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Vadose Zone Journal

Climate Variability Controls on Unsaturated Water and Chemical Movement, High Plains Aquifer, USA

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Responses in the vadose zone and groundwater to interannual, interdecadal, and multidecadal climate variability have important implications for groundwater resource sustainability, yet they are poorly documented and not well understood in most aquifers of the USA. This investigation systematically examines the role of interannual to multidecadal climate variability on groundwater levels, deep infiltration (3–23 m) events, and downward displacement (>1 m) of chloride and nitrate reservoirs in thick (15–50 m) vadose zones across the regionally extensive High Plains aquifer. Such vadose zone responses are unexpected across much of the aquifer given a priori that unsaturated total-potential profiles indicate upward water movement from the water table toward the root zone, mean annual potential evapotranspiration exceeds mean annual precipitation, and millennia-scale evapoconcentration results in substantial vadose zone chloride and nitrate reservoirs. Using singular spectrum analysis (SSA) to reconstruct precipitation and groundwater level time-series components, variability was identified in all time series as partially coincident with known climate cycles, such as the Pacific Decadal Oscillation (PDO) (10–25 yr) and the El Niño/Southern Oscillation (ENSO) (2–6 yr). Using these lag-correlated hydrologic time series, a new method is demonstrated to estimate climate-varying unsaturated water flux. The results suggest the importance of interannual to interdecadal climate variability on water-flux estimation in thick vadose zones and provide better understanding of the climate-induced transients responsible for the observed deep infiltration and chemical-mobilization events. Based on these results, we discuss implications for climate-related sustainability of the High Plains aquifer.

Abbreviations: AMO, Atlantic Multidecadal Oscillation; ANN, annual climate variability; CHP, central High Plains; ENSO, El Niño/Southern Oscillation; HDP, heat-dissipation probe; NAMS, North American Monsoon System; NAO, North Atlantic Oscillation; NHP, northern High Plains; PDO, Pacific Decadal Oscillation; RC, reconstructed component; SSA, singular spectrum analysis; SHP, southern High Plains; SST, sea-surface temperature; >PDO, climate cycles with periodicities greater than PDO.

Subsurface hydrologic response to natural climate variability is of particular interest in semiarid and arid regions, such as the western USA, where groundwater resource availability and sustainability are important (Hanson et al., 2004). The response of soil ecosystems in these regions to climate and precipitation variability, including carbon and nutrient biogeochemical cycling and timing of annual biological activity, is relatively well understood (Austin et al., 2004). However, knowledge of subsoil (vadose zone at depths >2 m below land surface) hydrodynamic responses to climate variability in these regions is limited because of a general lack of field observations over timescales long enough to document infiltration and percolation of infrequent recharge events. As such, a strong reliance on modeling approaches driven by stochastic parameterization of climatic forcings (Small, 2005) has prevailed to explore subsurface responses to precipitation variability. Subsurface processes of particular interest in these regions include deep percolation events leading to spatially diffuse or focused groundwater recharge (Small, 2005) and possible mobilization of large subsoil chemical reservoirs of evapoconcentrated pore-water Cl^- and NO_3^- (Hartsough et al., 2001; Walvoord et al., 2003, 2004; McMahon et al., 2006) that accumulated throughout the Holocene in response to a shift to a drier climate and an associated shift in vegetation (Phillips, 1994).

Although some climatic extremes, such as intense precipitation events, are dominantly episodic, there are low-frequency controls that modulate climate and associated precipitation and drought frequency on a variety of timescales (Dettinger et al., 1998). Recent studies in the western and southwestern USA have identified quasiperiodic variations in hydrologic time series of precipitation, groundwater levels, and streamflow that appear to reflect a wide range of quasiperiodic climate forcings on timescales ranging from days to multiple decades (Hanson et al., 2004; Gurdak and Hanson, 2005; Hanson and Dettinger, 2005; Hanson et al., 2006). Such quasiperiodic hydrologic variations represent teleconnections in response to recurrent and persistent global climatic variability (Hanson and Dettinger, 2005). The magnitude and phase relation of interannual, interdecadal, and multidecadal climate variability can result in extreme climatic forcings on these sensitive hydrologic systems. Understanding these climatic variations and correlations between different hydrologic time series can improve predictions of hydrologic system responses to future climate cycles (Ghil et al., 2002). Yet

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the hydroclimatic affects of interannual to multidecadal climate variability on most vadose zones and groundwater systems, such as the High Plains aquifer or other aquifers of the USA, are not well understood. Climate variability on these temporal scales is likely to have substantial impacts on groundwater systems located in arid and semiarid regions.

This paper presents an investigation of hydrologic and geochemical responses in thick vadose zones as well as fluctuations in groundwater levels to natural climate variability across the regionally extensive High Plains aquifer. The specific objectives of this paper are twofold. First, field evidence was used to document relatively deep infiltration and chemical (Cl^- and NO_3^-) displacement within thick (>15 m) vadose zones and to identify the link to likely episodic precipitation events from climatic variations. The observations were recorded using a vadose zone monitoring network established in 2000 by the USGS as part of the National Water-Quality Assessment (NAWQA) Program (McMahon et al., 2003, 2006; Bruce et al., 2005). Second, a novel application of spectral time-series analysis was used to quantify climatic variations on vadose zone water flux and transit times using lag correlations between fluctuations in precipitation time series and corresponding groundwater levels. From this analysis, we infer natural climatic forcings responsible for the observed deep infiltration and chemical displacement in the vadose zone of the High Plains aquifer.

Site Description

The High Plains aquifer underlies an area of about 450,000 km^2 in parts of eight western states (Fig. 1) of the Great Plains physiographic province. The study area is subdivided into the southern (SHP), central (CHP), and northern High Plains (NHP) subregions. In 2000 water use from this regional aquifer represented approximately 30% of all groundwater pumped in the USA (Maupin and Barber, 2005), and it is threatened by continued declines in water levels and deteriorating water quality (Dennehy et al., 2002).

Natural Climate Variability

The High Plains is characterized by a middle-latitude dry continental climate with temperatures ranging from 6 to 17°C, moderate amounts of precipitation (406–711 mm), frequent winds, low humidity, and a high rate of evaporation (1525–2768 mm) (Dennehy et al., 2002). The climate of the High Plains is known to contain several quasiperiodic modes of natural variability across a range of temporal scales, including annual (ANN) (<2 yr), interannual, interdecadal, and multidecadal climate variability (Gurdak and Hanson, 2005; McCabe et al., 2004). At least some of these modes of variability may manifest in a variety of climatic, hydrologic, and ecological impacts, most notably in the frequency and duration of enhanced precipitation (wet periods) and drought (dry periods). The interannual to multidecadal variations and resulting hydroclimatic impacts generally lack spatial uniformity across the High Plains (Gurdak and Hanson, 2005).

The interannual to multidecadal hydroclimatic variations of the High Plains are related to teleconnections from variability in oceanic and atmospheric conditions, which are often indexed using sea-surface temperature (SST) or atmospheric pressure anomalies (Gurdak and Hanson, 2005; McCabe et al., 2004; Woodhouse and Overpeck, 1998). The interannual climatic variations of the High Plains include cycles of 2 to 6 yr resulting from the well-studied ENSO (Wolter and Timlin, 1993, 1998). The definition and effects of interdecadal variations on the order of 10 to 25 yr were identified by Mantua and Hare (2002) as related to the PDO and have been identified using spectral analysis of hydrologic time series and tree-ring reconstructions across the High Plains (Gurdak and Hanson, 2005). The positive (warm) phase of ENSO and PDO variability is generally associated with warmer and wetter conditions for much of the western USA, while the negative (cold) phase of variability is associated with periods of cooler and drier climatic conditions (Mantua and Hare, 2002; Wolter and Timlin, 1993, 1998). Periods of climate variability greater than 25 yr also have been identified and are referred to as greater than PDO (>PDO)

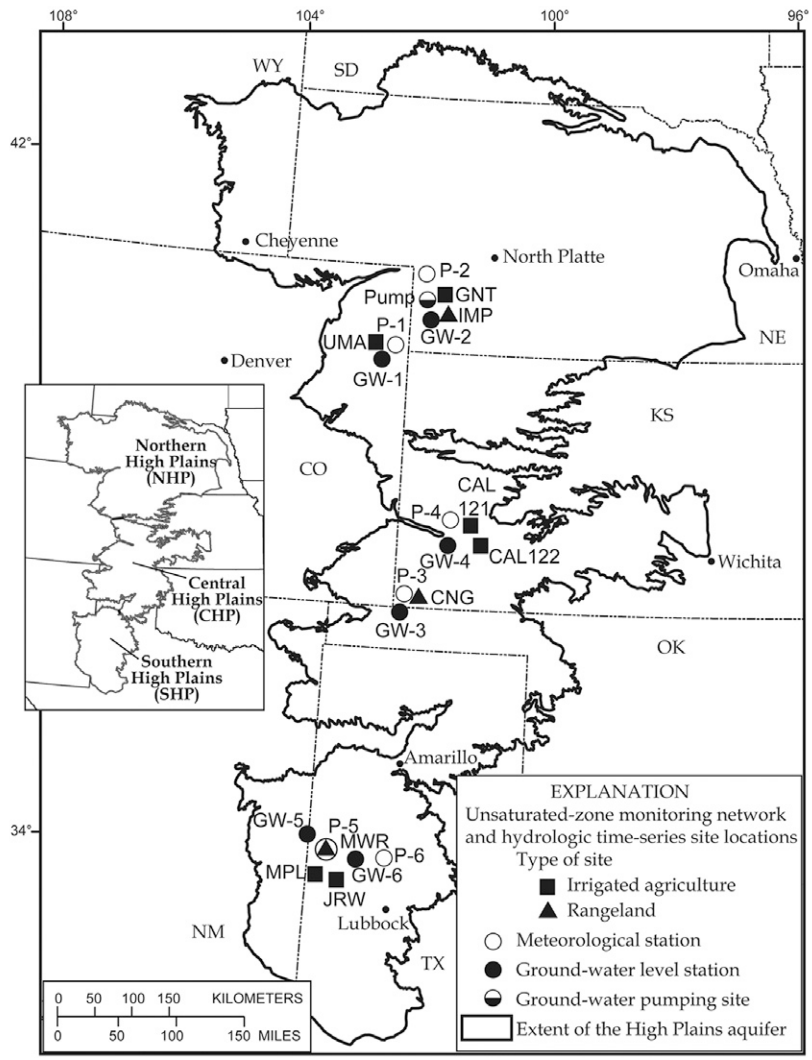

FIG. 1. Map showing the location and extent of the High Plains aquifer, unsaturated zone monitoring network sites, and selected groundwater level (GW), pumping, and meteorological (P) stations.

(Hanson et al., 2004; 2006). In addition, other cycles are estimated in hydrologic time series that fall between the ENSO and PDO cycles with variations on the order of 6 to 10 yr. These periodicities may be related to periodic persistence of northerly movement of subtropical moisture from the Gulf of Mexico or from the Pacific and may be related to the North American Monsoon System (NAMS). The specific cause of variations in precipitation with periodicities of 6 to 10 yr that may be linked to processes such as the NAMS remains uncertain (Hanson et al., 2004). For example, Fye et al. (2006) found strong statistical evidence from tree-ring reconstructions that a 7- to 8-yr periodicity in precipitation variability may originate from the North Atlantic Oscillation (NAO). However, Hanson et al. (2004) observed variability in June to September atmospheric water-vapor fluxes associated with the monsoon in the American Southwest on the 6- to 10-yr timescale. They concluded that monsoonal moisture flow from the tropical latitudes up through the Gulf of California may contribute to persistent wet or dry periods with periodicities of 6 to 10 yr in the southwestern USA. The >PDO cycles may be attributed to longer PDO fluctuations with periodicities of 50 to 70 yr (Minobe, 1997) or variations in the Atlantic Multidecadal Oscillation (AMO) (50–80 yr) (Kerr, 2000); yet the sources and stationarity of these forcings remains uncertain. The AMO is represented by an index of detrended SST anomalies averaged over the North Atlantic and has recently been identified as having a strong influence on summer rainfall and drought frequency, and may modulate teleconnections between ENSO and winter precipitation over the conterminous USA (McCabe et al., 2004). McCabe et al. (2004) suggested the current (2004) positive AMO conditions may lead to more above-normal frequencies of drought in the USA. Since 1995 AMO has been positive, preceded by negative (cold) phases during 1905–1925 and 1970–1990 and positive (warm) phases during 1860–1880 and 1930–1960 (McCabe et al., 2004).

Vadose Zone

In the High Plains aquifer, median vadose zone thicknesses range from about 15 to 40 m for rangeland and from 25 to 60 m for irrigated cropland. The vadose zone consists of unconsolidated clay, silt, sand, and gravel, with scattered cemented zones consisting of calcium carbonate and sometimes silica.

Water Movement

Previous studies of water movement in the vadose zone of the High Plains aquifer indicate that the direction and rate of water movement are controlled by land use (Scanlon et al., 2005), spatial patterns in climate (McMahon et al., 2006), geomorphology (Wood and Sanford, 1995; Scanlon and Goldsmith, 1997; Fryar et al., 2001), vegetation, and soils (Keese et al., 2005). Generally, previous observations of vadose zone water movement at the USGS vadose zone network within irrigated agricultural and rangeland settings of the NHP indicate the potential for downward water movement within the vadose zone, with little seasonal change that is consistent with gravity-driven quasi-steady flow below the root zone (McMahon et al., 2006). In contrast, rangeland of the SHP (MWR, Fig. 1) has total potentials that increased substantially with depth (McMahon et al., 2006), indicating the potential for upward water movement from the water table to the root zone, which is consistent with interplaya observations by Scanlon and Goldsmith (1997). Across the vadose zone monitoring network (Fig. 1), estimated downward water fluxes ranged from 0.2 to 111 mm yr^{-1} (McMahon et al., 2006). Irrigated agricultural vadose zone sites had larger fluxes (17–111 mm yr^{-1}) than rangeland sites (0.2–70 mm yr^{-1}). The largest water fluxes were observed at the NHP vadose zone sites (70 to 111 mm yr^{-1}) followed by CHP sites (5 to 54 mm yr^{-1}) and SHP sites (0.2–32 mm yr^{-1}), due in part to spatial climatic differences from north to south with lower evapotranspiration rates in the NHP than in the SHP. McMahon et al. (2006) suggested that the measured downward water flux (0.2 mm yr^{-1}) at the SHP rangeland site (MWR) represents past hydrologic conditions because upward hydraulic gradients were observed during the study.

Chemical Reservoirs

The relatively thick vadose zone overlying the High Plains aquifer, like many other thick vadose zones in semiarid and arid regions, contain pore-water reservoirs with large Cl^- and NO_3^- concentrations that generally follow a conservative solute-type profile due to natural evapoconcentration over thousands of years (Phillips, 1994; Walvoord et al., 2003) and from anthropogenic N input, primarily from agricultural fertilizers (McMahon et al., 2003, 2006). Walvoord et al. (2003) inferred that natural sources of NO_3^-, like Cl^-, can be mobilized from the surface soil to the subsoil and beyond the root zone during occasional deep-wetting events. Strong hydraulic gradients created by the plant roots and subsequent upward movement of water vapor is the primary process that concentrates natural Cl^- and NO_3^- in pore water of vadose zone. Cl^- and NO_3^- concentration profiles that increase with depth just below the root zone (Walvoord et al., 2003, 2004; Seyfried et al., 2005). These profiles characteristically display a bulge of high concentrations in the upper 10 m, with peak concentrations observed between depths of 2 and 5 m (Walvoord et al., 2004).

Groundwater quality in the High Plains aquifer is potentially vulnerable to these natural and anthropogenic Cl^- and NO_3^- reservoirs in the vadose zone (Gurdak and Qi, 2006; Gurdak et al., 2007). This groundwater vulnerability may be realized if other processes mobilize and transport the Cl^- and NO_3^- reservoirs to the water table; such as conversion of rangeland to irrigated and rain-fed agricultural land (McMahon et al., 2006; Scanlon et al., 2005). McMahon et al. (2006) attributed the downward displacement of some NO_3^- to the mobilization of naturally accumulated NO_3^- by irrigation return flow after rangeland was converted to irrigated agricultural land.

Materials and Methods

Beginning in 2000, a network of nine vadose zone monitoring sites (UMA, GNT, IMP, CAL121, CAL122, CNG, JRW, MPL, and MWR) were instrumented across the High Plains aquifer (Fig. 1) to assess processes and rates of water movement and the storage and transit times of chemicals (McMahon et al., 2003, 2006). Within each of the three subregions (NHP, CHP, and SHP), two vadose zone sites were located in irrigated agricultural settings and one site was located in rangeland (Table 1). Details about the installation and capabilities of this network are described by McMahon et al. (2003, 2006).

Vadose Zone Data Collection and Analysis

During installation of each vadose zone monitoring site, sediment cores and cuttings were collected for analyses of volumetric water content and water-extractable concentrations of Cl^- and NO_3^-. Concentrations of deionized-water extractable Cl^- and NO_3^- were quantified by ion chromatography (McMahon et al., 2003), with reporting limits of 0.8 $\mu g\ g^{-1}$ (as Cl^-) and 0.2 $\mu g\ g^{-1}$ (as N). To identify changes in the vadose zone chemical profiles resulting from deep infiltration events during 2003 and 2004, as described below, selected sites were cored in 2005 for additional analyses of water-extractable concentrations of Cl^- and NO_3^-. The selected sites were limited to the CHP and SHP because Cl^- and NO_3^- vadose zone inventories are much larger compared with those of the NHP (McMahon et al., 2006), which provides a stronger signal of possible chemical mobilization. All 2005 boreholes were located <0.5 m from original boreholes, which assumes no substantial spatial variability of vadose zone concentration profiles for Cl^- and NO_3^- within the 0.5-m distance. However, the maximum depths from which cores were collected in 2005 were substantially less (ranging 3.5–12 m below land surface) than during the original site installation (2000–2002). Cemented layers within the vadose zone and different drilling equipment limited the depth of the boreholes collected in 2005. It is unlikely that the drilling disturbed the 2000 or 2005 chemical profiles.

Table 1. Unsaturated zone (UZ) monitoring network and hydrologic time-series site information.

Site†	Site ID	Time-series network	Period of record
Northern High Plains (NHP)			
UMA	400443102383701	UZ	2002–2005
GNT	404427101421601	UZ	2002–2005
IMP	403713101334401	UZ	2002–2005
GW-1	400115102285700	groundwater levels	1967–2003
GW-2	403220101384001	groundwater levels	1944–2003
P-1	59243	precipitation	1890–2003
P-2	254110	precipitation	1890–2003
Pump	23518	groundwater pumping irrigation well	1980–2002
Central High Plains (CHP)			
CAL121	374625100490701	UZ	2002–2005
CAL122	374412100453201	UZ	2002–2005
CNG	370349101431001	UZ	2002–2005
GW-3	364450102190001	groundwater levels	1938–1995
GW-4	374100101270501	groundwater levels	1943–1998
P-3	340908	precipitation	1925–2003
P-4	144464	precipitation	1889–2003
Southern High Plains (SHP)			
JRW	334043102365501	UZ	2002–2005
MPL	334905102545001	UZ	2002–2005
MWR	335830102444201	UZ	2002–2005
GW-5	342006103134201	groundwater levels	1954–2004
GW-6	341010102240801	groundwater levels	1951–1997
P-5	416135	precipitation	1921–2005
P-6	417079	precipitation	1889–2003

†GW, groundwater station; P, meteorological station.

Heat-dissipation probes (HDPs; Model 229 Water Matric Potential Sensor, Campbell Scientific, Logan, UT) were installed at various depths at each site during original installation (McMahon et al., 2006) and provided an indirect measure of matric potential in the vadose zone. Unlike the NHP and SHP sites, the CHP sites lack HDPs within the upper 23 m of the vadose zone. The HDPs are capable of measuring the matric potential from approximately −0.01 to −100 MPa with a sensitivity that is proportional to the matric potential (Flint et al., 2002). Details about HDP installation and measurement capability are described by McMahon et al. (2006).

At the SHP sites, a 10-m aluminum neutron moisture-meter access tube (10 cm i.d.) was installed approximately 3 m from the vadose zone monitoring sites by using dry-drilling technology. These access tubes allowed for detailed measurements of volumetric water-content profiles of the upper 30 m using neutron moisture meter (Model 503 Hydroprobe Moisture Depth Gauge, CPN International [CPN, 1980], Pacheco, CA, with ^{241}Am-Be, 50 mCi source and boron trifluoride [BF3] detector tube). The neutron moisture meter was calibrated for volumetric water content, using field samples collected from SHP sites and methods described by CPN (1980) and Hignett and Evett (2002). Calibration equations were developed relating neutron counts linearly to volumetric water content of vadose zone profiles and were verified used the independent volumetric water contents as measured from sediment cores collected during site installation. Volumetric water-content profiles were measured at the SHP sites before and after the infiltration events described below as an independent means of testing the temporal changes in vadose zone moisture profiles as measured by the HDPs.

Hydroclimate Data Processing and Analysis

Frequency analysis of hydrologic time series allows for inference and better understanding of subsurface hydrologic responses to climate variability near the vadose zone monitoring network. The hydrologic time-series data processing and analysis generally follow methods described by Hanson et al. (2004).

The time series evaluated in this paper include groundwater levels from wells, precipitation data from meteorological stations, and a groundwater pumping record from an irrigation well (Table 1) that were compiled from a larger network of hydrologic time series in the High Plains (Gurdak and Hanson, 2005). These data were selected on the basis of the location of paired groundwater level wells and precipitation stations to individual sites of the vadose zone network within each subregion of the High Plains aquifer (Fig. 1), and the length and completeness of record. The water levels (1938–2006) were obtained from USGS National Water Information System (USGS, 2006). Monthly precipitation time series (1889–2005) were obtained from the NOAA (2006b). The short record length (<5 yr) of the matric potential time series (Fig. 2) measured by HDPs prevented direct evaluation of these data using the frequency analysis techniques described below.

The first data-processing steps (Fig. 3) were designed to remove the influence of zero values common to precipitation time series of semiarid and arid climates and to integrate any incomplete groundwater-level records. This step was performed by transforming all hydrologic time series into monthly cumulative departure series from the period of record using the monthly mean (Fig. 3b). Next, the residuals of the monthly cumulative departure series were obtained by subtracting a regression-fitted low-order (cubic) polynomial. The overall shape of the low-order polynomial represents temporal trends (or responses)

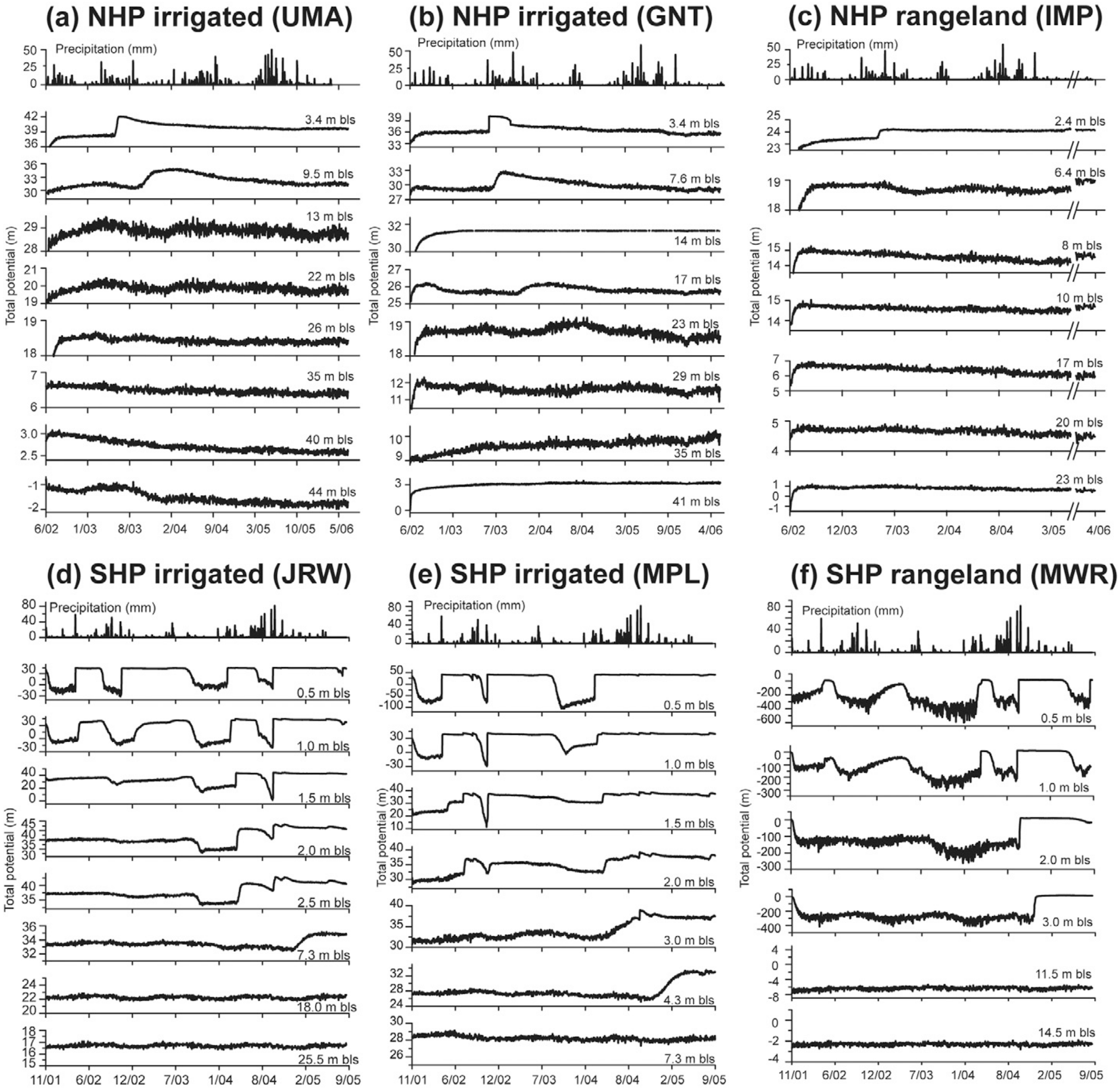

FIG. 2. Daily time series (2001–2006) of precipitation and total (matric plus gravimetric) potential for selected depths below land surface (bls) at (a) northern High Plains (NHP) irrigated site (UMA); (b) NHP irrigated site (GNT); NHP rangeland site (IMP); (d) southern High Plains (SHP) irrigated site (JRW); (e) SHP irrigated site (MPL); and (f) SHP rangeland site (MWR). Daily time series of precipitation are shown for meteorological stations (a) P-1; (b,c) P-2; (d,e) P-6; and (f) P-5.

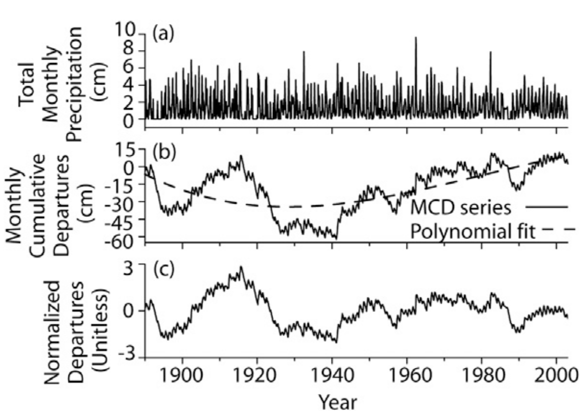

FIG. 3. Demonstration of data-processing steps using (a) total monthly precipitation from meteorological station P-1. The original series is transformed into a (b) monthly cumulative departure (MCD) series and is fitted with a low-order, cubic polynomial fit. The resulting residuals are normalized by the historic mean and referred to as the normalized departure series (c). Note the different *y*-axes for (b) and (c).

in the hydrologic time series to larger climatic cycles or periods of anthropogenic effects, and tends to be dominated by the lowest frequency containing the most variance (Hanson et al., 2004). The residuals are finally normalized by the historic mean to facilitate statistical comparisons between various data types and are referred to as normalized departures (unitless) (Fig. 3c). These data-processing steps effectively eliminate parts of the lowest-frequency cycles that would dominate the variance of the time-series analysis (Hanson et al., 2004). These steps are necessary to remove much of the long-term anthropogenic effects in the groundwater-level time series, such as implementation of improved irrigation technology, and annual anthropogenic effects, such as crop rotation and other irrigated-agricultural practices of the High Plains region.

Singular spectrum analysis (Vautard et al., 1992) was then applied during frequency analysis of the time-series residuals, using a modified approach developed by Hanson et al. (2004) for regional hydroclimatic assessment of the southwestern USA. The SSA method is a widely used form of principal-component analysis in lag-time domain that uses a data-adaptive signal-to-noise enhancement to detect periodic signals in short, noisy time series (Vautard et al., 1992). The data-adaptive filters help separate the time series into reconstructed components (RCs) that are statistically independent and thus can be classified into trends, oscillatory patterns, and noise (Ghil et al., 2002). Singular spectrum analysis has been shown to be optimal in the sense of capturing the maximum variance with the fewest independent RCs (Shun and Duffy, 1999). The variability in most hydrologic time series can be adequately described in terms of the first 10 RC and often takes the form of quasiperiodic or near-sinusoidal oscillations (Hanson et al., 2004). Additionally, the strength of the linear relationship between RCs of similar periodicities from two distinct hydrologic time series was quantified using lag correlation coefficients. Because the RCs have uniform time intervals from the cumulative monthly departure series, the phase lags were calculated as the average time interval between all lagged phase shifts. Conceptually, the calculated phase lags represent an average best-fit of the dependent time series to the independent time series over all observed phase shifts. The goal here is to estimate periodic or nearly periodic variance in RCs for hydrologic time series to evaluate the significance of detected oscillations from underlying noise. Thus, these estimates provide insight into natural climate variability responsible for observed and statistically significant dynamic response of vadose zone water movement, chemical displacement, and groundwater-level fluctuations to the High Plains aquifer.

Results and Discussion

Water Movement in the Vadose Zone

Daily time series of total (matric plus gravimetric) potential profiles at the NHP and SHP vadose zone monitoring sites are shown in Fig. 2. In 2003 and 2004, respective NHP and SHP total potential profiles indicate relatively sharp- and uniform-wetting fronts that reached depths previously unobserved during the period of monitoring (2001–2005). The propagation of these wetting fronts indicate predominantly pistonlike or matrix flow that reach depths ranging from 6.4 to 23 m in the NHP and 3.0 to 7.3 m in the SHP (Fig. 2). Before these two events, temporal fluctuations in total potential showed little change and was limited to the fluctuation zone of the upper 1 m at the SHP rangeland site (MWR) and upper 2 m at the SHP irrigated agricultural sites (JRW and MPL). The fluctuations in total potential within the upper 1 to 2 m are largely controlled by infiltration and subsequent evapotranspiration. The near-surface profiles at NHP sites indicate no appreciable fluctuations in total potential, except for the 2003 infiltration event.

Generally, the depths of propagation of the wetting fronts are greater at the irrigated agricultural sites (4.3 at MPL to 23 m at GNT) than the corresponding rangeland sites (3.0 at MWR to 6.4 m at IMP) (Fig. 2). These differences are likely related to site-specific antecedent moisture and potential conditions, which are controlled in part by native vegetation, land-use practices (crop vegetation), soil texture, lithology, and spatial patterns in climate. However, the onset of the wetting fronts at the three NHP sites and the three SHP sites are coincident to relatively dry conditions preceding intense and frequent precipitation in May 2003 (NHP) and throughout much of 2004 (SHP) (Fig. 2). Although no wetting fronts reached the water table, the current (2006) depth of the wetting fronts below the zone of fluctuation (>2–3 m below land surface) may indicate eventual recharge to the aquifer at the sites. In 2003 and 2004, similar deep infiltration and actual recharge likely occurred in areas of the NHP and SHP with smaller (<23 and <7.3 m, respectively) vadose zone thickness. In the CHP, HDPs were not installed within the upper 23 m that would be necessary to record any near-surface temporal fluctuations in total potential; thus, wetting fronts were not measured. However, substantial changes in the chemical profiles from 2000 to 2005 indicate that a partial flushing event likely occurred, discussed below.

Selected total-potential profiles on days before the infiltration events (Fig. 4) (NHP: 1 Jan. 2003; and SHP: 1 Jan. 2004) show a monotonic decrease in total potential with depth, indicating a downward gradient at all sites (Fig. 4a–4e) except the SHP rangeland site (MWR) (Fig. 4f). The 1 Jan. 2004 profile at MWR shows an increase in total potentials with depth, indicating the potential for upward water movement from the water table to the root zone. McMahon et al. (2006) previously attributed the spatial differences in total-potential profiles of the SHP sites to the conversion of rangeland to irrigated cropland, indicating the importance of land conversion and irrigation practices on recharge across the High Plains aquifer. The climatic events responsible for the deep infiltration events of 2003 and 2004 have similar, but transient, effects on total potential profiles in the SHP. Selected total-potential profiles during the infiltration events (NHP: 1 July 2003; and SHP: 1 Feb. 2005) show a slight decrease in total potential with depth at all sites (Fig. 4a–4f), including MWR, which indicates the potential for downward water movement to the deep vadose zone.

The independent measurements of volumetric water-content profiles at the SHP rangeland site (MWR site) using the neutron moisture meter between 2001 and 2002 were compared with the total potential profile measured in 2005 (Fig. 5). These independent measurements are evidence of the changes in vadose zone water storage. The 2005 neuron moisture-meter profile reveals a substantial increase in the volumetric water content profile, particularly within the fractured caliche zone between the 3-m and 11.5-m HDPs. This independently mea-

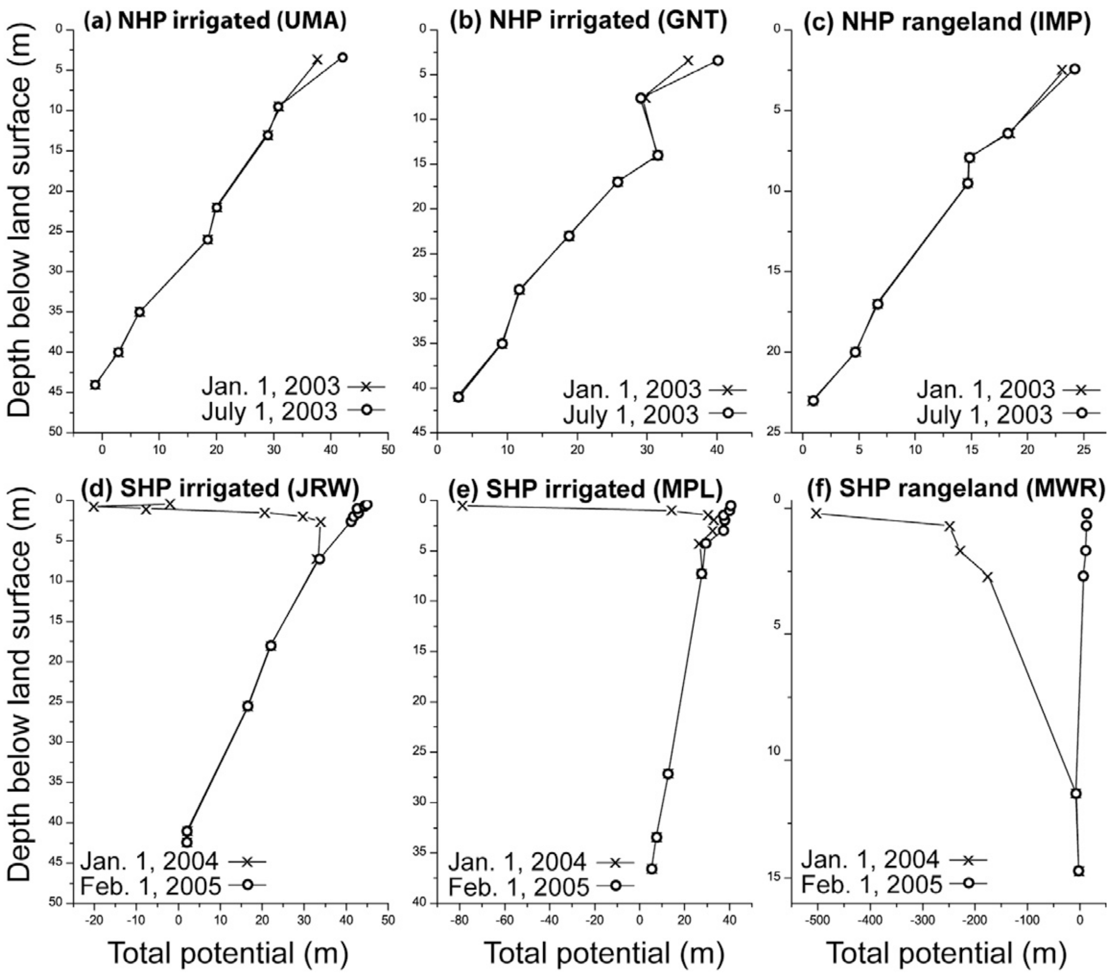

FIG. 4. Comparison of quasi-steady state total-potentials profiles for selected days before infiltration event (black crosses) and during infiltration events (white circles) at northern High Plains (NHP) irrigated sites (a,b) and rangeland site (c), and southern High Plains (SHP) irrigated sites (d,e) and rangeland site (f). These profiles extend to the water table (note different *y*-axis scales) and represent only 2 d from the complete record of total-potential time series shown in Fig. 2.

sured profile in 2005 represents a 4.9% depth-averaged increase in volumetric water content, which is equivalent to a 75-mm increase in unsaturated zone water storage.

Downward Movement of Chemicals in the Vadose Zone

The constructed Cl^- and NO_3^- concentration profiles in vadose zone pore water from cores collected in 2000 were compared with profiles collected in 2005 after the infiltration events (Fig. 6). At CNG, the Cl^- and NO_3^- profiles show that the Cl^- and NO_3^- peak concentrations moved downward about 1.1 m from 2000 to 2005 (Fig. 6a). This displacement is attributed to advective transport caused by the deep infiltration events of 2003 and 2004. The 1.1-m displaced 2005 Cl^- peak has maintained the general shape and relative concentrations as in 2000, which can be attributed to the conservative nature of Cl^- relatively homogeneous sand of the vadose zone material at CNG that would enable uniform, matrix flow during infiltration. However, the total NO_3^- inventory in 2005 (450 μg N g^{-1}) was approximately 50% smaller than in 2000 (910 μg N g^{-1}) (Fig. 6a). The available data does not support a clear explanation; however, the loss of NO_3^- in 2005 may be attributed to an incomplete flushing or denitrification of the NO_3^- measured in 2000. Similar evidence was documented for displacement of chemicals at the SHP sites (Fig. 6b–6d). Some of the 2005 pro-

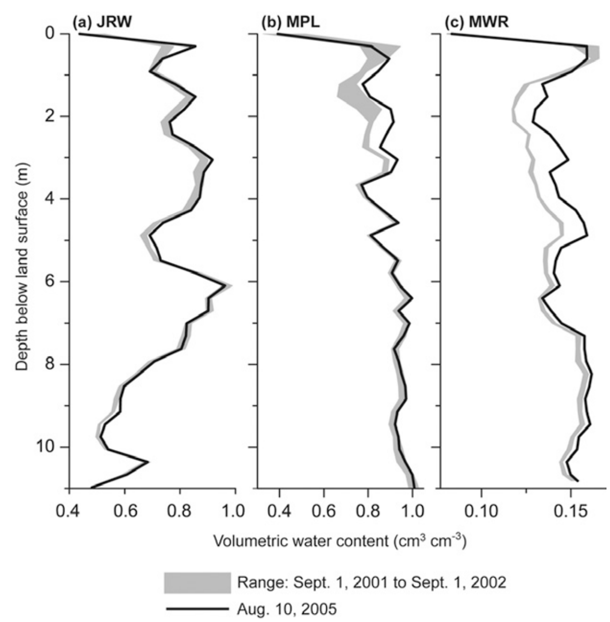

FIG. 5. Volumetric water-content profiles beneath the southern High Plains (SHP) irrigated sites JRW and MPL (a,b) and rangeland site MWR (c) that were obtained using neutron moisture meter. These profiles indicate 2005 wetting event reached depths that verify measurements from heat dissipation probes (HDPs), especially between 3 and 11 m below land surface at MWR site (HDPs were not located between 3 and 11 m at MWR).

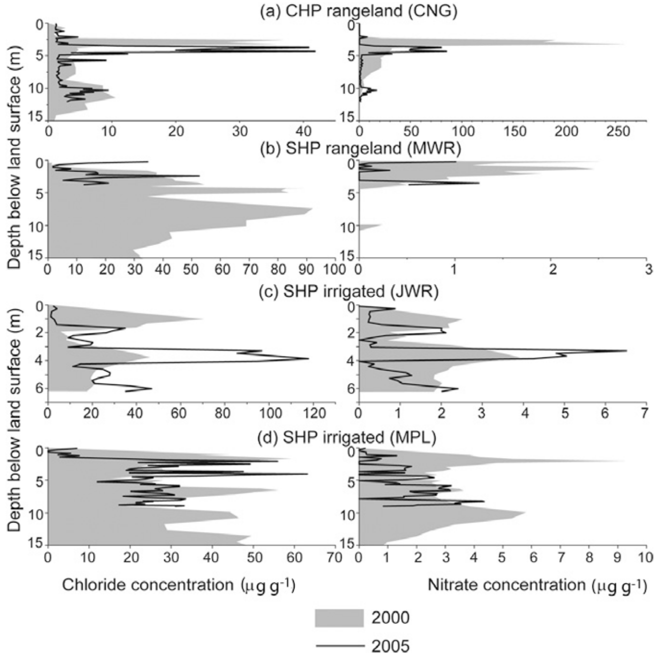

FIG. 6. Comparison of profiles for chloride and nitrate concentrations in the vadose zone at central High Plains (CHP) rangeland and southern High Plains (SHP) rangeland and irrigated sites, 2000 and 2005.

files are too shallow to resolve temporal differences. However, the profiles at JWR indicate that the uppermost Cl^- and NO_3^- peaks also moved downward approximately 1 m. Additionally, the profiles indicate that the majority of the chemical reservoir has shifted down approximately 3 m (Fig. 6c). Although there are some temporal differences in Cl^- profiles at MPL (Fig. 6d), the data do not indicate a distinct displacement, which may be related to the fractured-caliche lithology present near land surface at MPL. The overall shape of the CHP and SHP chemical profiles, including the depth of Cl^- and NO_3^- peaks, developed from natural evapoconcentration processes over thousands of years (Phillips, 1994; Hartsough et al., 2001; Walvoord et al., 2003). Therefore, the apparent >1 m downward and uniform displacement of both Cl^- and NO_3^- profiles at depths below the zone of total potential fluctuation at the rangeland vadose zone sites within a 1- to 5-yr period indicates the significance of this mobilization event.

Annual Precipitation Variability

Time series of total annual precipitation (Fig. 7) indicate that the onset of the deep infiltration events occurred during a relatively wet year (2004) at the SHP sites and, conversely, during a relatively dry year (2003) at the NHP sites. The total annual precipitation at site P-5 (MWR) in 2004 (899 mm) was more than double the long-term (1921–2004) mean (430 mm), making it the second-wettest year on record, behind 1941 (1105 mm) (Fig. 7b). The unusually large total annual precipitation observed at P-5 (MWR) in 2004 was not a spatially isolated occurrence (NOAA, 2006b). Much of the SHP received approximately 150% of the long-term (1951–2001) mean annual precipitation in 2004; similar trends extended across much of the CHP, which received as much as 140% of the long-term (1951–2001) mean annual precipitation in 2004. Total annual precipitation at the NHP site was less (300 mm) than the long-term (1900–2005) average (437 mm) during the 2003 onset of the deep infiltration at the NHP sites (Fig. 7a) and was only slightly greater than the long-term average in 2004.

Grissino-Mayer's (1996) reconstruction of the precipitation records for the past 2200 yr using tree-ring data located near the SHP in northern New Mexico provides perspective on the 2004 SHP precipitation totals and corresponding vadose zone chemical profile displacement. Grissino-Mayer (1996) suggested that the last 200 yr have been the wettest period for the past 2200 yr, with the previous 20 yr representing the wettest period of the entire 2200-yr record. The amount of precipitation in 2004 across the SHP and CHP, relative to the paleorecord of precipitation of the region, offers plausibility of the infrequent nature of the substantial (>1 m) infiltration event that mobilized chemicals under natural, rangeland settings in the CHP and SHP. Identifying subsurface hydrologic response to natural climate variability is a first step toward predicting possible changes in the frequency of such episodic events in response to global climate change. A number of predictions by general circulation models suggest the future climate of the High Plains will be warmer and drier, while others project a warmer and wetter greenhouse-effect climate in the High Plains. Yet there is a general agreement of an increase in the occurrence of extreme events, such a continued trend toward more high-intensity rainfall events (Merideth, 2001; Rosenberg et al., 1999; Woodhouse and Overpeck, 1998). General circulation model projections of increased precipitation intensity could have substantial implications for the frequency of downward displacement of solutes, increased water fluxes, and the future groundwater sustainability and quality of the High Plains aquifer.

Seasonally, the onset of deep infiltration events at NHP and SHP sites occurred under different precipitation regimes (Fig. 8), indicating that regional differences exist across the High Plains aquifer; thus, recharge will be variable to climate variability. The onset of the deep infiltration at the SHP vadose zone sites coincides with particularly large total monthly precipitation from June to December 2004, especially the near record-

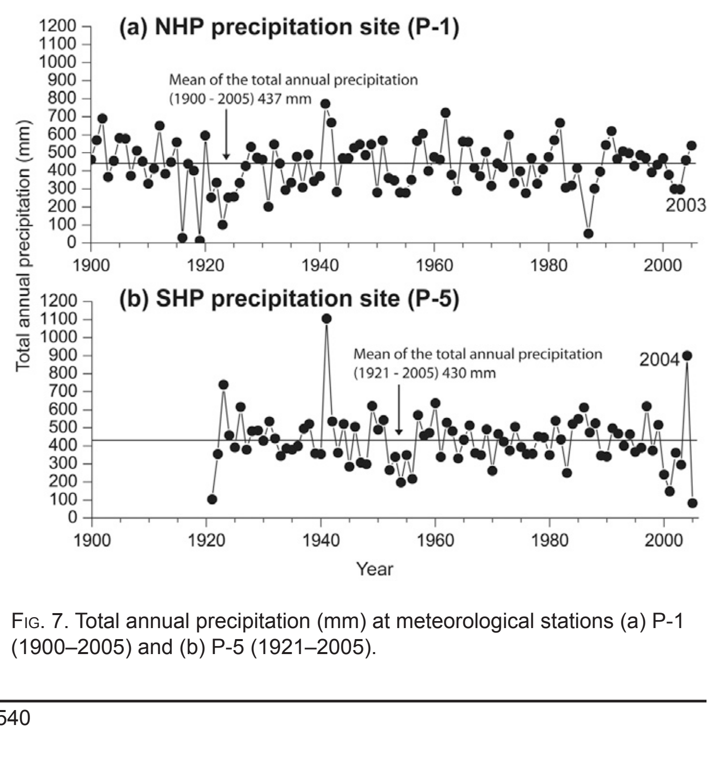

FIG. 7. Total annual precipitation (mm) at meteorological stations (a) P-1 (1900–2005) and (b) P-5 (1921–2005).

setting precipitation during November 2004 (Fig. 8b). The total monthly precipitation at P-5 was 350% of average for November, receiving 250 to 300 mm of wet snow on 2 Nov. 2004, followed by another substantial rain–snow event (50–100 mm as rain) 13–17 Nov. 2004 (NOAA, 2006a). NOAA (2006a) reported similar record precipitation accumulations across the SHP for all of November 2004: Lubbock, TX, was 937% of average, and Amarillo, TX, was 597% of average. In contrast, the apparent onset of the deep infiltration events in the NHP coincided with monthly totals of February to April that were only between the 50th and 75th percentiles (Fig. 8a) but followed 2 yr of below-average precipitation (Fig. 7a). The observed deeper infiltration to climate variability in the NHP and likely higher water-use efficiency of the SHP vegetation, which is manifested in the strong downward total potential gradient at the SHP sites (Fig. 4), indicates a relatively greater potential for deep infiltration and recharge at the NHP sites compared with the SHP sites.

Interannual to Interdecadal Climate Variability

The results of the SSA indicate that all hydrologic time series contain variations in the important interannual to interdecadal climatic cycles for the western USA (>PDO, PDO, NAMS, and ENSO) (Table 2).

Precipitation Correlation to Climate Variability

The majority of the variance in the precipitation time series was attributed to >PDO (>25 yr) and PDO periods (10–25 yr), capturing 51.4 to 64.2% and 4.52 to 78.4% of the variance, respectively (Table 2). Approximately 68 to 93% of the total variance in the RCs for each precipitation time series is attributed to the >PDO and PDO periods. These results indicate the importance of lower-frequency climatic forcings, with lesser controls by NAMS, ENSO, and ANN on precipitation variabil-

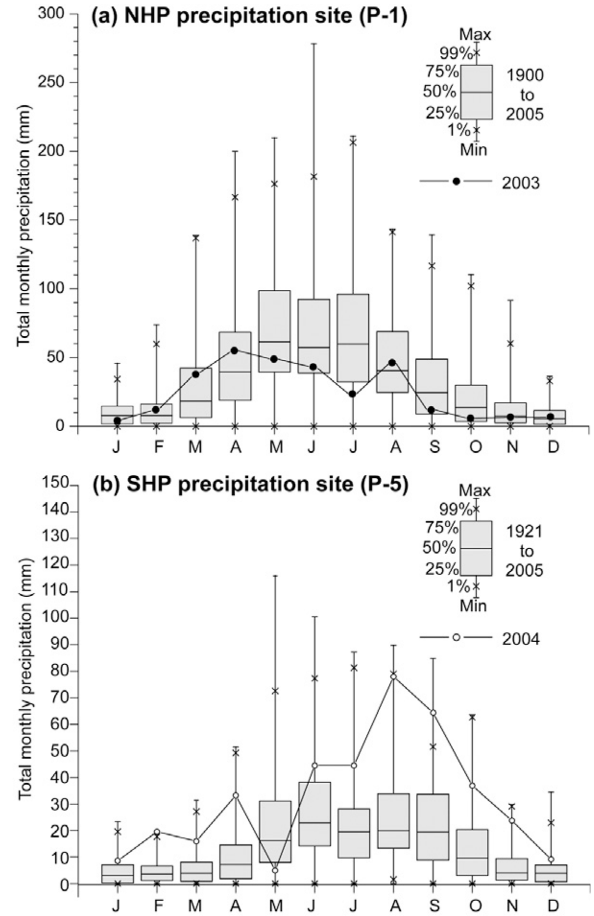

Fig. 8. Distribution of total monthly precipitation (mm) at meteorological stations (a) P-1 (1900–2005), highlighting monthly precipitation for 2003, and (b) P-5 (1921–2005), highlighting monthly precipitation for 2004.

Table 2. Summary of frequencies and variance for groundwater levels and precipitation time series at groundwater level (GW), pumping (Pump), and meteorological (P) stations.

Reconstructed component	Period (variance)												
	Northern High Plains					Central High Plains				Southern High Plains			
	P-1	GW-1	P-2	GW-2	Pump	P-3	GW-3	P-4	GW-4	P-5	GW-5	P-6	GW-6
	yr (%)												
1	38.0†	12.8‡	32.6†	20.8‡	6.7§	20.8‡	11.1‡	28.8†	18.5‡	21.1‡	20.8‡	38.3†	18.5‡
	(64.2)	(67.0)	(51.4)	(64.2)	(48.8)	(53.6)	(35.3)	(51.6)	(23.3)	(40.0)	(78.4)	(53.5)	(73.8)
2	22.8‡	6.4§	16.3‡	11.1‡	3.5¶	15.2‡	8.3§	17.7‡	6.9§	13.0‡	11.1‡	19.2‡	11.9‡
	(24.5)	(27.3)	(28.1)	(22.8)	(37.2)	(29.9)	(32.0)	(31.0)	(12.6)	(28.6)	(17.2)	(28.9)	(19.8)
3	10.4‡	3.3¶	9.9§	5.4¶	2.0¶	7.2§	4.3¶	10.5‡	4.5¶	7.0§	4.6¶	11.0§	4.4#
	(4.52)	(4.04)	(7.78)	(4.04)	(10.5)	(6.79)	(13.2)	(8.35)	(9.23)	(10.7)	(1.70)	(6.77)	(1.71)
4	7.6§	2.4¶	7.1§	3.6¶	1.4#	5.2¶	3.6¶	7.7§	3.2¶	5.5¶	3.4¶	7.7§	3.2¶
	(1.75)	(1.23)	(2.76)	(2.64)	(2.71)	(1.91)	(9.12)	(1.85)	(7.23)	(4.89)	(0.82)	(2.09)	(0.62)
5	5.7¶	1.8#	5.7¶	2.8¶	1.1#	1.0#	2.7¶	1.0#	2.6¶	1.0#	2.4¶	6.1§	2.3¶
	(0.84)	(0.28)	(1.75)	(2.39)	(0.61)	(1.36)	(2.86)	(1.11)	(5.46)	(2.37)	(0.42)	(1.65)	(0.43)
6	1.0#	1.5#	1.0#	2.3¶	0.9#	1.0#	2.2¶	1.0#	1.0#	1.0#	2.0¶	4.9¶	1.9#
	(0.71)	(0.11)	(1.69)	(1.55)	(0.14)	(1.34)	(1.35)	(1.10)	(4.78)	(2.22)	(0.34)	(1.26)	(0.35)
7	1.0#	1.2#	1.0#	2.0¶	0.7#	3.8¶	2.0¶	5.9¶	1.0#	3.3¶	1.6#	4.1¶	1.0#
	(0.70)	(0.04)	(1.41)	(0.46)	(0.03)	(1.04)	(1.11)	(0.80)	(4.74)	(2.08)	(0.25)	(0.83)	(0.19)
8	4.6¶	1.0#	4.5¶	1.7#	0.6#	3.1¶	1.3#	4.7¶	2.0¶	3.1¶	1.4#	1.0#	1.0#
	(0.60)	(<0.01)	(1.00)	(0.27)	(<0.01)	(0.84)	(0.75)	(0.51)	(4.14)	(1.90)	(0.19)	(0.79)	(0.18)
9	3.7¶	0.9#	3.8¶	1.1#	0.6#	2.6¶	1.5#	2.6¶	1.7#	2.5¶	1.2#	1.0#	0.7#
	(0.40)	(<0.01)	(0.84)	(0.26)	(<0.01)	(0.67)	(0.73)	(0.48)	(3.81)	(1.71)	(0.14)	(0.71)	(0.17)
10	3.2¶	0.8#	3.2¶	1.1#	0.5#	2.2¶	1.7#	2.6¶	1.5#	2.3¶	1.1#	3.6¶	0.7#
	(0.32)	(<0.01)	(0.56)	(0.25)	(<0.01)	(0.42)	(0.55)	(0.41)	(3.45)	(1.09)	(0.09)	(0.51)	(0.17)

† Cycles with periodicities greater than Pacific Decadal Oscillation (>PDO) (>25 yr).
‡ Pacific Decadal Oscillation (PDO) cycles (10–25 yr).
§ North American Monsoon System (NAMS) cycles (6–10 yr).
¶ El Niño/Southern Oscillation (ENSO) cycles (2–6 yr).
# Annual climate variability (ANN) (<2 yr).

Table 3. Summary of lag correlation coefficients and phase lags (in years) for hydrologic time-series at selected climate cycles at groundwater level (GW), pumping, and meteorological (P) stations. "—" indicates no reconstructed components of the dependent hydrologic time series were identified within the selected climate cycle.

Dependent hydrologic time series	AMO cycles (50–80 yr)†		PDO cycles (10–25 yr)		NAMS cycles (6–10 yr)		ENSO cycles (2–6 yr)		ANN cycles (<2 yr)	
	Lag correlation	Phase lag (years)	Lag correlation	Phase lag (years)	Lag correlation	Phase lag (years)	Lag correlation	Phase lag (years)	Lag correlation	Phase lag (years)
	(1) AMO‡	(1) AMO‡	(1) PDO§	(1) PDO§	(1) NAMS¶	(1) NAMS¶	(1) MEI#	(1) MEI#		
			(2) P-1	(2) P-1	(2) P-1	(2) P-1	(2) P-1	(2) P-1	(2) P-1	(2) P-1
			(3) P-2	(3) P-2	(3) P-2	(3) P-2	(3) P-2	(3) P-2	(3) P-2	(3) P-2
			(4) P-3	(4) P-3	(4) P-3	(4) P-3	(4) P-3	(4) P-3	(4) P-3	(4) P-3
			(5) P-4	(5) P-4	(5) P-4	(5) P-4	(5) P-4	(5) P-4	(5) P-4	(5) P-4
			(6) P-5	(6) P-5	(6) P-5	(6) P-5	(6) P-5	(6) P-5	(6) P-5	(6) P-5
			(7) P-6	(7) P-6	(7) P-6	(7) P-6	(7) P-6	(7) P-6	(7) P-6	(7) P-6
							(8) Pump	(8) Pump	(8) Pump	(8) Pump
Precipitation										
P-1	(1) −0.30	(1) 5.0	(1) 0.34	(1) 2.6	(1) 0.64	(1) 5.3	(1) 0.30	(1) 3.8		
P-2	(1) −0.54	(1) 21	(1) 0.27	(1) 3.1	(1) 0.70	(1) 3.4	(1) 0.21	(1) 5.0		
P-3	(1) —	(1) —	(1) 0.67	(1) 5.1	(1) 0.52	(1) 1.3	(1) 0.36	(1) 3.8		
P-4	(1) −0.72	(1) 5.8	(1) 0.41	(1) 8.0	(1) 0.74	(1) 2.1	(1) 0.50	(1) 0.25		
P-5	(1) —	(1) —	(1) 0.83	(1) 2.0	(1) 0.75	(1) 2.9	(1) 0.24	(1) 0.60		
P-6	(1) −0.70	(1) 20	(1) 0.82	(1) 2.6	(1) 0.81	(1) 4.2	(1) 0.12	(1) 1.0		
Groundwater levels										
GW-1	(1) —	(1) —	(1) 0.90	(1) 22	(1) 0.38	(1) 2.0	(1) 0.45	(1) 1.0		
			(2) 0.92	(2) 21	(2) 0.20	(2) 1.5	(2) 0.29	(2) 4.6	(2) 0.32	(2) 2.3
GW-2	(1) —	(1) —	(1) 0.82	(1) 27	(1) —	(1) –	(1) 0.38	(1) 0.33		
			(3) 0.67	(3) 22	(3) —	(3) –	(3) 0.24	(3) 0.33	(3) 0.25	(3) 4.7
							(8) 0.66	(8) 4.8	(8) 0.42	(8) 3.5
GW-3	(1) —	(1) —	(1) 0.51	(1) 16	(1) 0.56	(1) 2.8	(1) 0.42	(1) 4.6		
			(4) 0.32	(4) 11	(4) 0.55	(4) 1.9	(4) 0.32	(4) 4.1	(4) 0.02	(4) 4.6
GW-4	(1) —	(1) —	(1) 0.31	(1) 40	(1) 0.57	(1) 4.8	(1) 0.39	(1) 0.5		
			(5) 0.58	(5) 28	(5) 0.01	(5) 2.8	(5) 0.31	(5) 0.25	(5) 0.01	(5) 5.0
GW-5	(1) —	(1) —	(1) 0.93	(1) 50	(1) —	(1) —	(1) 0.36	(1) 0.10		
			(6) 0.87	(6) 46	(6) —	(6)–	(6) 0.20	(6) 0.33	(6) 0.07	(6) 4.5
GW-6	(1) —	(1) —	(1) 0.90	(1) 32	(1) —	(1) —	(1) 0.30	(1) 1.7		
			(7) 0.99	(7) 29	(7) —	(7) —	(7) 0.29	(7) 5.0	(7) 0.19	(7) 5.0
Groundwater pumpage										
Pump	(1) —	(1) —	(1) —	(1) —	(1) 0.78	(1) 1.3	(1) 0.01	(1) 5.1		
			(2) —	(2) —	(2) 0.63	(2) 0.4	(2) 0.38	(2) 1.8	(2) 0.18	(2) 0.92

† AMO, Atlantic Multidecadal Oscillation; PDO, Pacific Decadal Oscillation; NAMS, North American Monsoon System; ENSO, El Niño/Southern Oscillation; ANN, annual climate variability.
‡ Unsmoothed Atlantic sea-surface temperature anomalies north of the equator (1951–2000) (Enfield et al., 2001).
§ Standardized PDO index of monthly sea-surface temperature over the North Pacific Ocean (1900–2006) (Mantua and Hare, 2002).
¶ Index of seasonal precipitation anomalies associated with the North American Monsoon System (1948–2006) (Gutzler, 2004).
# MEI, Multivariate ENSO Index (1950–2006) (Wolter and Timlin, 1993, 1998).

ity across the High Plains. The lag correlation coefficients ranging from −0.30 to −0.72 (Table 3) indicate relatively strong and inverse correlations between the AMO index (>PDO) (Enfield et al., 2001) and precipitation variability across the High Plains, with phase lags ranging from 5.0 to 21 yr (Table 3). Precipitation variability demonstrated stronger correlations to PDO (0.27 to 0.83), with phase lags of 2.0 to 8.0 yr (Table 3), further indicating the strength of correlations between lower-frequency climatic cycles and precipitation variability across the High Plains. The strength and direction of correlation between >PDO and PDO cycles and precipitation variability is consistent with the findings from McCabe et al. (2004), which indicate the majority of variance in drought frequency across the conterminous USA is attributed to positive AMO and negative PDO phases of variability.

Groundwater-Level Correlation to Climate Variability

Although precipitation time series are most influenced by >PDO variability, the variance captured (23.3–93.5%) by the RCs for the groundwater-level time series indicate the importance of PDO-like forcings on groundwater level fluctuations, with lesser influence by NAMS (12.6–32.0%), ENSO (1.70–13.2%), and ANN (<0.01–4.78%) (Table 2). Pacific Decadal Oscillation–like variations in groundwater levels accounted for a majority of the total variance at the NHP (67–87%) and SHP (93.6–95.6%) sites, while accounting for only 23.3 to 35.3% of the variance at the CHP sites. The regional differences in groundwater-level variance attributed to PDO may be a result of spatial patterns in recharge rates, which are indirectly influenced by physical processes related to climatic forcings and to vadose zone lithology and thickness near the wells.

Even though groundwater of the High Plains aquifer is influenced by pumping, the climatic variability that dominantly controls the groundwater levels also were successfully identified. The lag correlation coefficients (Table 3) of groundwater pumping to variability in precipitation and climate indices were mixed, with weak correlations (0.01–0.38) in the ANN and ENSO periodicities and moderate to strong correlations (0.63–0.78) in the NAMS periodicities. The influence of pumping on groundwater-level variability was quantified by moderate lag correlations (0.42–0.66) in the ENSO and ANN periods, with phase lags of 3.5 to 4.8 yr (Table 3). Yet no RCs from the pumping time series were identified within the PDO period, which was previously identified as having control on groundwater-level variability. As verification, the GW-2 series was clipped to the identical period of record as the groundwater pumping

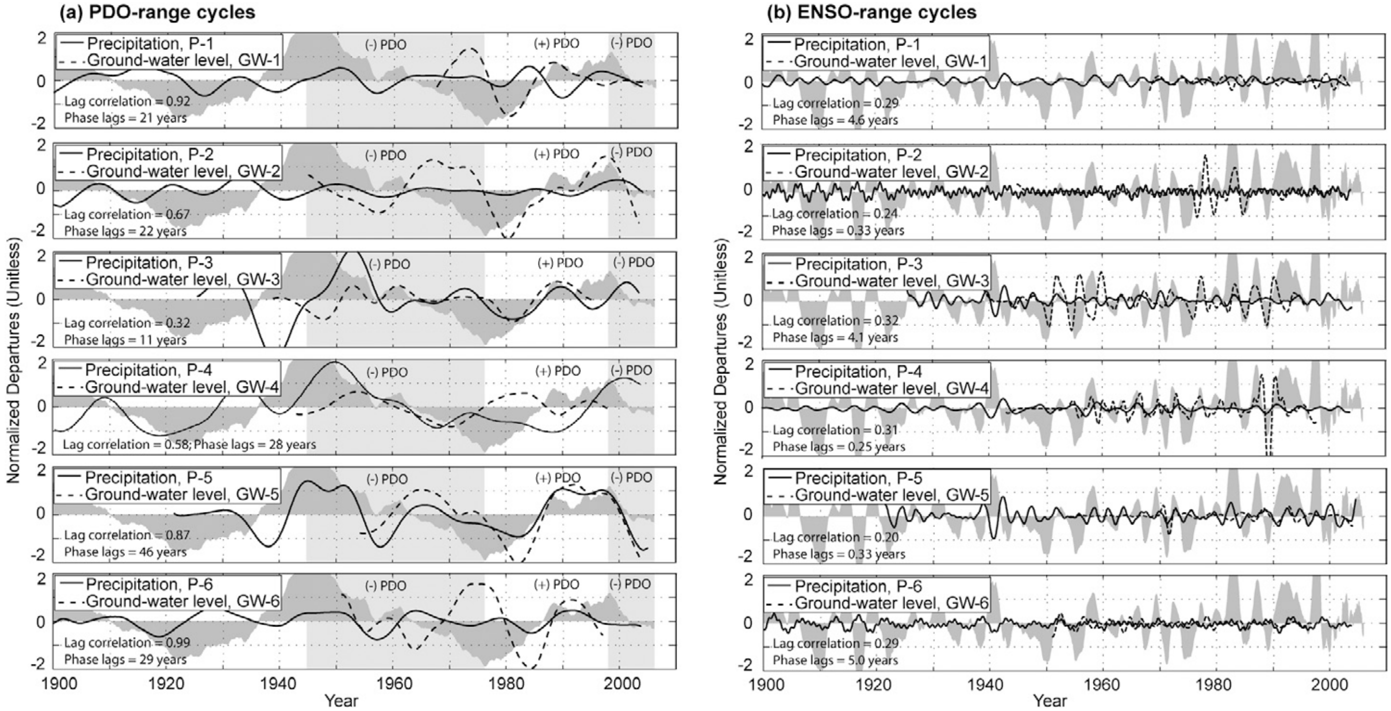

FIG. 9. The reconstructed components for paired precipitation and groundwater-level sites for (a) Pacific Decadal Oscillation (PDO)-range cycles (10–25 yr) and (b) El Niño/Southern Oscillation (ENSO)-range cycles (2–6 yr) are compared with respective PDO (Mantua and Hare, 2002) and multivariate ENSO (Wolter and Timlin, 1993) indices, shown in gray.

record and applied to SSA. The results of SSA identified PDO periodicities within the shortened GW-2 record, which suggests that the pumping record was substantially long enough to record any possible PDO signals. The lack of PDO periodicities within the groundwater pumping records indicates that anthropogenic responses are PDO independent. This suggests that the controlling cycles of groundwater-level variability were based on and driven by PDO-like climatic factors and not substantially influenced by pumping variability.

Additional evidence is demonstrated by comparing the relatively short phase lags (0.4–5.1 yr; Table 3) between pumping responses with precipitation and climate indices, which identifies a characteristically rapid anthropogenic response to wet–dry fluctuations in shorter-term climate variability (ANN, ENSO, and NAMS). Similarly, the phase lags between GW-2 response to pumping in the ENSO and ANN periodicities are small (3.5–4.8 yr; Table 3), suggesting that some portion of the anthropogenic signal remains in the groundwater-level variability within these interannual periods. Conversely, the phase lags between groundwater levels and variability in PDO-like precipitation and the PDO index are significantly longer (11–50 yr; Table 3). Conceptually, the decadal-scale (11–50 yr) phase lags in groundwater levels to precipitation variability represent responses to natural processes, such as vadose zone travel times, and further demonstrate that the majority of the anthropogenic signal has been removed from groundwater-level variability in the PDO period. This example of paired groundwater pumping and groundwater-level response illustrates the challenge in removing the entire anthropogenic signal from groundwater-level fluctuations in heavily pumped aquifers.

Groundwater-Level Correlation to Precipitation

The lag correlations (0.32–0.99; Table 3) for each pair of RCs from corresponding precipitation and groundwater-level time series for the PDO range result in a range of decadal-scale phase lags (11–46 yr; Table 3 and Fig. 9a). Other climatic forcings partially dampen or accentuate the PDO-like signal in groundwater levels out of phase with precipitation variability (Fig. 9a). The lag-correlated precipitation and groundwater levels (Fig. 9a) generally follow previously identified trends associated with negative (dry) PDO indices during the 1950s to 1970s and positive (wet) PDO phase shifts during the 1980s and 1990s (Mantua and Hare, 2002), including the renewed negative (dry) PDO index since 1999 that has been observed in hydrologic time series of the southwestern USA (Hanson et al., 2004; Schmidt and Webb, 2001). The negative PDO shift in 1999 is expressed as drier than normal climate and is specifically apparent in the negative phase shift of the groundwater levels of the NHP (GW-1 and GW-2) and SHP (GW-5) (Fig. 9a).

The lag correlation of groundwater levels with precipitation variability in the ENSO range is relatively weak, with correlation coefficients ranging from 0.20 to 0.32 and phase lags of 0.25 to 5.0 yr (Table 3 and Fig. 9b). Many paired precipitation and groundwater-level responses are in phase or slightly phase-lagged with ENSO (Fig. 9b). However, a number of the paired responses are dampened or accentuated out of phase (Fig. 9b) and may be attributed to groundwater pumping or other climatic forcings. It is interesting to note that the amplitude of the ENSO signal oscillations and corresponding paired precipitation and groundwater-level responses are generally dampened during the negative PDO periods (Fig. 9a, 9b). These dampened amplitudes in the ENSO oscillations, combined with generally larger lag correlation coefficients in the PDO range of variability (Table 3), suggest the importance of PDO controls on vadose zone water movement leading to groundwater-level fluctuations. The lower-frequency controls, especially during PDO negative (dry) phases of variability, apparently dampen the variability in groundwater levels from higher-frequency climatic controls, such as ENSO.

Vadose Zone Water and Chemical Response to Climate Variability

This paper documents empirical evidence that vadose zone chemical reservoirs are also mobilized by infrequent, episodic precipitation events linked to natural quasi-periodic climate variability. The interactions between the >PDO, PDO, NAMS, ENSO, and ANN climate periods are shown as individual RCs (Fig. 10) to further investigate their individual effect on precipitation variability at P-1 in 2003 and at P-5 in 2004 and, in turn, the deep infiltration responses at corresponding NHP and SHP vadose zone sites. The precipitation variability in P-1 in 2003 (Fig. 10b) and in P-5 in 2004 (Fig. 10d) were coincident with a negative >PDO (P-1) phase, a negative PDO phase, a negative NAMS-like phase, and a weak to moderately positive ENSO phase of variation in the latter half of 2004. The results from the lag-correlation analysis (Table 3; Fig. 9a) show that precipitation variability and groundwater levels are particularly sensitive to negative PDO phases. However, the deconstruction of RC for 2003 and 2004 data (Fig. 10b, 10d) suggests that annual variability, not interannual to multidecadal variability, may have forced the hydroclimatic conditions leading to the deep infiltration events. The apparent link between the observed deep infiltration and chemical mobilization events to annual variability has important implications for recharge and aquifer sustainability. The High Plains (Castro et al., 2001) and much of the USA (McCabe et al., 2004) could expect a greater frequency of long-term drought if a sustained period of negative (dry) PDO and positive (dry) AMO conditions occur following their respective 1999 and 1995 phase shifts. The findings from this study illustrate the importance of these longer-frequency climatic variations on long-term groundwater-level fluctuations and aquifer sustainability but also suggest that shorter-frequency climatic events can have substantial controls on vadose zone water and chemical flux to the groundwater of the High Plains aquifer, even during extended dry periods.

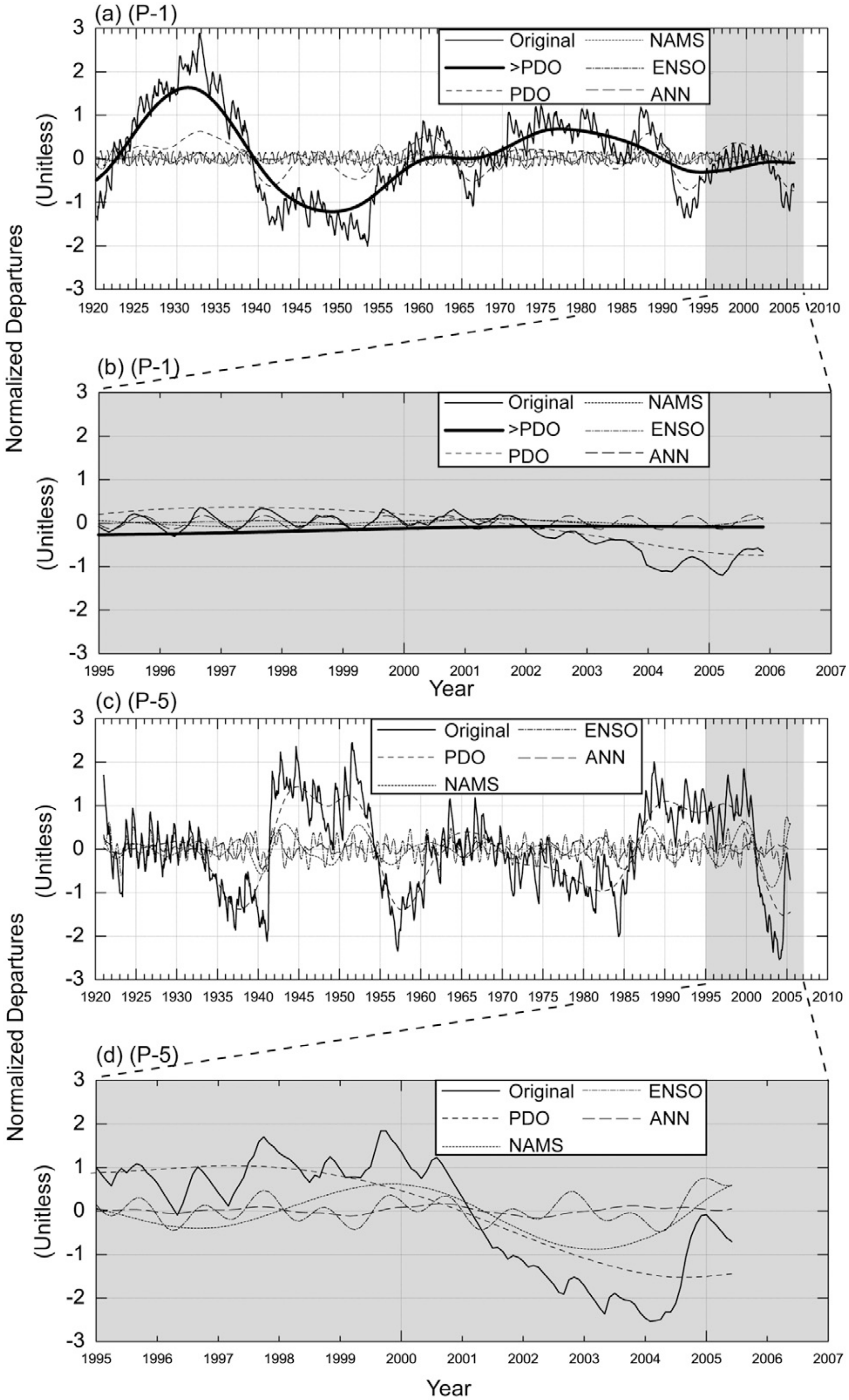

FIG. 10. Reconstructed components (RCs) for precipitation time series; meteorological stations (a,b) P-1 (northern High Plains) and (c,d) P-5 (southern High Plains). (PDO, Pacific Decadal Oscillation; >PDO, cycles with periodicities greater than PDO; NAMS, North American Monsoon System; ENSO, El Niño/Southern Oscillation; ANN, annual climate variability.)

Estimation of Climate Varying Vadose Zone Water Flux

Water levels in aquifers typically vary in response to time-varying rates of recharge (Dickinson et al., 2004), which suggests that the estimated phase lags, in years, between lag-correlated pairs of precipitation and groundwater-level time series (Table 3) have important conceptual considerations for understanding vadose zone water and chemical movement through the High Plains aquifer. We propose that the phase lags between the paired precipitation and groundwater levels for the PDO period are conceptually equivalent to approximate vadose zone transit times to the water table under the persistently wet phase of PDO variability. This conceptual model of transit times through the vadose zone of the High Plains aquifer can be implemented toward a generalized lag-based, hydrologic time series approach for estimating climate-varying water fluxes.

As a point of reference, the previously reported water flux and advective chemical transit times from the vadose zone monitoring network by McMahon et al. (2006) were estimated using the tritium method (Table 4). The tritium method integrates long-term (multidecadal) tritium migration in the vadose zone to estimated corresponding water flux. These estimates are used as the foundation to discuss the implications of interannual to mutlidecadal climate variability on long-term vadose zone water flux and transit-time estimation.

TABLE 4. Estimated water flux and transit times for tritium, hydrologic time series, and Cl^- displacement approaches. "—" indicates estimates are not applicable.

Site†	Depth averaged volumetric water content	Unsaturated zone thickness	Water flux (transit time)		Cl^- displacement¶
			Tritium approach‡	Hydrologic time-series approach for the PDO period§	
	mm mm^{-1}	m	——— mm yr^{-1} (yr) ———		
Northern High Plains (NHP)					
UMA	0.258	47	111 (77–112)	—	—
GNT	0.253	45	102 (84–112)	—	—
IMP	0.240	28	70 (86–96)	—	—
GW-1	0.250#	40	—	476 (21)	—
GW-2	0.250#	40	—	455 (22)	—
Central High Plains (CHP)					
CAL121	0.101	45	54 (51 to 84)	—	—
CAL122	0.113	45	39 (49 to 130)	—	—
CNG	0.092	50	5 (2000)	—	101
GW-3	0.100#	43	—	390 (11)	—
GW-4	0.100#	55	—	196 (28)	—
Southern High Plains (SHP)					
JRW	0.128	46	32 (132 to 188)	–	128
MPL	0.151	43	17 (302 to 373)	–	—
MWR	0.148	15	0.2 (10,500)	–	—
GW-5	0.140#	66	—	200 (46)	—
GW-6	0.140#	41	—	198 (29)	—

† GW, groundwater station; P, meteorological station.
‡ Water flux and advective chemical transit times estimated using tritium method, reported by McMahon et al. (2006).
§ Water flux and advective chemical transit times estimated using phase lag (Table 3) as transit time and Eq. [1]. PDO, Pacific Decadal Oscillation.
¶ Water flux estimated using chloride peak displacement method and Eq. [1].
# Estimated as the average of values from the UZ sites in each subregion.

The water fluxes that contributed to the observed advective displacement of the chemical profiles were calculated using the method of Cl^- peak displacement (McMahon et al., 2006; Stonestrom et al., 2003) and Eq. [1]:

$$q = \frac{\theta z}{t} \qquad [1]$$

where, q is the water flux (L T^{-1}), θ is the depth-weighted average volumetric water content (L^3 L^{-3}), z is the depth of the displaced Cl^- peak with respect to the original Cl^- peak (L), and t is the time elapsed from the original Cl^- peak to the sampling time of the displaced Cl^- peak (T). Equation [1] assumes that flow was one dimensional and vertically downward. The resulting estimated water flux for CNG ($\theta = 0.092$ mm mm^{-1}; $z = 1100$ mm; and $t = 1$) and JRW ($\theta = 0.128$ mm mm^{-1}; $z = 1000$ mm; and $t = 1$) are 101 mm yr^{-1} and 128 mm yr^{-1}, respectively. These transient water-flux estimates reflect the temporal "event scale" (Scanlon et al., 2002) and represent a 400 to >2000% increase over the long-term (multidecadal) water-flux estimates (32 and 5 mm yr^{-1}, respectively) using the tritium method (Table 4) (McMahon et al., 2006).

The proposed lag-based, hydrologic time-series approach was implemented using a modification of Eq. [1], where t represents transit times through the vadose zone and is equivalent to the phase-lag estimate from correlated paired precipitation and groundwater levels in Table 3 (T), θ is the depth-weighted average volumetric water content (L^3 L^{-3}), and z is the vadose zone thickness (L). This hydrologic time-series approach provides estimates of climate-varying water fluxes in response to persistently positive (wet) phases of PDO variability. The water fluxes calculated using this approach represent an intermediate temporal-scale between the event scale (days to weeks) and long-term water-flux estimates using the tritium method. To demonstrate this method, the climate-varying water fluxes were estimated for the strongly lag-correlated pairs of groundwater-level response to precipitation within the PDO period (Table 4).

Water-flux estimates from the hydrologic time-series approach range from 455 to 476 mm yr^{-1} for the NHP, 196 to 390 mm yr^{-1} for the CHP, and 198 to 200 mm yr^{-1} for the SHP, and are, on average, 5 (NHP), 9 (NHP), and 12 (SHP) times larger than the estimates using the tritium method (Table 4). These findings demonstrate a north-to-south trend of proportional increases in estimated water fluxes using the hydrologic time-series method over the tritium method estimates. This spatial trend indicates the importance of PDO variability within the conceptual model of water flux and chemical transport in the High Plains vadose zone. McMahon et al. (2006) proposed a conceptual model that posits a spectrum of slow to fast paths or transport mechanisms through the thick vadose zone of the High Plains. Using this model, a relatively large fraction of the High Plains landscape likely represents the slower end of the spectrum, where diffuse water flux (recharge) dominates. Water-flux estimates that represent the slow paths are observed at CNG and MWR sites using the tritium method (0.2–5 mm yr^{-1}, Table 4). A much smaller fraction of the High Plains likely represents fast-path zones and have been previously identified where surface water is concentrated, such as beneath dry streambeds, ditches, and shallow depressions or playas in the land surface (McMahon et al., 2006; Scanlon and Goldsmith, 1997; Wood and Sanford, 1995), ponding near irrigation wells (Walvoord et al., 2005), and beneath irrigated agricultural lands (McMahon et al., 2006; Scanlon et al., 2005). Previously published water-flux estimates beneath a fast-path zone of the SHP include 60 to 100 mm yr^{-1} beneath playas using the chloride mass-balance method; 77 to 120 mm yr^{-1} beneath playas using the tritium method; and ~200 to 600 mm yr^{-1} beneath playas receiving wastewater discharge (Scanlon and Goldsmith, 1997; Wood and Sanford, 1995). A comparison of the estimated water fluxes from the proposed hydrologic time-series approach to the conceptual model indicates that fast-path or preferential-flow mechanisms are likely controlling processes. Further investigations are needed to identify specific fast-path mechanisms that are integrated within the hydrologic time-series approach to water-flux estimation.

Implications of the relatively large estimated water fluxes (Table 4) with respect to previously identified strong lag correlations between precipitation and groundwater levels on the timescale of PDO variability (Table 3) are important because they indicate that water-level fluctuations of the High Plains aquifer, and ultimately aquifer sustainability, are dominantly controlled by fast-path recharge mechanisms. Furthermore, the observed proportional increase in water fluxes between water-flux estimation methods illustrates the temporal variability, from event-based to interdecadal climatic forcings, that is integrated when applying historical tracer techniques for water-flux or recharge estimation. For example, the tritium method used in the High Plains integrates vadose zone water and tracer movement that is influenced by the PDO variability of two distinct dry phases of variability (1953, which is the start of atmospheric testing of nuclear weapons and the tritium signal, to 1976; and 1998–2002) and two wet phases of variability (1977–1998; and 2002–2005). These findings demonstrate additional limitations for tracer-based techniques of estimating vadose zone water flux, as outlined by Scanlon et al. (2002). Additionally, these findings support the conclusion that climate-induced variations can have important implications for water-flux and recharge estimation, which ultimately effects policy and groundwater resource decision making (Hanson and Dettinger, 2005) on interannual to interdecadal timescales.

Conclusions

Natural climate variability (ENSO, NAMS, PDO, and AMO) has been identified as a fundamental and important hydrologic control on groundwater resources of the High Plains aquifer. This research provides valuable insight for improved understanding of interannual to interdecadal climatic forcings on vadose zone water and chemical movement of the High Plains aquifer. Better predictions of future recharge and vulnerability of groundwater quality is critical for management of aquifer sustainability in light of persistent and recurring natural climate-cycle forcings and potential global climate change. Hydrologic response to climate variability of a few years to decades is of particular importance because of the tangible implications for water-resource management on timescales within an individual human's lifetime. Continued long-term monitoring of vadose zone water and chemical-flux response is necessary for improved understanding of subsurface response to climate variability.

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References

Austin, A.T., L. Yahdjin, J.M. Stark, J. Belnap, A. Porporato, U. Norton, D.A. Ravetta, and S.M. Schaeffer. 2004. Water pulses and biogeochemical cycles in arid and semiarid ecosystems. Oecologia 141:221–235.

Bruce, B.W., J.J. Gurdak, P.B. McMahon, and R.T. Hanson. 2005. Response of deep percolation in the vadose zone to climate variability. AGU 86(52):Fall Meet. Suppl. Abstract H21E-1391.

Castro, C.L., T.B. McKee, and R.A. Piekle, Sr. 2001. The relationship of the North American monsoon to tropical and North Pacific sea surface temperatures as revealed by observational analysis. J. Clim. 14:4449–4473.

CPN. 1980. Operator's manual, 503 Hydroprobe moisture depth gauge. CPN International, Concord, CA.

Dennehy, K.F., D.W. Litke, and P.B. McMahon. 2002. The High Plains aquifer, USA: Groundwater development and sustainability. p. 99–119. *In* K.M. Hiscock, M.O. Rivett, and R.M. Davison (ed.) Sustainable groundwater development. Geological Society of London Special Pub. 193. Geological Society Publishing House, London, UK.

Dettinger, M.D., D.R. Cayan, H.F. Diaz, and D.M. Meko. 1998. North–south precipitation patterns in western North America on interannual-to-decadal timescales. J. Clim. 11:3095–3111.

Dickinson, J.E., R.T. Hanson, T.P.A. Ferré, and S.A. Leake. 2004. Inferring time-varying recharge from inverse analysis of long-term water levels. Water Resour. Res. 40:W07403. doi:10.1029/2003WR002650.

Enfield, D.B., A.M. Mestas-Nunez, and P.J. Trimble. 2001. The Atlantic multidecadal oscillation and its relation to rainfall and river flows in the continental U.S. Geophys. Res. Lett. 28:2077–2080.

Flint, A.L., G.S. Campbell, K.M. Ellett, and C. Calissendorff. 2002. Calibration and temperature correction of heat dissipation matric potential sensors. Soil Sci. Soc. Am. J. 66:1439–1445.

Fryar, A.E., W.F. Mullican, and S.A. Macko. 2001. Groundwater recharge and chemical evolution in the southern High Plains of Texas, USA. Hydrogeol. J. 9:522–542.

Fye, F.K., D.W. Stahle, E.R. Cook, and M.K. Cleaveland. 2006. NAO influence on sub-decadal moisture variability over central North America. Geophys. Res. Lett. 33:L15707. doi:10.1029/2006GL026656.

Ghil, M., M.R. Allen, M.D. Dettinger, K. Ide, D. Kondrashov, M.E. Mann, A.W. Roberston, A. Saunders, Y. Tian, F. Varadi, and P. Yiou. 2002. Advanced spectral methods for climate time series. Rev. Geophys. 40:1–41.

Grissino-Mayer, H.D. 1996. A 2129-year reconstruction of precipitation for northwestern New Mexico, USA. p. 191–204. *In* J.S. Dean et al. (ed.) Tree rings, environment, and humanity. Radiocarbon, Tuscon, AZ.

Gurdak, J.J., and R.T. Hanson. 2005. Correlation of climate variability with water quality in the High Plains aquifer. AGU 86(52):Fall Meet. Suppl. Abstract H31B-1303

Gurdak, J.J., J.E. McCray, G. Thyne, and S.L. Qi. 2007. Latin hypercube approach to estimate uncertainty in ground water vulnerability. Ground Water doi:10.1111/j.1745-6584.2006.00298.x.

Gurdak, J.J., and S.L. Qi. 2006. Vulnerability of recently recharged ground water in the High Plains regional aquifer to nitrate. USGS Scientific Investigations Rep. 2006-5050. USGS, Washington, DC.

Gutzler, D.S. 2004. An index of interannual precipitation variability in the core of the North American monsoon region. J. Clim. 17:4473–4480.

Hanson, R.T., and M.D. Dettinger. 2005. Ground water/surface water responses to global climate simulations, Santa Clara-Calleguas Basin, Ventura, California. J. Am. Water Resour. Assoc. 41:517–536.

Hanson, R.T., M.D. Dettinger, and M.W. Newhouse. 2006. Relations between climatic variability and hydrologic time series from four alluvial basins across the southwestern United States. Hydrogeol. J. 14:1122–1146.

Hanson, R.T., M.W. Newhouse, and M.D. Dettinger. 2004. A methodology to assess relations between climatic variability and variations in hydrologic time series in the southwestern United States. J. Hydrol. 287:252–269.

Hartsough, P., S.W. Tyler, J. Sterling, and M. Walvoord. 2001. A 14.5 kyr record of nitrogen flux from desert soil profiles as inferred from vadose zone pore waters. Geophys. Res. Lett. 28:2955–2958.

Hignett, C., and S.R. Evett. 2002. Neutron thermalization. p. 501–521. *In* J.H. Dane and G.C. Topp (ed.) Methods of soil analysis: Part 4. Physical methods. SSSA Series 5. SSSA, Madison, WI.

Keese, K.E., B.R. Scanlon, and R.C. Reedy. 2005. Assessing controls on diffuse groundwater recharge using unsaturated flow modeling. Water Resour. Res. 41:W06010. doi:10.1029/2004WR003841.

Kerr, R.A. 2000. A North Atlantic climate pacemaker for the centuries. Science 288:1984–1985.

Mantua, N., and S. Hare. 2002. The Pacific Decadal Oscillation. J. Oceanogr. 58:35–44.

Maupin, M.A., and N.L. Barber. 2005. Estimated withdrawals from principal aquifers in the United States, 2000. USGS Circular 1279. USGS, Washington, DC.

McCabe, G.J., M.A. Palecki, and J.L. Betancourt. 2004. Pacific and Atlantic ocean influence on multidecadal drought frequency in the United States. Proc. Natl. Acad. Sci. USA 101:4136–4141.

McMahon, P.B., K.F. Dennehy, B.W. Bruce, J.K. Böhlke, R.L. Michel, J.J. Gurdak, and D.B. Hurlbut. 2006. Storage and transit time of chemicals in thick unsaturated zones under rangeland and irrigated cropland, High Plains, United States. Water Resour. Res. 42:W03413. doi:10.1029/2005WR004417.

McMahon, P.B., K.F. Dennehy, R.L. Michel, M.A. Sophocleous, K.M. Ellett, and D.B. Hurlbut. 2003. Water movement through thick unsaturated zones overlying the central High Plains aquifer, southwestern Kansas, 2000–2001. USGS Water-Resour. Investigations Rep. 03-4171. USGS, Washington, DC.

Merideth, R. 2001. A primer on climatic variability and change in the southwest. Udall Center for Studies in Public Policy, Univ. of Arizona, Tucson.

Minobe, S. 1997. A 50–70 year climate oscillation over the North Pacific and North America. Geophys. Res. Lett. 24:683–686.

NOAA. 2006a. Climate of 2004, annual review, U.S. drought. Available at http://www.ncdc.noaa.gov/oa/climate/research/2004/ann/drought-summary.html (verified 21 Mar. 2007). National Climatic Data Center, NOAA, Asheville, NC.

NOAA. 2006b. United States Historical Climatology Network (USHCN). Available at http://www.ncdc.noaa.gov/oa/climate/research/ushcn/ushcn.html (verified 21 Mar. 2007). National Climatic Data Center, NOAA, Asheville, NC.

Phillips, F.M. 1994. Environmental tracers for water movement in desert soils of the arid American southwest. Soil Sci. Soc. Am. J. 58:15–24.

Rosenberg, N.J., D.J. Epstein, D. Wang, L. Vail, R. Srinivasan, and J.G. Arnold. 1999. Possible impacts of global warming on the hydrology of the Ogallala aquifer region. Clim. Change 42:677–692.

Scanlon, B.R., and R.S. Goldsmith. 1997. Field study of spatial variability in unsaturated flow beneath and adjacent to playas. Water Resour. Res. 33:2239–2252.

Scanlon, B.R., R.W. Healy, and P.G. Cook. 2002. Choosing appropriate techniques for quantifying groundwater recharge. Hydrogeol. J. 10:18–39.

Scanlon, B.R., R.C. Reedy, D.A. Stonestrom, D.E. Prudic, and K.F. Dennehy. 2005. Impact of land use and land cover change on groundwater recharge and quality in the southwestern US. Glob. Change Biol. 11:1577–1593.

Schmidt, K.M., and R.H. Webb. 2001. Researchers consider U.S. southwest's response to warmer, drier conditions. Eos Trans. AGU 82(41):475–478.

Seyfried, M.S., S. Schwinning, M.A. Walvoord, W.T. Pockman, B.D. Newman, R.B. Jackson, and F.M. Phillips. 2005. Ecohydrological control of deep drainage in arid and semiarid regions. Ecology 86(2):277–287.

Shun, T., and C.J. Duffy. 1999. Low-frequency oscillations in precipitation, temperature, and runoff on a west facing mountain front: A hydrogeologic interpretation. Water Resour. Res. 35:191–201.

Small, E.E. 2005. Climatic controls on diffuse groundwater recharge in semiarid environments of the southwestern United States. Water Resour. Res. 41: W04012, doi:10.1029/2004WR003193.

Stonestrom, D.A., D.E. Prudic, R.J. Laczniak, K.C. Akstin, R.A. Boyd, and K.K. Henkelman. 2003. Estimates of deep percolation beneath native vegetation, irrigated fields, and the Amargosa River channel, Amargosa Desert, Nye County, Nevada. USGS Open-File Rep. 03-104. USGS, Washington, DC.

USGS. 2006. National water information system for the nation. Available at http://waterdata.usgs.gov/nwis (verified 21 Mar. 2007). USGS, Washington, DC.

Vautard, R., P. Yiou, and M. Ghil. 1992. Singular-spectrum analysis: A toolkit for short, noisy chaotic signals. Physica D. 58:95–126.

Walvoord, M.A., F.M. Phillips, D.A. Stonestrom, R.D. Evans, P.C. Hartsough, B.D. Newman, and R.G. Striegl. 2003. A reservoir of nitrate beneath desert soils. Science 302:1021–1024.

Walvoord, M.A., D.A. Stonestrom, B.J. Andraski, and R.G. Striegl. 2004. Constraining the inferred paleohydrologic evolution of a deep unsaturated zone in the Amargosa Desert. Vadose Zone J. 3:502–512.

Walvoord, M.A., J.J. Gurdak, and P.B. McMahon. 2005. Focused flow near irrigation wells as a possible mechanism of regionally enhanced chemical migration in the High Plains, central U.S. Vadose EOS Trans. AGU 86(52):Fall Meet. Suppl. Abstract H131-1414.

Wolter, K., and M.S. Timlin. 1993. Monitoring ENSO in COADS with a seasonally adjusted principle component index. p. 52–57. *In* Proc. of the 17th Climate Diagnostic Workshop, Norman, OK. NOAA/N MC/CAC, NSSL, Oklahoma Clim. Survey, CIMMS. and the School of Meteor., Univ. of Oklahoma.

Wolter, K., and M.S. Timlin. 1998. Measuring the strength of ENSO: How does 1997/1998 rank? Weather 53:315–324.

Wood, W.W., and W.E. Sanford. 1995. Chemical and isotopic methods for quantifying ground-water recharge in a regional, semiarid environment. Ground Water 33(3):458–468.

Woodhouse, C.A., and J.T. Overpeck. 1998. 2000 years of drought variability in the central United States. Bull. Am. Meteorol. Soc. 79:2693–2714.