The annual pumpage employed in the transient model was calculated based on annual changes in crop acreage and crop-irrigation demand (derived from the soil-zone model). Irrigation return flow was estimated to be minimal. The numerical model was more sensitive to changes in pumpage than to hydrogeologic parameters and recharge, and thus the latter remained unchanged during the transient simulation.

C2. Lane and Scott counties, west-central Kansas.

Gutentag and Stullken (1976) studied the groundwater resources of Lane and Scott counties and presented a detailed water budget for Scott County for 1971 (a dry year) and 1972 (a wet year) as shown in Table II-10.

Table II-10[1]. 1971 and 1972 water budgets for Scott County[2] (values in inches/yr).

Budget item	Inflow		Outflow	
	1971	1972	1971	1972
Boundary flow	0.23	0.23	0.05	0.05
Recharge (from precipitation and streamflow losses)	0.36	0.62	---	---
Streamflow gains	---	---	0.05	0.05
Net well pumpage (total pumpage minus 20% irrigation return flow to aquifer	---	---	3.11	1.97
Loss of groundwater storage	2.62	1.22	---	---
Totals	3.21	2.07	3.21	2.07

1. Adapted from Gutentag and Stullken (1976)
2. Scott County area = 724 mi^2 = 463,360 acres

Recharge was considered to be 10% of the precipitation on irrigated land during the growing season and 1% of the precipitation on nonirrigated land during the growing season. Infiltration of streamflow (streamflow losses) is considered to be included with the recharge estimate from precipitation (Gutentag and Stullken, 1976). Thus recharge from precipitation during 1971 (a dry year) was estimated to have been 0.36 inch, and during 1972 (a wet year) 0.62 inch for Scott County. The corresponding values of estimated recharge for Lane County were 0.08 inch (1971) and 0.16 inch (1972), respectively. An additional source of recharge is return flow from irrigation. Gutentag and Stullken (1976), using figures experimentally derived by Meyer et al. (1953) for irrigated land in Finney County, assumed that 20% of withdrawal by wells subsequently returns to the groundwater reservoir (see section C5). The total amount of water pumped for irrigation in Scott County was 150,000 acre-feet (3.88 inches) in 1971 and 95,000 acre-feet (2.46 inches) in 1972. The part of the total water pumped that returns to the reservoir in Scott County was thus estimated to be 30,000 acre-ft (0.78 inch) in 1971 and 19,000 ac-ft (0.49 inch) in 1972. If the seepage of irrigation water is added to the recharge inflow and boundary inflows given in the budget for Scott County, the total recharge would be 1.37 inches in 1971 and 1.35 inches in 1972, indicating that total recharge remains relatively constant from year to year when there is little change in irrigated acreage (Gutentag and Stullken, 1976). Evapotranspiration from groundwater was considered negligible because the water table was well below the root zone nearly everywhere in the area (Gutentag and Stullken, 1976).

C3. Arkansas River valley in Hamilton and Kearny counties.

Barker et al. (1983) applied a USGS 2-D finite element groundwater flow model (written by J. V. Tracy and documented in Dunlap et al., 1984) to nearly 110,000 acres of the Arkansas River valley between the Colorado-Kansas state line and the Bear Creek Fault Zone in southwestern Kansas. The model was calibrated for both steady-state (pre-1970, considered as averaged 1951-1969 conditions) and transient conditions (1970-79). Monthly pumpage data were estimated from energy-consumption records. The simulated hydrologic fluxes for two periods, 1970-74 and 1975-79 are presented in the water budget Table II-11 below.

Table II-11[1]. Simulated water budget for Arkansas River alluvium between Colorado-Kansas state line and Bear Creek Fault Zone, Kearny and Hamilton counties, Kansas.[2]

	1970-74 transient simulation		1975-79 transient simulation	
Budget item	Inflow	Outflow	Inflow	Outflow
Boundary flow	1.15	1.59	1.02	1.35
Net recharge from precip. & irrigation (recharge–groundwater ET)	1.95	---	4.21	---
Stream leakage	1.50	0.54	1.67	0.08
Pumpage	---	3.10	---	6.23
Loss of groundwater storage	0.62	---	0.77	---
Totals	5.22	5.23	7.67	7.66

1. Adapted from Barker et al. (1983)
2. Model area approximately 110,000 acres

Recharge to the aquifer from incident water (precipitation plus irrigation) amounted to 22 to 26% of that total water, although approximately 15% of that recharged water was lost though groundwater evapotranspiration.

C4. Unconsolidated aquifer system of Kearny and Finney counties.

Dunlap et al. (1985) applied the 3-D USGS finite difference model (Trescott, 1975; Trescott and Larson, 1976) to simulate the High Plains, Arkansas River valley, and sandhills aquifer system in Kearny and Finney counties. Three vertical layers were used in the model: The top layer represented the valley and upper aquifers; the middle layer, the confining zone; and the bottom layer, the lower aquifer, which is the principal water-source aquifer in the area. The model was calibrated for both predevelopment (1940) or steady-state conditions, and transient (1974-1980) conditions. The pumpage from 1974-1980 was estimated over 4-month periods from 1) a soil-zone model (Lappala, 1978), calculating crop water demand for the major crops in the area, and 2) irrigated acreage data.

Predevelopment recharge for the Arkansas River valley and sandhills area from precipitation and irrigation was estimated to be 11% of the normal rainfall at Garden City, or 2.08 inches/yr (the same value was also used for transient conditions), and was estimated at less than 0.5 inch/yr for the High Plains aquifer (that was considered negligible in the model application). In the Dunlap et al. (1985) study, irrigation return flow was found insignificant in comparison to other recharge (such as precipitation recharge and river and canal seepage) and pumpage. Seepage losses from the Arkansas

River are a major source of recharge to the aquifer system. The simulated river and canal seepage and boundary inflow averaged 1.13 inches/yr (or 36,200 ac-ft/yr) over the 1974-1980 simulation period.

The simulated water budget for 1980 is shown in Table II-12, which shows that the lower aquifer is recharged by 1) leakage from the confining zone; 2) lateral, subsurface inflow; and 3) canal seepage. The major source of water for pumpage (approximately 58%) is from downward leakage of water from the overlying upper and valley aquifers. Approximately 42% of the groundwater pumped from the lower aquifer came from storage in the lower aquifer (Table II-12).

Table II-12[1]. Simulated water budget for unconsolidated aquifer system of Kearny and Finney counties[2], 1980 (values in inches/yr).

Budget item	Valley and upper aquifer		Lower aquifer	
	Inflow	Outflow	Inflow	Outflow
Boundary flow	0.37	0.90	0.26	0.44
Recharge from precipitation	2.08	---	0	---
River and/or canal seepage	1.10	---	0.19	---
Leakage to confining zone	---	13.37	---	---
Leakage from confining zone	---	---	6.61	---
Well pumpage[3]	---	0	---	11.32
Loss of groundwater storage	10.71	---	4.70	---
Total	14.26	14.27	11.76	11.76

1. Adapted from Dunlap et al. (1985)
2. Model area of valley and upper aquifer ≅ 603 mi^2 = 385,962 acres; model area of lower aquifer ≅ 1,227 mi^2 = 785,280 acres
3. All pumpage was assumed to come from the lower aquifer, except when the crop-irrigation demand could be met by surface water from irrigation canals. The amount of surface water available for irrigation was subtracted from the pumpage in the appropriate grid blocks

C5. Upper Arkansas River corridor in southwestern Kansas

Whittemore et al. (2001) applied and calibrated a two-layer MODFLOW model for the entire corridor of the Upper Arkansas River from the Colorado-Kansas state line to the Crooked Creek-Fowler Fault Zone in eastern Ford County. A rectangular area of 29-mile width and 126-mile length was employed to incorporate the extent of the alluvial trough in Hamilton and western Kearney counties, all of the ditch irrigation service areas, and the High Plains aquifer to the north and south of the river valley. A uniform grid cell size of 0.5-mi by 0.5-mi (0.25 mi^2) was employed, resulting in an active-cell model area of approximately 2,328 square miles. Two layers were employed in the model, one for the alluvial aquifer along the Arkansas River valley, and the other for the High Plains aquifer and the older alluvial aquifer underlying the sand dunes south of the river floodplain in Hamilton and western Kearny counties. A zone of low hydraulic conductivity clays and silty clays underlies much of the Arkansas River coarse sand and gravel alluvium and slows the downward movement of shallow groundwater into the underlying High Plains aquifer. The model was calibrated for predevelopment (1938-1942), steady-state conditions, and for 1991-2000 conditions using 10-year averaged water level measurements. Because of various artifices introduced in the 1990's model simulation conditions, only the predevelopment simulated water budget will be highlighted here. The model calibration involved trial-and-error, cell-by-cell adjustments of hydraulic conductivity and recharge values using the GIS ArcView environment to minimize computed head error. This procedure resulted in a simulated mean head error of less than 1 ft with a standard deviation of less than 3.28 ft (Whittemore et al., 2001). According to the authors, the computer model simulated the net gain in Arkansas River from Garden City to Dodge City as well as that from Syracuse to Garden City "well" (Fig. II-7).

The model-estimated total recharge from precipitation, irrigation canal seepage losses, and irrigation return flows was 1.04 inches/yr over the total active model cell area of 2,328.25 mi^2 (1,490,080 acres). The simulated steady-state water budget is given in Table II-13 (Whittemore, D.O., and Perkins, S.P., personal communication, January 2003), and shows that approximately 81% of the total recharge came from precipitation and irrigation return flows, approximately 15% came from river leakage, and less than 4% from boundary inflow. The major discharge from the alluvium and High Plains aquifer system went to Arkansas River baseflow (45%), with approximately 28.7% being artificially discharged by mostly irrigation wells, and approximately 26% being discharged as boundary outflow.

Table II-13[1]. Water budget from steady-state (1940s) model simulation of Upper Arkansas River corridor in southwestern Kansas[2] (values in inches/yr).

Budget item	Inflow	Outflow
Boundary flow	0.05	0.34
Recharge from precipitation, irrigation canals and irrigation return flows	1.04	---
River seepage	0.19	0.58
Well pumpage	---	0.37
Totals	1.28	1.29

[1.] Whittemore and Perkins, personal communication, January 2003
[2.] Active model area 1,490,080 acres

Whittemore et al. (2001) estimated (by taking into account gaging stations' streamflow differences, streamflow diversions, estimated evaporation, and phreatophyte use) that the average net recharge from the Arkansas River to the alluvium followed by leakage into the underlying High Plains aquifer during the decade 1989-1998 was about

7.4 inches/yr over the entire area of the Arkansas River alluvial valley from Hartland to Dodge City (which is equivalent to 118,000 acres; most of the flow losses between the Kansas-Colorado state line and Garden City occur from the western edge of the High Plains aquifer underlying the river valley near the former town of Hartland). During 1995-2000, when the river flows and recharge were greater, Whittemore et al. (2001) estimated that recharge rates to the river alluvium would be over 1 ft/yr.

C6. Finney County

Meyer et al. (1970) conducted a detailed study of recharge from streams, precipitation, return flow from irrigation, and inflow through the aquifers in Finney County, southwest Kansas. They developed a detailed water budget for all of Finney County minus the panhandle area (a total of 24 townships or 552,960 acres) using the following equation:

$$\textit{Total Recharge} = \textit{Change in Storage} + \textit{Total Discharge}, \qquad \text{(II-3)}$$

where all elements of the right-hand side of the equation were observed or estimated, and *Total Discharge = well pumpage + (lateral groundwater outflow − inflow) + streamflow seepage* (positive for baseflow, negative for streamflow losses). Data from the 24-year period (1940-1964) were used in the calculations. Observed values of the components on the right-hand side of the water-balance eqn. (II-3) are summarized in Table II-14.

Table II-14[1]. Water-balance components for the Finney County study area (552,960 acres). Data for 1940-1964. Values in inches/yr.

Budget item	Development conditions	Predevelopment conditions	1922-1930 predevelopment conditions
Change in storage	−0.511	0	0
Well pumpage	3.192	0	0
Outflow − inflow	0.043 (2000 ac-ft/yr)	0.043 (2000 ac-ft/yr)	0.043 (2000 ac-ft/yr)
Net seepage loss from Arkansas River	−0.037 (1700 ac-ft/yr)	−0.037 (1700 ac-ft/yr)	−0.456 (21,000 ac-ft/yr)
Recharge (from precip. and irrigation return)	2.69	0.006	0.413

[1] Data from Meyer et al. (1970)

Thus, the average recharge indicated by the water-balance equation (II-3) for the 24-year period (1940-1964) was 124,000 ac-ft/yr or 2.7 inches/yr over the area. If the groundwater system were considered to be under near-equilibrium conditions before pumping began, the components in the water balance equation would be as shown under the column "Predevelopment conditions" of Table II-14. However historic streamflow records for the period 1922-1930 indicate a seepage loss from the Arkansas River between Syracuse and Garden City of 21,000 ac-ft/yr or 0.46 inches/yr over the study area. (Because the Arkansas River gains and loses flow in about equal proportions as it passes through Hamilton and Kearny counties, most of the 21,000 ac-ft/yr loss during 1922-1930, like the 2000 ac-ft/yr loss during 1940-1964, is believed to be directly contributed to the groundwater reservoir within Finney County. (This loss would be reduced by about a 300-ac-ft/yr gain in flow of the river east of Garden City.) The long-term average predevelopment recharge to the aquifer was apparently less than 0.5 inch/yr (based on 1922-1930 conditions) or even less that 0.05 inch/yr (based on the 1940-1964 conditions).

Thus, according to Meyer et al. (1970), the development-conditions recharge rate of 2.7 inches/yr probably reflects an additional recharge resulting from recycled groundwater from irrigation and an accompanying increase in effective recharge from precipitation on the irrigated land. The change in land use from native grassland to cropland is a factor contributing to increased recharge. As the water table is lowered by pumping, the evapotranspiration losses are also reduced and the effective recharge to the aquifer is increased. Thus, Meyer et al. (1970) list four factors as probably having a role in increasing recharge during the 1940-1964 period: 1) increased precipitation (the period 1940-1951 was marked by precipitation rates significantly above average), 2) irrigation return flow, 3) change in land-use practices, and 4) evapotranspiration reduction due to lowered water tables.

Impacts of irrigation on recharge

With regard to irrigation-related return flow to the aquifer in Finney County, Meyer et al. (1953) conducted detailed ditch-loss studies and irrigation efficiency experiments at the Garden City Experiment Station. The ditch-loss studies showed losses of 10 percent occurring in a quarter-mile length of a farm supply ditch. Experiments on irrigation efficiencies in the area showed that deep percolation losses (those penetrating more than 6 ft) in a well-drained system are frequently 20 percent or more. With these facts in mind, Meyer et al. (1970) concluded that "it seems reasonable to assume that 15 percent of the water applied to irrigated fields could return to the groundwater reservoir, and that another 10 percent would return from ditch leakage."

According to Meyer et al. (1970), of the total 1940–March 1964 pumpage of 3,530,000 ac-ft, 25% or 882,000 ac-ft returned to the groundwater reservoir, and an additional 219,000 ac-ft returned from surface water. This would be a total contribution of about 1,100,000 ac-ft for the 24-year period, or an average annual irrigation return contribution of approximately 45,700 ac-ft, which is equivalent to 1 inch/yr over the area.

C7. Ford County

Spinazola and Dealy (1983) evaluated the hydrologic conditions in the Ogallala aquifer in Ford County during 1980 and 1981. They produced an approximate water budget for 1980 conditions for that part of Ford County principally underlain by the

Ogallala aquifer, an approximately 700-square-mile area that includes the four northwestern townships and all townships south of the Arkansas River (Table II-15).

Recharge to the aquifer, 0.6 in/yr, was considered as 3% of annual precipitation. Groundwater evapotranspiration along the 48-mile-long reach of the Arkansas River valley was considered to be the same as that calculated along the Arkansas River valley in Hamilton and Kearny counties between 1970-79 by Barker et al. (1983).

Table II-15[1]. 1980 water budget for Ford County underlain by the Ogallala aquifer[2] (values in inches/yr).

Budget item	Inflow	Outflow
Boundary flow	0.45	0.16
Recharge	0.58	---
Streamflow losses	0	---
Streamflow gains		0.14
Groundwater evapotranspiration	---	0.08
Well withdrawals	---	3.23
Loss of groundwater storage	2.58	---
Totals	3.61	3.61

1. Adapted from Spinazola and Dealy (1983)
2. Area considered = 700 mi^2 = 448,000 acres

C8. Pawnee Valley (Sophocleous, 1980, 1981)

In a hydrogeologic study of Pawnee Valley, western Kansas, Sophocleous (1980, 1981) estimated regional groundwater recharge in that valley using two different methods:

i) Interpretation of streamflow records at the discharge end of the flow system and of pumpage data over the 1925-1945 period, where near-equilibrium conditions could be assumed. Thus the long-term average recharge to the alluvial aquifer was assumed to equal the long-term groundwater outflow during the 1925-1945 period. Such analysis resulted in a recharge rate of 0.6 inch/yr over the alluvial aquifer area.

ii) Analysis of the soil-moisture budget based on hydrometeorological and soil data of the composite Pawnee Watershed. This analysis, based on 20 years of hydrometeorological data (1959-1978) using the Thornthwaite method for calculating potential evapotranspiration in conjunction with the modulated soil moisture technique of Holmes and Robertson (1959) to arrive at actual evapotranspiration and moisture surplus and deficit, resulted in a value for regional groundwater recharge of 0.4 inch/yr.

Thus, the average estimated regional groundwater recharge for the Pawnee Valley was about 0.5 inch/yr, which represents 2.2% of the 1959-1978 average annual precipitation of 22.7 inches. The Sophocleous (1980) study indicated that by 1978-79, the Pawnee Valley had been depleted by 37% compared to 1945-47. It is also interesting to note that during 1978-79, the groundwater appropriations in the Pawnee Valley alluvial aquifer (that reached at least 84,000 ac-ft) amounted to about 11 times the amount of estimated natural groundwater replenishment for the Pawnee Valley.

C9. Wet Walnut Creek basin, west-central Kansas

Koelliker et al. (1999) integrated basin modeling study

Koelliker et al. (1999; see also Ramireddygari et al., 2000, and Sophocleous et al., 1998) developed an integrated watershed and groundwater model by combining the watershed model POTYLDR (Koelliker et al., 1982) with the 2-D USGS MODFLOW model, and applied and calibrated it to the Wet Walnut Creek basin in west-central Kansas. The basin was divided into 78 subbasins and the Wet Walnut Creek valley aquifer was discretized into 0.5-mile × 0.5-mile-square cells. Using 1960 initial conditions (assumed to be near-equilibrium conditions), the model was run in monthly steps for the period 1960-1996. The data for the period 1960-1990 were used to calibrate the model, whereas the data for the 1991-96 period were reserved for verifying the calibrated model. The mean residual during the 1960-1990 calibration period was 1.51 ft with a standard deviation of 6.59 ft. The mean residual for the 1991-96 period was 0.53 ft with a standard deviation of 8.22 ft. The mean and median 1960-1996 water budget components for the Wet Walnut Creek valley aquifer are shown in Table II-16. The relatively large contrast of the mean and median values are indicative of the high variability of water budget components in the Wet Walnut valley.

Table II-16. Mean and median of the 1960-1996 water budget components for the Wet Walnut Creek valley aquifer[1] (in inches/yr).

Budget item	Mean	Median
Net boundary inflow	-0.04	0.003
Total recharge (from precipitation, irrigation & pond seepage)	1.89	1.08
Net streamflow loss	1.91	2.15
Well pumpage	4.02	4.39
Net loss of groundwater storage	0.28	1.43

[1]Aquifer area simulated = 174.63 mi^2 = 111,764 acres

The average 1990-96 recharge to the aquifer (which also includes pond seepage from the watershed structures and irrigation return flow in addition to precipitation percolation) was estimated at 1.9 inches/yr; however, it ranged widely from less than 0.05 inch/yr to more than 16 inches/yr during the 1993 flood year. In comparison, the average pumpage from the aquifer over the same period was 4.0 inches/yr. The average net stream seepage to the aquifer (i.e. streamflow loss minus streamflow gain or baseflow) was estimated at 1.9 inches/yr (Table II-16). Wet Walnut Creek was a net-gaining stream up to the mid to late 1960s, but since the late 1960s it became a net-losing stream to the alluvial aquifer. By the early to mid-1980s, the stream network system became the major source of recharge to the aquifer, even exceeding precipitation percolation- and pond-seepage-based recharge.

Other Wet Walnut Creek valley groundwater studies

Gillespie and Slagle (1972), in a groundwater recharge study of Wet Walnut Creek valley from Bazine to Albert, estimated the average annual recharge to the aquifer for 1965-69 to be 13,000 acre-ft/yr. If we approximate the study area to be 64,000 acres (100 mi^2), this recharge value converts to an estimate of 2.4 inches/yr.

Nuzman (1990) conducted a numerical modeling study of the Walnut Creek valley from near Ness City to Great Bend, an area of 124,160 acres (194 mi^2), using the USGS MODFLOW program (McDonald and Harbaugh, 1988) and a grid of 1-mi × 1-mi. A trial-and-error calibration resulted in a recharge estimate of 10% of an average 22 inches of annual precipitation (i.e., 2.2 inches/yr).

Finally, as a result of the 1990-91 public hearings on the designation of an Intensive Groundwater Use Control Area (IGUCA) in Barton, Rush, and Ness counties, Kansas, the Chief Engineer concluded, regarding long-term groundwater recharge in the Walnut valley, "that the long-term sustainable yield of the aquifer within the boundaries of the proposed control area as set forth in Conclusion No. 8 (i.e., an area of 348,800 acres) is no more than approximately 22,700 acre-ft per year" (Division of Water Resources, 1992, p. 96). This translates to 0.8 inch/yr. However, the declared IGUCA encompasses areas beyond the Walnut Creek alluvium. Considering that the Walnut Creek alluvial aquifer area from near Ness City to the confluence at Great Bend is approximately 128,000 acres (200 mi^2), the sustainable yield figure of 22,700 acre-ft/yr translates to a long-term recharge estimate of 2.1 inches/yr.

C10. Solomon River basin, Kansas

Jorgensen and Stullken (1981) studied the North Fork Solomon River valley from Kirwin Dam to Wakonda Lake (Glen Elder Dam) in north-central Kansas. They applied the USGS 2-dimensional finite-difference model (Trescott et al., 1976) to the area using the 1946 water-level conditions as approximately steady-state conditions, and employing a rectangular 0.25-mile × 0.50-mile grid network. The estimated effective recharge from precipitation (i.e. aquifer gain from precipitation minus evapotranspiration) was 2.55 inches/yr. The simulated steady-state hydrologic budget is shown in Table II-17 below.

According to Jorgensen and Stullken (1981), the simulated net leakage of approximately 2.5 inches/yr (i.e., leakage to the river minus leakage from the river, Table II-16) was consistent with the estimated gain in baseflow of the river within the area modeled.

Table II-18[1]. North Fork Solomon River valley[2] from Kirwin to Glen Elder dams used to simulate steady-state (1946) conditions (values in inches/yr).

Budget item	Inflow	Outflow
Boundary flow	0.08	0.11
Effective recharge	2.55	---
Leakage from river	0.47	---
Leakage from alluvial tributaries and into terrace deposits	0.56	---
Leakage to river	---	2.98
Well pumpage	---	0.02
Riparian evapotranspiration	---	0.53
Totals	3.66	3.64

1. Adapted from Jorgensen and Stullken (1981)
2. Simulated area: 55,040 acres

Burnett and Reed (1986) applied the same USGS model and grid network used by Jorgensen and Stullken (1981) to the South Fork Solomon River valley from Webster Reservoir to Waconda Lake (Glen Elder Dam) in north-central Kansas. They calibrated the model for transient 1970-79 conditions using two pumping patterns per year—one pumping period simulating the nonirrigation season (September through May) of each year, and another period simulating the irrigation season (June through August) of each year. The simulated hydraulic heads were within ± 5 ft of the measured hydraulic head at 49% of the sites, and were within ± 10 ft of the measured hydraulic head at 82% of the sites. Table II-18 shows the average simulated hydrologic budget for the area.

Average annual recharge to the aquifer from precipitation was determined to be 1.7 inches/yr for 1970-78. Average annual recharge from the Osborn irrigation canal was estimated to be 0.9 inch/yr over the same period. The 1970-78 average annual discharge from the aquifer to the river (baseflow) was estimated to be 2.3 inches/yr, whereas total well pumpage over the same period averaged about 1 inch/yr.

Sophocleous et al. (1990; see also McClain et al., 1995) estimated groundwater recharge for the Solomon River basin using two different methods: 1) interpretation of early streamflow records (1960-61) at the North Fork Solomon River gaging station near Glade (Phillips Co.), and 2) using a monthly soil-moisture budget for the Solomon River basin over the 25-yr period 1964-1988.

The long-term average recharge to the alluvial aquifer was assumed to equal the long-term average groundwater outflow during the early times of the Solomon watershed irrigation development. Such an equilibrium condition existed in the watershed until the early 1960s (Weston, 1979). During 1960 and 1961, the average amount of groundwater appropriated in the ~395,674-acre area drained by the North Fork Solomon above Glade was 13,860 acre-ft/yr, which amounts to 0.42 inch/yr over that subwatershed area (water appropriation data from Division of Water Resources, Kansas State Board of Agriculture). The average annual baseflow during the period 1960-61, as derived from the streamflow data at Glade, was ~10,200 acre-ft/yr, which amounts to 0.31 inches of water over the same subwatershed area. Thus the total groundwater outflow (baseflow plus pumpage) for 1960-61 was 0.73 inch/yr, which, under the assumption of equilibrium, represents the amount of groundwater recharge. Groundwater outflow through evapotranspiration was presumed negligible and therefore was not considered in the calculations.

The second method for estimating regional groundwater recharge in the Solomon River basin is the moisture-budget technique. If the 25 years of record (1964-1988) for all 19 weather stations covering portions of the Solomon River basin (Sophocleous et al., 1990) are considered representative of the average conditions in the Solomon River watershed, moisture budgets indicate that the total annual average moisture surplus or average potential annual groundwater replenishment in this watershed for predominant 12-inch soil-moisture-capacity soils varies from 0 inches to 3.8 inches. For the 1960-61 period, and based on the climatic data from the Kirwin Dam station (the closest station to the Glade streamgaging station and centrally located within the entire watershed) and the predominant soil-moisture capacity of 12 inches, precipitation totaled 27.84 inches, Thornthwaite potential evapotranspiration and actual evapotranspiration totaled 27.79 inches and 25.98 inches, respectively, and moisture surplus totaled 1.89 inches, which is above normal compared to the 25-year average of 1.16 inches for the same conditions.

During the 1960-61 period, the average total streamflow at Glade was 34,720 ac-ft/yr and the average baseflow was 10,200 ac-ft/yr, resulting in a direct surface runoff (the difference between total streamflow and baseflow) of 24,520 ac-ft/yr (0.74 inch/yr). The moisture surplus must, however, satisfy both the surface runoff and the groundwater recharge. This surface runoff figure, when subtracted from the average 1960-61 moisture surplus of 1.89 inches, based on the Kirwin station, results in a value for regional groundwater recharge of 1.15 inches. This value is of similar magnitude as the recharge value (0.73 inch/yr) calculated from baseflow and groundwater pumpage data.

Thus, assuming that the more than 395,000-acre subwatershed above Glade is typical of the entire Solomon watershed, the average estimated regional groundwater recharge for the Solomon watershed is 0.94 inches (based on the two previously mentioned recharge estimation methods), which represents only 4 percent of the average annual precipitation (23.29 inch/yr). During 1980-81, the groundwater appropriations in the Glade subwatershed, which reached 146,182 ac-ft, compared to 13,860 ac-ft in 1960-61, amounted to more than 4.7 times the amount of estimated natural groundwater replenishment for that subwatershed.

South Fork Solomon River basin above Webster Reservoir

Weston (1979) computed an annual groundwater budget for the South Fork Solomon River basin above Webster Reservoir for the period 1947-1976. By assuming equilibrium conditions prior to 1966, and by estimating long-term average outflow during that period (consisting of estimated baseflow, and total appropriated pumpage minus 20% to reflect net withdrawal since a portion of what is pumped is returned to the groundwater system), Weston (1979) estimated total inflow or recharge for the South Fork Solomon River basin above Webster Reservoir (an area of 1044 mi^2) as 0.56 inch/yr for the predevelopment, steady state period 1947-1965. Average annual recharge for the entire 1947-1976 study period was estimated at 0.51 inch/yr.

C11. Arkansas River valley from Kinsley to Great Bend

Sophocleous et al. (1993) modeled the Arkansas River valley from Kinsley to Great Bend using the MODFLOW model in two dimensions in conjunction with the parameter estimation model MODINV (Doherty, 1990). They employed a regular, rectangular 1 mi × 1 mi grid, oriented in a southwest to northeast direction incorporating the Arkansas River from Kinsley to Great Bend, and calibrated the model for both predevelopment, steady-state (1955) and transient (1955-1990) conditions using annual time steps and pumpage data from DWR water rights data.

The calibrated steady-state recharge was estimated as 1 inch/yr with a standard error of 0.02 inch/yr, and the average 1955-1990 recharge was estimated as 1.8 inch/yr with a standard error of 0.06 inch/yr. A summary of all inflows and outflows in the region is presented in the predevelopment and 1985-1990 development-period water budgets (Table II-19). The ratios $s/\Delta h$ (square root of error variance over the difference between the highest and lowest value of head in the model region) are relatively small (0.009 and 0.015 for the 1955 and 1990 water budgets, respectively), indicating satisfactory model fit.

Table II-18[1]. Average (1970-78) hydrologic budget for the Sourth Fork Solomon River valley[2] from Webster to Glen Elder dams (values in inches/yr).

Budget item	Inflow: Irrigation season (90 days)	Inflow: Nonirrigation season (275 days)	Annual inflow total	Outflow: Irrigation season (90 days)	Outflow: Nonirrigation season (275 days)	Annual outflow total
Boundary flow	0.06	0.19	0.25	0.01	0.04	0.05
Recharge from precipitation	0.46	1.21	1.67	---	---	---
Recharge from groundwater irrigation return	0.08	0	0.08	---	---	---
Recharge from surface water irrigation return	0.11	0	0.11	---	---	---
Leakage from Osborn irrigation canal	0.89	0	0.89	---	---	---
Net river leakage	---	---	---	0.36	1.93	2.29
Well pumpage	---	---	---	0.81	0.15	0.96
Groundwater evapotranspiration	---	---	---	0.16	0.11	0.27
Loss of groundwater storage	---	0.83	0.83	0.26	---	0.26
Totals			3.83			3.83

1. Adapted from Burnett and Reed (1986)
2. Active-node simulated area: 64,000 acres

Table II-19[1]. Simulated steady-state and transient groundwater budgets of Arkansas River valley[2] from Kinsley to Great Bend (values in inches/yr).

Budget item	1955 steady-state simulation		1985-90 transient simulation	
	Inflow	Outflow	Inflow	Outflow
Boundary flow	0.18	0.66	0.33	0.43
Recharge	0.97	---	1.78	---
Streamflow loss	0.32	---	1.30	---
Streamflow gain	---	0.59	---	0.20
Well pumpage	---	0.23	---	4.19
Net loss of groundwater storage	0	---	1.41	---
Totals	1.47	1.48	4.82	4.82

1. Adapted from Sophocleous et al. (1993)
2. Simulated active-node model area: approx. 288,000 acres

The predevelopment budget shows that the major input to the aquifer is groundwater recharge and the major outputs from the system are lateral outflow from the model boundary near Great Bend and stream baseflows (streamflow gain). In contrast, the major inputs to the aquifer in recent times come from both streamflow losses and recharge, and the major outflow is well pumping, whereas streamflow gains from baseflows decreased significantly. Groundwater pumpage for irrigation is now balanced by an increase in recharge (mainly from irrigation return flow and conversion of grasslands to croplands), a decrease in natural discharge (mainly decreases in baseflow), and a net loss in aquifer storage (as manifested by long-term groundwater-level declines).

C12. Rattlesnake Creek watershed, south-central Kansas

Sophocleous and Perkins (1993 a,b) studied the lower Rattlesnake Creek watershed from west of the Macksville stream-gaging station near the southwest Stafford County boundary to the confluence with the Arkansas River in Rice County, an area of approximately 560 square miles. They applied and calibrated the USGS MODFLOW finite-difference model to the study area using one-square mile grid cell network in two dimensions and employed parameter estimation techniques to optimize model parameters for both steady-state (mid-1950s conditions) and transient (1955-1991) conditions using 3-yr stress periods and annual time steps. The parameter estimation program MODINV (Doherty, 1990) was employed to optimize model parameters. The predevelopment (c. 1955) recharge was estimated as 1.3 inches/yr with a standard error of 0.4 inch/yr, whereas the average development-period recharge (1955-1990) was estimated as 1.9 inches/yr with a standard error of 0.3 inch/yr. The predevelopment and development water budgets are displayed in Table II-20. The ratio of the square root of the error variance in the parameter estimation model over the difference between the highest and lowest value of head in the region was 0.0094 to 0.0096, a relatively very small value for both the steady-state and transient simulations, indicating that errors in the model were considerably less than the model response as indicated by the maximum head loss of 335 ft.

Table II-20[1]. Predevelopment (c.1955) and development (1988-1990)[2] water budgets for the lower Rattlesnake Creek watershed[3] (values in inches/yr).

Budget item	1955 steady-state simulation Inflow	1955 steady-state simulation Outflow	1988-1990 transient simulation Inflow	1988-1990 transient simulation Outflow
Boundary flow	0.32	0.05	0.41	0.05
Recharge	1.27	---	1.89	---
Streamflow loss	0.07	---	0.09	---
Streamflow gain	---	0.83	---	0.59
Groundwater evapotranspiration	---	0.72	---	0.66
Well pumpage	---	0.06	---	2.30
Loss of groundwater storage	0	0	1.21	0
Totals	1.66	1.66	3.60	3.60

1. Adapted from Sophocleous and Perkins (1993a)
2. Last stress period of the transient 1955-1990 simulation
3. Active node model area is approximately 535 mi^2 = 342,400 acres

In contrast to the 1950s water budget, where the largest outflows from the aquifer were baseflow to streams and evapotranspiration losses mainly to the Quivira marsh region, the present day (1990) dominant outflow is groundwater pumpage for irrigation. This superimposed discharge to the aquifer system is balanced by an increased in recharge (mainly through irrigation return flow and conversion of grasslands and dryland farming to irrigated agriculture), a decrease in discharge (mainly through decreased baseflow contributions to streams and decreased evapotranspiration), and a loss from aquifer storage (as manifested by long-term groundwater-level declines).

Sophocleous et al. (1997, 1999) also developed and applied an integrated watershed and groundwater model to the entire Rattlesnake Creek basin (approximately 1,317 square miles) by combining the USDA Soil and Water Assessment Tool (SWAT) watershed model (Arnold et al., 1993) with the USGS MODFLOW model. The basin was divided into 35 topographic subbasins (Fig. II-10) and the stream-aquifer system was modeled for both predevelopment or steady-state conditions (pre-1960), and development (1955-1994) or transient conditions. The calibration period spanned from predevelopment to 1980; the data for the post 1980 period were reserved for verifying/corroborating the calibrated model. The parameter estimation program (PEST; Doherty et al., 1994) was employed to optimize the aquifer parameters. The predevelopment average recharge was estimated at 1.4 inches/yr, whereas the average 1990-94 period recharge was estimated at 2.2 inches/yr but ranged from 0.5 inch/yr during 1994 to 5.0 inches/yr during the flood year of 1993. Table II-21 displays the water budget for the Rattlesnake Creek basin during predevelopment and present-day (1991, 1992, 1993, and 1994) conditions, representing a transition from extreme dryness to extremely wet conditions.

The average calibrated recharge for the entire transient simulation period (1955-1994) in each subwatershed in the Rattlesnake Creek basin is shown in Fig. II-10. Because the effective recharge to the aquifer is largely a function of the soil type and land use, given similar precipitation patterns, a wide variation in recharge can be observed. For example, subwatersheds 1 through 4 in the far upstream boundary of the watershed near the Kiowa-Edwards-Ford counties border (Fig. II-10), consisting predominantly of the low hydraulic-conductivity Harney soil, have recharge averaging 0.4 to 0.8 inch/yr. On the other hand, subwatersheds 33 and 35 in southwest and northeast Stafford County, respectively (Fig. II-10), that consist predominantly of the highly permeable Tivoli soils, show much higher recharge, averaging more than 4 inches/yr. (This more detailed and up-to-date study supercedes the recharge estimates based on the soil-water budget pilot study of the Rattlesnake Creek basin by Sophocleous and McAllister, 1987, 1990. However, the impact of soil, plant, and land-use factors on recharge from that study are presented in Part I, section 8.1.) The overall area-weighted average recharge during the 1955-1994 simulation period was 2.1 inches/yr.

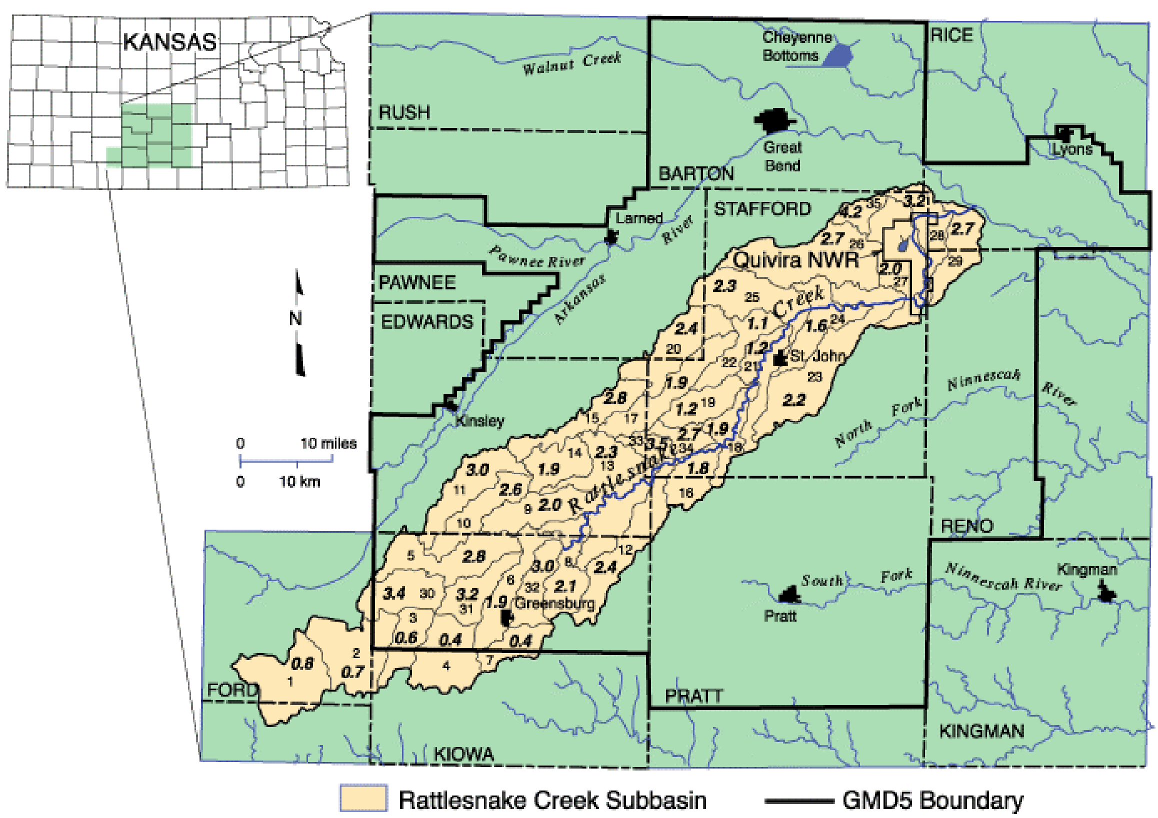

Figure II-10. Rattlesnake Creek subwatersheds and simulated average (1955-1994) recharge rate (inches/yr) in each sub-watershed. Bold numbers are recharge rates; smaller numbers are subwatershed identification numbers (adapted from Sophocleous et al., 1997).

C13. Water budgets for the major Kansas wetlands of Cheyenne Bottoms and Quivira National Wildlife Refuge

The two major Kansas wetlands of Cheyenne Bottoms and Quivira National Wildlife Refuge are shown in Figs. II-10 and II-13. Sophocleous and Shapiro (1987) employed the Versatile Soil-moisture Budget (VB) model of Baier et al. (1979) to calculate the water budget of the nonwater-covered area of the Drummond-Tabler soil association (more than 86% of total area of approximately 60 square miles), which encompasses the Cheyenne Bottoms. The water budget was run for two water years (Oct. 1, 1982-Sept. 30, 1984). The vegetation cover of this area consisted of 44% small grains, 40% grasses, and 12% cattails. The average hydrologic budget for that area is shown in Table II-21 with an estimated deep drainage of 2 inches/yr. This deep drainage value is practically equal to groundwater recharge because of the shallow water table in the area.

Sophocleous and Ma (1993) also analyzed the hydrologic budget of a portion of the Rattlesnake Creek watershed encompassing the Quivira National Wildlife Refuge (NWR) in south-central Kansas (145 mi^2). They employed the Versatile Soil-moisture Budget (VB) model, supplemented with additional surface runoff routines (Sophocleous and Ma, 1993), to estimate the daily hydrologic budget of the study area during the 8-yr period 1985-1992. The 8-yr average hydrologic budget for the nonwater-covered area pertaining to the Quivira NWR is estimated as shown in the following Table II-23.

Note that the deep drainage is highly variable (high coefficient of variation) and is practically equal to groundwater recharge because of the relatively shallow depth to water table in the Quivira NWR.

Table II-21[1]. Predevelopment (pre-1960) and present-day (1991-94) annual water budgets for the Rattlesnake Creek basin (values in inches/yr).

Budget item	pre-1960 predevelopment simulation		1991 transient simulation		1992 transient simulation		1993 transient simulation		1994 transient simulation	
	Inflow	Outflow	Inflow	Outflow	Inflow	Outflow	Inflow	Outflow	Inflow	Outflow
Boundary flow	0.29	0.87	0.10	0.63	0.14	0.39	0.20	0.39	0.13	0.42
Recharge	1.44	---	0.69	---	3.11	---	5.03	---	0.52	---
Streamflow loss	0.17	---	0.15	---	0.34	---	0.40	---	0.13	---
Streamflow gain	---	0.72	---	0.19	---	0.20	---	0.36	---	0.31
Well pumpage	---	0.16	---	3.27	---	1.96	---	2.20	---	3.06
Outflow to Quivira marshes	---	0.16	---	0.16	---	0.16	---	0.20	---	0.18
Loss of groundwater storage	0	0	3.35	0.05	1.43	2.31	1.28	3.76	3.33	0.14
Totals	1.90	1.91	4.29	4.30	5.02	5.02	6.91	6.91	4.11	4.11

[1] Data from Sophocleous et al. (1997)

Table II-22[1]. Average 1983 and 1984 water-year budget for the Drummond-Tabler soil association encompassing the Cheyenne Bottoms.

Hydrologic component	Average value (inches/yr)
Precipitation	21.76
Actual evapotranspiration	21.24
Soil-profile (60-inches) moisture deficit	2.25
Surface runoff	0.55
Deep drainage	2.01

1. Adapted from Sophocleous and Shapiro (1987)

Table II-23[1]. 1985-1992 hydrologic budget components for Quivira National Wildlife Refuge (based on Hudson NOAA climatic station).

Hydrologic component	Average value (inches/yr)	Std. deviation
Precipitation	25.99	6.84
Actual evapotranspiration	24.09	5.22
Soil-profile (60-inches) moisture deficit	3.84	2.03
Surface runoff	0.90	0.70
Deep drainage	0.87	1.37

1. Adapted from Sophocleous and Ma (1993)

C14. Equus Beds aquifer modeling

a) Green and Pogge (Green et al., 1973) study

Green and Pogge (Green et al., 1973; see also Knapp et al; 1975) developed a comprehensive basin hydrology simulator by combining the Kansas Water Budget Model (Smith and Lumb, 1966), a model similar to the Stanford Watershed Model, with a 2-D finite difference groundwater model. They field tested this simulator on the Little Arkansas River basin that encompasses the Equus Beds aquifer for the 1946-1970 period. Groundwater recharge data are only presented for subbasin 5 (within the drainage basin of Little Arkansas River), which includes 139 mi^2 in portions of townships 23S to 25S and ranges 1W to 2W, including the towns of Halstead and Sedgwick, Kansas.

Figure II-11 from that study shows the percent of annual precipitation falling on that subbasin that percolates to groundwater for the period 1946 to 1970. The amount of percolation recharge is highly dependent on the actual hydrologic conditions that exist during each year. It is clear from that figure, that use of a simple fraction of the precipitation would result in significant errors in the amount of percolation recharge.

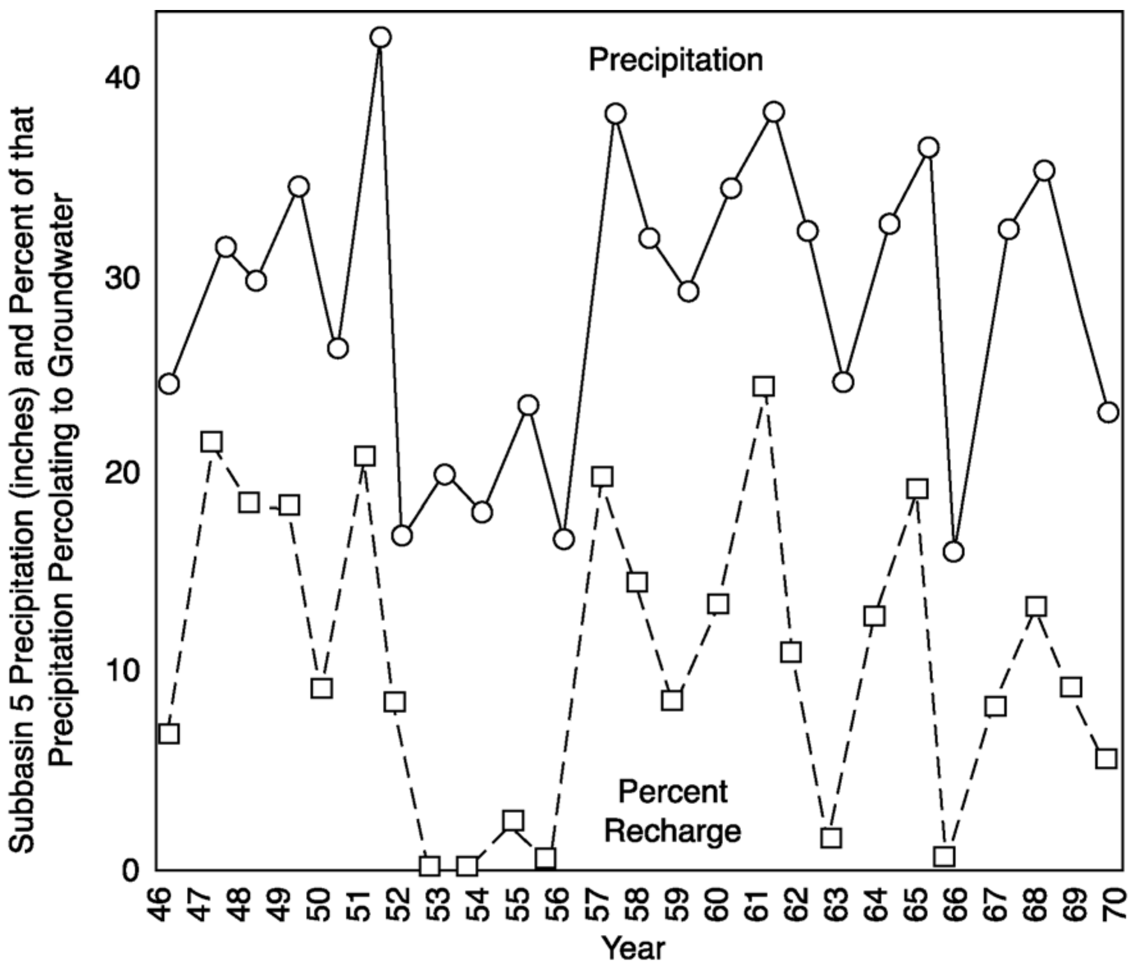

Figure II-11. Percent of annual precipitation for the period 1946-1970 falling on subbasin 5 within the drainage basin of Little Arkansas River (encompassing Halstead and Sedgwick, Kansas) that percolates to groundwater (adapted from Green et al., 1973).

b) Sophocleous (1982) study

Sophocleous et al. (1982; see also Sophocleous, 1984) employed a finite difference (1-mi^2 grid cell) parameter estimation model (Cooley, 1977, 1979) to a 240-square-mile area of the Equus Beds aquifer encompassing portions or all of townships 22S-25S and ranges 1W-4W, including the towns of Burrton, Halstead, and Sedgwick, as well as the Wichita well field area. Steady-state conditions (existing during the early 1940s) were employed in optimizing model parameters. Two recharge zones were considered in that model, one in the sand dune area north of the town of Burrton, and the other encompassing the rest of the area. A detailed uncertainty analysis of the model results is presented in that study. The "optimized" recharge estimate for the sand dune region north of Burrton was 6.5 inches/yr, with a standard error of 1.2 inches/yr, whereas the rest of the model region, including the Wichita well field, had a recharge of 1.65 inches/yr with a standard error of 0.5 inch/yr. The ratio of the square root of the error variance in the parameter estimation model over the difference between the highest and lowest value of head in the region was 3.6 ft/150 ft = 0.02, a relatively small value, indicating that errors in the model were considerably less than the model response as indicated by the maximum head loss of 150 ft. {Sophocleous et al., 1982 (see also Sophocleous, 1984) employed three different solute transport models to predict the extent and concentration of the brine plume near Burrton vis-à-vis the Wichita well field.}

c) Spinazola et al. (1985) study

Spinazola et al. (1985) employed the 3-dimensional USGS MODFLOW model to simulate the Equus Beds (upper layer) and Wellington (lower layer) aquifers in Sedgwick, Harvey, Reno, McPherson, and Marion counties using a regular, 1-mi^2 cell grid. The model was run for both steady-state (1940) and transient conditions (1940-1979). Calculated recharge to their model was considered to be a function of 1940 precipitation, the soil type, and thickness of clay in the unsaturated zone. The combination of soil types and unsaturated clay thicknesses were used to define a "recharge factor" for each cell of the model grid, that when multiplied by the average annual precipitation resulted in the recharge estimates employed in the modeling. Figure II-12 displays the 1940 initial-condition simulation recharge estimate, with values ranging from 0.1 to 5.5 inches/yr. The transient simulation consisted of five stress periods between 1940 and 1979 (1940-1952, 1953-58, 1959-1963, 1964-1970, and 1971-79) corresponding to uniform trends in withdrawal from the aquifer, where the well withdrawal was averaged for the length of the stress period. Recharge changed for each stress period as the product of the "recharge factor," mentioned previously, and average precipitation during the stress period. The simulated water budget for the Equus Beds aquifer for the end of 1940 initial-condition simulation and for the periods 1964-1970 and 1971-79 are shown in Table II-24.

Between 1940 and 1964-1970, withdrawals by wells increased 1,630%. Streamflow gain decreased by 54%, whereas streamflow loss increased by 760%. Between 1964-1970 and 1971-79, withdrawal by wells increased by about 42%. Recharge increased by about 13% during this period; however, the rate of decrease in storage was about 26%, resulting in lower water table elevations, which in turn resulted in decreased baseflow to streams by about 21%, while loss from streams to the aquifer increased by about 57%. Declining water levels also resulted in a small decrease in groundwater transpiration and boundary flow (Spinazola et al., 1985).

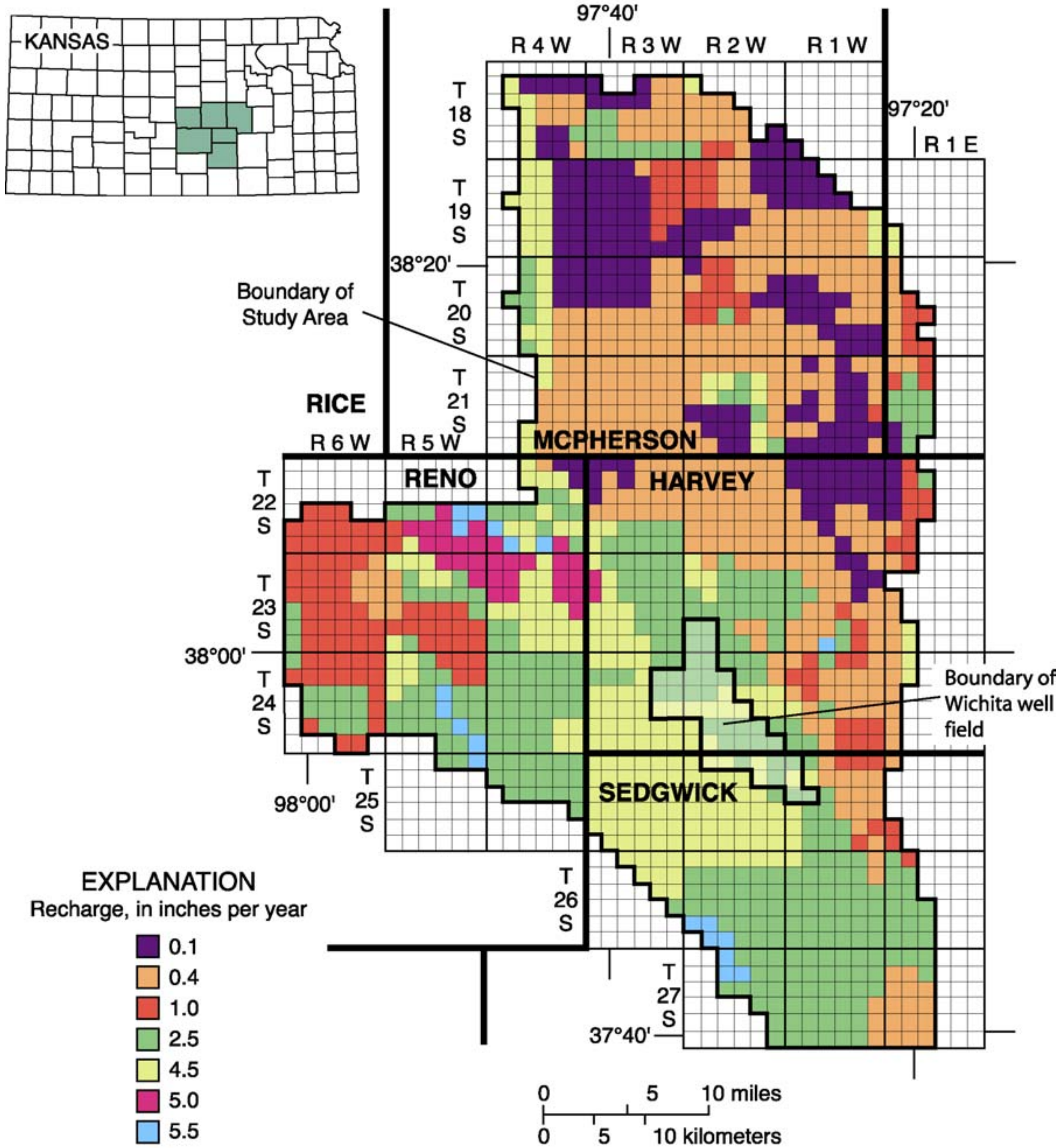

Figure II-12. Simulated predevelopment (1940) recharge distribution in the Equus Beds aquifer, Kansas (adapted from Spinazola et al., 1985).

d) Myers et al. (1996) study

Myers et al. (1996) also employed the 3-dimensional USGS MODFLOW model to study interactions between the Arkansas River and the Equus Beds aquifer in parts of Reno, Harvey, and Sedgwick counties, an area of approximately 690 mi^2, that includes the cites of Hutchinson, Newton, and Wichita. The model area was divided into three layers, and was calibrated for both steady-state (pre-1940) and transient conditions (1940-1989). A variable-spacing model grid was laid out with rows parallel to the Arkansas River, with a smaller grid spacing (1,000 ft × 5,000 ft) near the river. The same recharge distribution developed by Spinazola et al. (1985) for 1940 was assumed valid for the steady-state pre-1940 model (representing 1934-39 conditions). The mean absolute difference between measured hydraulic heads for 235 individual wells and their corresponding middle-layer simulated hydraulic heads was 3.20 ft.

The transient model used the same five stress period as the ones used by Spinazola et al. (1985) plus a sixth stress period (1980-89), characterized by marked fluctuations in the volume of agricultural pumpage. Recharge was based on the mean precipitation at climatic stations at Hutchinson, Mount Hope, and Wichita for each stress period, and was estimated as follows (Myers et al., 1996): 1) the recharge specified for each steady-state model cell was divided by the mean annual precipitation for the pre-1940 period represented in the steady-state model; 2) the resulting quotient for each model cell was then multiplied by the study-area mean annual precipitation for each stress period in the transient model. The mean absolute difference between hydraulic heads for 232 individual wells and their corresponding middle-layer model cell at the end of 1989 was 4.67 ft. Streamflow that was exceeded 70% of the time at each gaging station, assumed to represent baseflow, was compared to model-simulated flow in the stream reach where the gaging station was located. Because streamflows specified for the starting stream reach in the model were held constant for each stress period, the model did not simulate the annual seasonal or short-term variation of measured streamflow (Myers et al., 1996).

The simulated steady-state and transient groundwater budgets for the periods 1964-1970, 1971-79, and 1980-89 are shown in Table II-25.

Table II-24[1]. Simulated water budget for the Equus Beds aquifer[2] (values in inches/yr).

Budget item	1940 initial-condition simulation		1964-1970 transient simulation		1971-79 transient simulation	
	Inflow	Outflow	Inflow	Outflow	Inflow	Outflow
Lateral boundary flow	0.39	0.24	0.41	0.23	0.41	0.22
Recharge	1.70	---	1.69	---	1.91	---
Streamflow loss	0.02	---	0.14	---	0.23	---
Streamflow gain	---	1.38	---	0.74	---	0.61
GW evapotransp.	---	0.41	---	0.27	---	0.25
Well pumpage	---	0.08	---	1.25	---	1.78
Loss of groundwater storage	0	0	0.26	0.009	0.35	0.027
Totals	2.11	2.11	2.50	2.50	2.90	2.89

1. Adapted from Spinazola et al. (1985)
2. Simulated area = 1,406 mi^2 = 899,840 acres

In general, from 1940 to 1989 there were appreciable increases of boundary inflow, streamflow loss, and well pumpage; and decreases in boundary outflow, streamflow gain, and groundwater evapotranspiration. In response to the declining groundwater levels, streamflow gains decreased in the Arkansas River within the model area (from 21 cfs in 1940 to a simulated baseflow loss of about 52 cfs by the end of 1989) and in the Little Arkansas River (from 67 cfs in 1940 within the model area to about 27 cfs by the end of 1989). During 1940-1989, the quantity of chloride discharged from the Arkansas River to the Equus Beds aquifer increased in direct proportion to the volume of water loss from the river (Myers et al., 1996). On the basis of simulated streamflow and assuming that the chloride concentration in river water that moves into the aquifer is 630 mg/L (which is the median concentration in Arkansas River water collected during 1988-1991), the chloride-load discharge from the river to the aquifer was estimated to be about 21 ton/day in 1940 and about 100 ton/day by 1989.

C15. Other county-scale recharge studies

a) Grant and Stanton counties, southwest Kansas

Fader et al. (1964) estimated recharge from precipitation from a 1939-1942 water level contour map of Grant and Stanton counties, assuming negligible well pumpage at that time period. By carefully selecting a study area of 160 square miles between the towns of Johnson and Ulysses, where it could be assumed that leakage inflow from the underlying sandstone aquifers was negligible, they estimated the recharge rate to be about 0.3 inch/yr, which is about 2% of the annual precipitation. This recharge rate was applied to the rest of the Grant and Stanton unconsolidated aquifers in their study. Fader et al. (1964) also estimated the recharge of the unconsolidated aquifers in the area from upward leakage from the underlying sandstone aquifers to be of approximately the same magnitude as the precipitation recharge.

b) Seward County

Byrne and McLaughlin (1948) estimated recharge from precipitation in Seward County based on 1941-44 well hydrograph data from three upland wells that showed an average rise of 0.22 ft/yr and assuming an aquifer specific yield of 0.15 to come up with a recharge estimate of about 0.4 inch/yr.

Table II-25[1]. Simulated steady-state and transient groundwater budgets for all three model layers of the Equus Beds aquifer[2] (values in inches/yr).

Budget item	pre-1940 steady-state simulation		1964-1970 transient simulation		1971-79 transient simulation		1980-89 transient simulation	
	Inflow	Outflow	Inflow	Outflow	Inflow	Outflow	Inflow	Outflow
Boundary flow	0.58	0.73	0.69	0.70	0.69	0.67	0.75	0.64
Recharge	2.58	---	2.32	---	2.59	---	2.53	---
Streamflow loss	0.26	---	0.73	---	0.90	---	1.24	---
Streamflow gain	---	2.01	---	1.18	---	1.02	---	0.68
GW evapotranspiration	---	0.62	---	0.38	---	0.34	---	0.24
Well pumpage	---	0.06	---	1.59	---	2.20	---	3.15
Change in storage	0	0	0.11	0.005	0.06	0.006	0.19	0.001
Totals	3.42	3.42	3.85	3.85	4.24	4.24	4.71	4.71

1. Adapted from Myers et al. (1996)
2. Simulated active-node model area is approximately 692 mi^2 = 442,950 acres

c) Meade County

Frye (1942) made an inventory of water discharged from the artesian water-bearing beds of Meade County by upward leakage through the confining beds, along faults, through springs and flowing and nonflowing wells, and determined that the total annual discharge of artesian water was about 10,000 acre-feet. He concluded that nearly all the recharge to the artesian water-bearing beds occurs to the west and north of the county in southwestern Gray County, parts of Haskell County, northeastern Seward County, and the southernmost part of Finney County, an area of the order of 685 square miles. Assuming equilibrium conditions, where the recharge is assumed to be equal to the discharge, Frye (1942) estimated an average recharge of about 0.27 inch/yr, which represents about 1.5% of average annual precipitation in the area (considered as 18 inches/yr). According to Frye (1942), in the area of sand dunes in southern Finney County and northern Haskell County that percentage is probably much greater, and in certain other parts of the area it is probably much less, and locally there may be none at all.

d) McPherson moratorium area, Equus Beds aquifer

McElwee et al. (1979) conducted a water-budget analysis of the McPherson moratorium area (56 mi^2) of the Equus Beds aquifer based on January 1978 water-table levels. Recharge was estimated as 2 inches/yr (Table II-26).

Table II-26[1]. 1978 water-budget analysis of the McPherson moratorium area[2], Equus Beds aquifer.

Budget item	Inflow	Outflow
Boundary flow	1.40	0.47
Recharge	2.00	---
Well pumpage	---	5.02
Loss of groundwater storage	2.09	---
Total	5.49	5.49

[1.] Adapted from McElwee et al. (1979)
[2.] Area considered = 56 mi^2 = 35,840 acres

D. Field-based experimental recharge studies

Field-experimental groundwater recharge studies in Kansas are highlighted in this section.

D1. Movement of moisture in the unsaturated zone in a dune area, southwestern Kansas (Prill, 1968)

Prill (1968, 1977) conducted a study to investigate the requirements necessary for deep percolation under three types of vegetative conditions: 1) a barren area, 2) a sagebrush-grass community, and 3) a grass community, over a period from the fall of 1964 through 1966. This period depicts maximum moisture changes because 1965 was a year when precipitation was one of the highest on record (29.07 inches in Garden City) and was preceded and succeeded by years when precipitation was below normal (12.23 inches in 1964 and 12.04 inches in 1966).

The study site is located in the extensive area of dune sand immediately south of the Arkansas River near Garden City in southwestern Kansas. At each site, neutron probe access holes containing 2-inch aluminum tubing for moisture-content logging were drilled through the dune sand into the underlying alluvial deposits. The water table in this area was about 30 feet below the top of the alluvial deposits.

Even though the period of study included a year when precipitation was nearly the highest on record, built-up moisture under a sagebrush-grass community penetrated to a depth of only 14 feet, whereas the zone of evapotranspiration extended to at least 17 feet (Prill, 1968). Under a grass community where the zone of evapotranspiration extended to about 11 feet only a small amount of moisture (2 inches) moved as deep percolation. Under a barren area, where most of the loss by evaporation occurred in the upper 1 foot, large quantities of moisture moved as deep percolation.

Prill (1968) estimated that the average annual recharge in the vegetated area is about 0.5 inch. Prill (1968) pointed out that the periods when conditions are favorable for recharge are few and usually occur when precipitation is considerably above average during the nongrowing season and the early part of the succeeding growing season. The high rate of evapotranspiration all but eliminates the possibility of recharge during the summer months.

D2. Pilot recharge assessment at two sites in south-central Kansas (Great Bend Prairie and Equus Beds aquifers) (Sophocleous and Perry, 1985)

Sophocleous and Perry (1984, 1985, 1987) experimentally assessed groundwater recharge at two instrumented sites in south-central Kansas (one site near Zenith in Reno County (GMD5), and one near Burrton in Harvey County (GMD2) over an approximately 19-month period during 1982 and 1983. Although the two instrumented sites were located in sand-dune environments in areas characterized by shallow water table and subhumid continental climate, a significant difference was observed in the estimated effective recharge. The estimates ranged from less than 0.1 inch at the Zenith site to approximately 6.1 inches at the Burrton site during the major recharge period from February to June 1983. The main reasons for this large difference in recharge estimates were the greater thickness of the unsaturated zone and the lower moisture content in that zone resulting from lower precipitation and higher potential evapotranspiration for the Zenith site. Effective recharge took place only during late winter and spring. No summer or fall recharge was observed at either site during the observation period of this study.

D3. Recharge assessment for the Great Bend Prairie aquifer in GMD5 (Sophocleous, 1992, 2000c)

Recharge-related variables were monitored in the field on a year-round basis at 10 sites distributed throughout the GMD5 area (fig. II-13; Sophocleous, 1991, 1992, 1993a, 1993b, Sophocleous and Stern, 1993). The methodology used in quantifying recharge for the region consisted of combining the hydrologic or soil-water balance on a storm-by-storm year-round basis with the resulting water table rises. Each recharge assessment site was equipped with a weighing and recording rain gage, a neutron-probe access tube for measuring the soil-profile water content, a water table well with a water-level recorder, and two deeper piezometers. Two of the sites were also equipped with weather stations that recorded solar radiation, air temperature, relative humidity, barometric pressure, and wind speed. Using the data collected at these sites and detailed weather data from the Sandyland Experiment Station, just south of St. John (fig. II-13), the soil-water balance for each recharge-producing storm period was calculated. By associating the result with the consequent water table rise, which was tied to specific precipitation events, reliable effective recharge values for different storm periods were obtained (Sophocleous, 1991).

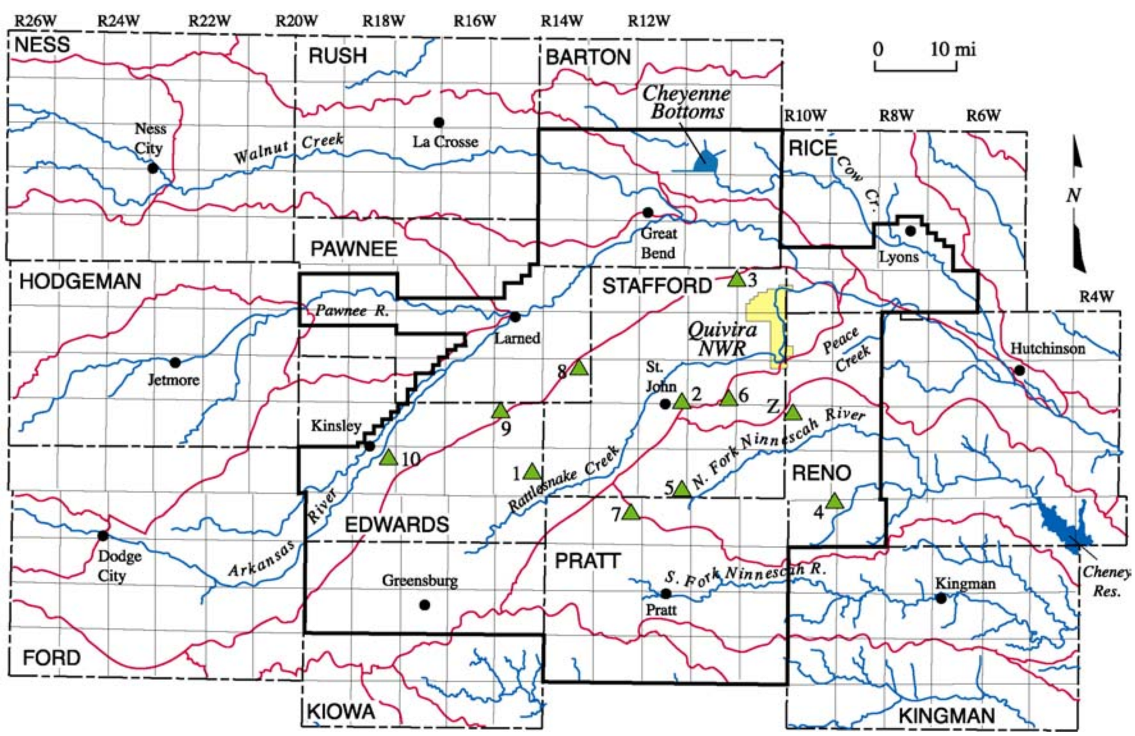

Figure II-13. Groundwater Management District No. 5 boundary (heavy black line) with groundwater recharge assessment sites (triangles). River basins are outlined with red lines.

Table II-27 gives the average values of measured precipitation, depth to water table, and estimated annual recharge for all the monitored sites in GMD5. For the original recharge sites 1 through 5 data have been collected from 1985 to 1992, whereas for sites 6 through 10 data were collected during the 1988-1992 period. The unusually high recharge estimates for site 4 in Reno County, which received the highest precipitation among all sites, were due to the site being located on the streambank of a tributary to Wolf Creek where the depth to the water table is very shallow, approximately 2-4 ft. Sites 8, 9, and 10 received no detectable recharge during the period 1988-1992.

During the flood year of 1993, average precipitation in Stafford County was approximately 36.6 inches (which is approximately double the long-term average) and estimated average recharge (based on sites 1, 2, 3, and 5) was 5.9 inches. This amount of estimated recharge caused an average county-wide water-table rise of 5.4 ft (Sophocleous et al., 1996).

Sophocleous (1992, 1993a, 2000c) using a combination of statistical (forward stepwise regression) analysis and GIS overlay analysis, identified the portion of the GMD5 area that each recharge site or cluster of sites represents (fig. I-11), and derived an area-weighted average recharge for the GMD5 of 1.4 inches/yr based on the 1985-1990 recharge site data, as shown in Table II-28.

Additional details on the regionalization methodology for the Great Bend Prairie recharge are presented in Part I, Section 8.2. Additional information on GMD5 recharge can be found in Sophocleous (1992, 1993a, 1993b, 2000c).

Table II-27[1]. 1985-1992 site-specific groundwater recharge estimates for GMD5.

Site number	Location	Total precipitation (inches/yr)	Minimum and maximum depth to water table (ft)	Estimated groundwater recharge (inches/yr)
1	Edwards County sec. 13, T25S, R16W 8-yr avg. (1985-1992) and (standard error)	23.36 (2.0)	17.2-20.6	1.4 (0.6)
2	Stafford County sec. 36, T23S, R13W 8-yr avg. (1985-1992) and (standard error)	22.75 (1.9)	24.6-27.4	1.6 (0.4)
3	Stafford-Barton counties sec. 7, T21S, R11W 8-yr avg. (1985-1992) and (standard error)	23.22 (1.7)	18.5-23.1	0.9 (0.3)
4	Reno County sec. 1, T25S, R9W 8-yr avg. (1985-1992) and (standard error)	27.91 (2.2)	2.1-5.1	6.5 (1.0)
5	Stafford-Pratt counties sec. 36, T25S, R13W 8-yr avg. (1985-1992) and (standard error)	25.47 (2.0)	10.6-15	2.8 (0.6)
6	Stafford County sec. 36, T23S, R12W 5-yr avg. (1988-1992) and (standard error)	20.90 (2.6)	11.4-23.4	0.9 (0.3)
7	Pratt County sec. 11, T26S, R14W 5-yr avg. (1988-1992) and (standard error)	22.95 (2.5)	21.9-28.3	2.4 (0.8)
8	Pawnee County sec. 14, T23S, R15W 5-yr avg. (1988-1992) and (standard error)	20.93 (2.1)	26.9-28.0	0.0
9	Edwards County sec. 5, T24S, R16W 5-yr avg. (1988-1992) and (standard error)	19.73 (2.3)	32.5-34.0	0.0
10	Edwards County sec. 1, T25S, R19W 4-yr avg. (1988-1991) and (standard error)	18.64 (2.3)	49.6-51.2	0.0
	Arithmetic avg., sites 1-10 and (standard error)	23.03 (0.71)		1.9 (0.30)

[1]. Adapted from Sophocleous (1993a)

Table II-28[1]. Recharge zonation of GMD5 based on GIS overlay analysis.

Recharge zone	Approximate area within GMD5 (mi^2)	Percentage of GMD5 area	Recharge sites within zone	1985-1990[a] average annual recharge (inches) from within zone
1	1,313	33.3	8, 9, 10	0[b], 0.5[c]
2	830	21.1	3, 6	1 (0.3)[d]
3	1,398	35.4	1, 2, 7	2 (0.4)[d]
4	401	10.2	5	3 (0.7)[d]
5	2	0.1	4	7 (1.2)[d]
Area-weighted average recharge: 1.4 inches/yr				

[1.]Adapted from Sophocleous (1992)
[a]1988-1990 for sites 6-10
[b]1988-1992 average based on recharge sites
[c]Twenty-year average based on Pawnee River valley study (Sophocleous, 1981)
[d]Standard error of zonal recharge (inches/yr)

D4. Deep vadose zone study of groundwater recharge in the High Plains aquifer of southwest Kansas (Sophocleous et al., 2002)

Recent improvements in technology make it possible to study the deep vadose zone, well below the rooting depths of plants. Such technology has been employed to monitor, on a continuous basis, the deep pore-water fluxes en route to the High Plains aquifer (HPA). This proof-of-concept pilot study provided, for the first time, information on the quantity and pattern of water fluxes in the deep vadose zone that impact the management of both the quantity and quality of the HPA in Kansas. The Sophocleous et al. (2002) preliminary investigation evaluated the use of heat-dissipation sensors and advanced tensiometers for measuring pore-water pressures in deep boreholes at two irrigated land-use sites (Sites 1 and 2 located in southern Finney County, Fig. II-7) and one natural grassland site (Site 3 located in the Cimarron National Grassland in Morton County, Fig. II-7). Continuous time series data obtained from the heat-dissipation sensors (installed at 116 ft at the irrigated sites and at 137 ft at the grassland site) revealed constant pore-water pressures with time over the May 2000–September 2001 monitoring period.

The observed time-series patterns of pore-water pressure head in the deep vadose zone of the High Plains imply homogenization at depth of near-surface, temporally varying water fluxes, resulting in much lower-intensity but nearly constant (steady) recharging fluxes to the High Plains aquifer.

The measured average hydraulic-head gradients for all sites, together with the estimated recharge fluxes based on Darcy's law, are shown in Table II-29. It is interesting to note that the hydraulic head gradients between the deepest heat-dissipation sensor and the groundwater level at each site were approximately 0.75 for Sites 1 and 2, whereas for Site 3 it was nearly zero.

Darcian methodology was used to obtain estimates of recharge at all the three investigation sites. Estimated recharge rates for the irrigated land-use sites (Sites 1 and 2) were appreciably higher (0.12 inch/yr and 0.02−0.04 inch/yr, respectively—Table II-28) than the estimated recharge rate for the natural grassland site (Site 3; less than 0.01 inch/yr—Table II-29). In all cases the estimated annual recharge values were less than one percent of annual precipitation. Although the large uncertainty associated with these estimates and the small number of study sites precludes using these flux estimates alone to draw firm conclusions regarding present-day recharge in the region, the irrigated and natural grassland sites selected are representative of irrigated and grassland areas overlying the High Plains aquifer in southwestern Kansas.

The U.S. Geological Survey conducted several chemical and tracer analyses at the aforementioned sites. Chloride profile analysis (see Appendix B of Part I on tracers for recharge estimation) of the grassland site indicated that the recharge flux below the root zone ranged from less than 0.1 to 0.39 inch/yr, and that estimate is considered to be a long-term (on the order of hundreds of years) estimate of recharge at the grassland site (P. B. McMahon, USGS, personal communication, September 2002). Both irrigated sites had bomb tritium (see Appendix B of Part I) and pesticides detected in both the unsaturated and saturated zones (P. B. McMahon, USGS, personal communication, 2001), which implies a much higher recharge rate at those sites than at the grassland site.

The Darcian-based water-flux estimation aspect of this High Plains aquifer program was a pilot study of recharge assessment. That study showed that deep-vadose-zone hydrology, a mostly unexplored frontier due to technological obstacles, can be monitored and analyzed. More instrumented sites, similar to the ones employed here (and taking advantage of the experience gained during this study), in combination with additional methodologies, are needed to assess the deep vadose zone water (and chemical) fluxes reaching the High Plains aquifer.

Table II-29[1]. Measured hydraulic-head gradients and estimated water fluxes in deep, unsaturated High Plains sediments based on heat-dissipation sensors. Site locations are shown in Fig. II-7 (adapted from Sophocleous et al., 2002).

Site ID (1)	County and legal location (2)	Land use (3)	Heat dissipation-sensor depth locations (ft) (4)	Average depth to water level (ft) (5)	Hydraulic head gradient[2] (dimensionless) (6)	Direction of flow (6)	Estimated Darcian water fluxes[3] (inches/yr) (7)
1	Finney 26S-32W-21adc	irrigated cropland	66; 116	148.5	1.01	downwards	0.12
2	Finney 26S-31W-31ccd	irrigated cropland	31; 116	149.5	0.86	downwards	0.02–0.04
3	Morton 34S-41W-25ada	native grassland	77; 137	164.0	0.47	downwards	0.004–0.01

1. Adapted from Sophocleous et al. (2002)

2. Gradient taken between the two heat dissipation (HD)-sensor depth levels indicated in column 4.

3. Assuming arithmetic average of the two unsaturated hydraulic conductivities at the HD sensor depth levels shown in column 4. The hydraulic conductivities are based on laboratory-measured saturated hydraulic conductivity values and water retention curves, all determined from collected cores by the U.S. Geological Survey laboratory in Sacramento, CA. The RETC fitting program (van Genuchten et al., 1998) was used to quantify the hydraulic properties of the collected cores. The relative water-flux error was estimated by Sophocleous et al. (2002) to be at least 102%.

PART III. CONCEPTUALIZATION OF THE KANSAS HIGH PLAINS AQUIFER AND ITS RECHARGE CHARACTERISTICS, INCLUDING SUGGESTIONS FOR APPROPRIATE RECHARGE-QUANTIFICATION TECHNIQUES

Conceptualizing High Plains aquifer recharge

Understanding the sources of recharge and the spatial and temporal variability in recharge is basic to developing a conceptual model of recharge. Potential sources of recharge of the Kansas High Plains aquifer include precipitation, surface water (rivers, streams, ponds, playas, lakes, floods), return flow from irrigation, lateral groundwater flow into the aquifer from outside areas (for example, lateral inflows from the Colorado High Plains aquifer to Kansas), and cross-formational flow from adjacent aquifers (for example, the Dakota aquifer, as shown in section B4 of Part II).

Western Kansas has a semi-arid continental climate with moderate precipitation, low humidity, and high evaporation. Winters are relatively moderate and summers are often hot. The mean annual precipitation ranges from less than 17 inches to the west near the Colorado border to more than 30 inches in the easternmost extent of the High Plains aquifer (Equus Beds aquifer region). About three-fourths of the precipitation falls during the growing season (April through September). Average free water surface evaporation ranges from 52 inches in the easternmost extent of the High Plains aquifer to more than 68 inches in southwestern Kansas (Sophocleous, 1998b). The Kansas High Plains is characterized by flat to gently rolling terrain, which, in combination with the semiarid climate of western Kansas results in minimal surface runoff. The mean annual surface runoff in western Kansas ranges from less than 0.1 inch to about 1.1 inches (Sophocleous, 1998b).

Recharge generally increases with increased precipitation. The seasonal distribution in precipitation may be more important than the average annual precipitation because winter precipitation is more effective in recharging groundwater than summer precipitation. As Scanlon, Dutton, and Sophocleous (2002) also pointed out, 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 data and time steps no larger than daily because precipitation at such smaller time scales can greatly exceed evapotranspiration and result in effective recharge. In addition to climatic factors, recharge is affected by soil texture/structure and hydraulic conductivity (coarse grained soils generally result in higher recharge rates than fine-grained soils), land cover (croplands generally result in higher recharge than grasslands and shrublands), and land use (irrigated lands result in higher recharge than drylands) (Sophocleous and McAllister, 1987; Sophocleous, 1992).

The High Plains aquifer of Kansas consists mainly of a heterogeneous sequence of unconsolidated deposits of sand, gravel, silt and clay of principally alluvial origin deposited during the Tertiary (the only stratigraphic unit of Tertiary age identified in western Kansas is the Ogallala Formation of Pliocene age—Gutentag, 1963) and Quaternary periods and unconformally overlies Permian, Jurassic-Triassic, and Cretaceous formations. The type and degree of cementation within the aquifer varies. Lime-cemented and silica-cemented beds of silty and sandy gravel (mortar beds) and sandy silt (caliche) occur throughout the aquifer and at the outcrop form ledges or caprock.

The Ogallala Formation, which makes up the main part of the High Plains aquifer in western Kansas, was deposited primarily by easterly flowing aggrading streams carrying debris from the Rocky Mountains. A vast plain of braided streams and coalesced alluvial fans was formed. Ogallala sediments filled paleovalleys eroded into the pre-Ogallala surface. However, more recent studies in the southern High Plains (Gustavson, 1996) indicate that the Ogallala in Texas and eastern New Mexico consist of alluvial sediments that partly fill paleovalleys and widespread thick eolian sediments that cap both paleo-uplands and most fluvial sections. These strata, apparently deposited under mostly semiarid to subhumid climatic conditions, do not constitute coalescing or overtopping wet alluvial fans (Gustavson, 1996). Deposition of the High Plains aquifer in some areas was contemporaneous with dissolution of underlying Permian salt beds, resulting in additional ground-surface subsidence and increased accumulation of High Plains sediment. The lower part of the formation in paleovalley-fill alluvium tends to have more coarse-grained sediment and thus greater hydraulic conductivity than the upper part, although Breyer (1975) concluded that the distribution of sediment types within the Ogallala Formation is largely random. The major identifying feature of braided streams is the coarsening-upward as well as fining-upward sequences of alluvial deposits (Gutentag et al., 1984). This process of coarsening and fining of the alluvial deposits gives a random distribution of sediments in the High Plains aquifer, suggesting that the aquifer may behave as homogeneous on a regional scale. Test holes drilled within a 160-acre tract often show a predominance of clays and silts at one site and of sand and gravel nearby (Stullken et al., 1985).

The Quaternary deposits are of Pleistocene and Holocene age. Considerable thicknesses of both alluvial and eolian deposits occur at the surface of the High Plains in Kansas. Quaternary alluvium (stream-laid clay, silt, sand, and gravel) is the predominant type of Cenozoic deposit in most of western Kansas. Pleistocene loess mantles much of the upland areas in western Kansas, and Pleistocene and Holocene dune sands cover a significant portions of the High Plains area. Because of the similarity in composition, the contact between the Ogallala Formation and the overlying Pleistocene deposits is difficult to determine from drillers' logs, gamma-ray logs, and some test-hole logs.

Figure III-1 shows an east-west cross section depicting the stratigraphy from western Stanton County to the Gray County line across township 28. The lithology of the wells and test holes has been simplified to show the aquifers (sand and gravel), aquitards (silt), and quasi-aquicludes (clay and caliche). Mixtures of materials were designated as to their major constituent for clarity of illustration. The slopes on the eroded bedrock and Ogallala surfaces in the eastern part of the area are moderate as opposed to steep slopes in the western part. The sediments are thickest where the slopes are moderate (Gutentag, 1963).

The configuration of the bedrock surface is a composite of subaerial erosional surfaces of several ages (Merriam and Frye, 1954). This surface also has been affected by structural movement and by subsidence associated with the solution of evaporites from Permian rocks (Gutentag et al., 1981). The pre-Ogallala surface south of the Arkansas River has also been modified by post-Ogallala erosion. The irregular bedrock surface in southwest Kansas between the Bear Creek and the Crooked Creek-Fowler faults (Gutentag et al., 1981) generally slopes at about 13.5 feet per mile – a gradient of 0.0026) to the east-southeast from 3,500 feet above sea level near the Colorado State line in southwest Stanton County to about 2,000 feet above sea level near the town of Meade in Meade County, Kansas. The Bear Creek and Crooked Creek-Fowler faults in southwest Kansas are attributed to dissolution of halite and gypsum from the Blaine Formation and Flower-pot Shale of the Lower Permian Nippewalla Group.

The High Plains aquifer ranges in saturated thickness from 0 to more than 550 ft (as of 2000), just south and west of Liberal in Seward and Stevens counties. Generally, the greatest saturated thickness is where the unconsolidated deposits overlie the deepest channels in the bedrock. In some areas the High Plains aquifer is hydraulically connected to the overlying alluvium, such as along the Arkansas and Cimarron River valleys. The Lower Cretaceous Dakota aquifer is also hydraulically connected to the High Plains aquifer in some locations−that is, the Ogallala Formation is not separated from the Dakota Formation by shale, clay, or other low-permeability units. General flow within the aquifer is eastward. Streams affect local flow patterns as they become discharge or recharge points for the aquifer. Based on average values of hydraulic gradient and aquifer characteristics, the velocity of water moving through the aquifer is about 1 ft/day (Gutentag et al., 1984), which is typical of sand and gravel aquifers.

Many areas of the aquifer have been irrigated since the 1940s. Average annual withdrawal for irrigation was greatest during the 1980s, but during the 1990s the total rate of irrigation withdrawal decreased. Irrigation inefficiency probably was high during the 1940s and 1950s but decreased during the past few decades. Luckey and Becker (1999) estimated that irrigation inefficiency decreased from 24% during the 1940s and 1950s to less than 4% by the 1980's.

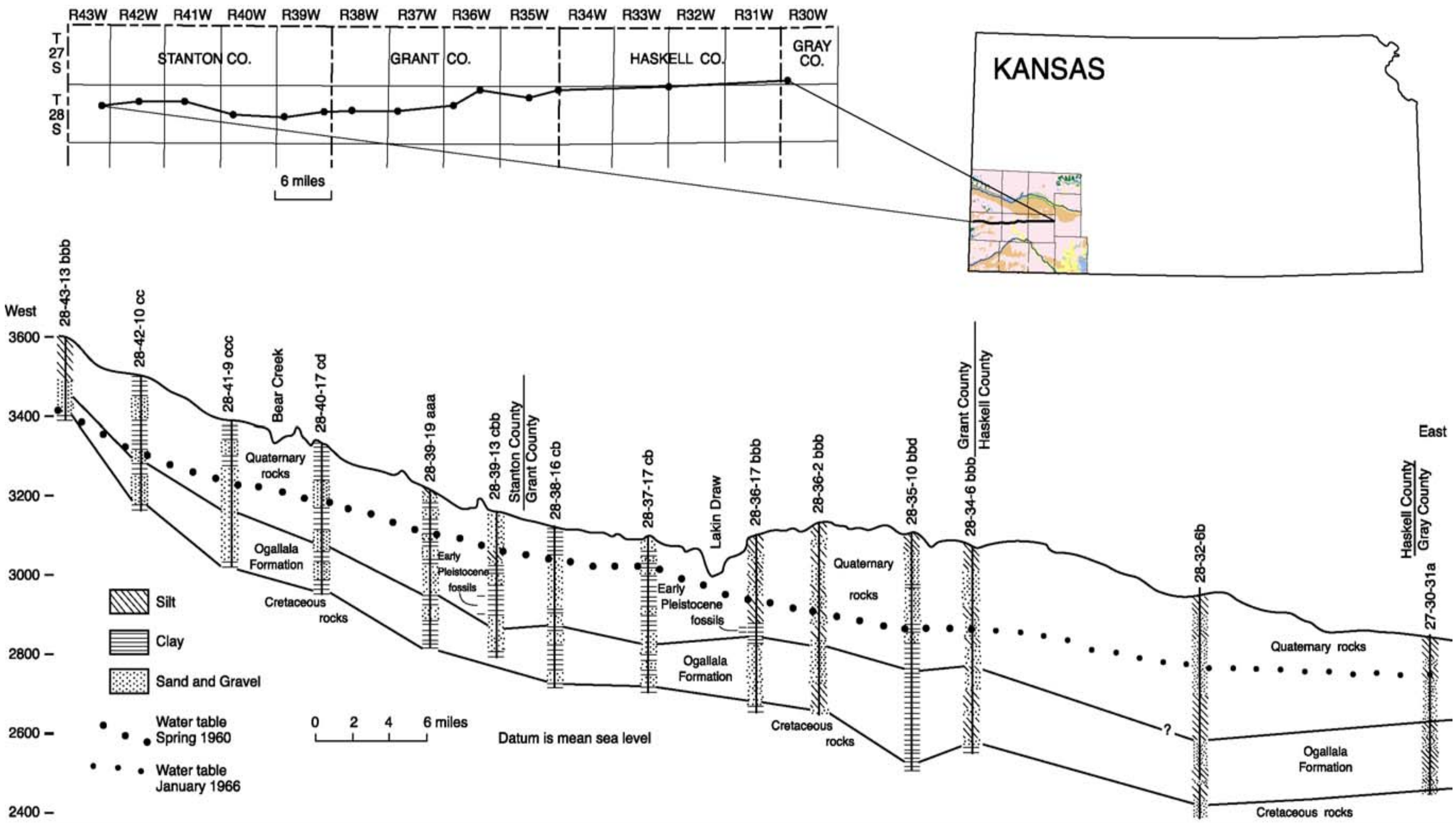

Figure III-1. East-west cross section through the Ogallala Formation and Quaternary deposits in Stanton, Grant, and Haskell counties, southwestern Kansas (cross section in Stanton and Grant counties from Gutentag, 1963).

Estimated recharge is much less than the water quantity extracted from the High Plains aquifer in western Kansas, resulting in significant long-term water table declines as well as streamflow declines (Kromm and White, 1992; Sophocleous, 2000a,b; Schloss et al., 2000). Prior to heavy irrigation development, the Arkansas River received baseflow from the High Plains aquifer and the connected alluvial deposits. Under these conditions, groundwater naturally flowed towards the river. At the present time, however, the water table has declined below the streambed so that the flowing river may be a recharge source for the underlying sediments.

Irrigation return flow may contribute significant amounts of recharge to the High Plains aquifer. The amount of return flow depends on irrigation rate, irrigation inefficiency, soil type, depth to water, and rate of downward movement (or velocity) of water from the root zone to the water table. Return flow may reach the water table much later than the year or the decade in which irrigation was applied, and the delay or lag time may increase as depth to water increases. The velocity of water moving downward through the unsaturated zone is an important, although poorly constrained, variable (Scanlon, Dutton, and Sophocleous, 2002). 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 reached 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 (Scanlon, Dutton, and Sophocleous, 2002).

Appropriate techniques for quantifying recharge in the High Plains aquifer

As we have seen in Part II, the main techniques that have been used for estimating recharge in the High Plains aquifer in Kansas are Darcy's Law, annual water table fluctuation analyses in combination with estimates of aquifer specific yield, groundwater modeling, soil-water budget modeling, and base-flow analyses. Although a number of techniques for quantifying recharge in the High Plains aquifer of Kansas have been used, it is apparent from the review of existing recharge estimates that additional recharge studies are required to better quantify recharge. As we also mentioned in Part I, section 9, one of the difficulties of determining appropriate techniques for quantifying recharge in the High Plains is that many techniques are restricted to measuring recharge rates within a certain range, which may not be known a priori before the recharge study is undertaken. Therefore, only different approaches that are likely to provide the most quantitative estimates of recharge can be suggested. Results provided by initial studies should be used as platforms for additional data to optimize the techniques and refine the recharge estimates. An iterative approach will be required to accurately quantify recharge rates, and a variety of approaches should be applied because of uncertainties in recharge estimates (see also section 9 of Part I). Results from the various techniques can be compared to determine uncertainties in recharge estimates.

Following similar recommendations on appropriate techniques for quantifying recharge for the major Texas aquifers (Scanlon, Dutton, and Sophocleous, 2002), we offer the following suggestions, especially in view of the general lack of tracer-based methodologies for recharge quantification in Kansas. Because of the generally thick unsaturated zone of the High Plains aquifer in western Kansas, many of the techniques for estimating recharge to the Ogallala aquifer may 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 streams using channel-water budgets (differential streamflow measurements) or other techniques such as heat tracers and seepage meters.

Appropriate unsaturated-zone techniques may include the use of chloride concentrations in soil water. Low chloride concentration beneath playas in Texas suggest high recharge rates, whereas high chloride concentrations in interplaya settings suggest low recharge rates. Such studies have not yet been reported in Kansas. 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. However, it would be difficult to use chloride to quantify recharge rates in irrigated regions because of uncertainties in the chloride input to the system (Scanlon, Dutton, and Sophocleous, 2002). 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. However, such unsaturated-zone techniques have rarely been used in Kansas, especially beyond the plot-size scale.

Saturated-zone methods provide a more spatially averaged recharge rate than the point estimates provided by unsaturated-zone techniques. Water table fluctuations may be used in areas of shallow water table (such as the Equus Beds and Great Bend Prairie portion of the High Plains aquifer) to quantify recharge rates. Wood and Sanford (1995) used chloride concentrations in groundwater to estimate recharge in the north half of the southern Ogallala in Texas. Anthropogenic substances such as pesticides and CFCs may provide qualitative indicators of high recharge rates. Tracers such as ^{3}H, $^{3}H/^{3}He$ can be used to quantify recharge in sandy areas 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.

PART IV[1]. COUNTY-BY-COUNTY AND DISTRICT-WIDE TABULATED RECHARGE VALUES AND RELATED STATISTICS (IN EXCEL SPREADSHEETS) FOR THE KANSAS GROUNDWATER MANAGEMENT DISTRICT REGIONS BASED ON KGS BULLETINS AND OTHER PUBLICATIONS

The following set of five EXCEL spreadsheets [by Groundwater Management District (GMD) region] contain county by county recharge estimates and related information based predominantly on three major sources of information: the Kansas Geological Survey (KGS) Bulletins, the 1967 Kansas Water Resources Board (KWRB) Irrigation in Kansas 701-project report, and the USGS 1991 potential natural recharge in Kansas report, although additional sources of information were also used. In addition, summary recharge estimates for each Kansas region (approximately corresponding to each GMD) and related statistics are also given. Based on the aforementioned three major sources of information, the following recharge estimates for the entire Ogallala aquifer of western Kansas (approximately comprised of the regions occupied by GMDs 1, 3, and 4) and their related statistics are given in Table IV-1 below. The average of those three mean recharge estimates for the Ogallala aquifer of western Kansas is 0.37 inch/yr.

[1] This part was written with the assistance of KGS student assistant Mr. Anish Pradhananga, and it is also available as a separate KGS Open-file Report 2003-11 by Sophocleous and Pradhananga (2003).

Table IV-1. Western Kansas Ogallala aquifer recharge estimates and related statistics based on three-agency estimates: Kansas Geological Survey (KGS), Kansas Water Resources Board (KWRB), and United States Geological Survey (USGS).

No.	County	Area (mi^2)	Recharge[1] (in/yr)	Recharge[2] (in/yr)	Recharge[3] (in/yr)
1	Wallace	910	0.2500	0.1030	0.2576
2	Greeley	788	0.1000	0.1428	0.2641
3	Wichita	717	0.1000	0.2615	0.4603
4	Scott	724	0.5000	0.2331	0.4299
5	Lane	720	0.2500	0.2240	0.2487
6	Hamilton	992	*	0.1000	0.1800
7	Kearny	861	*	0.2400	0.5600
8	Finney	1302	*	0.2300	0.5600
9	Gray	873	*	0.3200	0.9400
10	Ford	1082	0.5000	0.6000	0.9400
11	Stanton	685	0.3000	0.3200	0.3900
12	Grant	571	0.3000	0.3000	0.7300
13	Haskell	580	*	0.3100	0.9800
14	Morton	720	*	0.3100	0.4200
15	Stevens	731	*	0.3100	0.7500
16	Seward	643	0.4000	0.2900	0.9900
17	Meade	979	0.2700	0.2800	0.9600
18	Cheyenne	1027	*	0.2227	0.4236
19	Rawlins	1080	*	0.1823	0.7257
20	Decatur	900	0.5000	0.3646	0.7500
21	Norton	880	0.3700	0.3281	0.8757
22	Sherman	1055	0.1000	0.2399	0.2417
23	Thomas	1070	0.2500	0.3627	0.3820
24	Sheridan	893	0.2500	0.2940	0.7391
25	Graham	891	0.5000	0.1473	0.6839
26	Logan	1073	0.1667	0.0874	0.1380
27	Gove	1070	*	0.1928	0.2033
28	Trego	900	0.2100	0.2083	0.2708
Total		24717			
Average			**0.2954**	**0.2573**	**0.5534**
Standard deviation			0.1396	0.1028	0.2803
Area-weighted avg.			**0.2936**	**0.255**	**0.5412**
90% Conf. interval upper limit			0.3176	0.2704	0.5891
90% Conf. interval lower limit			0.2732	0.2442	0.5176

1 : Ogallala aquifer recharge estimates based on KGS Bulletins (* indicates counties for which recharge had not been quantified).

2 : Ogallala aquifer recharge estimates based on KWRB "Irrigation in Kansas" 1967 report.

3 : Ogallala aquifer potential recharge estimates based on USGS-WRIR 87-4230 (Hansen, 1991).

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WEST CENTRAL KANSAS (GMD 1 Region)

No.	County	Area mi²	Reference Bulletin & publ. yr	Aquifer	Avg. Precip. (in/yr)	Avg. Temp F	Recharge from ac-ft/yr (in/yr) precip.	stream	lateral inflow	irrig.	Total Recharge ac-ft/yr (in/yr)	Method	Reference	Soils	Topog	Avg. Depth to Water Table	Vegetation & Land use	Comments
1	**Wallace**	910	KGS B 18	Alluvium [1d]	17.03	55	[1a]not known but	most of the					[1a] no reference	silty and	High Plains	[1e]12 ft-30 ft at alluvium	crops and	[1e]KGS B 161
			1931				studies in	streams are					source mentioned	silty and	section of	[1e]30 ft-100 ft at Ogallala	grasses	
			KGS B 161				other similar	influent so					Hodson, W.G.,1963	loamy	the Great		farm land	
			1963	Ogallala: [1d]			High Plain areas	contribute							Plains		& pasture	
				(present in approx.			indicate **1/4 in**	appreciable	[1b]estimated to	[1b]about **10%** of		author's estimate	[1b] Hodson,W.G.,1963		physiographic			
				2/3 of the county)			of recharge.	recharge. But	be approx.	applied water.					province			
							More in southern	less in northern	**36500 ac-ft/yr.**	Relatively high in								
							part than in	part of the county	Inflow occurs	alluvial valleys								
							northern part of	due to nearness	along the									
							the county.	of shale to the	western edge		5000 [1c]	-------	[1c] KWRB' 67					
								land surface	of the county		(**0.1 in/yr**)							
				Alluvium			1040 [1d]					soil-water budget	[1d] USGS' 91					
							(**0.02 in/yr**)											
				Ogallala			12500 [1d]											
							(**0.26 in/yr**)											
2	**Greeley**	788	KGS B 108	Ogallala Fm.	16.98	52.5	[2a]**0.1 in/yr**	considerable,		considered low		author's	[2a] Prescott, G.C., et al.,	silty	High Plains	[2e]25 ft-40 ft at alluvium	crops and	[2e] ref. KGS B 108
			1954	(unconsolidated sand				but unknown		due to relatively		estimate	1954		section of the	[2e]20 ft-80 ft at Ogallala	grasses	
			KGS Irr.	and gravel):						impermeable					Great Plains		agriculture	
			Ser.2	as much as 300 ft thick,			[2b]4000 ac ft/yr			loess cover		(author's estimate)	[2b] Slagle and Weakly,		physicgraphic			[2b] For Greeley and Wichita
			1976	of which as much as			(**0.05 in/yr**) during					10% of pcp. in	1976		province.			combined.
				145 ft was saturated.			1971-72					irrigated & 1% of			Flat to gently			
				as of 1971-72								pcp. in nonirrigated			rolling upland			
				Yields 100 to 2000 gpm.								land. (163000 ac			plains.			
				Dakota Fm.								irrigated & 390000						
				(sandstone aquifer):								ac of nonirrigated						
				400-550 ft thick. Similar								land in Greeley and						
				aquifer in adjacent								Wichita) -1971-72						
				counties yield 30-300						[2b]44000 ac-ft/yr		20% of irr. applied						
				gpm.						(**0.17 in/yr**)		(0.22" in irrigated &						
												0.05" in nonirr. land)						
									[2a]estimated to			Darcy's Law						Lateral Inflow,subsurface
									be about 5000									outflow & total recharge are
									ac-ft/yr.		[2b]27000	(irr&nonirr. land)+Lat. in.						for Wichita and Greeley
									(**0.06 in/yr**)		(**0.28 in/yr**)	(0.17+.05)+.06						combined
											[2c]6000		[2c] KWRB' 67					
											(**0.14 in/yr**)							
				Ogallala			[2d]11100					soil-water budget	[2d] USGS' 91					
							(**0.26 in/yr**)											
3	**Wichita**	717	KGS B 108	Ogallala is the	18.63	52.5	[3a]**0.1 in/yr**	44 ac-ft of loss		negligibly low		author's	[3a] Prescott, G.C., et al.,	silty	High Plains	[3g]15 ft-40 ft at alluvium	crops and	[3g] ref. KGS B 108
			1954	principal water bearing				per mile as per				estimate	1954		section of the	[3g]10 ft-80 ft at Ogallala	grasses	
			KGS Irr.Ser.2	fm. Small amount is				the study in							Great Plains		agriculture	
			1976	also obtained from				Whitewoman	[3a]estimated to			Darcy's Law			physicgraphic			
				alluvium and Niobrara				Creek in July	be about 5000						province.			
								1972	ac-ft/yr.		[3b]27000	(irr&nonirr. land)+Lat.	[3b] Slagle and Weakly,		Flat to gently			[3b] For Greeley and Wichita
				Ogallala Fm.					(0.06 in/yr)			(0.17+.05)+.06	1976		rolling upland			combined.
				(unconsolidated					occurs at west-						plains.			
				sand and gravel)			[3b]4000ac-ft/yr		ern boundary			(author's estimate)						
				as much as 300 ft			(**0.05 in/yr**)		of the county.			10% of pcp. in						
				thick, of which as			during 1971-72					irrigated & 1% of						
				much as 145 ft is								pcp. in nonirrigated				In the period of		
				saturated. Yields								land. (163000 ac				1948-1972 water level		
				100 to 2000 gpm.								irrigated & 390000				declined from 10-50		Lateral Inflow,subsurface
												ac of nonirrigated				ft. in places where		outflow & total recharge are
				Dakota Fm.								land in Greeley and				irrigation is practiced.		for Wichita and Greeley
				(sandstone aquifer)								Wichita) 1971-72						combined
				400-550 ft thick.														
				Similar aquifer in			[3c]**0.28 in/yr**					mathematical model	[3c] Dunlap, L.E., et al.,					[3c] based on the study in an
				adjacent counties								(steady state)	1980					area of 12 mi² in northwesterr
				yield 30-300 gpm.														Wichita ,west-central
							[3d]1-10% of annual						[3d] Gutentag & Stullken,					Kansas referred to by the
							pcp.(120-1200						1976					authors as "intensive study
							ac-ft) in intensive				[3e]10000		[3e] KWRB' 67					area."
							study area				(**0.26 in/yr**)							
				Ogallala			[3f]17600 ac-ft/yr					soil-water budget	[3f] USGS' 91					
							(**0.46 in/yr**)											

WEST CENTRAL KANSAS (GMD 1 Region)

No.	County	Area mi²	Reference Bulletin & publ. yr	Aquifer	Avg. Precip. (in/yr)	Avg. Temp F	Recharge from ac-ft/yr (in/yr) precip.	stream	lateral inflow	irrig.	Total Recharge ac-ft/yr (in/yr)	Method	Reference	Soils	Topog	Avg. Depth to Water Table	Vegetation & Land use	Comments
4	**Scott**	724	KGS B 66	Alluvium: Thickness	18.61	53.3	[4a]probably less						[4a] Theis, C. V., 1937	silty soils	High Plains	Based on depth of	Agriculture	
			1947	ranges from few feet			than **1/2 inch**						(KGS B 66)	and loamy	section of the	water table Scott Co.	38% crop-	
			KGS Irr. Ser.1	to as much as 200 ft										soils	Great Plains	may be divided into:	land & 62%	
			1976	in the southern part.											physiographic	1. Shallow water area	for grazing	
				Yields from 250-1500											province.	depth between <25 ft	as of 1939	
				gpm.											85% upland	to 75 ft		
															plains & 15%		crops and	
				Ogallala Fm. is the											stream flood	2. Intermediate	grasses	
				major aquifer and											plains &	depth between <50 ft		
				thickness ranges from											intermediate	to 100 ft		
				few feet to 215 ft in											slopes			
				central Scott County.												3. Deep water area		
				Yield ranges from												depth >100 ft		
				100-1000 gpm.												(as of 1947)		
							[4b]14000 ac-ft				[4b]53000	assuming 10%	[4b] Gutentag & Stullken,					total recharge including
							(**0.36"**) in 1971				(**1.4in/yr**)	& 1% of the	1976					seepage of irrigation
							[4b]24000 ac ft				in1971	precipitation to						water.
							(**0.62"**) in 1972				[4b]52000	percolate in irrigated						1972 being a wet year,
				Chalk aquifer							(**1.3 in/yr**)	and nonirrigated						infilteration is high but
				(Niobrara Fm.)Yields							in 1972	land, respectively						irrigation recharge low due to
				from 500-1000 gpm								during the growing						low application
												season.						
				Sandstone aquifer												[4g] 15 ft-20 ft at alluvium		[4g] ref. KGS B 66
				(Dakota Fm.) 430 ft						[4b]30000 ac-ft		based on pumpage				[4g] 30 ft-120 ft at Ogallala		
				in southwestern Scott						(0.78") in 1971		figures						
				to 710 ft in thickness						19000 ac-ft								
				in southwestern Lane					[4b]9000 ac-ft/yr	(0.49") in1972		Darcy's Law						lateral inflow along the
				county. Yields from					(**0.23 in/yr**)			A=5834400 sq ft						Scott-Wichita county line
				30-300 gpm.								(0.2093 sq mi)						
												I=0.0029				In the period from		Water table depletion: ref.
												K=64 ft/day				1940-48 to 1973, in		KGS B 27, KGS B 93
																about half of the area		and comparing the data
										[4c]about 20% of		experiment in	[4c] Meyer, W.R., et al.,			in Scott and Lane		with similar data collected
										irr. applied		Finney County	1953			water level declined		in 1973
																from <10 ft to 10 ft , in		
																about 10% of the area		
																decline was >30 ft and		
																the decline of as much		
											[4e]9000	.----------	[4e] KWRB' 67			as 50 ft was also observed		
											(**0.23 in/yr**)							
				Ogallala			[4f]16600 ac-ft/yr					soil-water budget	[4f] USGS' 91					
							(**0.43 in/yr**)											
5	**Lane**	720	KGS B 93	Alluvium: Thickness	18.77	53.6	[5a]340					soil-water budget	[5a] USGS' 91					
			1951	ranges from few feet			(**0.009 in/yr**)									10% to 60% of the sat.		
			KGS Irr. Ser.1	to as much as 200 ft												thickness has been		
			1976	in the southern part.												reduced throughout most		
				Yields from 250-1500												of the main body of the		
				gpm.												unconsolidated aquifer.		
																1940-48 to 1973		
				Ogallala Fm. is the			[5g]9550											
				major aquifer. Thickness			(**0.25 in/yr**)											
				ranges from few feet to														
				215 ft in central			[5b]about **1/4 inch**					From study in	[5b] Frye, J.C., 1942	silty soils			agriculture	
				Scott County. The								southern High Plains		and silty			and pasture	
				median thickness is								Estimation based		and loamy			460800 ac	
				110 ft. Yield ranges from								on assumption that		soils			farmland.	
				100-1700 gpm.								10% of pcp. In					(1946 census)	
							[5c]3000 ac-ft					irrigated and 1% in	[5c] Gutentag & Stullken,					
							(**0.08 in**) in 1971	negligible, as all				nonirrigated land	1976				crops and	
							6000 ac-ft	streams in Lane				percolates down to					grasses	
							(**0.16 in**) in 1972	are ephemeral.				groundwater table						
				Sandstone aquifer												[5f]25 ft-50 ft at alluvium		[5f] ref. KGS B 93
				(Dakota Fm.) 430 ft						[5c]6000 ac-ft		based on pumpage				[5f]<25 ft-75 ft at Ogallala		
				in southwestern Scott						(**0.16"**) in 1971		figures						
				to 710 ft in thickness						5000 ac-ft								
				in southwestern Lane						(**0.13"**) in 1972	18000 in 1971 [5c]	pcp+ lat. infl+irr						
				County. Yields from							(**0.46 in/yr**)							
				30-300 gpm.							20000 in 1972 [5c]	pcp+ lat. infl+irr						
											(**0.52 in/yr**)							
				Chalk aquifer														
				(Niobrara Fm.)Thick-						[5d]about 20% of		experiment in	[5d] Meyer, W.R., et al.,					
				ness of about 400 ft.						irr. applied		Finney County	1953					
				Not an important														
				aquifer as yield is small.														
											8600 [5e]		[5e] KWRB' 67					
											(**0.22 in/yr**)							

WEST CENTRAL KANSAS (GMD 1 Region)

No.	County	Area mi²	Reference Bulletin & publ. yr	Aquifer	Avg. Precip. (in/yr)	Avg. Temp F	Recharge from ac-ft/yr (in/yr) precip.	stream	lateral inflow	irrig.	Total Recharge ac-ft/yr (in/yr)	Method	Reference	Soils	Topog	Avg. Depth to Water Table	Vegetation & Land use	Comments
	Summary										**0.33 in/yr** [a]	[1]arithmetic avg.	[a] USGS' 91					[1]all averages are calculated
	(District avg.[1])										(std. dev. **0.10**)							considering only the counties
											0.19 in/yr [b]	[1]arithmetic avg.	[b] KWRB' 67					included in the GMD.
											(std. dev. **0.07**)							
											0.24 in/yr [c]	[1]arithmetic avg.	[c] KGS Bulletins					
											(std. dev. **0.16**)							

Note:

KWRB'67: Irrigation in Kansas. Kansas Water Resources Board,1967. Report no. 16e.
USGS' 91: Hansen, C.V., 1991. Estimates of freshwater storage and potential natural recharge for principal aquifers in Kansas. U.S. Geological Survey, Water-Resources Investigations Report 87-4230.
KGS B: Kansas Geological Survey Bulletin
std. dev.: Standard deviation

EQUUS BEDS AQUIFER (GMD 2 Region)

No.	County	Area mi²	Reference Bulletin & publ. yr	Aquifer	Avg. Precip. (in/yr)	Avg. Temp F	Recharge from ac-ft/yr (in/yr) precip.	stream	lateral inflow	irrig.	Total Recharge ac-ft/yr (in/yr)	Method	Reference	Topography & Soils	Avg. Depth to Water Table	Vegetation & Land use	Comments
1	**McPherson**	896	KGS B 79 1949		28.91	56					[1a]20% of pcp. **7 in/yr** or 365 ac-ft/yr/mi² for the period of record 1938-1943 inWichita well field area (85 mi²) **6 in/yr** or 320 ac-ft/yr/mi2 for normal pcp. in Wichita well field area (0.2*28.91=5.8 ≅6 in/yr for norm. pcp.)	[1a]Water level fluctuation and sp. yield 20%	[1a]Williams & Lohman 1949	nearly level to gently sloping silty and clayey soils on uplands	10 ft-110 ft	farm and pasture -1945	area covered in B 79 = 2340 mi² includes McPherson,and parts of Marion, Harvey, Reno and Sedgwick counties
											[1b]138000 (**2.89in/yr**)		[1b]KWRB' 67				
				Alluvium			[1c]9220 (**0.19 in/yr**)					soil-water budget	[1c]USGS' 91				
				Equus Beds			[1c]70700 (**1.47 in/yr**)				[1d]**0.1-4.5 in/yr**	[1d]3-D,finite diff, groundwater flow model	[1d]Spinazola, J.M., et al., 1985				[1d] recharge for predevelopment period (before 1940)
2	**Reno**	1262	KGS B 79 1949		28.53				quantity not known but small compared to that from pcp.		[2a]20% of pcp. 0.2*28.53=**5.7 in/yr**	[2a]groundwater level fluctuation and sp. yield (20%)	[2a]Williams & Lohman 1949	nearly level to moderately sloping loamy soils on uplands; nearly level, loamy and sandy soils on floodplains	<10 ft-50 ft	farmland and pasture	area covered in KGS B 79 = 2340 mi² includes McPherson,and parts of Marion, Harvey, Reno and Sedgwick counties
			KGS B 64 1946		27.83	54.3					[2b]20% of pcp. 300 ac-ft/sqmi/yr in the Arkansas River valley (**5.56 in/yr**)	[2b]water table fluctuation records and sp. yield estimates	[2b] Williams, C. C. 1946				
									[2c]500 ac-ft/mi into Mcpherson Fm. .=**1.07 in/yr** about 20000ac-ft/yr .=**3.75 in/yr** from Arkansas Valley upstream into the Hutchison area (study area of 100 mi²)			[2c]permeability, water table gradient, well logs (Darcy's Law)	[2c] Williams, C. C. 1946				[2c]500*11.4=5700 ac-ft/yr =1.07 in/yr in 100 mi² of study area in vicinity of Hutchinson; inflow= 500 ac-ft/mi/yr 11.4 mi=length of inflow boundary for the study area.(approximate) (see pl.1,KGS B 64, part 5)
											[2d]276400 (**4.1 in/yr**)		[2d]KWRB' 67				
				Alluvium			[2e]3880 (**0.06 in/yr**)					soil-water budget	[2e]USGS' 91				
				Equus Beds			[2e]139000 (**2.07 in/yr**)				[2f]**0.1-5.5 in/yr**	3-D finite diff, groundwater flow model	[2f]Spinazola, J.M., et al., 1985				[2f] recharge for predevelopment period (before 1940)

EQUUS BEDS AQUIFER (GMD 2 Region)

No.	County	Area mi^2	Reference Bulletin & publ. yr	Aquifer	Avg. Precip. (in/yr)	Avg. Temp F	Recharge from ac-ft/yr (in/yr) precip.	stream	lateral inflow	irrig.	Total Recharge ac-ft/yr (in/yr)	Method	Reference	Topography & Soils	Avg. Depth to Water Table	Vegetation & Land use	Comments
3	**Harvey**	540	KGS B 79								[3a]from 1938-1943	[3a]change in storage	[3a]Williams & Lohman	nearly level to	10 ft-40 ft	farmlands	[3a]avg. pcp. For the period
			1949								about 20% of	and pumpage data	1949	gently sloping		and pastures	29.7 in-- from the data for
											pcp. Or 320	estimated area of		loamy and sillty			Wichita, Sedgwick Co.
											ac-ft/sq mi/yr	W/T affected-100		soils on uplands;			
											in the well	sq. mi.		nearyl level,			
											field area			loamy and			
											≅ **6 in/yr**			sandy soils on			
											(well field area			floodplains			
											85 sq mi)						
			KGS B 119		33.69						[3b]**3.75in/yr**	[3b]change in storage	[3b] Stramel, G. J., 1956				The area covered in this
			1956		(1940-48)						from1940-48	sp. yield 20%					report includes parts of
					34.18						[3b]**8.8 in/yr**	[3b]change in storage					Harvey and Sedgwick
					(1948-52)						from 1948-1952	sp. yield 20%					counties, referred to as
					33.62						[3b]**6 in/yr** in avg	[3b]change in storage					well field area (378 mi^2)
					(1938-52)						from 1938-1952	sp. yield 20%					Equus Beds area: 2340 mi^2
					pcp.data												
					for Wichita						[3c]88700		[3c]KWRB' 67				
					Sedgwick Co.						**(3.08 in/yr)**						
				Equus Beds			[3d]61700					soil-water budget	[3d]USGS' 91				
							(2.14 in/yr)										
			[3e]USGS		29.05	13.4^0C	[3e]**6.06 in/yr** at					[3e]Darcy's Law and	[3e] Sophocleous & Perry,				
			WRIR		(1966-82)	56.12	Burrton from					mass balance	1987				
			87-4097				Feb to Jun										
			(Sophocleous				1983										
			and Perry				≅ 154 mm										
			1987)								[3f]**0.1-5.0 in/yr**	[3f]3-D,finite diff,	[3f]Spinazola, J.M., et al.,				[3f] recharge for
												groundwater flow	1985				predevelopment period
												model					(before 1940)
4	**Sedgwick**	1000	KGS B 79[4a]		30.2	57					[4a,a']20% of pcp.	[4a,a']groundwater level	[4a]Williams & Lohman 1949	nearly level to	<10 ft-30 ft	farmland and	[4a]Area covered in this report
			1949		(1917-44)						approx.**6 in/yr**	fluctuation and		moderately sloping		pasture	is 2340 sq mi.includes
			KGS B 119[4a']								320 ac-ft/yr/mi^2	specific yield (20%)	[4a'] Stramel, G. J., 1956	loamy soils on			McPherson and part of
			1956											uplands			Marion, Reno, Sedgwick
														nearly level,			and Harvey
			[4b]USGS		28.6	56.3	[4b]avg from	[4b]streams are	[4b]no inflow from north		**[4ba]0.4-5.5 in/yr**	[4ba]3-D,finite diff,	[4ba]Spinazola, J.M., et al.,	loamy and sandy			[4b]The study area includes
			WRIR		(1888-1985)		**0.1-8.8 in/yr**	in equilibrium	but gains from east		in Arkansas	groundwater flow	1985	soils on floodplains			Sedgwick and parts of
			88-4225-1989				depending on	with G/W or	and west but		River valley of	model					Reno, Kingman, Harper,
							local condition.	are gaining	outflow from		Sedgwick Co.						Harvey, Summer, Marion
							(from summary;		southern boundry		and adj. areas						Butler and Cowley counties
							no source)		is probably equal								
									so net effect on		[4c]193000		[4c]KWRB '67				
									recharge is		**(3.59 in/yr)**						
									insignificant								
				Alluvium			[4d]41900					soil-water budget	[4d]USGS' 91				
							(0.78 in/yr)										
				Equus Beds			[4d]36500										
							(0.68 in/yr)				**[4e]0.4-5.5 in/yr**	[4e]3-D,finite diff,	[4e]Spinazola, J.M., et al.,				[4e] recharge for
											(2.5-5.5 in/yr in	groundwater flow	1985				predevelopment period
											Arkansas River	model					(before 1940)
											valley of Sedgwick						
											Co. and adj.areas)						

EQUUS BEDS AQUIFER (GMD 2 Region)

No.	County	Area mi²	Reference Bulletin & publ. yr	Aquifer	Avg. Precip. (in/yr)	Avg. Temp F	Recharge from ac-ft/yr (in/yr) precip.	stream	lateral inflow	irrig.	Total Recharge ac-ft/yr (in/yr)	Method	Reference	Topography & Soils	Avg. Depth to Water Table	Vegetation & Land use	Comments
	Summary	11523		Lower Arkansas Unit	< 22->32	57	[a]20% of pcp.	streams are	some from		precise estimate		[a] KWRB, 1960				includes
				that includes the			in Equus Beds	effluent	adjecent areas		cannot be made						Rice, McPherson, Reno
				Equus Beds aquifer					to the west but		based on available						Harvey Sedgwick, and
									insignificant.		data(1960). Because						Ellsworth counties
											of the variable local						
											conditions annual						
								`			recharge could range						
											from practically zero						
											to as much as 50%						
											of annual pcp. in						
											some areas.						
	Equus Beds	**240**	[b]KWRRI	Equus Beds	30		[b]Non-sand dune					**[b]**groundwater flow	[b]Sophocleous, M.A., et al.,				
	modeling area	(area	1982				Equus Beds					model(steady-state	1982				
		selected	(Sophocleous,				model area:					groundwater flow					
		in the	et al.,)				4.2cm/yr or					in two dimensions)					
		report as					**1.65 in/yr**					parameters used:					
		indicated					Sand dune area					transmissivity,					
		in cited					north of Burrton					areal recharge,					
		reference)					16.3 cm/yr or					leakance of the					
							6.4 in/yr					stream beds and					
												specified head.					
		1406	[c]USGS		[c]30.37	56.6	[c](**1.70 in/yr**)	**0.02 in/yr**	[c](**0.14 in/yr**)		**1.86 in/yr**	[c]A modular 3-D	[c]Spinazola, J.M., et al.,				[c]parts of Harvey, Marion
			WRIR	Equus Beds	preceding		1940 (predev.)	1940 (predev	1940(predev.)			finite-diff, groundwater	1985				McPherson,Reno and
			85-4336		1940		(**1.91 in/yr**)	**0.22 in/yr**	(**0.19 in/yr**)		**2.32 in/yr**	flow model					Sedgwick counties.
			(Spinazola		30.58		1971-79 dev. P	1971-79 dev	1971-79 dev.period			(Mc Donald and					
			and others)		(1951-80)							Harbaugh,1984)					
							[ca]20% of pcp.					[ca]from fluctuation of	[ca]Williams & Lohman,				
												water table and	1949				
												specific yield					
							[cb]**3.75 in.** (during					[cb]Change in storage	[cb]Stramel, G.J.,1956				
							period Sep 1940										
							to Jan 48)										
							8.8 in. (during										
							period Jan										
							1948 to Jan 52										
	District avg.[1]	3698									[d]**5.92 in/yr**	[1]arithmetic avg.	[d] KGS Bulletins				[1]all averages are calculated
											(std. dev. **0.15**)						considering only the counties
											[e]**3.41 in/yr**	[1]arithmetic avg.	[e]KWRB' 67				included in the GMD.
											(std. dev. **0.54**)						
											[f]**1.59 in/yr**	[1]arithmetic avg.	[f]USGS' 91				[f]recharge for Equus Beds
											(std. dev. **0.68**)						
											[g]**0.18-5.13 in/yr**	[1]arithmetic avg.	[g] Spinazola, et al.,1985				
											(std. dev. **0.15**)						

Note:

KWRB'67: Irrigation in Kansas. Kansas Water Resources Board,1967. Report no. 16e.
USGS' 91: Hansen, C.V., 1991. Estimates of freshwater storage and potential natural recharge for principal aquifers in Kansas. U.S. Geological Survey, Water-Resources Investigations Report 87-4230.
KGS B: Kansas Geological Survey Bulletin
std. dev.: Standard deviation

SOUTHWESTERN KANSAS (GMD 3 Region)

No.	County	Area mi^2	Reference Bulletin & publ. yr	Aquifer	Avg. Precip. (in/yr)	Avg. Temp F	Recharge from ac-ft/yr (in/yr)				Total Recharge ac-ft/yr (in/yr)	Method	Reference	Topography & Soils	Avg. Depth to Water Table	Vegetation & Land use	Comments
							precip.	stream	lateral inflow	irrig.							
1	**Hamilton**	992	KGS B 49 1943		17.67	53.9	High in sand-hills area due to porous soil and presence of basins without surface drainage Relatively low in upland area due to impermeable soil		flow occurs southeastwards inflow occurs from Prowers Co., Colorado and from Greeley Co.					nearly level to gently sloping silty soils on uplands; nearly level, loamy & sandy soils on flood- plains; rolling to hummocky sandy soils on uplands;	alluvium <10 ft-25 ft dune sand 25 ft-50 ft rest>50-200 ft		
							[1a]Ark. Valley area 1000 or >1000acft/yr					[1a]water balance inflow=outflow (neglecting the contribution of subsurface flow as it is low)	[1a]McLaughlin, T.G., 1943				
								[1b]**0.96in/yr** (1970-74) **1.59 in/yr** (1975-79)			[1b]**1.95 in/yr**(1970-74) **4.21 in/yr**(1975-79) (pcp.+irr.-ET)	[1b]USGS 2D finite element GW flow model	[1b]Barker, R.A., et al., 1983				[1b]in Ark. River valley in Hamilton & Kearny Co. A=110,000 acres
											[1c]5400 ac-ft /yr (**0.1 in/yr**)		[1c]KWRB' 67				
				alluvium			[1c]2980 (**0.06 in/yr**)					[1c]soil-water budget	[1c]USGS' 91				
				High Plains (Ogallala)			[1c]9410 (**0.18 in/yr**)										
2	**Kearny**	861	KGS B 49 1943		15.85	53.9	High in sandhills area due to porous soil and presence of basins without surface drainage Relatively low in upland area due to impermeable soil	[2a]19000 ac-ft/yr (**4.24 in/yr**) over the alluvial area of 53760 ac from Hartland to Garden City along Arkansas River	Inflow occurs in north from Wichita and in west from Hamilton			[2a]difference in discharge at different points along the stream	[2a,2b]McLaughlin, T.G., 1943	sloping silty soils on uplands; nearly level, loamy & sandy soils on flood- plains; rolling to hummocky sandy soils on uplands;	alluvium 10 ft-25 ft dune sand 25 ft-50 ft rest>50-200 ft		
							[2b]Ark. Valley area 1000 or >1000acft/yr					[2b]water balance inflow=outflow (neglecting the contribution of subsurface flow as it is low)					
								[2c]**0.96in/yr** (1970-74) **1.59 in/yr** (1975-79)			[2c]**1.95 in/yr**(1970-74) **4.21 in/yr**(1975-79) (pcp.+irr. -ET)	[2c]USGS 2D finite element GW flow model	[2c]Barker, R.A., et al., 1983				[2c]in Ark. River valley in Hamilton & Kearny Co. A=110,000 acres
							[2d]**2.08in/yr** (1974-80) for sandhills and Ark. River valley . **0.5in/yr** for High Plains	[2d]1.10in/yr (1974-80) for sandhills and Ark. River valley . 0.19in/yr for High Plains				[2d]USGS 3D model (Trescott, 1975)	[2d]Dunlap, L.E., et al., 1985				[2d]Model area of upper aquifer (Ark. River valley+Sand dunes area)=603mi^2 & for lower aquifer (main High Plains aquifer)= 1227 mi^2 in Kearny and Finney counties.
											[2e]10800 (**0.24 in/yr**)		[2e]KWRB' 67				
				alluvium			[2f]1750 (**0.04 in/yr**)					[2f]soil-water budget	[2f]USGS' 91				
				High Plains (Ogallala)			[2f]25700 (**0.56 in/yr**)										

SOUTHWESTERN KANSAS (GMD 3 Region)

No.	County	Area mi²	Reference Bulletin & publ. yr	Aquifer	Avg. Precip. (in/yr)	Avg. Temp F	Recharge from ac-ft/yr (in/yr) precip.	stream	lateral inflow	irrig.	Total Recharge ac-ft/yr (in/yr)	Method	Reference	Topography & Soils	Avg. Depth to Water Table	Vegetation & Land use	Comments
3	**Finney**	1302	KGS B 55 1944		20.22	54.7	[3a]**1.4 in/yr** (1940-52) period of above- normal pcp.					[3a]Change in water lvl. in obs. wells S=0.2 and delta h=7ft	[3abcd]Meyer, W.R., et al., 1970	Rolling to gently sloping to nearly level sandy, loamy and silty soils on the uplands; nearly level loamy & sandy soils on floodplains.			[3abcd]area under consideration 552960ac
									[3b]45000 ac-ft/yr **0.98 in/yr** over the area considered, from the west & north of Scott Co. But the discharge from the co. was 47000 ac/yr (area=552960 ac)	[3c]15% of irrig. applied 10% for ditches		[3b]using water lvl. map of Latta-1944 and Q=TIL [3c]Experiment on irr. efficiency and measurement of Q in 2 points of canal.					
											[3d]124000ac-ft/yr or **2.7 in/yr** (1940-64) **<0.5 in/yr** predev. period (1922-30) considering equilibrium of system **<0.05 in/yr** long term avg. (1940-64)	[3d]water balance equation					[3d]2.7 in/yr reflects additional recharge resulting from recycled G/W for irrigation & an accompanying increase in effective R from pcp. on the irrig. land.
											[3h]16000 ac-ft/yr (**0.23 in/yr**)		[3h]KWRB' 67				
								[3i]max.25300 ac-ft/yr from Ark.River between Hartland & Garden City, (alluvium area of 53760ac) for the period of (Oct 1922-Sep 1942) **5.65 in/yr**.	[3j]max 2100 ac-ft/yr in narrow part of Ark. River valley near Hartland (alluvium was 2250 ft wide and avg. thickness 33')			[3i]Flow difference between Syracure & Garden City	[3i]KGS B 55 [3j] Latta, B.F., 1944				[3i]channel width avg. 0.05 mi; length from Hartland to Garden City 22 miles; channel area where seepage occurs=1.1 mi²=704 ac
									[3k]approx.1000 ac-ft/yr near Hartland				[3k] McLaughlin, T.G., 1943				
				Alluvium and High Plains (Ogallala)			[3l]600 ac-ft/yr (**0.00006 in/yr**) [3l]38400 ac-ft/yr (**0.55 in/yr**)	[3m]1700ac-ft/yr from Ark. River				[3l]soil-water budget	[3l]USGS' 91 [3m]Meyer, W.R., et al., 1970				
								[3n]37646 ac-ft/yr \\ **8.4 in/yr** (1991-1998) for the valley area (53760 ac) between Syracuse & Garden City. 55745 ac-ft/yr \\ **10.45 in/yr** (1991-99) for the valley area (64000 ac) between Garden City and Dodge City.				[3n] Water balance of stream	[3n]Whittemore, D.O., et al., 2001				
							[3O]**2.08in/yr** (1974-80) for sandhills and Ark. River valley . **0.5in/yr** for High Plains	[3O]**1.10in/yr** (1974-1980) for sandhills and Ark. River valley . **0.19in/yr** for High Plains				[3O]USGS 3D Model (Trescott 1975)	[3o]Dunlap, L.E., et al., 1985				[3O]model area of upper aquifer & valley=603mi² & for lower aquifer .=1227 mi² in Kearny and Finney counties
							[3P]**0.5 in/yr** in the vegetated sand dune area but higher in the barren sand dune area				[3q]**0.02-0.04 in/yr** **0.12 in.yr** (for irrigated land	[3P]neutron probe [3q]heat dissipation sensors and Darcy's Law	[3P]Prill, R.C.,1968 [3q]Sophocleous, M., et al., 2002				[3P]study area located in dune sand area in the Ark. River valley near Garnden City, for the period 1964-66 (period of high rainfall)
4	**Gray**	873	KGS B 55 1944		21.43			Arkansas River is gaining throughout its course in the county.			[4a]14800ac-ft/yr (**0.32 in/yr**)		[4a]KWRB'67	nearly level to gently sloping silty soils on uplands; undulating sandy soils on uplands; nearly level loamy & sandy soils on floodplains.			
				High Plains (Ogallala)			[4b]43700 ac-ft/yr (**0.94 in/yr**)					[4b] soil-water budget	[4b]USGS' 91				

SOUTHWESTERN KANSAS (GMD 3 Region)

No.	County	Area mi²	Reference Bulletin & publ. yr	Aquifer	Avg. Precip. (in/yr)	Avg. Temp F	Recharge from ac-ft/yr (in/yr) precip.	stream	lateral inflow	irrig.	Total Recharge ac-ft/yr (in/yr)	Method	Reference	Topography & Soils	Avg. Depth to Water Table	Vegetation & Land use	Comments
5	**Ford**	1082	KGS B 43 1942		20.5	54.3	small; estimated to be about <**0.5 in/yr** due to low permeability high evaporation and high ET.	small as Ark. River was effluent (gaining) most of the times in a year and so were other small rivers.		some in places in the vicinity of irrigatioin ditches and fields				nearly level to gently sloping silty soils on uplands; undulating sandy soils on uplands;	10 ft->150 ft		
							Data not adequate for quantative estimate but considered to be low.				[5b]**0.6 in/yr** (1980-81)	[5b]water budget	[5b]Spinazola & Dealy, 1983	nearly level loamy & sandy soils on floodplains.			[5b]water budget for 1980 conditions over 700 mi² underlain by Ogallala aquifer of the Ark. River
											[5c]34700 (**0.6 in/yr**)		[5c]KWRB' 67				
				Alluvium			[5d]370 (**0.006in/yr**)					[5d] soil-water budget	[5d]USGS' 91				
				High Plains (Ogallala)			[5d]54400 (**0.94 in/yr**)										
6	**Stanton**	685	KGS B 168 1964		15.03		[6a]**0.3 in/yr**, equivalent to 2% of pcp. (over area of 160mi² between Johnson and Ulysses)					[6a]difference in flow at 2 diff. locations underground.	[6a]Fader, S.W., et al., 1964	nearly level to gently sloping silty and loamy soils on uplands	<25 ft-250 ft		
							[6b]< **0.5 in/yr** due to impermeable soil.	[6b]large quantity from Bear Creek and possibly Sand Arroyo after rains. But actual quantity not known	[6b]receives some form the rainwater that reaches underground reservoir in southeastern Colorado.				[6b] Theis, C.V., et al., 1935				
											[6c]11700 (**0.32 in/yr**)		[6c]KWRB' 67				
				High Plains (Ogallala)			[6d]14000 (**0.39 in/yr**)					[6d] soil-water budget	[6d]USGS' 91				
7	**Grant**	571	KGS B 168 1964		17.24	54.6	[7a]**0.3 in/yr**, equivalent to 2% of pcp. (over area of 160mi² between Johnson and Ulysses)					[7a]difference in flow at 2 diff. locations underground.	[7a]Fader, S.W., et al., 1964	nearly level to gently sloping silty, loamy and sandy soils on uplands and flood-plains	alluvium 10 ft-50 ft rest <50 ft to >200 ft		
							.--------	[7b]Gain some water from Bear Creek, Lakin Draw & Sand Arroyo, whereas Cimarron River is losing in some regions and gaining in some	[7b]inflow occurs from Stanton Co. on the west & Kearny Co.on the north.				[7b] McLaughlin, T.G., 1946				
											[7c]9300 (**0.3 in/yr**)		[7c]KWRB' 67				
				High Plains (Ogallala)			[7d]22300 (**0.73 in/yr**)					[7d] soil-water budget	[7d]USGS' 91				
8	**Haskell**	580	KGS B 61 1946		18.02	54.6	.------	[8a]Gain some water from Bear Creek, Lakin Draw & Sand Arroyo, whereas Cimarron River is losing in some regions and gaining in others.	[8a]inflow occurs from Grant Co. on the west & Finney Co.on the north.				[8a] McLaughlin, T.G., 1946	nearly level to gently sloping silty and loamy soils on uplands	70 ft-250 ft		
											[8b]9600 (**0.31 in/yr**)		[8b]KWRB' 67				
				High Plains (Ogallala)			[8c]30200 (**0.98 in/yr**)					[8c] soil-water budget	[8c]USGS' 91				
9	**Morton**	720									[9a]12000 (**0.31 in/yr**)		[9a]KWRB' 67	nearly level to gently sloping silty soils on uplands; nearly level to undulating loamy and sandy soils on uplands and flood-plains.			
				High Plains (Ogallala)			[9b]16300 (**0.42 in/yr**)					[9b] soil-water budget	[9b]USGS' 91				
											[9c]**0.004-0.01** for native grass land	[9c]heat dissipation sensors and Darcy's Law	[9c]Sophocleous, M., et al., 2002				
10	**Stevens**	731	KGS B 61 1946		17.87	54.6	.------	[10a]Gain some water from Bear Creek, Lakin Draw & Sand Arroyo.	[10a]inflow occurs from Morton Co. on the west				[10a] McLaughlin, T.G., 1946	nearly level to undulating loamy and sandy soils on uplands and flood-plains; nearly level to gently sloping silty soils on uplands.	alluvium <10 ft to 20 ft rest 80 ft-150 ft		
											[10b]12000 (**0.31 in/yr**)		[10b]KWRB' 67				
				High Plains (Ogallala)			[10c]29200 (**0.75 in/yr**)					[10c] soil-water budget	[10c]USGS' 91				

SOUTHWESTERN KANSAS (GMD 3 Region)

No.	County	Area mi^2	Reference Bulletin & publ. yr	Aquifer	Avg. Precip. (in/yr)	Avg. Temp F	Recharge from ac-ft/yr (in/yr) precip.	stream	lateral inflow	irrig.	Total Recharge ac-ft/yr (in/yr)	Method	Reference	Topography & Soils	Avg. Depth to Water Table	Vegetation & Land use	Comments
11	**Seward**	643	KGS B 69 1948		19.01		[11a]**0.4 in/yr**	[11b]some from Cimarron River in northwestern part of the county.	[11b]from Stevens Co. on the west and Haskell Co. on the north but quantity is not known.			[11a]change in water level of 0.22'(1941-44) Sy=15%	[11a,b]Byrne & McLaughlin, 1948	gently sloping loamy & sandy soils on uplands & floodplains; nearly level to gently sloping silty soils on uplands.	<10 ft-20 ft in alluvium rest 50 ft->200 ft		
											[11b]10000 (**0.29 in/yr**)		[11b]KWRB' 67				
				High Plains (Ogallala)			[11c]34100 (**0.99 in/yr**)					[11c] soil-water budget	[11c]USGS' 91				
12	**Meade**	979	KGS B 45 1942		18.43 (17 in. avg. snow)	56	small from the pcp. that falls on the area due to impervious soil.	small due to steep slope of streams		some, as the shallow water table is quite high	[12a]10000 ac-ft/yr (**0.27 in/yr**) (eqv. to 1.5% of pcp.)	[12a]equating recharge to the discharge from the aquifer.	[12a] Frye, J.C., 1942	nearly level to moderately sloping silty soils on uplands.	<10 ft- > 150 ft		recharge into the artesian aquifer including the upward leakage through the confining beds. Area involved=685 sq. mi (recharge area of artesian water includes Finney, Haskell, Gray,Seward and Meade counties)
											[12b]14800 (**0.28 in/yr**)		[12b]KWRB' 67				
				Alluvium			[12c]4430 (**0.08 in/yr**)					[12c] soil-water budget	[12c]USGS' 91				
				High Plains (Ogallala)			[12c]50300 (**0.96 in/yr**)										
	Summary						[a] **0 to 2 in/yr** (**0.24 in/yr** in avg.) [a]**0.25 in/yr** in dune sand area to the south & Ark. River valley					[a]2D finite diff USGS model (Trescott1976)	[a,b]Stulken, L.E., et al., 1985				
							[b]**0.24 in/yr**	[b]**0.08 in/yr**	[b]**0.07 in/yr**		[b]**0.39 in/yr**	[b]simulated steady-state water budget (pre-1950)					
							[c]**0.58 in/yr**	[c]**0.05 in/yr**	[c]outflow higher than inflow		[c]**0.63 in/yr** (pcp.+ stream)	[c]3D MODFLOW (for 1982 period)	[c] Watts, K.R., 1989				[c]active node model area for the High Plains aquifer: 2695680 ac
							[d]**0.34 in/yr** (**0.45 in/yr** in Meade, Seward and eastern Stevens but **0.23 in/yr** in Morton and westen Stevens)					[d]2D finite diff USGS model (Trescott1976)	[d] Havens & Christenson, 1984				[d]active node model area 14,208,000 ac (includes part of Okhaloma and Morton, Stevens, Seward and Meade Co. in Kansas)
							[e]**0.6 in/yr** over total area; 175000ac-ft**1.5in/yr** in irr area of 1400000ac and 35000ac-ft\\ **0.15in/yr** in nonirr. area of 2824000ac in 1975 based on 15 in. of pcp. during growing season		[g]inflow from north & west of co. 8400 ac-ft/yr **0.024 in/yr**; outflow 15300ac-ft/yr\\ **0.04 in/yr**-to east	[f]420000ac-ft/yr to 560000ac-ft/yr (**1.2-1.6 in/yr**) -1975	[e]period (1922-1930) considering equilibrium of system **<0.05 in/yr** long term avg. (1940-64)	[e]10% of pcp. in irr. land and 1% of pcp. in nonirr. land [f] 20% of applied irr. [g]Darcy's Law	[efg]Gutentag, E.D., et al., 1981				
											[h]165000 ac-ft/yr (**0.29 in/yr**)	.---------	[h]KS. Governor's Task Force report, 1977				
											[i]**0.57 in/yr** (**0.3 in/yr** in nonirr. land and **1.8 in/yr** in irr. land)	[i]based on previous recharge estimates from pcp. and irr. to Ogallala aquifer in different subregions in western KS.(1977)	[i]O'Connor & McClain, 1982				
	District avg.[1]										[j]**0.70 in/yr** (std. dev. **0.26**)	[1]arithmetic avg.	[j] USGS' 91				[j]airthemetic average for Ogallala
											[k]**0.30 in/yr** (std. dev. **0.11**)	[1]arithmetic avg.	[k] KWRB' 67				[1]all averages are calculated considering only the counties included in the GMD.
											[l]**0.35 in/yr** (std. dev. **0.09**)	[1]arithmetic avg.	[l] KGS Bulletins				

Note:

KWRB'67: Irrigation in Kansas. Kansas Water Resources Board,1967. Report no. 16e.
USGS' 91: Hansen, C.V., 1991. Estimates of freshwater storage and potential natural recharge for principal aquifers in Kansas. U.S. Geological Survey, Water-Resources Investigations Report 87-4230.
KGS B: Kansas Geological Survey Bulletin
std. dev.: Standard deviation

NORTHWEST KANSAS (GMD 4 Region)

No.	County	Area mi^2	Reference Bulletin & publ. yr	Aquifer	Avg. Precip. (in/yr)	Avg. Temp F	Recharge from ac-ft/yr (in/yr) precip.	stream	lateral inflow	irrig.	Total Recharge ac-ft/yr (in/yr)	Method	Reference	Soils	Topog	Avg. Depth to Water Table	Vegetation & Land use	Comments
1	**Cheyenne**	1027	KGS B 100 1953		18	52		----------	----------	---------	[1a]12200 (**0.22 in/yr**)	----------	[1a] KWRB' 67	silty soils and loamy and silty soils	flat to rolling upland plains	90-175 ft along southern border 35-90 ft along southwestern border.	cropland and pasture much of land is pasture	" Data on permeability and transmissibility are not adequate to permit estimating either subsurface flow into or out from the county." (KGS B 100)
				Ogallala			23200 [1b] (**0.42 in/yr**)				[1b]23200 (**0.42 in/yr**)	soil-water budget	[1b] USGS' 91					
2	**Rawlins**	1080	KGS B 117 1956		18.5	52.3		----------	--------------	--------	[2a]10500 (**0.18 in/yr**)		[2a] KWRB' 67	silty soils	gently rolling	<10 ft at valleys to >200 ft at upland areas	cropland and pasture	
				Ogallala			[2b]41800 (**0.73 in/yr**)					soil-water budget	[2b] USGS' 91				agriculture	
3	**Decatur**	900	KGS B 196 1969	Ogallala Fm. avg. thickness 200 ft avg.saturated thickness 45 ft Alluvial valleys: yield range from 300 -1450 gpm depending on location.	18.42	53.2	less than **1/2 inch**	----------	[3b]5000 ac-ft/yr at western county boundary (**0.10 in/yr**)	-------	[3a]17500 (**0.36 in/yr**)	[3b] based on sat. thickness of water bearing strata, water table gradient, and avg. K of 40.1 ft/day	[3a] KWRB' 67	silty soils	gently rolling uplands	about 10-40 ft from surface in valleys to >100 ft in most places & 200 ft or more in high parts Ogallala Fm. avg. tk. 200 ft avg.saturated thickness 45 ft (1962)	cropland and pasture	
				Ogallala			[3c]36000 (**0.75in/yr**)					soil-water budget	[3c] USGS' 91					
4	**Norton**	880	KGS B 81 1949	-------------	20.81	52.8	[4a]approx. **1/4-1/2 in**	---------	-------------	-------		-------------	[4a] Frye, J.C., 1942	silty soils	Plains border section of the Great Plains physiographic province	40 ft in valleys upto 175 ft in uplands <10 ft in alluvium	cropland and pasture agriculture	
											[4b]15400 (**0.33 in/yr**)		[4b] KWRB' 67					
				Alluvium			[4C]2180 (**0.05 in/yr**)					soil-water budget	[4c] USGS' 91					
				Ogallala			[4C]41100 (**0.86 in/yr**)											
5	**Sherman**	1055	KGS B 105 1953		18	51.9	Lower than that from lateral inflow	------	[5a]approx. 21480 ac-ft/yr from west and southwest (**0.38 in/yr**)	--------	**0.1 of an inch [5a]**	[5a] "estimate based on available data."	[5a] Frye, J.C., 1942	mostly underlain by deposits of tertiary Ogallala Fm. Silty soils and loamy and silty soils	High Plains section - consists of nearly flat to gently rolling upland plains	in upland areas generally>100 ft <10 ft in stream valleys avg slope is 15 ft/mi	cropland and pasture farmland & pasture	
											[5b]13500 (**0.24 in/yr**)		[5b] KWRB' 67					
				Ogallala			[5c]13600 (**0.24 in/yr**)					soil-water budget	[5c] USGS' 91					
6	**Thomas**	1070	KGS B59 1945		17.95	51.8	**[6a]1/4 inch**	--------	----------	------		--------------	[6a] Frye, J.C., 1942	silty soils	flat to gently rolling	>200 ft to only a few feet along valleys <100 ft in east central part	cropland and pasture farmland	
											[6b]20700 (**0.36 in/yr**)		[6b] KWRB' 67					
				Ogallala			[6c]21800 (**0.49 in/yr**)					soil-water budget	[6c] USGS' 91					

NORTHWEST KANSAS (GMD 4 Region)

No.	County	Area mi^2	Reference Bulletin & publ. yr	Aquifer	Avg. Precip. (in/yr)	Avg. Temp F	Recharge from ac-ft/yr (in/yr): precip.	stream	lateral inflow	irrig.	Total Recharge ac-ft/yr (in/yr)	Method	Reference	Soils	Topog	Avg. Depth to Water Table	Vegetation & Land use	Comments
7	**Sheridan**	893	KGS B 116 1956		19.35	53.6		----------		--------	**about 1/4 inch** [7a]	Author's estimate	[7a] Bayne, C.K., 1956	silty soils and silty and loamy soils	nearly flat to gently rolling	<10 ft to 160 ft	cropland and pasture agriculture	
											14000 [7b] (**0.29 in/yr**)		[7b] KWRB' 67					
											35600 [7c] (**0.75 in/yr**)	soil-water budget	[7c] USGS' 91					
				Alluvium			390 [7c] (**0.01 in/yr**)											
				Ogallala			35200 [7c] (**0.74 in/yr**)											
8	**Graham**	891	KGS B 110 1955		20.55	53.9		--------	-----------	-------	[8b]7000 (**0.15 in/yr**)		[8b] KWRB' 67	silty soils and silty and loamy soils	High Plains section of Great Plains physiographic province	few feet to 140 ft	cropland and pasture agriculture about 40% of land is pasture	[8a] avg. ET from free water surface in growing season is 11.5 inches per month, & 75% of pcp. occurs in the growing season, so recharge is low (KGS B 110 p31)
							[8a] probably **<1/2 inch**					Authors' estimate	[8a] Prescott, G.C., 1955					
				Alluvium			[8c]2710 (**0.06 in/yr**)					soil-water budget	[8c] USGS' 91					
				Ogallala			[8c]32500 (**0.68 in/yr**)											
9	**Logan**	1073	KGS B 129 1958		18.97	53.3	[9b] **1/6 inch** (0.85% of the total pcp.)	-------	[9a] approx. < 25ac-ft/yr in southern upland, negligible in northern upland	10% of applied		[9a] based on groundwater contour map and saturated thickness	[9a,9b,9c] Johnson, C.R., 1958	silty soils and silty and loamy soils	High Plains section & plains border secn. in east	>40 ft from surface in uplands	cropland and pasture agriculture	[9b] on the northern upland of the county. Based on Darcy's Law at a sec. across N-S direction near eastern border of the county.
											[9c]estimated 3000 in northern upland (calculated value 2600 from precipitation)	[9c] From Darcy's Law at N-S section near eastern county boundary.						
											[9d]5000 (**0.09 in/yr**)		[9d] KWRB' 67					
				Alluvium			[9e]1880 (**0.03 in/yr**)				[9e]9780 (**0.17 in/yr**)	soil-water budget	[9e] USGS' 91			20ft from surface in alluvium aquifer		
				Ogallala			[9e]7900 (**0.14 in/yr**)											[9e] avg. sat. thickness 30' K_{avg}=58.82 ft/day
10	**Gove**	1070	KGS B 145 1960		20.89	53.2		---------	------------	--------	[10a]11000 (**0.19 in/yr**)		[10a] KWRB' 67	silty soils and silty and loamy soils	High Plains section	50 ft-150 ft	cropland and pasture agriculture 684000 ac	
				Alluvium			[10b]1700 (**0.03 in/yr**)					soil-water budget	[10b] USGS' 91					
				Ogallala			[10b]11600 (**0.20 in/yr**)											
11	**Trego**	900	KGS B 174 1965		21.4		[11a]at least 10000 ac-ft/yr (**0.21 in/yr**)	-------	-------	-------		Authors' estimate	[11a] Hodson, W.G., 1965	silty and loamy	High Plains section	Alluvium : few feet to 20 feet Ogallala: <10 ft to 100 ft	cropland and pasture agriculture	
											[11b]10,000 (**0.21 in/yr**)		[11b] KWRB' 67					
				Alluvium			[11c]4350 (**0.1 in/yr**)					soil-water budget	[11c] USGS' 91					
				Ogallala			[11c]13000 (**0.27 in/yr**)											

NORTHWEST KANSAS (GMD 4 Region)

No.	County	Area mi²	Reference Bulletin & publ. yr	Aquifer	Avg. Precip. (in/yr)	Avg. Temp F	Recharge from ac-ft/yr (in/yr) precip.	stream	lateral inflow	irrig.	Total Recharge ac-ft/yr (in/yr)	Method	Reference	Soils	Topog	Avg. Depth to Water Table	Vegetation & Land use	Comments
	Summary	8050	USGS	Alluvium	[β] 16-21				-----------	-------			[β]Jenkins & Pabst,			Alluvium is as	cropland	[β] mainly during 6
			OFR 4-75	has yield of		51-78	**0.25 in/yr**	most					1975	loessal	flat to gently	much as 105 ft	and pasture	months of growing
			1975	as much as			or about	streams							rolling	thick but =< 65 ft		season
				1500 gal/min			100,000	in western								is saturated		(area is different in Open-
							ac-ft/yr	part lose										file Report from the total
				Ogallala Fm.				water by								Ogallala Fm. has		area , as it reflects only
				has yield of				infiltration.								saturated thick-		the area in which the
				500-1200				Runoff is								ness ranging		study was carried out.)
				gal/min				only 1-2%								from 0-270 ft		
								of total pcp.										
				Dakota Fm.												Dakota Fm. lies		
				has yield of a												600-2600 ft		
				few gal/min												below ground		
																surface and its		
	District avg.[1]	9059									[a]**0.48 in/yr**	[1]arithmetic avg.	[a] USGS' 91			thickness ranges		[a]for Ogallala
											(Std. dev. **0.25**)					from 200-300 ft		
											[b]**0.23 in/yr**	[1]arithmetic avg.	[b] KWRB' 67					[1]all averages are calculate
											(Std. dev. **0.09**)							considering only the count
											[c]**0.29 in/yr**	[1]arithmetic avg.	[c] KGS Bulletins					included in the GMD.
											(Std. dev. **0.17**)							

Note: KWRB'67: Irrigation in Kansas. Kansas Water Resources Board,1967. Report no. 16e.
USGS' 91: Hansen, C.V., 1991. Estimates of freshwater storage and potential natural recharge for principal aquifers in Kansas. U.S. Geological Survey, Water-Resources Investigations Report 87-4230.
KGS B: Kansas Geological Survey Bulletin
Std. dev.: Standard deviation

GREAT BEND PRAIRIE (GMD 5 Region)

No.	County	Area mi²	Reference Bulletin & publ. yr	Aquifer	Avg. Precip. (in/yr)	Avg. Temp F	Recharge from ac-ft/yr (in/yr) precip.	stream	lateral inflow	irrig.	Total Recharge ac-ft/yr (in/yr)	Method	Reference	Topography & Soils	Avg. Depth to Water Table	Vegetation & Land use	Comments
1	**Barton**	892	KGS B 88		24.18		.----	.----	.-----	.-----				nearly level to	<10 ft->30 ft	farmland	
			1950								[1a]83100		[1a]KWRB' 67	gently sloping			
											(**1.72in/yr**)			silty soils on			
														uplands			
				GBP aquifer			[1b]33800					soil-water budget	[1b]USGS' 91				
							(**0.71 in/yr**)							nearly level silty			
														soils on flood-			
											[1c]22700	groundwater	[1c]Nuzman, C.,1990	plains			study area includes parts of
											(**2.19 in/yr**)	modeling					Ness, Rush and Barton Co.
											for Wallnut Creek						
											valley alluvium						
											[1d]60000	modification	[1d]Fader & Morton, 1972				
											(**1.26 in/yr**)	from KWRB' 67					
														nearly level to	20 ft		
2	**Rice**	721	KGS B 85		25.86	56	.-----	.-----	.-------	.--------	.-------			gently sloping		farmland	
			1950		(1898-1942)						[2a]74800		[2a]KWRB' 67	silty soils on			
											(**1.93 in/yr**)			uplands			
				GBP aquifer			[2b]43000					soil-water	[2b]USGS' 91	nearly level loamy			
							(**1.11 in/yr**)					budget		& sandy soils on			
														floodplains			
											[2c]75000	modification	[2c]Fader & Morton, 1972				
											(**1.95 in./yr**)	from KWRB' 67					
3	**Pawnee**	755	KGS B 80		23.48		.------	mainly in upland	from west	.-----				nearly level to	dune sand:	agriculture	
			1949					areas & the dun	and southwest		[3a]52600		[3a]KWRB' 67	gently sloping	<10 ft-50 ft		
								sand areas of			(**1.30 in/yr**)			silty & loamy	alluvium:10 ft-30 ft		
								Arkansas River						soils on uplands	terrace deposits:		
				GBP aquifer			[3b]18800					soil-water	[3b]USGS' 91	Nearly level silty	<20 ft-60 ft		
							(**0.47 in/yr**)					budget		loamy & sandy	Ogallala:		
														soils on flood-	<20 ft->100 ft		
											[3ca]0.6 in/yr	[3ca]mass balance	[3ca] Sophocleous, M. A.,	plains			
											for area of 325 mi²	(equilibrium of	1981				
											in Pawnee River	inflow and					
											valley	outflow)					
											[3cb]0.39 in/yr	[3cb]soil-moisture	[3cb] Sophocleous, M. A.,				
												budget	1981				
											[3cc]0.5 in/yr	avg. of 3ca &3cb	[3cc] Sophocleous, M. A.,				
												above	1981				
4	**Stafford**	794	KGS B 88		24.58		.-----	.-------	plate 1 shows	.------				undulating to	<20 ft-40 ft	farm land	
			1950						that the GW flows		[4a]187500		[4a]KWRB' 67	gently sloping			
									into the county		(**4.42 in/yr**)			sandy, loamy &			
									from Pawnee and					silty soils on			
				GBP aquifer			[4b]80400		Edwards Co.			soil-water	[4b]USGS' 91	uplands			
							(**1.9 in/yr**)					budget					
											[4c]190000	modification	[4c]Fader & Morton, 1972				
											(**4.49 in./yr**)	from KWRB' 67					

GREAT BEND PRAIRIE (GMD 5 Region)

No.	County	Area mi²	Reference Bulletin & publ. yr	Aquifer	Avg. Precip. (in/yr)	Avg. Temp F	Recharge from ac-ft/yr (in/yr) precip.	stream	lateral inflow	irrig.	Total Recharge ac-ft/yr (in/yr)	Method	Reference	Topography & Soils	Avg. Depth to Water Table	Vegetation & Land use	Comments
5	**Reno**	1262	KGS B 79		28.53		.-----	.-------	quantity not known	.-------	[5a]20% of pcp.	[5a]GW level	[5a] Williams & Lohman,	nearly level to	<10 ft-50 ft	farmland	area covered in B79=
			1949						but small compared		0.2*28.53=**5.7 in/yr**	fluctuation and	1949	moderately		and pasture	2340 mi2
									to that from pcp.			Sp. Yield (20%)		sloping loamy			includes McPherson,and
														soils on uplands;			parts of Marion, Harvey,
			KGS B 64		27.83	54.3					[5b]20% of pcp.	[5b]WT fluctuation	[5b,5c] Williams, C.C., 1946	nearly level, loamy			Reno and Sedgwick
			1946								300 ac-ft/sqmi/yr	records and sp.		and sandy soils			counties
											in the Arkansas	yield estimates		on floodplains			
											River valley						
									[5c]500 ac-ft/mi into		**(5.56 in/yr)**	[5c]permeability,					[5c]500*11.4=5700 ac-ft/yr
									Equus Beds from dune sand			WT gradient,					=1.07 in/yr in 100
									northeast of Hutchinson			well logs					mi² of study area in
									.=**1.07 in/yr**			(Darcy's Law)					vicinity of Hutchinson;
																	inflow= 500 ac-ft/mi/yr
									about 20000ac-ft/yr								11.4 mi=length of inflow
									.=**3.75 in/yr** from		[5d]276400		[5d]KWRB' 67				boundary for the study
									Arkansas Valley		**(4.1 in/yr)**						area.(approximate)
									U/S into the								(see pl1,Bul 64 part5)
				Alluvium +			[5e]143000		Hutchinson			soil-water	[5e]USGS' 91				
				Equus Beds			**(2.12 in/yr)**		(study area of			budget					
				& GBP aquifers					100 mi²)								
											[5f]0.1-5.5 in/yr	[5f]3-D,finite diff,	[5f]Spinazola, J.M., et al.,				[5f] recharge for
												groundwater flow	1985				predevelopment period
												model					(before 1940)
											[5g]270000	modification	[5g] Fader & Morton, 1972				
											(**4.01 in./yr**)	from KWRB' 67					
6	**Edwards**	619	KGS B 80		22.44						[6a]84000		[6a]KWRB' 67	undulating to	dune sand:	agriculture	
			1949								(**2.54 in/yr**)			gently sloping	<10 ft-50 ft		
														sandy & loamy	Alluvim:10 ft-30 ft		
				GBP aquifer			[6b]33100					soil-water	[6b]USGS' 91	soils on uplands	terrace deposits:		
							(**1 in/yr**)					budget			<20 ft-60 ft		
															Ogallala:		
											[6c]50000	modification	[6c] Fader & Morton, 1972		<20 ft->100 ft		
											(**1.51 in/yr**)	from KWRB' 67					
7	**Kiowa**	720	KGS B 65		22.15	56	[7a]in sandhills	streams are	water enters			change in	[7a] Latta, B.F., 1948	gentle to modera	Dune Sand:	agriculture	
			1948				2.2% of pcp.	influent but the	Meade and			storage		slopes & nearly	10 ft- 70 ft	farming &	
							0.58 in/yr	amount of recha	Ogallala Fms. Of			porosity-20%		flat surfaces with	Kingsdown Silt:	stock raising	
								is not known.	this area from					sandy, loamy &	>100 ft		
									Ford & Clark					silty soils.	Meade and Ogallala		
									counties.						<20 ft-60 ft		
															Ogallala≈ 20 ft		
									a part of water								
									in Meade and		[7b]99600		[7b]KWRB' 67				
									Ogallala Fms. is		(**2.6 in/yr**)						
									obtained from								
									Dakota Formation								
				Alluvium			[7c]50070					soil-water	[7c]USGS' 91				
				and			(**1.3 in/yr**)					budget					
				GBP aquifer													
									[7d]500-1000			Darcy's Law	[7d] Fader & Stullken, 1978				
									(**0.013-0.026 in/yr**)								
											[7e]50000	modification	[7e] Fader & Morton, 1972				
											(**1.3 in/yr**)	from KWRB' 67					
8	**Pratt**	729	KGS B205		24.04	avg.monthly	[8a]5-10% of pcp.	.-----		5-10% of			[8a] Layton & Berry, 1973	nearly level to	aluvium:<10'-12'	agriculture	
			1973			82 in Jul.	1.2-2.4 in/yr		[8a]**0.98 in/yr**	applied		[8a]Darcy's Law		gently sloping	< 10 ft- 12 ft		
						34 in Jan.	avg **1.6 in/yr**	negligible	across the western					silty and loamy			
							higher recharge		boundary					soils on uplands	dune sand:		
							if for dune sand								<20 ft-50 ft but in		
							area.		[8a]about 1500 ac-ft/yr			measurement			southeastern part		
									of saline water			of chemical			>50 ft-80 ft		
									leaks upward from			constituents					
									Permian rocks;			in streamflow			Loveland and Crete		
									actual amount is not						Fm.: 40 ft->100 ft		
									known.		[8b]169000		[8b]KWRB' 67				
											(**4.35 in/yr**)				Sappa and Grand		
				GBP aquifer			[8c]77800					soil-water	[8c]USGS' 91		Island Fm.:		
							(**2 in/yr**)					budget			20 ft-60 ft		
											[8d]150000	modification	[8d] Fader & Morton, 1972				
											(**3.86 in./yr**)	from KWRB' 67					
											[8e]3.5 in/yr	STREAM-AQUIFER	[8e] Moya, P., 1985				[8e] q= natural recharge rate
											from model calibrate	(Kemblowski,1982)					k= hydraulic conductivity
											ratio $q/k=8*10^{-6}$	a model utilizing					
											for k=100 ft/day	the integrated					
												finite diff. method					

GREAT BEND PRAIRIE (GMD 5 Region)

No.	County	Area mi^2	Reference Bulletin & publ. yr	Aquifer	Avg. Precip. (in/yr)	Avg. Temp F	Recharge from ac-ft/yr (in/yr) precip.	stream	lateral inflow	irrig.	Total Recharge ac-ft/yr (in/yr)	Method	Reference	Topography & Soils	Avg. Depth to Water Table	Vegetation & Land use	Comments
9	**Kingman**	864	KGS B 144		29.28	57.9	.-----	streams are	some from Pratt	negligible	not known			lies in the	dune sand:	agriculture	
			1960					effluent	Co. on the west.	(1955-56)				Great Bend	<10 ft-20 ft	farm and cattle	
														physiographic	alluvium: <10 ft	raising	
											[9a]201600		[9a]KWRB' 67	province.	Loveland and		
											(**4.47 in/yr**)				Crete Fm.:10 ft-20 ft		
														nearly level to	Ogallala:10 ft-20 ft		
				Alluvium			[9b]62700					soil-water	[9b]USGS' 91	moderately slopin	Sappa and Grand		
				and			(**1.36 in/yr**)					budget		loamy soils on	Island Fm.: 10 ft-70 ft		
				GBP aquifer										uplands	Fullerton and		
											[9c]150000	modification	[9c] Fader & Morton,1972		Holdrege Fm.:		
											(**3.26 in/yr**)	from KWRB' 67			20 ft-40 ft		
10	**Barber**	1146	KGS OFR		24.89	57.3	.-----	.-----	.------	.----				High Plains in			
			29-1								[10a]59100		[10a]KWRB' 67	northern and			
			1929								(**0.97 in/yr**)			western parts			
				alluvium			[10b]18280					soil-water	[10b]USGS' 91	& Plains border			
				and			(**0.3 in/yr**)					budget		in the eastern			
				GBP aquifer										part of the co.			
											[10c]40000	modification	[10c] Fader & Morton,1972				
											(**0.65 in/yr**)	from KWRB' 67		moderately			
														sloping to nearly			
														level clayey &			
														loamy soil			
11	**Summary**	5400	KGS IR4	GBP aquifer	22.5 at the		**2 in/yr**					water level	Fader & Stullken, 1978				Counties included: All of
	Great Bend		1978		western		5-10% of pcp.					fluctuation in					Kiowa, Kingman, Pratt &
	Prairie				border to		(1951-71)					wells					Stafford & parts of Barber,
					31.5 at the												Barton, Edwards, Pawnee,
					eastern												Reno, and Rice
											0.75 in/yr	USGS -2D	Cobb, P.M., et al.,1983				
											(1950-75)	finite-difference					
												flow model					
											0.28 in./yr	regional flow	Luckey, R.L., et al.,1986				
											(1950-80)	model for the					
												High Plains					
												aquifer(USGS					
												2-D finite diff.					
												flow model)					
											4.3 in/yr	daily soil-	Sophocleous & McAllister,				
											(1982-83)	moisture	1990				
											(Rattlesnake Cr.	budget	(KGS G/W Series 11)				
											basin)						
	District avg.[1]										**2.87 in/yr**	[1]arithmetic avg.	KWRB' 67				[1]all averages are calculated
											(Std. dev. **1.25**)						considering only the counties
																	included in the GMD.
											2.62 in/yr	[1]arithmetic avg.	Fader & Morton,1972				
											(Std. dev. **1.43**)						
											1.33 in/yr	[1]arithmetic avg.	USGS' 91				
											(Std. dev. **0.62**)						
											1.9 in/yr (1985-92)	arithmetic avg.	Sophocleous, M.,1992				
											(Std. dev. **0.71**)						
											1.4 in/yr (1985-90)	area-weighted					
												avg.					
												field measured					
												variables employing					
												Darcy's method					
												and water					
												budget analysis.					

Note:

KWRB'67: Irrigation in Kansas. Kansas Water Resources Board,1967. Report no. 16e.
USGS' 91: Hansen, C.V., 1991. Estimates of freshwater storage and potential natural recharge for principal aquifers in Kansas. U.S. Geological Survey, Water-Resources Investigations Report 87-4230.
KGS B: Kansas Geological Survey Bulletin
Std. dev.: Standard deviation