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Field data and flow system response in clay (vertisol) shale terrain, north central Texas, USA

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Abstract:

The water budget in clay shale terrain is controlled by a complex interaction between the vertisol soil layer, the underlying fractured rock, land use, topography, and seasonal trends in rainfall and evapotranspiration. Rainfall, runoff, lateral flow, soil moisture, and groundwater levels were monitored over an annual recharge cycle. Four phases of soil–aquifer response were noted over the study period: (1) dry-season cracking of soils; (2) runoff initiation, lateral flow and aquifer recharge; (3) crack closure and down-slope movement of subsurface water, with surface seepage; (4) a drying phase. Surface flow predominated within the watershed (25% of rainfall), but lateral flow through the soil zone continued for most of the year and contributed 11% of stream flow through surface seepage. Actual flow through the fractured shale makes up a small fraction of the water budget but does appear to influence surface seepage by its effect on valley-bottom storage. When the valley soil storage is full, lateral flow exits onto the valley-bottom surface as seasonal seeps. Well response varied with depth and hillslope position. FLOWTUBE model results and regional recharge estimates are consistent with an aquifer recharge of 1·6% of annual precipitation calculated from well heights and specific yield of the shale aquifer. Copyright © 2005 John Wiley & Sons, Ltd.

KEY WORDS water budget; clay-vertisol; groundwater; recharge; lateral flow; FLOWTUBE

INTRODUCTION

In many parts of North America, clay shale deposits are typically considered confining layers because of their fine-grained nature and assumptions of low permeability and concomitant restricted water movement (Neuzil, 1994). However, when these sediments are exposed at the surface they are subject to processes such as unloading and fracturing, shrink–swell volume changes, and animal burrowing and roots, all of which can significantly alter their permeability (Mills and Capuano, 1995). In addition to changes in the geological substrate, overlying residual clay soils are subject to seasonal variations in permeability due to cracking and related volume changes due to moisture variations.

The hydrogeology of these terrains is important, in that they are often chosen as burial sites for toxic metals, radionuclides, and landfill operations, as they are thought to store little groundwater and contribute little groundwater to streams or wells (Hanor, 1993). The interaction between the vadose zone and groundwater is important in understanding seasonal variations in soil water content, surface runoff, soil cracking and swelling, and recharge processes, which are important in septic system design and agricultural drainage (Ruland *et al.*, 1991; Harris *et al.*, 1994). Understanding the mechanisms of groundwater recharge and discharge is critical to assessing the potential of this terrain to transfer pollutants to the water table or streams and in assessing long-term stability of shale slopes affected by pore water pressure (Van Asch and Buma, 1997; Angeli *et al.*, 1998).

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Few studies have analysed the rainfall–recharge relationships in clay (vertisol)–shale bedrock terrain where the water table is at depths greater than several metres (George and Conacher, 1993). Although many studies have modelled and analysed water movement through cracking clay soils (Bevan and Germann, 1982; Hendricks, 1989; Booltink *et al*., 1993; Bronswijk *et al*., 1995; Favre *et al*., 1997), few studies have linked this recharge on a watershed scale to the underlying groundwater system within a fractured bedrock (Gburek and Folmar, 1999) in a subhumid climate regime. The purpose of this study is to assess the seasonal hydrogeology of such clay shale terrain. Data from this study are used to describe empirically the monitored water budget elements, including rainfall, surface runoff, soil moisture and soil flow, and groundwater fluctuations over a year-long monitoring cycle and, where possible, relate each to a holistic view of watershed dynamics.

Study site

The Blackland Prairie Province of Texas is an area of more than $4{\cdot}5 \times 10^6$ ha that contains over 40% of the state's population, including the cities of Dallas, Waco, Austin and San Antonio. Owing to the juxtaposition of cities and expansive soils, this area is reported to have annual repair costs to lightly loaded foundations and roads upwards of US$100 000 000 (Allen and Maier, 1993). The Blackland Prairie Province is a major agricultural region. Present-day agricultural land use consists of cattle production on pasture and rangeland, and corn, grain sorghum, and oat production under a wide range of tillage and management operations. Long, hot summers and short, mild winters characterize the climate in the Blackland Prairie. The growing season lasts, on average, from mid March to mid November. A majority of rainfall is associated with the passage of Canadian continental and Pacific maritime fronts (Knisel and Baird, 1971). Convective thunderstorms during the warmer months contribute intense, short-duration rainfall events. Tropical hurricanes can also contribute substantial rainfall, but their occurrence is rare. Freezing rain, sleet, and snow occur occasionally, but they do not contribute significant moisture. Long-term records collected at the site indicate an annual mean rainfall of 890 mm, with relatively wet spring and fall seasons and drier summer and winter seasons (Harmel *et al*., 2003). On average, 72 days per year experience rainfall greater than 0·76 mm, and 10 days have rain amounts greater than 25 mm.

The study site is located within the USDA-ARS Grassland, Soil and Water Research Laboratory watershed network near Riesel, TX (Figure 1). The watersheds were originally established in 1937 as the Blacklands Experimental Watershed.

The study watershed, Y2, drains 53·4 ha and includes three smaller upland subwatersheds Y6 (6·6 ha), Y8 (8·4 ha), and Y10 (7·5 ha). Total relief in the watershed is 12 m. Houston Black clay soil (fine, smectitic, thermic, udic Haplustert) dominates the site. This soil series consists of very deep, moderately well-drained soils formed from weakly consolidated calcareous clays and marls and generally occurs on 1–3% slopes in upland areas. This soil is very slowly permeable when wet (saturated hydraulic conductivity: $1{\cdot}52$ mm h^{-1}). However, when dry, preferential flow associated with soil cracks contributes to rapid infiltration rates. Because of its highly expansive clay, which shrinks and swells with changes in moisture content, Houston Black soil is recognized throughout the world as the classic vertisol. This soil typically has a particle size distribution of 17% sand, 28% silt, and 55% clay.

Regional hydrogeology

The research watersheds are underlain by marls and chalks of the Taylor Group of the Gulf Series, Cretaceous age (Figure 2). These units dip gently toward the east–southeast at approximately 17 m km^{-1}. The general site is within the axial trend of the Balcones Fault Zone. Within this zone the faults are normal and down-dropped to the east. Faults and fractures are difficult to see in outcrop due to rapid weathering of the bedrock. However, regional lineament studies have shown a strong regional influence of the Balcones Fault system, which has been locally overprinted by nontectonic-related fractures produced by release of overburden pressure and fractures due to weathering of the smectitic shale.

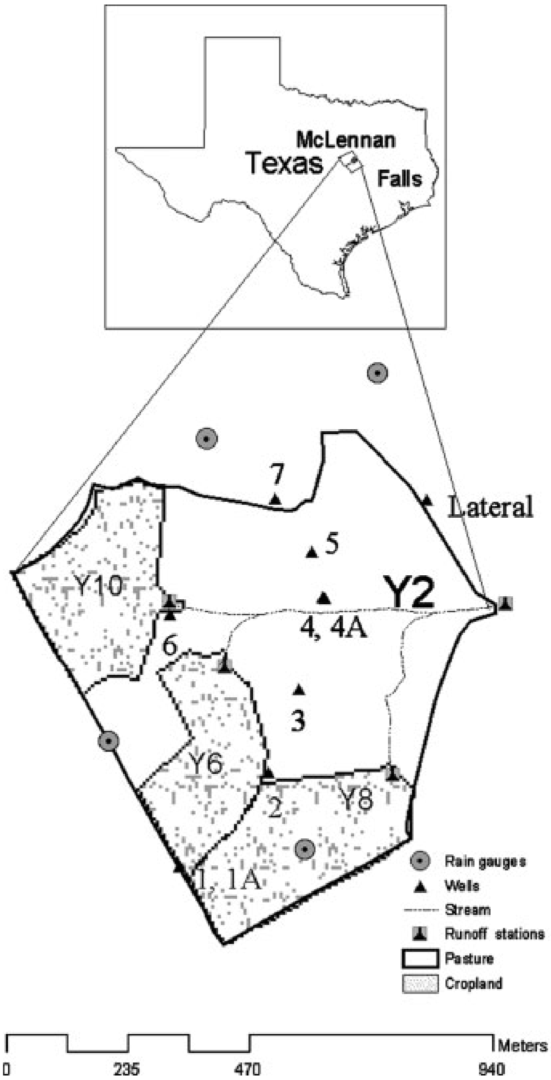

Figure 1. Location and instrumentation of Grassland, Soil and Water Research Laboratory Watersheds, Riesel, TX

Shallow groundwater in this region has been evident since the early farming days of the late 19th Century, as shown by the abundance of shallow hand-dug wells across the landscape. This shallow system has been studied and shown to follow local topography at an average depth of 3 m (watertable depth: 0·3 m, elevation +1·6 m; Osburn and Werner, 1992; Figure 2). Recharge occurs through aerial infiltration at the outcrop and deep lateral flow from the adjacent Austin Chalk.

SITE CHARACTERIZATION AND INSTRUMENTATION

Rainfall and surface water

Rainfall for watershed Y2 was measured at five locations with a tipping-bucket rain gauge (Hydrologic Services Pty, Sydney, Australia) connected to a Campbell Scientific CR10 datalogger (Campbell, Logan, UT).

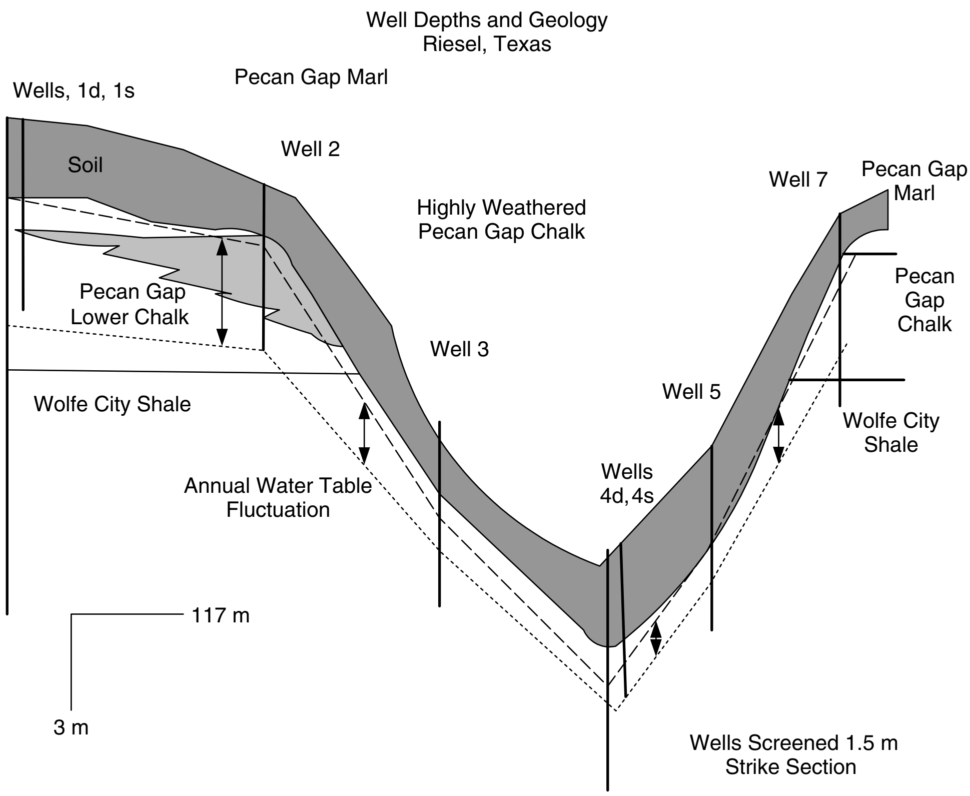

Figure 2. Geologic cross-section, well location and relative depths

Sub-daily rain data are recorded at 10 min intervals (sensitivity 0·254 mm). A standard rain gauge is also used at each site as a backup and calibration device. Daily rainfall values used in calculations are weighted based on the representative Thiessen area of each gauge.

The runoff station at the Y2 consists of a Columbus A-1 deep-notch concrete weir and a 4·6 m Parshall flume. This dual weir structure, installed in the late 1930s, was designed to measure both small and large flows accurately. During the study, each stilling well at Y2 was instrumented with a KPSI (Keller Pressure Systems, Inc., Hampton, VA) pressure transducer connected to Campbell Scientific CR10 datalogger and a float gauge with a chart recorder. Discharge measurements are made by continuously recording the flow level in each stilling well and converting measured depth to flow rate with an established stage–discharge relationship.

Soil and subsurface site characterization

In 1966, a drain trench was constructed by Knisel and Baird (1971) to analyse hillslope seepage rates on a terraced slope adjacent to the study watershed. The drainage trench was constructed parallel to the slope contours about midway down the slope. The trench is 0·46 m wide by 1·52 m deep by 61 m long. The trench is backfilled with 0·91 m of washed gravel (0·64 to 3·8 cm). Building felt was placed on top of the gravel to prevent vertical infiltration, and the trench was backfilled with topsoil. A metal collection box with a 90° v-notch weir receives flow from the trench through a 10 cm diameter PVC pipe. Flow at the weir is monitored with a pressure transducer and datalogger.

A series of groundwater monitoring wells were installed between 1997 and 2001. Wells were located at the interfluve (divides), crest or upper hillslope, lower hillslope, and valley bottom (Figure 2). All wells were screened (~60 cm) and backfilled with a sandpack to the top of the screened interval. This was followed by a bentonite seal and backfill. Wells were installed to measure shallow aquifer water levels and recharge properties of the site. Each well is instrumented with a KPSI pressure transducer (Keller Pressure Systems, Inc., Hampton, VA) attached to a CR10x datalogger (Campbell Scientific, Inc., Logan, UT) to provide a continuous record of piezometric head. The water level in each well is also monitored twice weekly with a hand-held 'e-line' water-depth gauge to provide backup, quality control, and calibration data.

A reference well (1) was drilled (NX core) to a depth of 13·7 m. This well penetrated the entire geologic section affecting flow in the watershed. The first 3·0 m were obtained with Shelby tubes (pushed), and the remainder of rock samples were obtained by rotary coring. Core recovery was good to excellent (ranging from 82 to 100%).

Non-vertical fractures in the core were noted. Fracture frequency decreased with depth. This relationship is illustrated by a more extensive study in the same geologic units by Alexander (1998), who quantified fracture frequency with depth (Figure 3). Results of detailed fracture studies in the Taylor shale bedrock performed to the north of the study area using inclined borings show the same general trend (Table I).

Shale bedrock in the north central Texas can be divided into four weathering zones. From the surface, these are: (1) the soil zone; (2) the highly oxidized (weathered) zone; (3) the transition zone; (4) the unweathered zone. These are similar to designations used by Ruland *et al.* (1991) for zones observed in weathered and fractured clayey till. The soil zone consists of 0·9–1·8 m of dominantly smectitic clays that shrink and swell on a seasonal basis. Major cracking occurs during the summer; cracks reach up to 2 m in depth. The shale in zone 2 is noted by the presence of laminar bedding, but the rock matrix is completely oxidized. Zone 3 is noted by highly oxidized fractures but a relatively unoxidized matrix. Zone 4 is noted by unoxidized fractures and matrix. At the study site, only zones 1 to 3 are observed. Zone 1 ranges from 1·8 m on the interfluves to 0·9 m along the hillslopes, to 1·8–3·0 m in the valley bottom. Zone 2 ranges from 1·2 to 1·5 m on the hillslopes to greater than 6·1 m in the valley bottom. Zone 3 averages 1·5 m. Typically, weathering is deeper on the divides than the valley bottoms (Blank *et al.*, 1952). In this area, the divides represent an older landscape and are covered with scattered ancient quartzite cobbles. The more recent drainage and incision (Plio-Pleistocene) has created the present gently undulating landscape, with relief from the valley bottom to the divide ranging from 18 to 30 m.

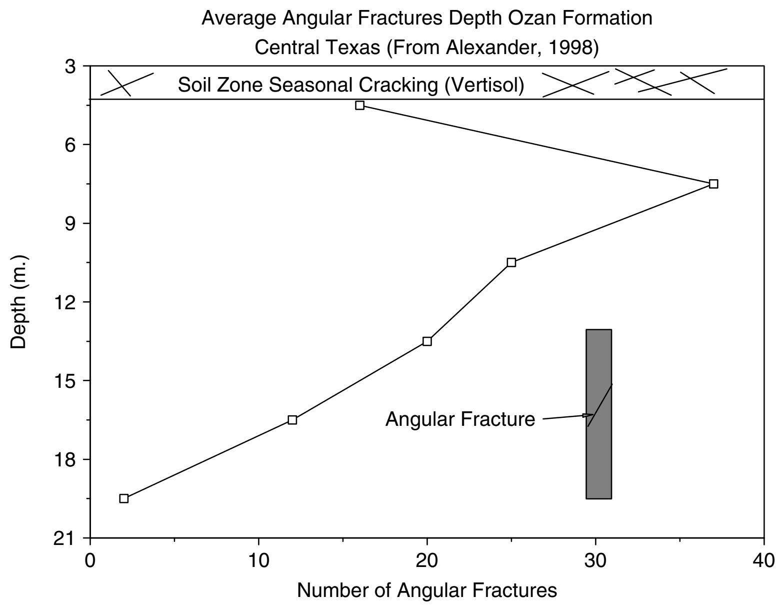

Figure 3. Bedrock fracture assessment for study area

Table I. Aquifer and soil characterization for the study area

	Soil zone	Weathered rock	Unweathered rock	
Thickness (m)	1–3, soil	1–2, highly fractured	>15, moderate–low fractured	
Seismic velocity (m s^{-1})	190–251	2100	6120	
Fractures/cracks	Cracks to 2 m	16·4/m bedding 0·42/m joint	20·5/m bedding 0·62/m joint	
K (m day^{-1})	<.06 to 10	0·78 mean	0·46 mean	
Storage	Variable with cracks Matrix 0·15	0·035–0·017	<0·017 to 9×10^{-4}	
Grain Size	5% sand 50% silt 45% clay	1·6% fine sand 25–45% silt 40–56% clay	Same as weathered rock	23–77·2% $CaCO_3$
Depth (cm)	Soil zone			
	0·1 bar	0·3 bar	15 bar	K (m day^{-1})
12·0	44·64	41·48	30·15	0·1463
24·5	40·23	39·67	30·78	0·2194
48·3	40·85	39	32·55	0·1463
99·0	42·97	42·45	34·4	0·067

The lowest stratum at the site is the Wolf City member of the Taylor, which locally consists of sandy marl that grades upward into a silty marl. Overlying the silty marl is a stratum of chalk, about 4·0 m thick, belonging to the Pecan Gap Member of the Taylor. This chalk is hard enough to have some effect on the topography and local soil depth and may locally influence the groundwater flow system (Blank *et al.*, 1952). The chalk is transitional upwards into a highly calcareous marl that underlies the soil zone in the uplands. Mechanical and partial chemical analysis of the outcropping units are shown in Table I. Of note is the increase in silt and decrease in carbonate content, which represents the major difference in the mapped geologic units.

Geophysics

Resistivity and EM-31 surveys were performed at the study site. Seismic refraction studies were conducted in the outcrop of the shale bedrock in central Texas. General results of the regional seismic refraction studies are shown in Table I. The weathered upper bedrock material has lower seismic velocities than the unweathered bedrock. This upper, highly resistive interval, termed the 'active zone' by local engineers, corresponds to the approximate zone of seasonal moisture variation and related shrink–swell of the soils and bedrock. Based on past studies in the area, resistivity soundings (Wenner array) give a better indication of the soil zone depth and the highly oxidized and fractured upper zone than seismic refraction (Paniszczyn, 1989) and may be a useful tool in studies of the vadose zone and surficial geotechnical properties (Font, 1980; Frolich and Parks, 1989). Resistivity soundings were taken at the height of the dry season (August), which correlates with the period of maximum crack openings. Resistivity profiles are shown over each well site (Figure 4). The soundings illustrate the depth of the highly resistive upper zone over the entire watershed. This highly resistive zone is thought to reflect the area of dominant crack development. This zone corresponds to the depth of the lowest water table observed during the monitoring period. EM-31 profiles were obtained seasonally over the site during dry and wet periods. Preliminary results (in preparation) indicate high seasonal flux in the uplands and lower flux in the lower hillslope and valley.

Hydraulic properties

Soil infiltration rates were determined from ring infiltrometer tests in the local area (Guillette, 1988). These tests were also conducted on the Heiden and Houston Black soils mapped at the research site. The duration

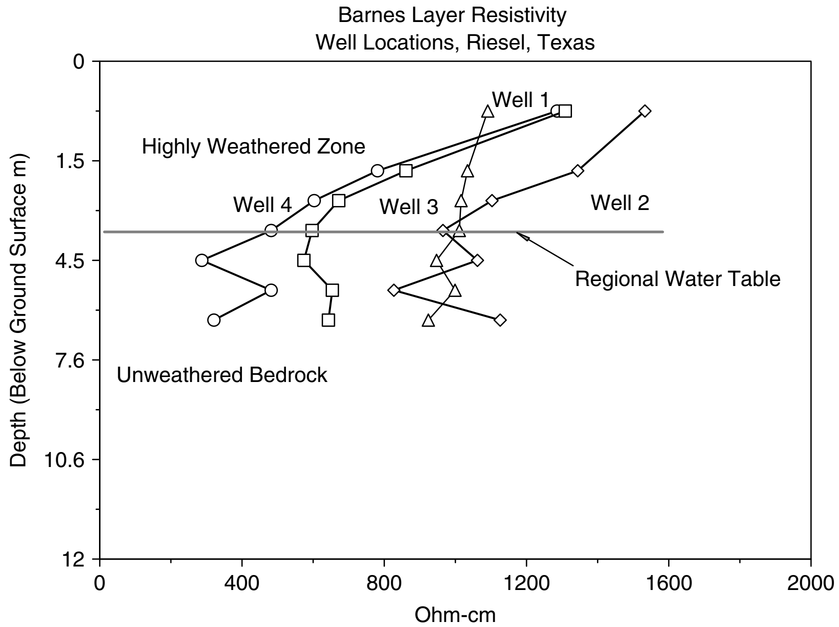

Figure 4. Resistivity soundings and water table at study site

of each test was 72 h, and an average of three trials was used to calculate infiltration rates. Rates for the Heiden soil, mapped on the interfluves, were consistently one order of magnitude higher than the Houston Black soil, which is mapped on the lower slope and valley bottom. This is attributed to the blockier nature and increased fracture density of the Heiden soil. In addition, laboratory tests have been completed on the saturated permeability and soil tension of the soils at the site. Results are included in Table I (Holtan *et al.*, 1968). Tensiometers were not installed at the study area owing to the expansive nature of the soils, which results in seasonal cracking and frequent loss of an adequate seal for proper tensiometer operation. Soil moisture sensors were installed at the upper hillslope position in order to assess the annual moisture changes in the soil associated with shrinking (cracking) and swelling of the soil. These sensors (Theta probes) were installed at depths of 0·15, 0·25, 0·45 and 0·95 m. The dielectric sensors operate using the frequency domain reflectometry approach, in which an oscillator generates an AC field that is applied to the soil in order to detect changes in the soil dielectric properties, which are linked to variations in volumetric soil water content. In addition, the shrinking and swelling of the interfluve soils were monitored at this site (Arnold *et al.*, 2005) to determine the correlation between seasonal soil moisture change and soil cracking and runoff at the site. Crack volume was determined by noting the differential movement of heave settlement points installed at depths up to 2·44 m, after Bauer *et al.* (1993) and Cheng and Pettry (1993). Crack volume was calculated as the difference between the three-dimensional volume change and change in layer thickness. This was done for each individual soil layer monitored (0·15, 0·45, 0·91, 1·52, and 2·44 m). By adding the values of each layer, the three-dimensional volume change and crack volumes of the entire soil profile were obtained (Arnold *et al.*, 2005).

The average hydraulic conductivity K of the bedrock was derived from four sources: local constant-rate pumping tests (Guillette, 1989), laboratory permeameter tests, packer tests (Osburn and Werner, 1992), and slug tests (Plant, 2000). The results of the tests are given in Table I.

In the pumping tests, wells were drilled to a depth of 6·1 m and completed with 0·7 m Tri-Loc slotted screens. The well array utilized in the test consisted of a pumping well and three observation wells at different distances from the pumping well. Two observation wells were placed at right angles to the third observation well to test for horizontal aquifer anisotropy. The observation wells were oriented parallel and perpendicular to the flow direction, which was assumed to be aligned with the transmissivity tensor. The drawdown data indicate that drainage, slopes and observation well placement were near to the transmissivity tensor alignment. Groundwater flow in the marl is heterogeneous and anisotropic, as it is controlled by a heterogeneous fracture network. Horizontal flow along the bedding planes is more dominant than vertical flow. A change in slope of the drawdown curve was noted during the pump tests and was attributed to several different fracture networks in the shallow groundwater system. Hydraulic conductivities achieved from laboratory permeameter tests on reconstituted samples of the bedrock found at the site typically range from 1×10^{-8} to 1×10^{-10} cm s^{-1}. (McAtee, 2003, personal communication).

Hydraulic conductivity values from packer tests were completed as part of an overall hydrogeologic assessment of shale and chalk bedrock in the former Superconducting Super Collider site north of the study area carried out by Osburn and Werner (1992). The median conductivity value of the tests in both the shales and chalk was 2×10^{-7} cm s^{-1}. The median value in fault zones was significantly higher ($3{\cdot}5 \times 10^{-6}$ cm s^{-1}). They found no correlation of hydraulic conductivity on the basis of formation (chalk versus shale), fractured versus unfractured rock, or depth.

Aquifer testing was performed at the site by Plant (2000) by using both additive and subtractive slug testing. These tests confirmed the vertical heterogeneity of the site. Tests on the shallow wells provide values of hydraulic conductivity ranging from 5×10^{-3} cm s^{-1} (wells 1, 2, and 7) to 6×10^{-4} cm s^{-1} (wells 3 and 5); the deep well (1) had a computed conductivity value of 3×10^{-6} cm s^{-1}.

PRELIMINARY RESULTS

Interpretation of the dynamics of the clay terrain is best examined by analysing seasonal variations in each flow system between periods of recharge and discharge (Christen *et al.*, 2000; Jennings *et al.*, 2001; Jyrkana *et al.*, 2002). The period chosen for this comparison is from September 2000 to August 2001. Because the beginning and end dates correspond with dry periods before major recharge, net annual soil moisture change was assumed negligible in water budget calculations for this period.

Rainfall, runoff and soil moisture

The basin response to rainfall is related to hillslope position and soil moisture status. To assess the entire recharge–discharge cycle, results are shown from the onset of recharge in late fall to the beginning of recharge the next fall. During the study period, from September to August, the total rainfall was 1152 mm. This is 29% greater than the annual average for the watershed of 890 mm.

Soil moisture increased rapidly near the surface when cumulative rainfall reached 63 mm (day 290), or about 5% of annual rainfall. At 0·31 m soil depth, recorded volumetric moisture increased from 20 to 38%. At 0·51 m, soil moisture increased from 30 to 40%. Deeper soil moisture at 0·95 m began to respond 21 days later (310), when rainfall totals had reached 273 mm or about 23% of total annual rainfall (Figure 5). Volumetric soil moisture jumped from 30 to 45% at 0·95 m at this time.

Runoff from the watershed began when 249 mm of cumulative rainfall had fallen on the watershed (day 307; Figure 6). Runoff commenced after soil flow had begun. Total streamflow recorded for the year-long study period was 291 mm. This amount represents surface runoff and baseflow. Baseflow was separated from total flow using a filtering technique after Arnold and Allen (1999). Baseflow, recorded at the lower weir, accounted for approximately 11% of total flow, or 32 mm, during the study period.

Soil flow

Lateral flow through the upper 1·5 m of the soil began when rainfall totals reached 116 mm, or 9·6% of total rainfall (Figure 7). Lateral flow peaked 4 days later with 13 mm additional rainfall, for a total of 12·7%

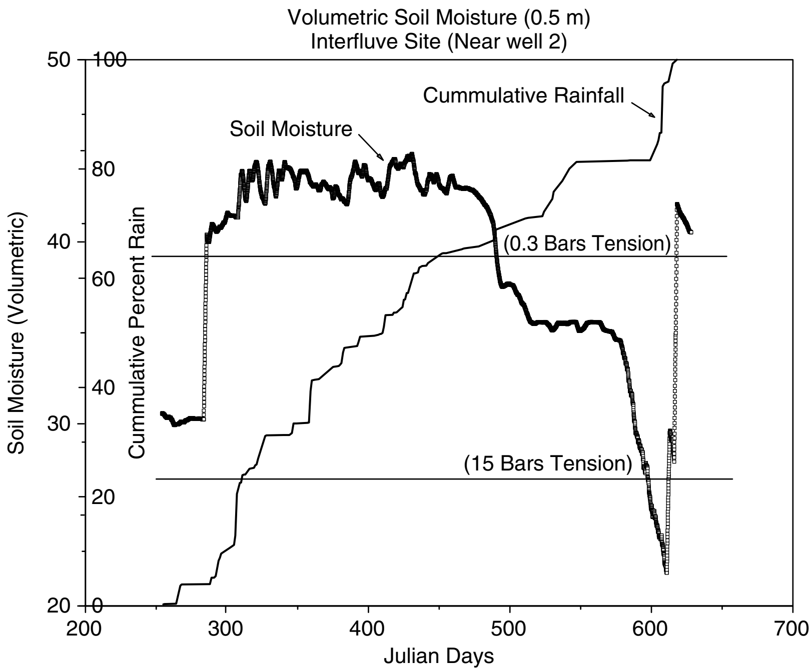

Figure 5. Volumetric soil moisture at 0·5 m over the study period and cumulative rainfall (soil moisture taken near well 2)

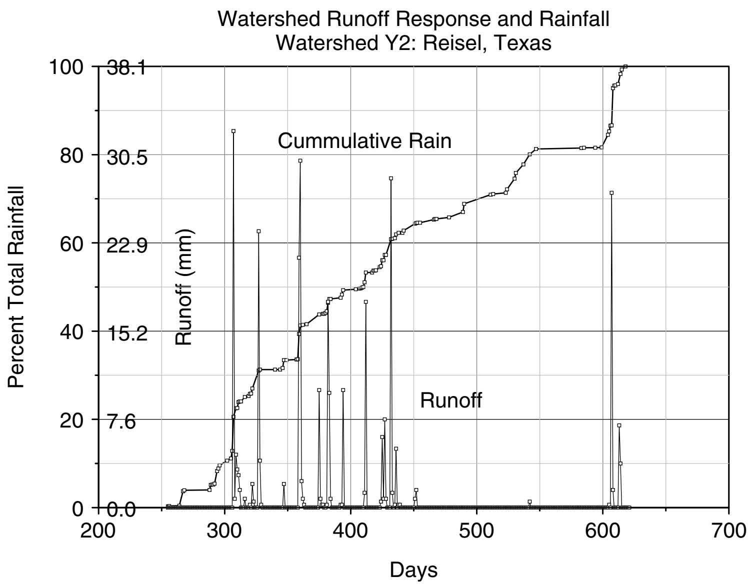

Figure 6. Cumulative rainfall and watershed runoff

of annual precipitation. More pronounced lateral flow occurred when cumulative rainfall reached 249 mm (20·5% of annual rainfall). Assuming that the same rates occur in the main watershed as were measured in the seepage trench (Figure 1), it is estimated that a total of 154 mm of water passed downslope through the soil zone during the monitoring period.

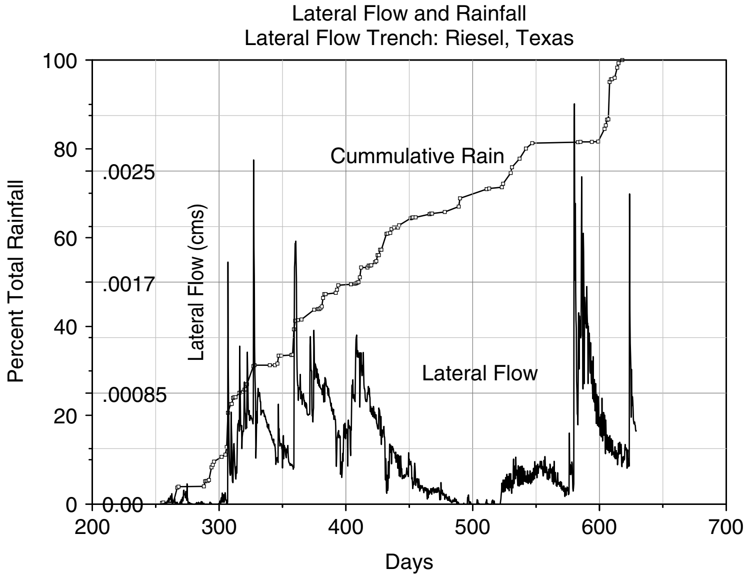

Figure 7. Cumulative rainfall and lateral soil flow

Well response

Well response varied with depth and hillslope position. Interfluve wells responded before hillslope wells, and hillslope wells responded prior to valley-bottom wells, and shallow wells responded before deep wells (Figure 8). The shallow interfluve well (1s) began to respond on day 306, with 156 mm of cumulative rainfall, and reached its peak on day 313, with 292 mm of cumulative rainfall. The total well response was 2·5 m in less than 7 days. The upper hillslope well (well 2) began to respond on day 334, with 379 mm of cumulative rain. This well continued to rise until day 446, for a total well response of 2·7 m. The lower hillslope well (well 3) began to rise on day 370, with 503 mm of rainfall. This well peaked 192 days later, with 985 mm of rainfall. The total well height response for well 3 during recharge was 1·2 m. The valley-bottom well began to respond on day 271, with 48 mm of rainfall, and peaked on day 439, for a total well height variation of 0·14 m. The deeper interfluve well (1d) began to respond on day 334, with 379 mm of rainfall, and peaked on day 502, for a total flux of 0·7 m. From late May to September, the deep well (well 1d) water level is higher than the shallow well (1s), and this may be indicative of confined conditions (Figure 8).

Water-level response is related to topographic position in the watershed. A correlation matrix was calculated for daily well heights in the watershed. The coefficient is a simple measurement of the degree of correlation between two variables in the sample. The product moment correlation coefficient can vary between $-1{\cdot}0$ and $+1{\cdot}0$, with 0·0 indicating no correlation. A value of 0·905, therefore, indicates a very strong correlation between the well heights of the two wells on the basis of the annual sample data. Conversely, a high negative value would indicate that as the water level in one well goes up, the water level in the other well goes down at a similar rate. No correlation indicates poor relationships in water-level response. A strong positive relationship occurs between interfluve and upper hillslope wells (1s, 2, 7), between the lower hillslope wells (3, 5, 6), and between the valley-bottom wells (4s, 4d), as shown in Table II.

Preliminary estimates of recharge to the bedrock were determined by calculating the overall fluctuation in the groundwater levels from the beginning to end of the rainfall period, after Christen *et al.* (2000) and Jennings *et al.* (2001). Since precipitation is highly seasonal, shallow groundwater levels, as recorded in the wells, rise and fall on a seasonal basis. The net recharge can then be assessed for a period by subtracting the later water surface from the earlier one. The resultant flux can be used in conjunction with estimates

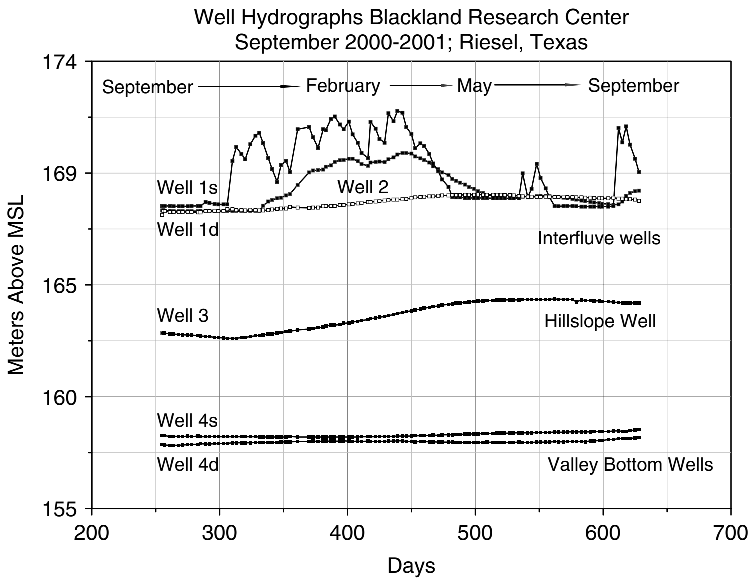

Figure 8. Well hydrographs for study period

Table II. Correlation matrix for well heights within the watershed

Wells	1deep	1	2	3	4deep	4	5	6	7
1deep		−0·26	0·135	**0·882**	0·326	0·119	0·664	**0·852**	−0·027
1s			**0·726**	−0·374	−0·238	−0·007	−0·481	0·032	**0·828**
2				−0·080	−0·182	−0·069	−0·352	0·442	0·593
3					0·643	0·482	**0·895**	**0·777**	−0·342
4deep						**0·905**	**0·763**	0·313	−0·157
4s							0·593	0·206	−0·002
5								0·563	−0·363
6									0·043
7									

of soil/rock effective porosity to assess volumetric changes in groundwater. Surfaces were contoured for the higher water table in March and the lowest seasonal water table in October. The volume change between the surfaces was determined by subtracting the differences between the two surfaces and contouring and computing the volume. This volume was then multiplied by the effective porosity (assumed to be 0·017, after Knisel and Baird (1971)) to calculate the volume of groundwater. The results indicate that, over approximately 53 ha, the net recharge to the groundwater system is 18·6 mm, or about 1·6% of precipitation. Dutton *et al.* (1994), working on similar lithologies to the north of the study area, reported a recharge to the weathered bedrock of 24·4 mm a year, or approximately 3% of precipitation. They computed recharge by summing the fluxes from a groundwater model output divided by the area of the land surface.

Test well response and recharge: FLOWTUBE

A simple two-dimensional cross-sectional groundwater model, FLOWTUBE, was used as a second test of the appropriateness of the chosen aquifer recharge and conductivity values compared with the monitored well heights at the site after work by Christen *et al.* (2000) and George and Conacher (1993). Dawes *et al.* (2000) described the FLOWTUBE model in detail. The model underlying the FLOWTUBE program is based on a finite difference solution to the one-dimensional Darcy law for saturated flow in a semi-confined aquifer. The model assumes a no-flux boundary at the uphill end of the flowtube and an outlet boundary condition where groundwater intersects the soil surface or where a stream is present, representing a constant piezometric head. The model runs over a selected time step. Recharge is chosen to represent long-term values. The model assumes a three-layer system, with a conducting aquifer underlain by a nonconducting layer and overlain by a semi-confining upper layer. These boundary conditions appeared suitable for the site and were thought to be an appropriate test for the assumed recharge values to the bedrock.

The FLOWTUBE model was calibrated by comparing the computed steady-state piezometer heads with the seasonal average water table in the monitored watershed using recharge estimates derived from contouring

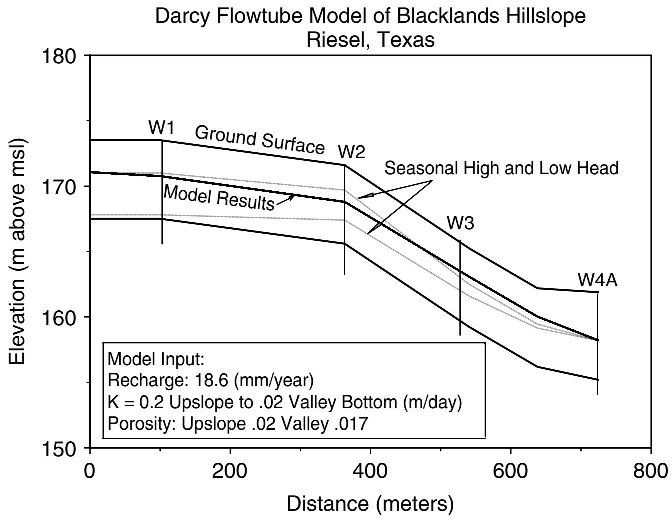

Figure 9. FLOWTUBE model results for hillslope segment

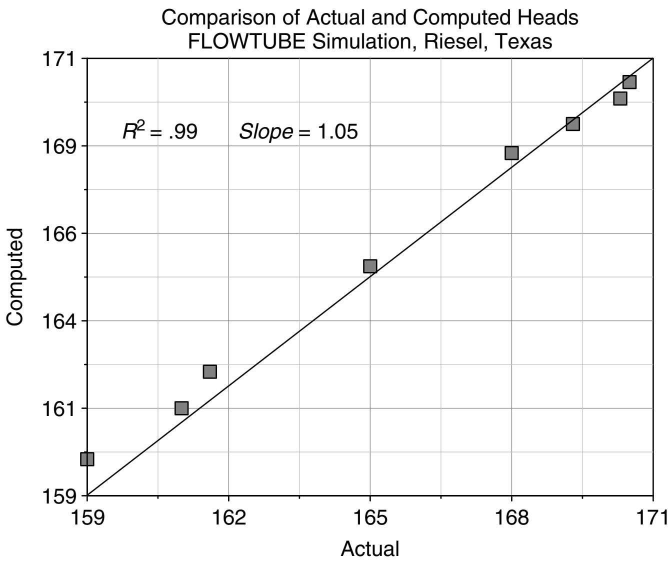

Figure 10. Actual head versus predicted head for FLOWTUBE model

the seasonal well levels, and assuming aquifer properties based on local tests (Figures 9 and 10). Recharge of 18·6 mm was derived from the seasonal change in well heights as described. Minor differences in hillslope and valley-bottom permeability and porosity were used to reflect observed changes in weathering of the units.

DISCUSSION OF RESULTS

Water budget

Well response to precipitation is a function of growing season, evapotranspiration (ET), soil crack volume, depth and hillslope position. Arnold *et al*. (2005) have demonstrated that crack volume at this site is a product of soil moisture status, vegetation type/land use, and soil depth.

The hypothesized seasonal recharge cycle is summarized in Figures 5 and 11. Of the 1151 mm of rainfall, total surface runoff measured at the lower weir accounted for 292 mm, or about 25·3% of total rainfall. Long-term annual surface runoff for the watershed averages 21%. Lateral soil flow accounts for about 154 mm of rainfall. It is estimated that 30 mm of this water reaches the surface as seeps during late winter and early spring and enters the stream. This was determined from the baseflow separation of total stream flow recorded at the weir. The remainder of lateral soil flow either migrates downslope and is absorbed by the soils within the valley or is evaporated from the ground surface prior to arriving at the stream channel. Lateral flow quantities and rates are consistent with field tests in similar soils and bedrock by Ritchie *et al*. (1972). They found saturated hydraulic conductivities averaged 2·5 to 3·5 cm day^{-1}. They observed uniformity of K throughout the 175 cm profile. During their testing, the water table did not rise appreciably, indicating water was moving laterally through the soil. High values of lateral flow were also observed by McKay *et al*. (1993) in a highly weathered and fractured clayey till. They measured lateral groundwater velocities of as much as 24 m day^{-1} from field experiments. ET accounts for the remaining 65–75% of the annual water budget. Soil storage is considered negligible on an annual basis, except in very wet years (such as the study period), in which case some storage can be carried over into the second year.

The small amount of groundwater furnished to the stream from shale terrain is supported by analysis of shale watershed response in north central Texas. For 10 US Geological Survey gauging stations encompassing shale and limestone watersheds in north central Texas, shale watersheds were found to have about two times less baseflow than comparably sized limestone watersheds (Figure 12). This was accomplished by separation of storm flow from baseflow from daily stream flow using filter methods after Arnold and Allen (1999). Baseflow days derived from the same analysis are the number of days it takes the hydrograph discharge to fall one log cycle. In the case of shale terrain, the above analysis found this period to be less than 4–5 days. This indicates that most of the water discharged to the stream in this terrain is probably coming from the soil zone. The small quantity of water from the groundwater system that reaches the valley bottom is thought to be taken up directly by the valley-bottom vegetation or flows slowly down valley.

Results of FLOWTUBE model runs are consistent with regional MODFLOW results reported by Dutton *et al*. (1994). They calculated a regional recharge rate for chalk and shale terrain of 2·4 cm year^{-1} (less than 1% of precipitation) by summing the fluxes into the model and dividing by the land surface area. Of note is their observation that 99% of the groundwater flows through the upper weathered zone. This is consistent with the findings of the FLOWTUBE model and water balance results in this study. The calculated fracture porosity with depth, coupled with the low overall fracture porosity, is consistent with the large changes in water table with short-duration precipitation events. The fact that water table flux on the divides and interfluve zones was greater than the valley bottom suggests that the fractures in this zone are more extensive.

Descriptive model of watershed dynamics

Phase I. Mid-summer to fall (Figure 11). At the dry condition, the surface soils are extensively cracked to depths of more than 4·6 m. Maximum crack volume approaches 120 mm (over the watershed), with over

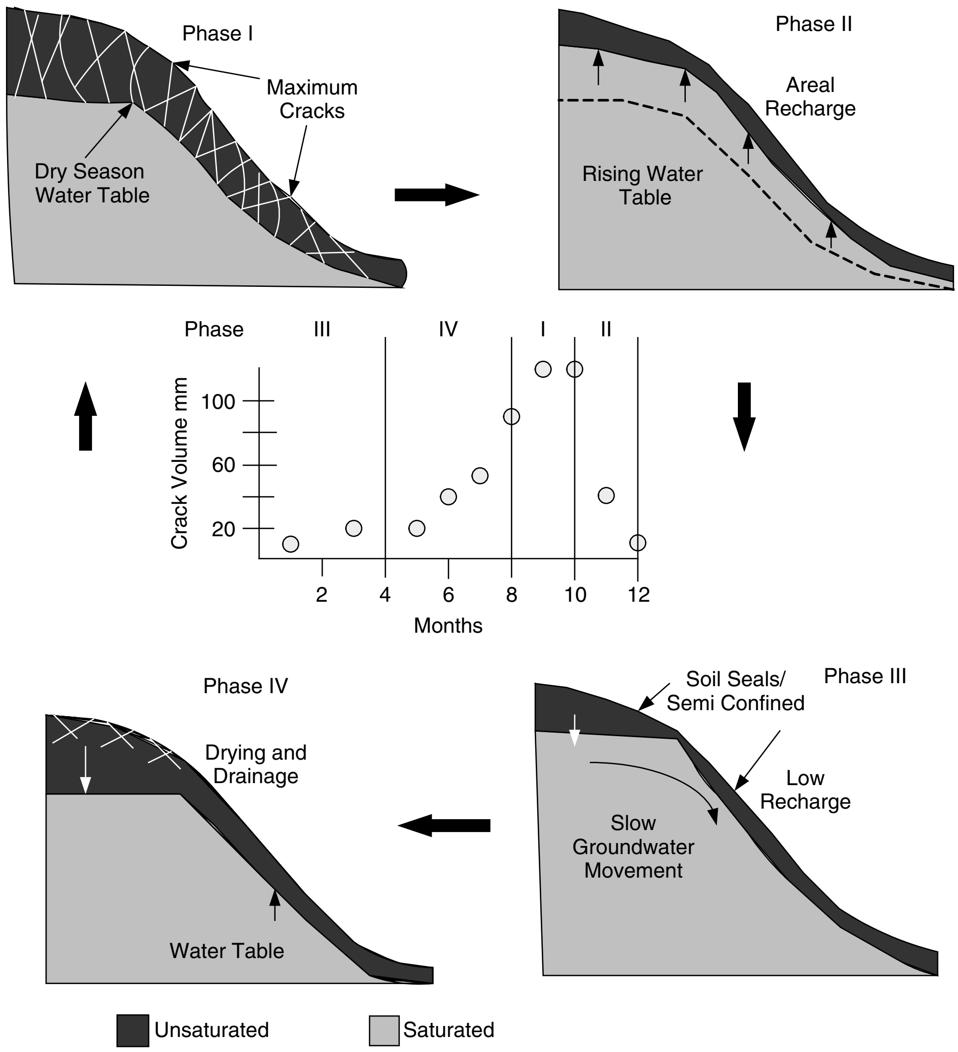

Figure 11. Seasonal phases of soil–aquifer interrelationships for study area

90% of the crack volume occurring in the upper 1·2 m (Arnold *et al.*, 2005). Crack depth is deepest on the divides and shallowest on the valley bottoms. The depth of cracks is thought to be related to capillary fringe and soil structure (Gillham, 1984). Observations in other local clay–shale watersheds support this assumption. Fall storms coupled with a slow reduction in ET lead to an increase in soil moisture and closing of surface shrinkage cracks. From dry conditions, it is estimated that available water capacity on the interfluve and upper hillslope is greater than 124 mm based on the average precipitation recorded at the watershed prior to runoff generation over a 3 year time period. Well levels are at their seasonal low during this phase. Throughout the fall, storms increase and rainfall begins to infiltrate the dry, cracked soils. Infiltrated water flows down the cracks between the peds through the A to the C horizons, with matrix wetting occurring in the surface soil and in the C horizons. Depending on rainfall intensity, wetting occurs from the lower to upper horizons (Blake *et al.*, 1973). Soil moisture readings and the rapid water table response of interfluve and upper hillslope wells support this assumption. Crack volume decreases rapidly within the soil profile, with crack closure occurring to a depth of approximately 2·4 m over a 50 day time period.

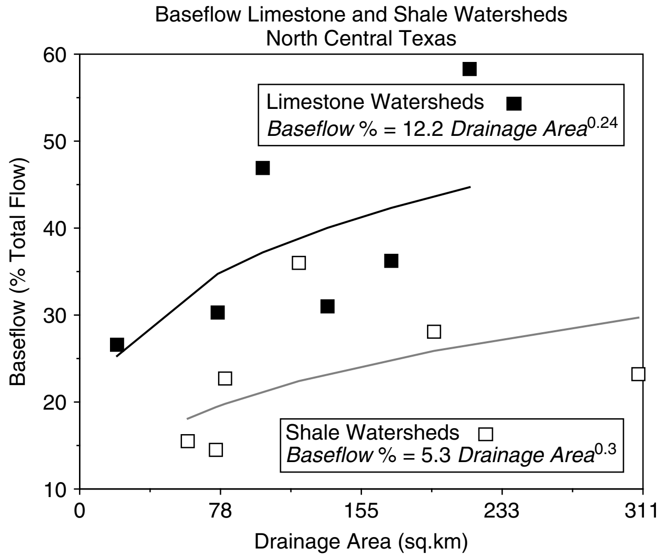

Figure 12. Baseflow (as percentage of total flow) for bedrock-controlled streams in central Texas

Visual observations, as well as work by Favre *et al.* (1997), show that surface cracks close first. Rapid recharge is related to bypass flow. Even when the soil surface approaches field capacity, the remaining cracks are sufficient to support flow rates of 2·5 cm day^{-1} around the soil structural units (Ritchie *et al.*, 1972).

Phase II. Late fall to winter (Figures 11 and 13). With the crack volume reduced due to soil expansion, surface runoff and lateral soil flow commence. Soil moisture has now reached field capacity. Peak runoff and peak lateral flow occur with soil moisture at or above field capacity during this phase (Figures 5–7). All wells respond to recharge during this period, with the highest response occurring on the interfluve and upper hillslope sites (Figure 8). All wells continue to rise during this period, which lasts around 150 days. Approximately 597 mm of rain fell during this recharge phase. Volumetric differences in contoured water table heights before and after recharge indicate a total flux in groundwater in the watershed of 18·6 mm. ET for this period is estimated to approach 287 mm based on Penman calculations (Dugas and Ainsworth, 1983).

Phase III. Late winter to late spring (Figure 11). From winter, with low ET (2·9–3·6 mm day^{-1}) and continued rain, soil cracks are almost totally closed at depth. Even with crack closure, soil flow continues preferentially through the closed macropore system, as shown in (Figure 7) and supported by studies by Innoue (1993). The well heights at the interfluve well (well 1s) begin to fall as water moves laterally downslope (Figure 11). This is a product of the saturated and 'sealed' soil zone, as well as the horizontal/vertical anisotropies inherent with the shale (horizontal flow greater than vertical flow). Lateral groundwater flow is also noted by increased well response in the upper hillslope well (well 2) as the water moves down gradient (Figure 8).

Phase IV. Late spring to summer. From early spring through to summer, despite increases in rainfall, ET continues to dominate the system; potential ET rates increase from 3·9 mm day^{-1} in February to 10 mm day^{-1} in July. Soil storage decreases below field capacity, runoff ceases, cracking increases, and lateral flow is substantially reduced (Figure 11). Downslope water movement continues with corresponding increase in water level in the lower hillslope (Figure 8). As the summer progresses, soil moisture continues to be depleted below the wilting point (Figure 5), and well heights decrease over the entire watershed (Figure 8). Surface runoff and lateral soil flow slow and then cease (Figures 6 and 7).

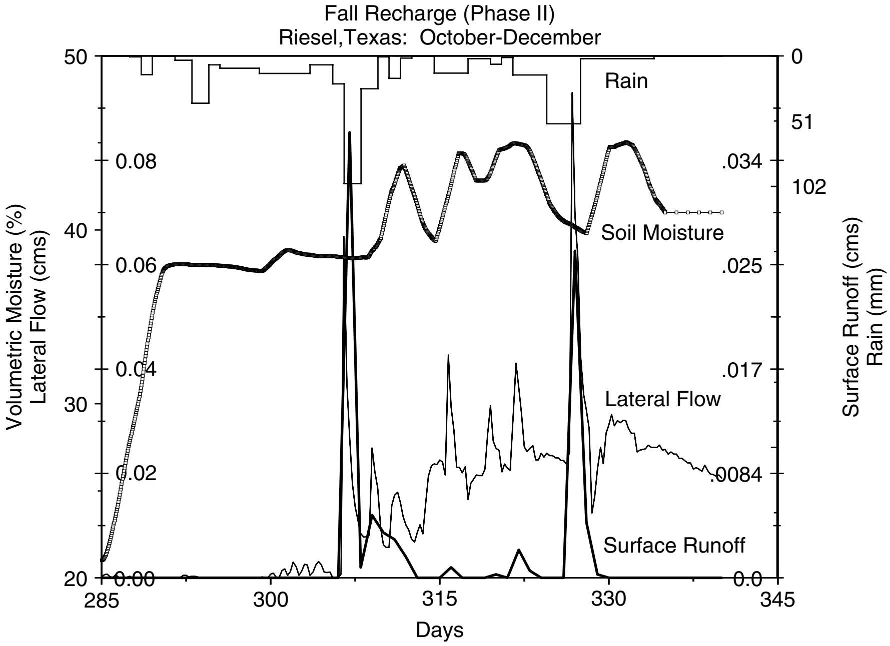

Figure 13. Phase II: fall recharge and rainfall, soil moisture, surface runoff and lateral flow

CONCLUSIONS

The water budget at the Riesel study area is controlled by a complex interaction between a vertisol soil layer, underlying fractured rock, land use, topography, and seasonal trends in rainfall and ET. These relationships are summarized in four phases that reflect how each system interacts seasonally. The vertisol soil moisture regime, controlled by land use, climate and topography, is viewed as the primary regulator of recharge, surface runoff and soil flow within this watershed and terrain. Cracks in the more weathered uplands allow rapid infiltration into the soil. Recharge through the cracks allows rapid and relatively deep wetting of an otherwise impermeable clay. When the soil storage reaches capacity, both surface runoff and lateral soil flow increase. In terms of total discharge, surface flow predominates, with lateral soil flow making up most of the remaining discharge. The persistence of the lateral flow component during the year is an important finding and is different from past reports by Favre *et al.* (1997), who reported the cessation of bypass flow in hours after rainfall. The linkage of cracks, bypass flow, and lateral soil flow demands more research in clay shale terrain. Groundwater flow appears to make up a small fraction of the annual discharge to the stream system. However, the groundwater system is still important in the mechanics of watershed response to precipitation in its apparent effects on the soil flow system. The slow downslope movement of the groundwater in the bedrock appears to keep the valley bottom moist during the summer months. This has a direct impact on deep-rooted vegetation in the valley bottom and may have an impact on watershed response to precipitation. The complexity of recharge in such fractured terrain has been noted by Haria *et al.* (2003). They suggest that the depth to the water table is a major parameter in determining well response to precipitation; when the water table is less than 4 m below the ground surface, they observed both preferential flow and matrix flow. Their studies indicated approximately 0·1 to 0·2% of applied pesticides could reach the shallow groundwater system. This study supports those findings.

Seepage or lateral soil flow is a dominant component of the water budget in fractured terrain and will be important in assessing the NO_3-N concentrations in such watersheds (Kelly and Pomes, 1998). Although dissolved contaminants should be retarded relative to the water velocity by diffusion into the clay, McKay *et al*. (1993) suggest that transport through the fractures carrying colloidal-sized contaminants, such as viruses, should be evaluated, as they could be transported more rapidly. In addition, the results suggest that the small fracture porosity could cause even small spills of immiscible liquids to spread over large areas and contaminate tens of cubic metres of fractured clay. The results of this study indicate significant amounts of pollution could reach the valley bottom and might be taken up by deep-rooted plants or shallow water wells.

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