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Large-scale water cycle perturbation due to irrigation pumping in the US High Plains: A synthesis of observed streamflow changes

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SUMMARY

We explore the influence of long-term, large-scale irrigational pumping on spatial and seasonal patterns of streamflow regimes in the High Plains aquifer using extensive observational data to elucidate the effects of regional-scale human alterations on the hydrological cycle. Streamflow, groundwater and precipitation time series spanning all or part of the period of intensive irrigation development (1940–1980) in the region were analyzed for trend and step changes using the non-parametric Mann–Kendall test and the parametric Student's *t*-test, respectively. Based on several indicators to evaluate the streamflow–groundwater connection degree over the High Plains aquifer, we found a systematic decrease in the hydraulic connection between groundwater and streamflow from the Northern High Plains to Southern High Plains. Trends and step changes are consistent with this regional pattern. Decreasing trends in annual and dry-season (mean July–August) streamflow and increasing trends in the number of low-flow days are prevalent in the Northern High Plains with a gradual decrease in trend detection towards the south. Additionally, field significance of trends was assessed by the Regional Kendall's *S* test over the period of most intensive irrigation development (1940–1980). The step-change results imply that the observed decreases in streamflow are likely attributable to the significant declines in groundwater levels and unlikely related to changes in precipitation because the majority of precipitation data over the region did not reveal any significant changes. Thus, it is very likely that extensive irrigational pumping have caused streamflow depletion, more severely, in the Northern High Plains, and to a lesser extent in the Southern High Plains over the period of study.

1. Introduction

The terrestrial water cycle forms a vital link between natural ecosystems and the global climate through complex interactions among its components. Identification and quantification of linkages between the components of the water cycle is further complicated because each component is linked to every other, either in direct or indirect ways, via dynamic flux exchange across a wide range of spatial and temporal scales (Fig. 1a). Thus, any change in one of the storages will have a subsequent effect on the other parts of the water cycle and on the natural hydrological fluxes. However, our knowledge of the potential impacts of these changes on the other components of the water cycle, along with their spatial scales or regional significance, is still very limited yet crucial for future climate variability prediction and water resources management.

Recent studies showed that, besides natural processes, human activities distinctly alter the hydrological cycle by disturbing the natural circulation of water over the continent (Costa et al., 2003; Foley et al., 2005; Nilsson et al., 2005; Huntington, 2006; Zhang et al., 2007; Adam and Lettenmaier, 2008; Barnett et al., 2008; Sahoo and Smith, 2009). One major cause of these disturbances is irrigation (Alpert and Mandel, 1986; Vorosmarty and Sahagian, 2000; Milly et al., 2005; Haddeland et al., 2006b, 2007; Milliman et al., 2008; Gerten et al., 2008; Rost et al., 2008; Wisser et al., 2009), which accounts for nearly 85% of the global water consumption (Gleick, 2003). In fact, the primary use of water worldwide is to irrigate the agricultural areas, which cover 40% of the land surface (Asner et al., 2004). As the demand for food increases along with the growing population, irrigated areas continue to expand with an actual expansion of 70% in the last 40 years (Gleick, 2003), and consequently, surface water and groundwater resources are being substantially exploited to comply with the corresponding increase in water demand. Lately, the global use of groundwater has surpassed surface water use as the primary source of irrigation (Healy et al., 2007; Giordano and Villholt, 2007), such that the total groundwater withdrawals for irrigation have increased from 23% of

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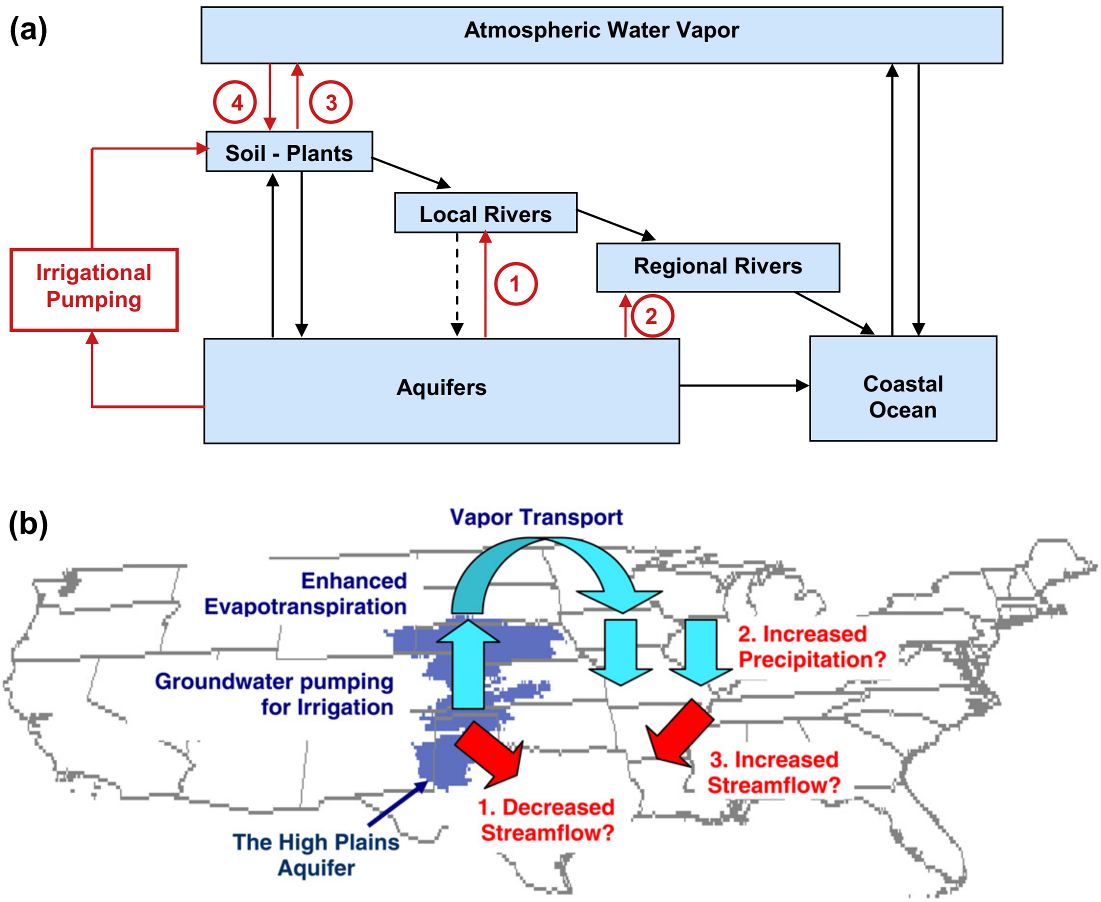

Fig. 1. (a) A simplified version of the terrestrial water cycle showing its reservoirs and the complex dynamic interactions among them (red arrows indicate fluxes most directly affected by pumping; numbers 1–4 indicate impacts of pumping on local river flow, regional river flow, ET, and precipitation, respectively), and (b) objectives of this study, showing the three components of the irrigation-induced water cycle and focus of the paper (filled area represents the High Plains aquifer).

total withdrawals for irrigation in 1950 to 42% of that in 2000 for the conterminous USA (Hutson et al., 2004). Most of the water extracted from aquifers for irrigation is lost into the atmosphere by evapotranspiration (ET) after it is applied to the land surface, while the rest either runs off to a stream or infiltrates through the soil zone becoming groundwater again. Due to the interactions among the reservoirs of the hydrological cycle, this disturbance will have subsequent effects on local and regional river flow (fluxes 1 and 2 in Fig. 1a), on ET (flux 3), and consequently on precipitation (flux 4). Accordingly, extensive pumping of groundwater leads to depleted subsurface storages, especially in arid and semi-arid regions where the natural aquifer recharge rates are very low. Over the last century, groundwater levels across the United States declined substantially, generally during the dry season and in semi-arid regions, as a result of increased groundwater usage for irrigation (Bartolino and Cunningham, 2003). Furthermore, groundwater mining is a growing problem throughout the world which adversely affects major aquifer systems as well as local areas (Konikow and Kendy, 2005). One well-known case is the High Plains aquifer system of the US Great Plains, where large-scale irrigational pumping induced a depletion of more than 330 km^3 in the stored volume of water, a net decrease of 8.5% of the pre-development water in storage, from pre-development (about 1950) to 2007 (McGuire, 2009).

One direct effect of groundwater irrigation is the significant reduction of surface water availability, also known as "streamflow depletion", due to decreased groundwater discharge to streams and wetlands caused by excessive and prolonged pumping (Winter et al., 1998; Sophocleous, 2002; Kollet and Zlotnik, 2003). The impact can be large especially in areas where groundwater and surface water systems are closely-connected, since groundwater is the principal source of streamflow in such places. For example, many perennial streams in western Kansas running across the High Plains aquifer in 1961 became shorter or disconnected, or disappeared by 1994 as a result of large groundwater withdrawals (Sophocleous, 2000). Additionally, the flow of streams in some parts of Kansas, Oklahoma and New Mexico has decreased to half of the initial recorded flow over time (Brikowski, 2008). A trend detection study by Wahl and Wahl (1988) identified decreasing trends in the annual mean flow, annual baseflow, and annual peak discharge of the Beaver River in the Oklahoma Panhandle from 1938 to 1986 while precipitation records showed no trend for the same period. Thus, they concluded that increased groundwater pumping from the underlying High Plains aquifer was the main mechanism generating the observed decreases in streamflow. Szilagyi (1999) examined the changes in the annual mean flow of Republican River basin where significant streamflow depletion is observed since the late 1940s. Analyzing eight US Geological Survey (USGS) gauging stations, he verified significant decreasing trends in the whole river basin that cannot be explained by precipitation variability. Subsequently, his modeling study (Szilagyi, 2001) showed that the observed streamflow depletion in the same river basin has resulted from human-induced changes such as irrigation, land cover changes and reservoir construction. Similarly, Burt et al. (2002) applied a multiple regression model to annual streamflow data from a single gauging station in the Republican River basin to evaluate the effect of groundwater irrigation on streamflow during the period 1936–1998 and found a strong inverse relationship between annual streamflow and the number of irrigation wells, in addition to a 75% decline in the mean annual flow over the same period. In a more comprehensive study, Wen and Chen (2006) searched for trends in streamflow using data from 110 gauging stations in eight major river basins throughout Nebraska during 1948–2003 and detected decreasing trends at the majority of gauges in the Republican River basin but only at a few in the eastern part. Without any significant changes in precipitation and temperature for the same period, their study concluded that groundwater withdrawal for irrigation was the primary factor leading to depletion of streamflow in Nebraska. Also, Buddemeier et al. (2003) reported that after the onset of extensive groundwater pumping, portions of major rivers crossing the High Plains aquifer experienced decreases in annual flow during the last few decades with the Arkansas River exhibiting the greatest flow depletion among the others.

Besides depleting the groundwater storage and reducing the baseflow to rivers, irrigation dramatically increases soil moisture during the warm season which may instigate indirect effects on the key components of regional climate including increases in ET, cooling of surface temperatures and enhancement of precipitation (the fourth link in Fig. 1a) (Eltahir and Bras, 1996; Eltahir, 1998; Vorosmarty and Sahagian, 2000; Pielke, 2001; Kanamitsu and Mo, 2003; Betts, 2004; Haddeland et al., 2006a). Several modeling studies showed that an increase in soil moisture induces higher ET and atmospheric moisture content which further contributes to the formation of local convective storms via enhanced moisture recycling over or downwind of the irrigated (or wetted soil) regions (e.g., Segal et al., 1989; Small, 2001; Pal and Eltahir, 2002; Koster et al., 2004; Dominguez et al., 2009). One study investigated the effect of land use changes on the regional climate of the irrigation-dominated northern Colorado plains (Chase et al., 1999). Their model results demonstrated that the magnitude of forcing induced by irrigational practices were strong enough to affect the regional temperature, cloud cover, precipitation and surface hydrology. Other regional studies showed significant differences in the heat and moisture fluxes between the irrigated (wet) and non-irrigated (dry) areas over India (Douglas et al., 2006), and Nebraska (Adegoke et al., 2007). Despite the intricacy of this mechanism, few observational studies detected a signal of irrigation–precipitation link over the High Plains aquifer. One study identified an irrigation-related increase in June precipitation during 1930–1970 over and near the heavily-irrigated regions in the Texas panhandle when synoptic conditions allowed low-level convergence and uplift (Barnston and Schickendanz, 1984). Another one observed an additional summer rainfall of 6–18% about 90 km downwind of the Texas panhandle during 1996 and 1997 (Moore and Rojstaczer, 2002). A third study by Adegoke et al. (2003) found cooler surface temperatures in summer within the densely-irrigated areas in Nebraska verified by both simulations and data analysis.

All of these earlier studies underline that irrigation significantly influences the climate and hydrology patterns not only at local scales but also at regional scales (Fig. 1b). Therefore, in this study, we aim to develop a comprehensive analysis of the regional impacts of irrigational pumping on the hydrological cycle to investigate whether an anthropogenic regional water cycle is embedded into the natural and continental-scale water cycle. Our research will be reported in a series of three papers. In this first paper, we investigate the direct effect of groundwater irrigation: streamflow depletion. In a second study, we analyze observed precipitation over the central US searching for signals of irrigation-enhanced precipitation downwind of the High Plains (DeAngelis et al., 2010). In a third report, we examine the observed groundwater and streamflow downwind of the High Plains where enhanced precipitation has been observed (Kustu and Fan, in preparation). We emphasize that all three studies rely on long-term observations in groundwater, streamflow and precipitation, and that our attention is on the regional-scale hydrologic and climatic linkages and feedbacks.

The focus of this paper is to determine the long-term, large-scale irrigational pumping effects on the spatial and seasonal patterns of streamflow regimes over the High Plains aquifer. There have been numerous observational and theoretical studies that investigated the groundwater–surface water interactions; however their focus of interest are the changes in small watershed scales (e.g., Hewlett and Hibbert, 1963; Dunne and Black, 1970a,b; Tanaka et al., 1988; De Vries, 1994, 1995; Eltahir and Yeh, 1999; Marani et al., 2001; Nyholm et al., 2003; Chen and Chen, 2004; Chen et al., 2008; Zume and Tarhule, 2008). Likewise, the aforementioned studies on streamflow trends in the High Plains aquifer concentrated at one to a few river basins, used different streamflow gauges and analysis methods, over different time periods, and, thus, lack a region-wide, methodologically consistent picture of where and when streamflow depletion is significant. No systematic effort yet has been made to understand the regional significance of groundwater pumping on streamflow despite the large-scale groundwater depletion observed in the aquifer since the 1930s. Hence, this paper will tie the scattered evidence together and establish the regional pattern of streamflow depletion, based on streamflow observations in conjunction with precipitation and water table data using all available records in the USGS archive.

Moreover, detection of abrupt (step) and gradual changes in hydrologic variables and comprehension of their likely causes are critical for long-term water management and assessment of future changes. The attribution of these changes to correct causes is more crucial than ever under the presence of long-term, CO_2-induced climate change trends. Most trend analysis studies attribute the observed changes in streamflow to the variations in climate (e.g., Lins, 1985; Dery and Wood, 2005; Miller and Piechota, 2008). Here, we hypothesize that large-scale human activities, such as the irrigation development in the High Plains region, may induce drastic, regional-scale changes in the hydrological cycle in a similar magnitude as caused by climate variability.

The specific objectives of this study are: (1) to examine the climatic, geologic, and hydrologic variabilities across the High Plains; patterns emerging from this analysis will shed light on where, along the climatic and hydrologic gradient, streamflow is most likely affected by groundwater pumping, (2) to examine the degree of hydraulic connection between the groundwater and streamflow across the climatic–hydrologic gradient; patterns emerging from this analysis will further pinpoint regions/settings where groundwater pumping is most likely to affect streamflow, (3) to quantify the streamflow depletion annually and seasonally over selected regions along the climatic–hydrologic gradient, using trend and step-change analysis tools, (4) to assess the field significance of detected trends, and (5) to attribute the observed streamflow changes to likely causes, i.e., changes in rainfall or in groundwater storage. The results of this study will improve our understanding and quantification of the impact of human modifications to the water cycle at regional scales during the second half of the last century.

The following sections first provide the background information on the study area, followed by the description of data sources and an outline of the methodology. Then, we discuss the observed changes in streamflow across the High Plains region for the period of intensive irrigational development using several indicators. We conclude with a geographic synthesis of regional variations in streamflow depletion caused by irrigational groundwater pumping.

2. The High Plains aquifer system

The High Plains aquifer, a subregion of the Great Plains, is the largest regional aquifer system in the US, and extends under parts of eight states from southern South Dakota to northwestern Texas with a surface area of 450,000 km^2 (Fig. 2a). Flat to gently-sloping vast plains formed by stream-deposited sediments transported eastward from the Rocky Mountains characterize the region (Dennehy, 2000). The aquifer consists of several hydraulically-connected geologic units of Tertiary or Quaternary age. The Brule Formation, the Arikaree Group and the Ogallala Formation constitute the upper Tertiary rocks. The Oligocene-aged Brule Formation, a low-permeable massive siltstone with layers of sandstone and volcanic ash, underlies parts of Nebraska, Colorado and Wyoming and is considered as part of the aquifer only in areas where its permeability is increased by secondary porosity. Overlying the Brule Formation is the Miocene- to Oligocene-aged Arikaree Group which is composed of massive fine-grained sandstone with local beds of volcanic ash, silt and clay underlying large parts of Nebraska, South Dakota and Wyoming. Over the Arikaree Group lies the Miocene–Pliocene Ogallala Formation of unconsolidated clay, silt, sand and gravel. The Ogallala Formation is the principal geologic unit of the aquifer covering 77% of the system's area. Unconsolidated alluvial deposits of Quaternary age overlie the Ogallala Formation on the east and constitute part of the aquifer in areas where they are in hydraulic connection with the Tertiary deposits. Most of the gravel, sand, silt and clay in the alluvial deposits are reworked material derived from the Ogallala Formation in the form of sand dunes, windblown loess and valley-fill deposits along the stream channels (Gutentag et al., 1984; Weeks et al., 1988). In general, the thickness of the aquifer decreases from north to south and from central to east. The High Plains aquifer is generally underlain by Permian- to Tertiary-aged evaporites such as anhydrite, gypsum, halite, limestone and dolomite.

The High Plains region has a typical mid-latitude dry continental climate with a high rate of evaporation, limited precipitation and abundant sunshine changing from arid to semi-arid from the Texas panhandle to western Kansas, and to sub-humid in some parts of central Kansas and eastern Nebraska (Gutentag et al., 1984). The region is characterized by natural climate gradients from east to west and north to south. Located at the center of a transition zone, a wetter to drier precipitation gradient from east to west, and a colder to hotter temperature gradient from north to south prevail across the region (Fig. 2b). These precipitation (east–west) and temperature (north–south) gradients produce a distinctive climate condition that varies substantially from hourly to decadal time scales. The average annual precipitation throughout the region is 500 mm with a range of 300 mm (Rodell and Famiglietti, 2002). Most of the precipitation falls as rain during the growing season, from April to September, however large variations in rainfall are observed both spatially and temporally due to the common thunderstorms and extreme weather events (Weeks et al., 1988). As a result of limited precipitation, naturally-occurring fertile soils with grassland vegetation cover the region (Kromm and White, 1992). The evapotranspiration rates are high, because of persistent winds and high summer temperatures, and annually average from 1500 mm in the north to 2700 mm in the south (Weeks et al., 1988) (Fig. 2b).

The High Plains is an unconfined blanket sand-and-gravel type aquifer with a general groundwater direction of west to east at a rate of 0.30 m/day. The water table reaches the surface near the rivers that are hydraulically-connected to the aquifer such as the Platte and the Arkansas Rivers. The saturated thickness of the aquifer varies from zero in the depositional areas of unconsolidated alluvial deposits to 300 m in north-central Nebraska, with an average of 60 m (Weeks et al., 1988). In 1980, the depth to water table was less than 30 m in about half of the aquifer, less than 60 m under most of Nebraska and Kansas, and between 60 and 90 m in parts of western and southwestern Nebraska and southwestern Kansas. In local areas of prolonged irrigational pumping, the water table could be found at 120 m or more below the ground (Miller and Appel, 1997). The aquifer is recharged mainly by precipitation and locally by seepage from streams. High evapotranspiration rates lower the aquifer recharge rates to less than 13 mm/yr in most parts, ranging from 0.6 mm/yr in Texas to 150 mm/yr in south-central Kansas, except in areas such as Nebraska Sandhills, where rainfall infiltrates quickly through the highly permeable sand to replenish the groundwater system (Gutentag et al., 1984). Groundwater naturally discharges to streams and springs and directly to the atmosphere by evapotranspiration in areas where the water

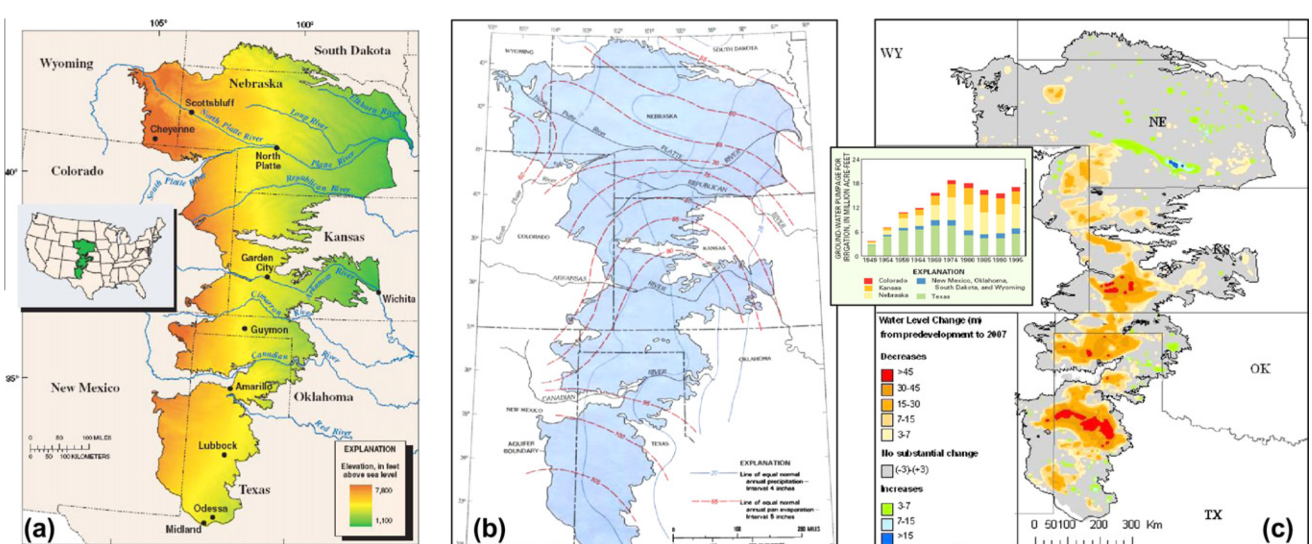

Fig. 2. (a) Location and topography of High Plains regional aquifer system (from Qi et al. (2002)), (b) Normal annual precipitation (blue) and Class-A pan evaporation (red) in the High Plains during 1951–1980 (from Kastner et al. (1989)), and (c) water-level changes in the High Plains from pre-development to 2007 (reproduced from McGuire (2009)), where insert shows volume of groundwater pumped for irrigation from the High Plains aquifer by State for selected years between 1949 and 1995 (from McGuire et al. (2003)).

Table 1
Total number of groundwater and streamflow sites examined for this study.

Type of sites	Number of sites in each state that has parts in the High Plains aquifer								Total number of sites
	SD	WY	NE	CO	KS	OK	TX	NM	
Groundwater	7	132	56	279	205	110	183	68	1040
Streamflow	18	40	193	18	111	22	25	7	431

table is near the surface. However, most of the discharge from the High Plains aquifer occurs by pumping for irrigational use, which results in an imbalance between the discharge and the natural recharge, changing the volume of storage. The total volume of drainable water in storage was estimated to be about 4010 km^3 in 1980, 65% of which is in Nebraska where the recharge rate is the greatest (Gutentag et al., 1984).

Due to the ideal topography and productive soils, High Plains is one of the major agricultural regions in the world, consisting of approximately 20% of the irrigated land in the US, with the aquifer supplying nearly 30% of the groundwater used for irrigation across the United States (Luckey et al., 1986; Sophocleous, 2005). In the region, water for irrigation is principally supplied from the aquifer (81% in 1995); however surface water is also used for irrigational use to a limited extent (19% in 1995), especially the Platte River in Nebraska, which supplies nearly all the surface water for irrigation (85%) (Dennehy, 2000). Towards the south, use of groundwater increases (~92%) (Dennehy, 2000) due to the scarcity of surface water resources (Buchanan et al., 2009). The development of groundwater irrigation started in the region in the 1930s in response to a drought and expanded rapidly from South to North by the 1960s with the invention of center-pivot irrigation systems (Miller and Appel, 1997). The groundwater irrigation developed first in New Mexico and Texas in 1930s, later in Oklahoma and Kansas in 1940s, and finally in Colorado, Nebraska and Wyoming during the 1950s and 1960s (Luckey et al., 1981). From 1940 to 1980, the total irrigated area in the region had increased from 8500 km^2 to about 56,000 km^2, which was irrigated with 22 km^3 of water by tapping approximately 170,000 wells that had been completed in the aquifer by 1980 (Weeks et al., 1988). This resulted in a depletion of 5% (~205 km^3) of the pre-development volume of stored water from the aquifer; 70% of which was in Texas and 16% in Kansas (Gutentag et al., 1984). As the groundwater withdrawals escalated from 5 km^3 to 23 km^3 from 1949 to 1974 (see insert in Fig. 2c), declines in water levels in the aquifer as much as 30 m were common in parts of Texas, Oklahoma and southwestern Kansas by 1980 (Gutentag et al., 1984). After 1980, the average rate of decline in water levels has decreased across the aquifer despite the continuous increase in the total irrigated area attributable, in large part, to the above-normal precipitation rates over the region between 1980 and 1994, and, in some part, to new pumping regulations and technologies in irrigation (Dugan and Sharpe, 1995). Water-level changes in the aquifer from pre-development to 2009 are shown in Fig. 2c.

3. Data and methods

3.1. Data sources

Stream gauge records in the High Plains were acquired from the USGS National Water Information System (NWIS) database (USGS, 2009; http://nwis.waterdata.usgs.gov/nwis/sw). The entire record, except for some gauges in Texas, is in the form of daily measurements starting from the early 1930s to the present. However, the record period of each stream gauge differs greatly such that some records extend back to the early 1900s while some others start in the late 1970s or even in 1980s. Most stations, especially the ones in Kansas and Texas, have interrupted records, but still no filling-in the data gaps is performed. Hence, the influence of limited data availability is noted in the evaluation of the results. Major dams and reservoirs throughout the High Plains are listed in the National Inventory of Dams by the US Army Corps of Engineers (USACE) (National Atlas, 2009; http://nationalatlas.gov/mld/dams00x.html) and their effects are considered in the analysis. Groundwater data come from two sources: the first is the USGS NWIS database (http://waterdata.usgs.gov/nwis/gw), which supplied the majority of the data, and the second is the Texas Water Development Board database (TWDB, 2009; http://www.twdb.state.tx.us/publications/reports/GroundWaterReports/GWDatabaseReports/GWdatabaserpt.htm), which is used to supplement the sparse USGS observations in Texas. Table 1 lists the total number of streamflow gauges (431) and groundwater monitoring wells (1040) explored for this study in the states of the High Plains aquifer. Out of 431 stream gauging stations, 64 gauges were selected for the trend analysis in this study (Table 2). These gauges are located in or downstream of the areas where significant water table decline (>7 m) has been observed (yellow, orange, and red[1] patches in Fig. 2c) and they have long and continuous data covering at least part of the period of intensive irrigation development (1940–1980). The record period of gauging stations varied from a minimum of 12 years to a maximum of 86 years. Of the 64, nine stream gauges that are located within each area of significant water table decline and have continuous daily measurements extending back to the 1940s were used in the step-change analysis. A total of 17 groundwater wells were used in this study, which were selected based on the highest number of measurements for the seasonal cycle analysis, the closest location to the stream gauges for the elevation analysis, and the longest period of record for the step-change analysis, all discussed in detail later. In addition to the streamflow and groundwater data, monthly precipitation totals at nine stations in the vicinity of the associated streamflow gauges was acquired from the Global Historical Climate Network (GHCN, 2009) station dataset (Vose et al., 1992) using the NOAA NCDC GHCN beta version 2, accessible via IRI/LDEO Climate Data Library (http://iridl.ldeo.columbia.edu/SOURCES/.NOAA/.NCDC/.GHCN/.v2beta/). Tables 3 and 4 list detailed information about the precipitation stations and groundwater wells used in this study, respectively. Fig. 3 shows the spatial distribution of all streamflow gauges, groundwater wells and precipitation stations considered for this study together with the dams in the High Plains.

3.2. Methodology

In this study, trend and step changes in time series of several hydrologic variables were analyzed in an effort to evaluate the impact of groundwater pumping on streamflow regimes in the High Plains region. While trend analysis has been applied widely in environmental sciences (e.g., Hirsch and Slack, 1984; Lins, 1985; Lettenmaier et al., 1994; Lins and Slack, 1999; Douglas et al., 2000; Zhang et al., 2001; Pilon and Yue, 2002), few studies

[1] For interpretation of color in Figs. 1–10, the reader is referred to the web version of this article.

Table 2
List of all stream gauges used in the trend and step-change analysis in this study.

	Stream sites	USGS ID number	Latitude	Longitude	State	Drainage area (km^2)	Record period	Dam effect	Dam constr. year	Number of records	Type of records
1	N1	6454500	42°27′35″	103°10′16″	NE	3626	1946–1994	No	x	17,533	Daily
2	N2	6455500	42°27′23″	103°04′08″	NE	3781	1946–1991	Yes	1945	16,437	Daily
3	N3	6457500	42°38′23″	102°12′38″	NE	11,111	1945–1991	Yes	1945	16,801	Daily
4	N4	6687000	41°20′13″	102°10′29″	NE	2326	1930–1991	No	x	22,281	Daily
5	N5	6823000	40°04′10″	102°03′03″	NE	6138	1935–2008	No	x	27,011	Daily
6	N6	6821500	40°01′45″	101°58′03″	NE	4403	1932–2008	No	x	28,008	Daily
7	N7	6823500	40°02′22″	101°52′00″	NE	445	1940–2008	No	x	24,904	Daily
8	N8	6824000	40°02′32″	101°43′41″	NE	61	1940–2008	No	x	24,904	Daily
9	N9	6824500	40°02′04″	101°32′34″	NE	12,639	1947–1994	No	x	17,440	Daily
10	N10	6828500	40°08′26″	101°13′47″	NE	21,238	1950–2008	No	x	21,321	Daily
11	N11	6829500	40°10′00″	101°02′52″	NE	21,600	1946–1993	Yes	1952	17,106	Daily
12	N12	6831500	40°25′54″	101°37′37″	NE	2719	1941–1994	No	x	19,631	Daily
13	N13	6832500	40°25′14″	101°30′44″	NE	2953	1946–1993	Yes	1950	17,381	Daily
14	N14	6834000	40°21′06″	101°07′25″	NE	3367	1950–2008	Yes	1970	21,355	Daily
15	N15	6835000	40°22′23″	101°07′01″	NE	3885	1949–1994	No	x	16,436	Daily
16	N16	6835500	40°14′05″	100°52′40″	NE	7744	1935–2008	Yes	1950	27,004	Daily
17	N17	6836000	40°14′10″	100°48′40″	NE	829	1946–1986	Yes	1987	14,732	Daily
18	N18	6827500	40°00′37″	101°32′31″	NE	7097	1937–2008	No	x	25,999	Daily
19	C1	6825500	39°34′32″	102°15′06″	CO	694	1950–1976	No	x	9632	Daily
20	C2	6825000	39°36′59″	102°14′32″	CO	3367	1950–1971	No	x	7805	Daily
21	C3	6826500	39°37′26″	102°09′47″	CO	4727	1946–1986	No	x	14,610	Daily
22	K1	6844900	39°40′37″	100°43′18″	KS	1155	1959–2008	No	x	18,039	Daily
23	K2	6846500	39°59′06″	100°33′35″	KS	4191	1946–2008	No	x	22,836	Daily
24	K3	6845000	39°48′47″	100°32′02″	KS	2813	1929–2006	No	x	28,472	Daily
25	K4	6873000	39°22′36″	99°34′47″	KS	2694	1945–2008	Yes	1959	23,266	Daily
26	K5	6858500	39°01′04.32″	101°20′50.90″	KS	1735	1946–1984	Yes	1964	13,819	Daily
27	K6	7138650	38°28′52″	101°29′16″	KS	1942	1966–1986	No	x	7213	Daily
28	K7	6859500	38°47′20″	100°52′10″	KS	3709	1951–1979	No	x	10,410	Daily
29	K8	6860000	38°47′41″	100°51′29″	KS	9207	1939–2008	No	x	25,274	Daily
30	K9	7156900	37°00′40″	100°29′29″	KS	22,108	1965–2008	No	1958	15,786	Daily
31	K10	7157500	37°01′57″	100°12′39″	KS	2997	1942–2008	No	x	24,187	Daily
32	K11	7139800	37°35′51.86″	100°00′53.79″	KS	191	1968–1990	No	x	8231	Daily
33	K12	7139000	37°57′21″	100°52′37″	KS	70,114	1922–2008	Yes	1969	31,594	Daily
34	K13	7139500	37°44′41″	100°01′57″	KS	79,254	1944–2007	Yes	1969	22,826	Daily
35	O1	7157000	36°58′33″	100°18′50″	OK	22,455	1942–1965	No	1958	8401	Daily
36	O2	7234100	36°38′42″	100°30′07″	OK	440	1965–1993	No	x	10,227	Daily
37	O3	7233000	36°38′38″	101°12′38″	OK	5095	1939–1964	No	x	9132	Daily
38	O4	7232500	36°43′17″	101°29′21″	OK	5540	1937–1993	Yes	1955	20,454	Daily
39	O5	7234000	36°49′20″	100°31′08″	OK	20,603	1937–2008	Yes	1978	26,042	Daily
40	O6	7236000	36°23′57″	99°37′22″	OK	4206	1942–1976	No	x	12,419	Daily
41	O7	7237000	36°34′00″	99°33′05″	OK	4504	1937–1993	No	x	20,458	Daily
42	O8	7316500	35°37′35″	99°40′05″	OK	2056	1937–2008	No	x	26,183	Daily
43	T1	7235000	36°14′19″	100°16′31″	TX	1805	1940–2008	No	x	24,946	Daily
44	T2	7233500	36°12′08″	101°18′20″	TX	2787	1945–2008	No	x	23,181	Daily
45	T3	7298000	34°33′34″	101°42′33″	TX	490	1939–1973	No	x	12,572	Daily
46	T4	7298200	34°32′36″	101°25′46″	TX	2978	1964–1986	Yes	1974	8096	Daily
47	T5	8080700	34°10′44″	101°42′08″	TX	3344	1939–2008	Yes	1975	25,428	Daily
48	T6	7295500	34°50′55″	102°10′32″	TX	5097	1939–2008	No	x	25,266	Daily
49	T7	7297500	35°00′38″	101°53′29″	TX	8726	1924–1949	Yes	1938	9391	Daily
50	T8	8082500	33°34′51″	99°16′02″	TX	40,243	1923–2008	Yes	1959	1318	Non-Daily
51	T9	8080500	33°00′29″	100°10′49″	TX	22,782	1922–2008	Yes	1960	1222	Non-Daily
52	T10	8082000	33°20′02″	100°14′16″	TX	13,287	1925–2008	Yes	1963	295	Non-Daily
53	T11	7297910	34°50′15″	101°24′49″	TX	10,906	1967–2008	Yes	1965	412	Non-Daily
54	T12	8123650	32°15′01″	101°29′26″	TX	24,136	1959–1979	Yes	1989	7578	Daily
55	T13	8124000	31°53′07″	100°28′49″	TX	39,645	1954–2008	Yes	1939	156	Non-Daily
56	T14	8123850	32°03′13″	100°45′42″	TX	38,617	1980–2008	Yes	1939	160	Non-Daily
57	T15	8120700	32°28′38″	100°56′58″	TX	10,132	1965–2002	Yes	1952	135	Non-Daily
58	T16	8121000	32°23′33″	100°52′42″	TX	10,272	1980–2008	Yes	1952	195	Non-Daily
59	T17	8123800	32°11′57″	101°00′49″	TX	25,387	1958–2008	Yes	1939	564	Non-Daily
60	T18	8133500	31°49′48″	100°59′36″	TX	5807	1939–1994	No	x	19,979	Daily
61	T19	7299890	34°56′08″	100°41′46″	TX	192	1968–2008	No	x	134	Non-Daily
62	T20	7301410	35°28′23″	100°07′14″	TX	743	1961–2008	No	x	17,213	Daily
63	T21	7301200	35°19′45″	100°36′32″	TX	1966	1967–1980	Yes	1939	4749	Daily
64	T22	7301300	35°15′51″	100°14′29″	TX	2802	1964–2008	Yes	1939	348	Non-Daily

searched for an abrupt step change in water resources data (McCabe and Wolock, 2002; Costa et al., 2003; Miller and Piechota, 2008; Kalra et al., 2008). Identification of a step change is equally important because it gives an estimate to quantify the amount of change caused by a certain factor over two different periods of time, especially when relatively sudden, step-like changes are expected.

In hydrologic trend studies, non-parametric methods that do not rely on any assumption about the underlying distribution of the data are preferred to the traditional parametric methods which assume that the data are drawn from a given probability distribution. This is because hydrological data are often strongly non-normal, typically show autocorrelation and/or spatial correlation, and usually consist of seasonal variations and, hence, do not usually

Table 3
List of the precipitation sites used in this study.

Precipitation sites	Site name	State	Latitude	Longitude	Record period	Elevation (m)
P1	Alliance 1 WNW	NE	42°06′36″	102°54′36″	1895–2003	1218
P2	Imperial	NE	40°31′12″	101°38′24″	1890–2005	1000
P3	Burlington Col USA	CO	39°17′59″	102°17′59″	1918–1989	1271
P4	Cheyenne Wells	KS	38°49′12″	102°20′59″	1900–2005	1296
P5	Liberal	KS	37°02′59″	100°55′12″	1907–2005	864
P6	Stratford	TX	36°21′36″	102°05′24″	1911–2005	1126
P7	Miami	TX	35°42′36″	100°38′24″	1905–2005	840
P8	Muleshoe 1	TX	34°14′24″	102°44′24″	1921–2005	1167
P9	Garden City 1 E USA	TX	31°53′59″	101°30′00″	1912–1989	802

Table 4
List of the groundwater wells used in this study. (SCA: Seasonal cycle analysis, EA: Elevation analysis, STC: Step-change analysis).

Wells	USGS Well ID number	State	Latitude	Longitude	Record period	Number of observations	Type of analysis
GW-N1	421505103051701	NE	42°15′05″	103°05′17″	1969–2008	259	SCA
GW-N2	403235101395501	NE	40°32′35″	101°39′55″	1964–2008	2353	SCA
GW-N3	420530103104001	NE	42°05′30″	103°10′40″	1968–2008	61	EA
GW-N4	403111101405301	NE	40°31′11″	101°40′53″	1970–1996	49	EA
GW-N5	420350102502501	NE	42°03′50″	102°50′25″	1946–1987	87	STC
GW-N6	402518101270301	NE	40°25′18″	101°27′03″	1946–1973	183	STC
GW-C1	393700102150000	CO	39°37′08″	102°14′55″	1956–1995	29	EA
GW-K1	392329101040201	KS	39°23′29″	101°04′02″	1947–2008	2137	SCA, STC
GW-K2	382013100583901	KS	38°20′13″	100°58′39″	1931–1998	1903	SCA, STC
GW-K3	374100101270501	KS	37°41′00″	101°27′05″	1958–1998	341	SCA
GW-K4	383046100594901	KS	38°30′46″	100°59′49″	1944–1998	123	EA
GW-K5	370857100234601	KS	37°08′57″	100°23′46″	1939–1989	218	EA, STC
GW-O1	363033101440701	OK	36°30′33″	101°44′07″	1956–1997	1754	SCA
GW-T1	TWDB-354401	TX	36°11′38″	101°20′29″	1951–2007	51	EA
GW-T2	TWDB-233905	TX	36°23′12″	102°52′42″	1937–2000	85	STC
GW-T3	TWDB-1023701	TX	34°38′36″	102°14′18″	1937–1998	71	STC
GW-T4	TWDB-2727301	TX	32°36′40″	102°38′32″	1937–1978	37	STC

conform to the assumptions (e.g., normality, independence, and linearity) of the standard parametric methods (e.g., *t*-test, analysis of variance, linear regression) (Helsel and Hirsch, 1992). Additionally, non-parametric methods are found to be more robust than their parametric equivalents, along with the advantages of having simpler and wider applicability, and being less sensitive to outliers in the data (Kundzewicz and Robson, 2004). While we acknowledge the more sophisticated statistical tools used in the detection of regional trends in hydrology (e.g., Katz et al., 2002; Renard et al., 2006), in this study, we will use the non-parametric Mann–Kendall test (Mann, 1945; Kendall, 1975) for its robustness, simplicity, and insensitivity to missing data.

3.2.1. Mann–Kendall test

The Mann–Kendall test is a rank-based approach that tests for randomness against trends in time-series data and has been widely used in hydrologic and climatic trend studies (e.g., Lins and Slack, 1999; Yue et al., 2003; Burn et al., 2004; Kahya and Kalayci, 2004; Dery and Wood, 2005; Aziz and Burn, 2006). The null hypothesis H_0 states that a sample of data $(x_1, x_2, \dots, x_n)$ consists of n independent and identically distributed random variables, whereas the alternative hypothesis H_1 is that a monotonic trend exists in the data. The test first ranks the entire observations according to time, and then successively compares each data value to all data values following in time by evaluating the Mann–Kendall test statistic, S, as:

$$S = \sum_{i=1}^{n-1} \sum_{j=i+1}^{n} \operatorname{sgn}(x_j - x_i) \tag{1}$$

where x_i and x_j are the sequential data values, n is the number of observations, and

$$\operatorname{sgn}(x_j - x_i) = \begin{cases} 1 & \\ 0 & \text{if} \quad x_j - x_i = 0 \\ -1 & \end{cases} \begin{matrix} x_j - x_i > 0 \\ \\ x_j - x_i < 0 \end{matrix} \tag{2}$$

The mean and variance of S, with the consideration of any possible ties (i.e., equal-valued members in a data set) in the x values are given by Kendall (1975) as:

$$E(S) = 0 \tag{3}$$

$$Var(S) = \frac{n(n-1)(2n+5) - \sum_{i=1}^{n} t_i i(i-1)(2i+5)}{18} \tag{4}$$

where t_i is the number of ties of extent i. Both Mann (1945) and Kendall (1975) show that when $n \geqslant 10$, the distribution of S tends to normality, and a standard normal Z-score based on the S statistic and the variance $Var(S)$ can be computed by:

$$Z = \begin{cases} \frac{S-1}{\sqrt{Var(S)}} & S > 0 \\ 0 \quad \text{if} & S = 0 \\ \frac{S+1}{\sqrt{Var(S)}} & S < 0 \end{cases} \tag{5}$$

Hence, H_0 should not be rejected, in a two-sided trend test, if $|Z| \leqslant z_{\alpha/2}$ where α is the size of the significance level. A positive value of Z indicates an upward trend, whereas a negative value indicates a downward trend. When no trend exists ($Z = 0$), Z becomes the standard normal distribution (Hirsch et al., 1982). In this study, a trend was considered to be in evidence when the null hypothesis is rejected at a significance level of 5% (i.e., $\alpha = 0.05$) for a two-tailed test. A robust estimate for the trend magnitude, determined by Hirsch et al. (1982), is given by the slope estimator (β):

$$\beta = Median\left[\frac{(x_j - x_i)}{(j - i)}\right] \quad \text{for all } j > i \tag{6}$$

where x_i and x_j are the data values at times i and j, respectively.

Concerns emerge for the application of the Mann–Kendall test under the presence of positive serial correlation and/or cross-correlation in the data series. It is recognized that both can increase the probability of detecting a trend when, in fact, there is no trend, leading to the incorrect rejection of the null hypothesis of no trend while it is true (Lettenmaier et al., 1994; von Storch and Navarra, 1995; Yue et al., 2002). Several approaches have been proposed to eliminate the possibility of overestimation caused by serial correlation in the hydrologic series. The most common approach is to "pre-whiten" the series prior to applying the trend test (von Storch and Navarra, 1995). However, opinion varies on the impacts of pre-whitening, and other approaches were suggested (Yue et al., 2002, 2003; Bayazit and Onoz, 2007; Hamed, 2009). Here, the effect of serial correlation is not considered, because we apply the trend test to annual data values which are approximately independent and, hence, do not exhibit serial correlation.

On the other hand, the effect of spatial correlation has generally been disregarded in most hydrologic trend studies, despite the fact that neglecting the presence of spatial dependence among sites in a specific region might lead to misleading results (Douglas et al., 2000; Yue and Wang, 2002; Renard et al., 2008; Khaliq et al., 2009). In this study, we use the Regional Kendall's S test developed by Douglas et al. (2000) to account for the effect of spatial correlation in streamflow data.

3.2.2. Regional Kendall's S test

Douglas et al. (2000) developed a new test statistic named as regional average Kendall's S ($\overline{S}_m$) to evaluate the field (regional) significance of trends rather than local (at individual sites) significance. The regional Kendall's S is calculated as the average of S values for all individual sites by:

$$\overline{S}_m = \frac{1}{m}\sum_{k=1}^{m} S_k \tag{7}$$

where S_k is Kendall's S for the kth station in a region with m stations. Under the presence of cross correlation, the variance of $\overline{S}_m$ becomes

$$Var(\overline{S}_m) = \frac{\sigma^2}{m}[1 + (m-1)\bar{\rho}_{xx}] \tag{8}$$

where $\bar{\rho}_{xx}$ is the average cross-correlation coefficient of the region,

$$\bar{\rho}_{xx} = \frac{2\sum_{k=1}^{m-1}\sum_{l=1}^{m-k}\rho_{k,k+l}}{m(m-1)} \tag{9}$$

and $\rho_{k,k+l}$ is the cross-correlation coefficient between stations k and $k + l$,

$$\rho_{k,k+l} = \frac{\sigma^2}{Cov(S_k, S_{k+l})} \tag{10}$$

Finally, the test statistic Z_m for correlated data series is evaluated as:

$$Z_m = \overline{S}_m / \sqrt{Var(\overline{S}_m)} \tag{11}$$

In this study, the field significance of trends in mean annual flow, mean dry-season flow, and number of low-flow days are evaluated at the 5% significance level (i.e., $\alpha = 0.05$) for a two-tailed test.

3.2.3. Student's t-test

The student's t-test, used here to detect step-changes, is a classical parametric test used to check if the means of two independent groups are statistically different. The null hypothesis H_0 is that the means of two groups are equal; whereas the alternative hypothesis H_1 is that the means are not equal. Basically, the test assumes that the data are normally-distributed and the time of change is known (Kundzewicz and Robson, 2000). For two groups with unequal variances the test statistic, t, is given by (Helsel and Hirsch, 1992):

$$T = \frac{(\overline{x_1} - \overline{x_2})}{\sqrt{\frac{s_1^2}{n_1} + \frac{s_2^2}{n_2}}} \tag{12}$$

where x_1, s_1 and n_1 are the mean, the sample standard deviation, and the number of observations of the first group, respectively, and x_2, s_2 and n_2 are the mean, the sample standard deviation, and the number of observations of the second group, respectively. Also, the degrees of freedom, df, is calculated approximately as (Helsel and Hirsch, 1992):

$$df = \frac{\left(s_1^2/n_1 + s_2^2/n_2\right)^2}{\frac{\left(s_1^2/n_1\right)^2}{(n_1-1)} + \frac{\left(s_2^2/n_2\right)^2}{(n_2-1)}} \tag{13}$$

All step-change results herein are evaluated at the 5% significance level (i.e., $\alpha = 0.05$) for a two-tailed test. For sample sizes larger than 40 ($n > 40$), the z-test statistic is calculated instead of a t-test statistic. For the purpose of step-change analysis, streamflow time series are divided into two parts: a 10-year long period (1941–1950, pre-irrigation) and a 20-year long period (1961–1980, post-irrigation). The first period is only 10 years due to the lack of groundwater records before 1941, and the need to select common periods across all stations for spatial comparison. Even so, only nine wells with sufficient data could be found near the stream gauges for this analysis. The interval 1951–1960 is the transition period and was discarded to allow for a less ambiguous step-change detection. To attribute the observed changes in streamflow to either changes in precipitation or in groundwater, monthly precipitation and daily water table data nearby were also analyzed by the same approach. The streamflow, groundwater and precipitation sites used in the step-change analysis are shown in Fig. 4.

4. Results and discussion

4.1. Regional patterns of groundwater–surface water connection

The greatest impact of irrigational pumping is likely to be observed in areas where streams are in hydraulic connection with the groundwater system since, in such areas, streams receive a significant portion of their inflow from the groundwater. The amount of groundwater contribution to streamflow varies depending on the hydrogeologic and climatic conditions. The key is whether a stream is predominantly surface runoff- or groundwater-fed. In arid regions with isolated summer thunder storms, surface runoff is the primary source for streamflow, and the water table is below the stream bed. In humid climates with frequent rain, infiltration is favored, which recharges the groundwater and enter the streams as baseflow long after the rain events. Controlling this partition (surface runoff vs. infiltration) is also terrain slope and soil permeability. The hydro-climatic conditions across the High Plains exhibit a north–south increase in temperature, a west–east increase in annual precipitation, a north–south and a central-east decrease in aquifer thickness, and a heterogeneous and anisotropic distribution of horizontal hydraulic conductivity. Thus, it is likely that there are significant spatial variations in the degree of hydraulic connectivity between groundwater and streamflow. There are several indicators that can tell us whether a stream is primarily fed by surface runoff (locally or upstream) or by groundwater inflow, based on simple analyses of precipitation, water table and streamflow. Streamflow stations used here were selected out of 64 stations listed in Table 2 based on the following criteria: (1) all have continuous daily measurements, (2) all record the flows from approximately the same size of drainage area (±15%), and (3) none

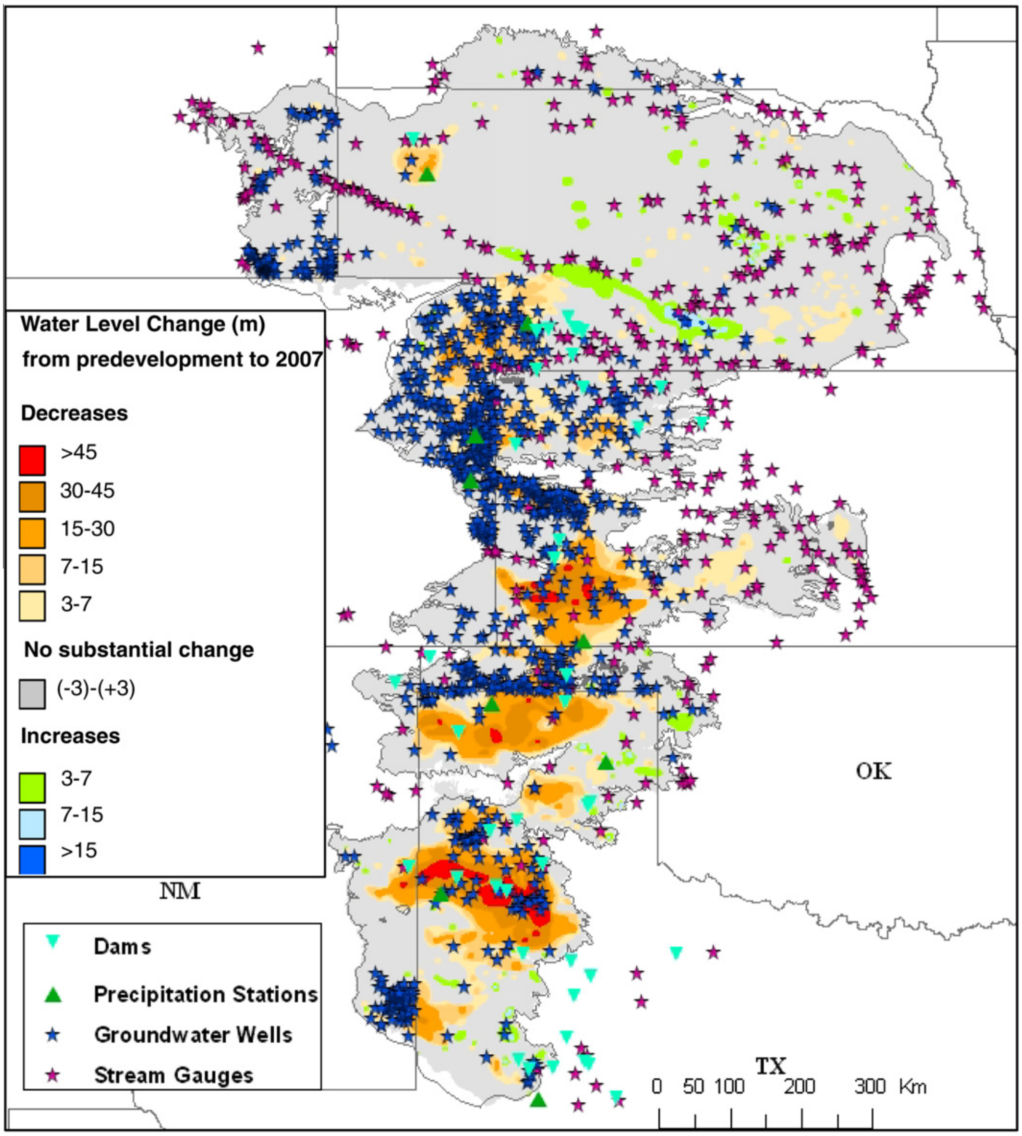

Fig. 3. Map with all the hydrologic sites examined for this study. Base map (McGuire, 2009) shows the water-level changes in the High Plains aquifer from pre-development to 2007.

are affected by dams. The water table data belonged to the well with the most number of observations closest to the associated stream gauges. The locations of the streamflow, groundwater and precipitation sites used in the analysis of groundwater–surface water connection are shown in Fig. 5.

First we examine the phase relationship between the seasonal cycle of streamflow and that of the local rainfall and water table. Local rainfall is a good surrogate for surface runoff and should have similar seasonal patterns. If the peak of streamflow leads the peak of rainfall, then the latter is not likely the main source. The phase relationships of the seasonal cycle between local rainfall and streamflow for selected sites are shown in Fig. 6 (first column). From north to south (a–f), a pattern seems to emerge; in the north (Nebraska and Colorado), streamflow peaks before local rainfall, a clear indication that the latter is not the main source for streamflow, and there is another mechanism causing discharge to increase in early spring. The peak of rainfall in late spring/early summer is typical since much of precipitation occurs in the form of local thunderstorms during the growing season (April–September) (Weeks et al., 1988). However, the streamflow peak occurs much earlier, in the spring, suggesting that the flow regime is controlled by the groundwater which is sourced in the Rockies to the west and responds strongly to seasonal snowmelt (Gutentag et al., 1984). Large-scale west–east groundwater flow in the highly permeable Ogallala formation of the aquifer is well documented (Gutentag et al., 1984; Weeks et al., 1988; Miller and Appel, 1997). This suggests that, in the northern part of the High Plains, groundwater is the primary source of streamflow, and, therefore, changes in groundwater storage will affect rivers more significantly. This is not surprising since Nebraska is recognized as one of the regions with the highest groundwater contribution to streams (up to 90%) across the USA, due to the highly-permeable sandy soils underlying the Nebraska Sand Hills that provide important recharge areas for the aquifer (Winter et al., 1998; Chen et al., 2003; Kollet and Zlotnik, 2003; Wen and Chen, 2006). Moving southward, rainfall becomes gradually more in phase with streamflow, indicating the increasing contribution from surface runoff in response to local rainfall events.

Second, we examine the phase relation between seasonal water table and streamflow. If the streamflow peak more or less coincides with the water table peak, there is further evidence that the water table is the main source. The seasonal cycle plots of streamflow vs.

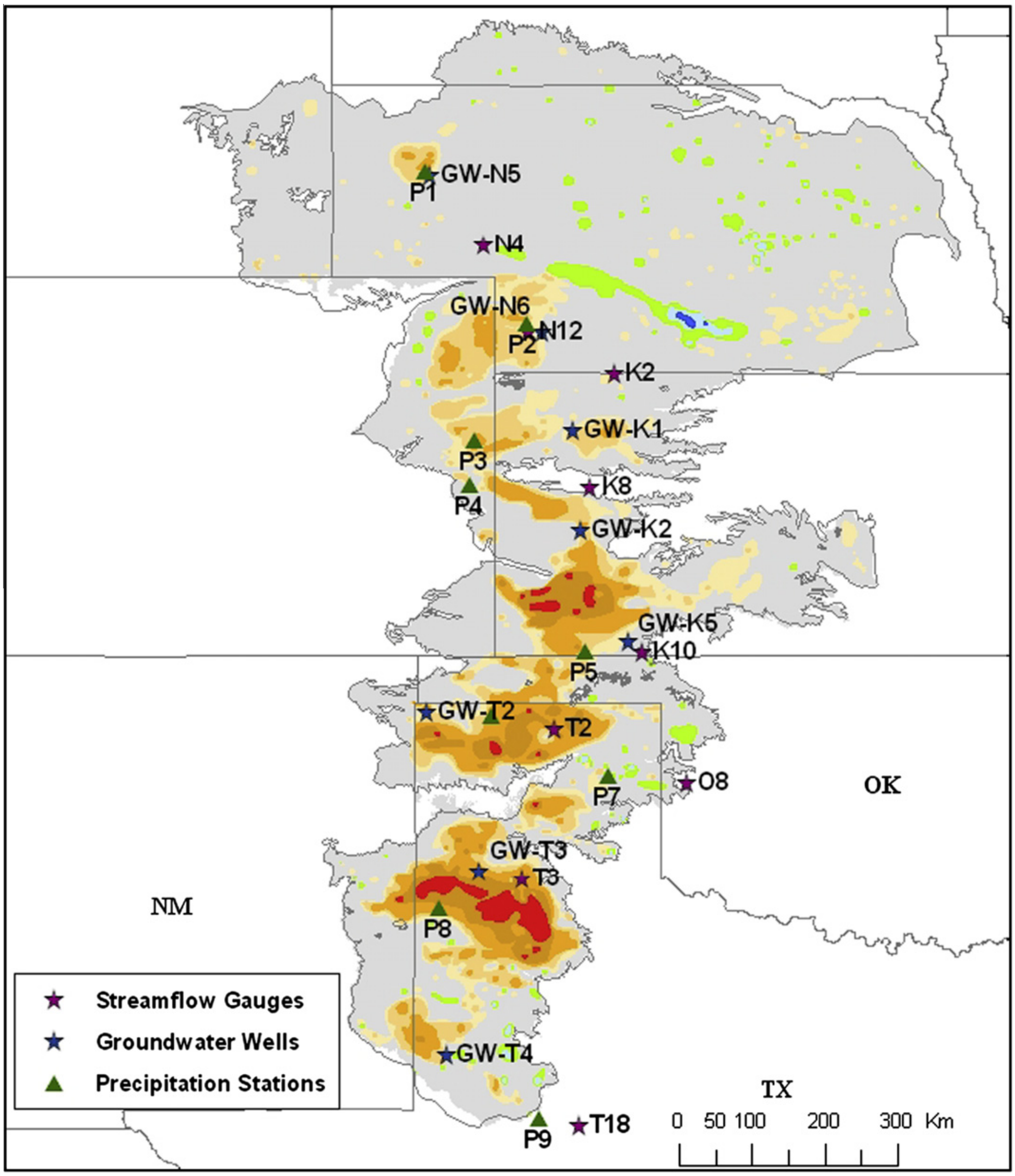

Fig. 4. Locations of the streamflow, groundwater and precipitation sites used in the step-change analysis.

water table elevation are shown in Fig. 6 (second column). The poor quality of the groundwater time series prevents a clear analysis, but a similar pattern can be discerned. In the north (Nebraska and Colorado), the seasonal water table is in phase with that of streamflow, suggesting close relationship between the two; in the south, the water table appears to lag behind streamflow, suggesting that the rivers are leaking and recharging the groundwater.

A third indicator of the relative importance of local rainfall vs. groundwater contribution to streamflow is the temporal persistence or memory of the latter. Streams fed by groundwater are expected to exhibit less temporal variability at the shorter time scales but more persistence or autocorrelation. Surface runoff-fed streams, on the other hand, are expected to show more temporal variability but less autocorrelation. The autocorrelation plots are shown in Fig. 6 (third column) for the six streamflow time series. According to this analysis, an autocorrelation plot would display either a smoothly-decaying curve for a stream that is groundwater-fed (slow deterministic event), or a sudden-declining curve for a stream that is dominated by surface runoff (quick random event). Again the varying data quality prevents a clear interpretation, but the general pattern is that streams in the north (Nebraska and Kansas) exhibit a slower decay in the autocorrelation than in the south, suggesting a more stable source of inflow characteristic of groundwater contributions.

Finally, we examine the relative elevation between the water table and the adjacent stream bed along the six streams from north to south. If the water table is higher than the stream bed, it is a clear indicator that the former is flowing into the latter; the lower streams function as sink drains for the groundwater. The elevation comparisons are shown in Table 5, with the locations of the groundwater wells, which are the closest ones to the associated stream gauges, shown in Fig. 5 (green stars). For each well, the average water table depth is calculated based on the period of record available. This simple and crude analysis suggests that the streams in the north (Nebraska, Colorado, and Kansas) are most likely to be receivers of local groundwater. Note that even at sites where the groundwater is lower than the adjacent stream bed, groundwater may still be a source further down the drainage gradient, feeding regional rivers and wetlands. Many rivers in Texas

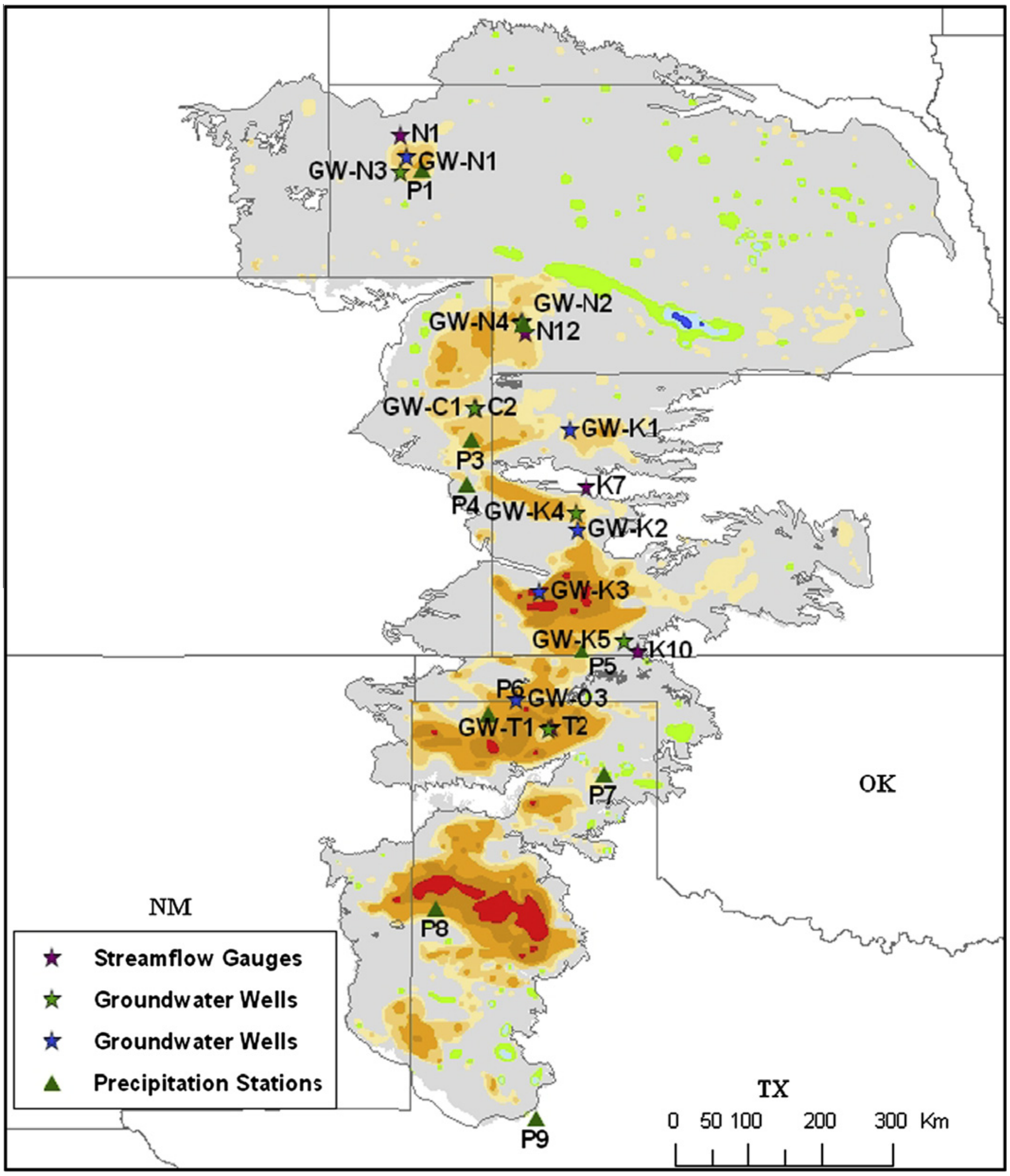

Fig. 5. Locations of the streamflow, groundwater and precipitation sites discussed in the analysis of groundwater–surface water connection. (Blue and green stars indicate the groundwater wells used in the seasonal cycle and elevation analysis, respectively.)

leak into the groundwater in the high lands, but receive groundwater in the lowlands and near the coastal regions (Schaller and Fan, 2009). It should also be noted that this analysis largely depends on the judgment of the user since the exact elevation of a streambed is difficult to establish. An elevation map was used on which an arbitrary point for the stream bed elevation was chosen based on the best judgment.

In conclusion, all the indicators we used to determine the degree of groundwater–streamflow connection in different hydro-climatic settings over the High Plains reveal a systematic decrease from north to south. Results from these analyses agree that the strongest connection is observed in Nebraska, and the weakest is in Texas, while parts in Colorado and Kansas act as a transition zone connecting the two end-members. The apparent N–S trend points out the regions susceptible to the expected effect of groundwater pumping on streamflow. Nevertheless, it should be noted that these results are constrained by the scarcity of groundwater data, and the main purpose of this analysis is to qualitatively determine the phase relationships between hydrologic variables to assess a general pattern in the strength of groundwater–streamflow connections.

4.2. Streamflow change analysis

4.2.1. Changes in annual mean streamflow

Trend analysis was first conducted on the mean annual streamflow of 64 gauging stations throughout the High Plains by using the Mann–Kendall test. The results are summarized in the 4th column of Table 6 and their spatial distribution is shown in Fig. 7a. Decreasing trends significant at 5% level are detected at 36 stream gauges of which 18 (50%) are in Nebraska, 1 (3%) in Colorado, 11 (31%) in Kansas, 4 (11%) in Oklahoma, and 2 (6%) in Texas. All the stations (100%) in Nebraska exhibit decreasing trends suggesting reduced annual mean streamflow over the period of record followed by 85% of the stations in Kansas, 50% in Oklahoma, 33% in Colorado, and 9% in Texas. The majority of stream gauges in Nebraska are located in the Republican River basin, where significant declines in water table resulting from groundwater pumping are

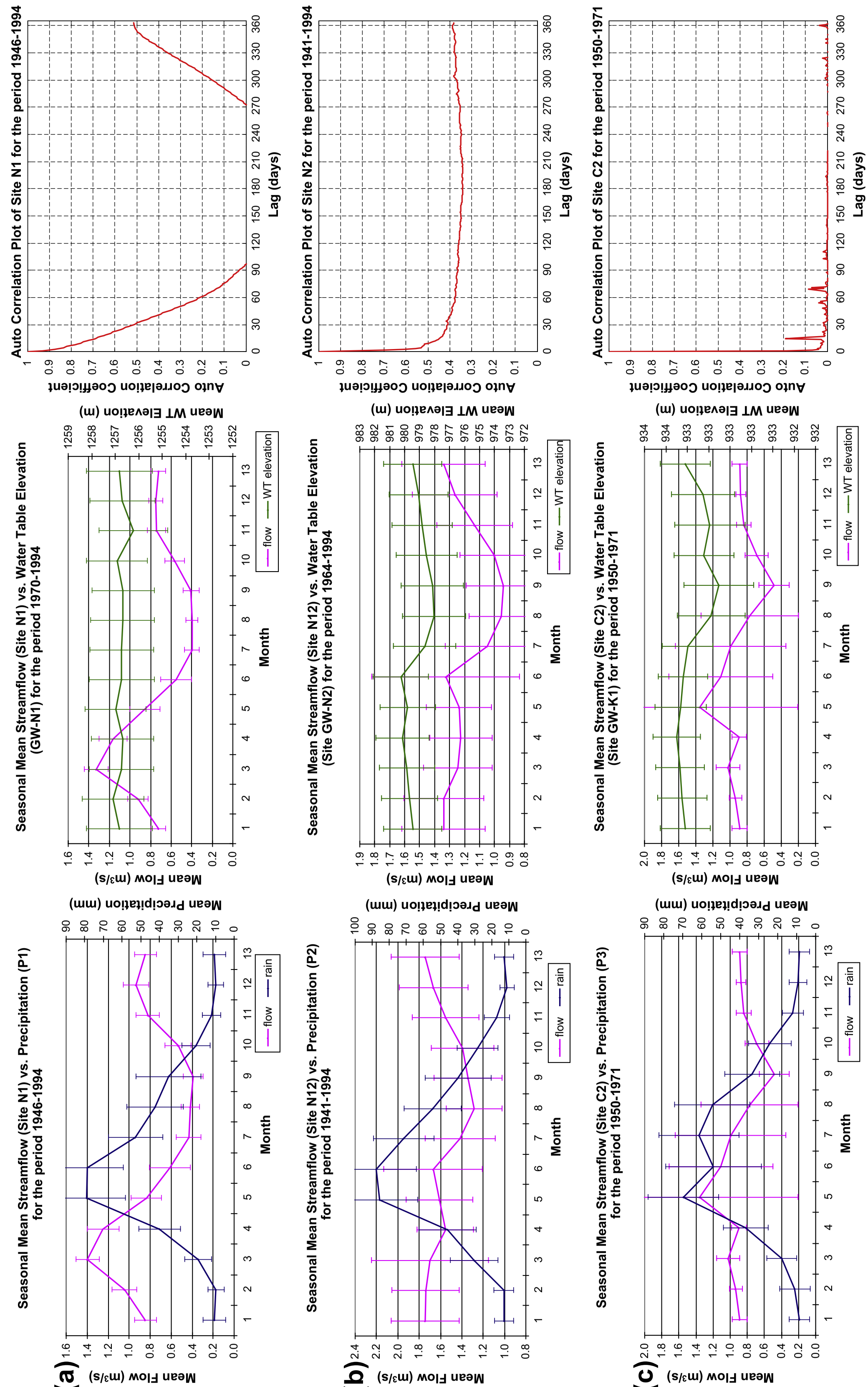

Fig. 6. Mean seasonal cycles of streamflow vs. local precipitation, streamflow vs. groundwater table elevation, and autocorrelation plots for the analyzed sites. (Error bars represent one standard deviation.)

observed in parts of Nebraska and in the adjacent parts of Colorado and Kansas. Other stream gauges in the same river basin in parts of Kansas (K1, K2, and K3) also show decreasing trends; however, out of three gauges in Colorado, two have insignificant trends, likely

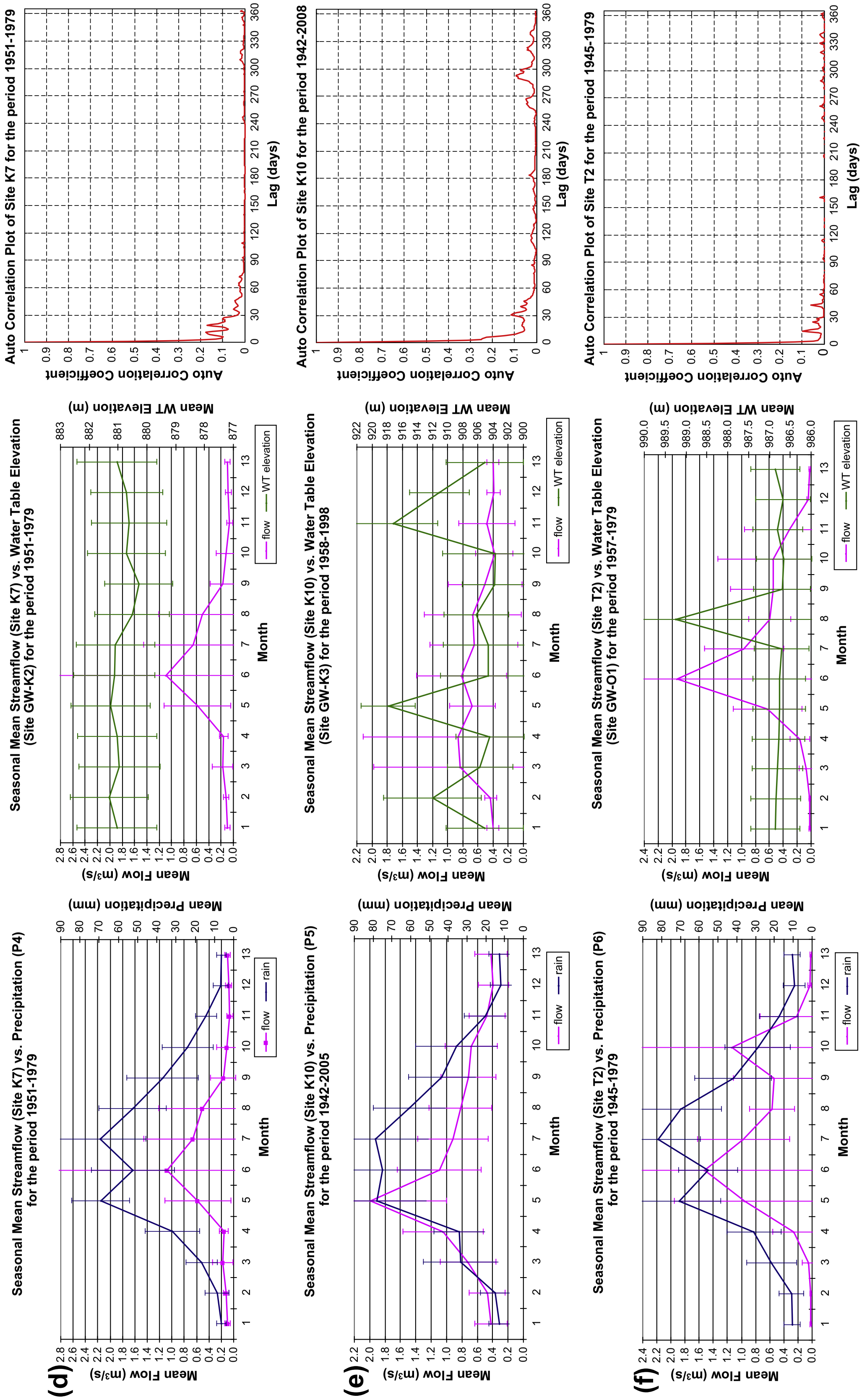

Fig. 6 (*continued*)

due to their short record period. The trend results for the Nebraska stream gauges, except for three (N4, N11, and N17), generally agree (83%) with those of Wen and Chen (2006), who analyzed the entire USGS stream gauges in Nebraska. Most of the stations with a signif-

Table 5
List of the streambed and mean water table elevations and their connection status.

Stream gauge	Well ID number	Well name	Mean WT elevation (m)	Streambed elevation (m)	Connection to stream
N1	420530103104001	GW-N3	1247	1223	Yes
N12	403111101405301	GW-N4	981	954	Yes
C2	393700102150000	GW-C1	1126	1122	Yes
K7	383046100594901	GW-K4	901	804	Yes
K10	370857100234601	GW-K5	718	660	Yes
T2	TWDB-354401	GW-T1	898	903	No

Table 6
Trend test results of mean annual flow, dry-season flow and number of low-flow days. (Stream sites in bold represent the ones under the dam effect.)

	Stream sites	Record period	Annual mean flow trends	Dry-season mean flow trends	Low-flow days trends
1	N1	1946–1994	Decreasing	Insignificant	Decreasing
2	**N2**	1946–1991	Decreasing	Insignificant	Decreasing
3	**N3**	1945–1991	Decreasing	Decreasing	Insignificant
4	N4	1930–1991	Decreasing	Decreasing	Insignificant
5	N5	1935–2008	Decreasing	Decreasing	Insignificant
6	N6	1932–2008	Decreasing	Decreasing	Insignificant
7	N7	1940–2008	Decreasing	Decreasing	Increasing
8	N8	1940–2008	Decreasing	Decreasing	Insignificant
9	N9	1947–1994	Decreasing	Insignificant	Insignificant
10	N10	1950–2008	Decreasing	Decreasing	Increasing
11	**N11**	1946–1993	Decreasing	Decreasing	Decreasing
12	N12	1941–1994	Decreasing	Decreasing	Increasing
13	**N13**	1946–1993	Decreasing	Insignificant	Increasing
14	**N14**	1950–2008	Decreasing	Decreasing	Increasing
15	N15	1949–1994	Decreasing	Decreasing	Increasing
16	**N16**	1935–2008	Decreasing	Decreasing	Increasing
17	N17	1946–1986	Decreasing	Insignificant	Decreasing
18	N18	1937–2008	Decreasing	Decreasing	Increasing
19	C1	1950–1976	Insignificant	Insignificant	Insignificant
20	C2	1950–1971	Insignificant	Insignificant	Increasing
21	C3	1946–1986	Decreasing	Insignificant	Insignificant
22	K1	1959–2008	Decreasing	Insignificant	Insignificant
23	K2	1946–2008	Decreasing	Decreasing	Increasing
24	K3	1929–2006	Decreasing	Insignificant	Insignificant
25	**K4**	1945–2008	Decreasing	Decreasing	Insignificant
26	**K5**	1946–1984	Insignificant	Insignificant	Increasing
27	K6	1966–1986	Decreasing	Insignificant	Insignificant
28	K7	1951–1979	Decreasing	Insignificant	Insignificant
29	K8	1939–2008	Decreasing	Decreasing	Increasing
30	K9	1965–2008	Decreasing	Decreasing	Increasing
31	K10	1942–2008	Decreasing	Decreasing	Decreasing
32	K11	1968–1990	Decreasing	Decreasing	Increasing
33	**K12**	1922–2008	Insignificant	Insignificant	Insignificant
34	**K13**	1944–2007	Decreasing	Decreasing	Increasing
35	O1	1942–1965	Insignificant	Insignificant	Decreasing
36	O2	1965–1993	Insignificant	Insignificant	Insignificant
37	O3	1939–1964	Insignificant	Insignificant	Insignificant
38	**O4**	1937–1993	Decreasing	Decreasing	Increasing
39	**O5**	1937–2008	Decreasing	Decreasing	Insignificant
40	O6	1942–1976	Decreasing	Insignificant	Decreasing
41	O7	1937–1993	Decreasing	Decreasing	Decreasing
42	O8	1937–2008	Insignificant	Insignificant	Decreasing
43	T1	1940–2008	Insignificant	Insignificant	Insignificant
44	T2	1945–2008	Decreasing	Insignificant	Increasing
45	T3	1939–1973	Insignificant	Insignificant	Increasing
46	**T4**	1964–1986	Insignificant	Decreasing	Increasing
47	**T5**	1939–2008	Insignificant	Insignificant	Insignificant
48	T6	1939–2008	Insignificant	Insignificant	Insignificant
49	**T7**	1924–1949	Insignificant	Insignificant	Insignificant
50	**T8**	1923–2008	Insignificant	Insignificant	–
51	**T9**	1922–2008	Insignificant	Insignificant	–
52	**T10**	1925–2008	Insignificant	Insignificant	–
53	**T11**	1967–2008	Insignificant	Insignificant	–
54	**T12**	1959–1979	Insignificant	Insignificant	Decreasing
55	**T13**	1954–2008	Insignificant	Insignificant	–
56	**T14**	1980–2008	Insignificant	Insignificant	–
57	**T15**	1965–2002	Decreasing	Insignificant	–
58	**T16**	1980–2008	Insignificant	Insignificant	–
59	**T17**	1958–2008	Insignificant	Insignificant	–
60	T18	1939–1994	Insignificant	Insignificant	Insignificant
61	T19	1968–2008	Insignificant	Insignificant	–
62	T20	1961–2008	Insignificant	Insignificant	Insignificant
63	**T21**	1967–1980	Insignificant	Insignificant	Insignificant
64	**T22**	1964–2008	Insignificant	Insignificant	–

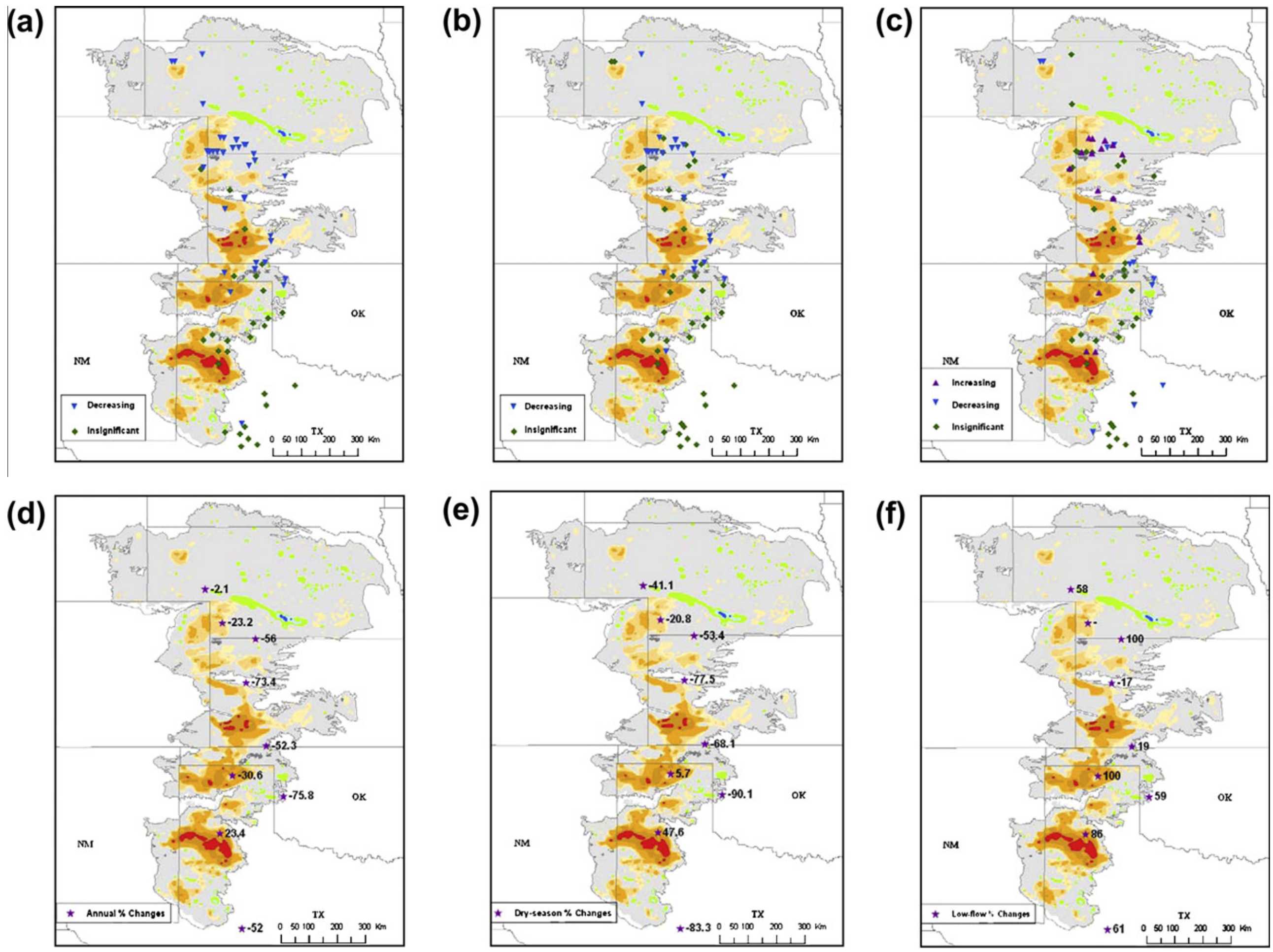

Fig. 7. Spatial distribution of trend analysis based on (a) mean annual streamflow, (b) mean dry-season streamflow, (c) number of low-days, and step-change analysis based on (d) long-term streamflow, (e) dry-season streamflow, (f) number of low-flow days (▼: stream gauge with decreasing trend, ▲: stream gauge with increasing trend, ⧫: stream gauge with no trend, and ★: stream gauge with % change).

icant decreasing trend are located in the Republican River basin, which coincides with the results of Szilagyi (1999), who also observed streamflow depletion in the same basin. Likewise, the trends detected at the gauges in the Oklahoma panhandle (O2, O3, O4, and O5) support the results of the Wahl and Wahl (1988) study. Texas is the state with the most insignificant trends, which is expected since rivers in this region are primarily fed by summer surface runoff as shown earlier.

We analyzed the step-changes in the monthly discharge time series by using Student's *t*-test. Detailed results are shown in Table 7, and Fig. 7d illustrates the percent change in streamflow at each gauge from period 1 (1941–1950) to period 2 (1961–1980). The rate of streamflow change varied from 23% more flow at gauge T3 to 76% less flow at gauge O8 between the two periods. The only stream gauge displaying increased streamflow from period 1 to period 2 is T3; but it does not have a substantial number of measurements for the second period. Gauge O8 in western Oklahoma shows a significant step-change and the largest decrease in streamflow; however no significant long-term trend could be detected by the Mann–Kendall test. This is because the rate of decline in annual streamflow is very steep from the 1940s to the 1970s, but has leveled off since.

The observed changes in streamflow can be related to either changes in precipitation or in groundwater inflow or both. Table 8 summarizes the step-change results of monthly mean precipitation, streamflow and groundwater data grouped for the same region. Although precipitation did not change significantly between the two periods, streamflow in the Republican River basin (gauge N12), in the Smoky Hill River basin (gauge K8) and in the Cimarron River basin (gauge K10) decreased between the pre-irrigation and post-irrigation periods. In contrast, groundwater data in the same regions exhibit significant decreases between the two periods implying that pumping is the major cause of the observed streamflow depletion in these regions. In fact, the decline in water levels is significant at all groundwater sites analyzed, but the attribution is not apparent in all cases. For example, although both discharge (gauge K2) and water table elevation decreased significantly in the Beaver Creek, a tributary of the Republican River basin, precipitation also decreased from the first period to the second, hence the main cause of reduced streamflow is unclear. Despite the significant reduction in groundwater levels, no statistically significant trends could be detected at the Texas stream gauges, which confirms our earlier findings that these rivers are not connected to the groundwater system. This is reasonable, since Texas was one of the states where irrigational pumping had started in as early as 1900s with a rapid increase between the mid-1940s and 1959, followed by a much slower rate of increase between 1959 and 1980. The area of irrigated land in 1980 on the High Plains of Texas was approximately equal to the 1959 level as a result of reduced groundwater availability in the Southern High Plains (Ryder, 1996). Therefore, the connection of groundwater with the local river system was already lost by the 1960s, so that pumping did not exert further influence on streamflow after that time. Nonetheless, this does not rule out that streamflow farther down the gradient,

Table 7
Step change test results of monthly mean streamflow.

Stream sites	Period 1 (1941–1950)			Period 2 (1961–1980)			Z statistic	p-Value	Two-tail test Trend (5%)	Change in means (%)
	Mean	Variance	Number of Observations	Mean	Variance	Number of Observations				
N4	1.979	0.909	120	1.938	1.0227	240	0.376	0.7067	Insignificant	−2.1
N12	2.050	0.099	120	1.575	0.302	240	10.402	0.0000	Significant	−23.2
K2	0.726	1.121	56	0.320	0.760	240	2.670	0.0076	Significant	−56.0
K8	1.176	9.692	120	0.312	1.2319	240	2.947	0.0032	Significant	−73.4
K10	1.672	9.147	96	0.797	2.665	240	2.683	0.0073	Significant	−52.3
T2	0.694	9.597	60	0.481	3.459	225	0.508	0.5619	Insignificant	−30.6
O8	1.638	9.231	120	0.397	0.684	240	4.394	0.0000	Significant	−75.8
T3	0.068	0.065	120	0.084	0.283	156	−0.326	0.7459	Insignificant	23.4
T18	0.346	2.205	120	0.166	0.575	240	1.247	0.2124	Insignificant	−52.0

Table 8
Summarized step-change test results of monthly mean streamflow, precipitation, and water table elevation.

Precipitation sites	Annual trend results (5%)	Stream sites	Annual trend results (5%)	Groundwater sites	Annual trend results (5%)
P1	Insignificant	N4	Insignificant	GW-N5	Significant
P2	Insignificant	N12	Significant	GW-N6	Significant
P3	Significant	K2	Significant	GW-K1	Significant
P4	Insignificant	K8	Significant	GW-K2	Significant
P5	Insignificant	K10	Significant	GW-K5	Significant
P6	Insignificant	T2	Insignificant	GW-T2	Significant
P7	Significant	O8	Significant	–	
P8	Insignificant	T3	Insignificant	GW-T3	Significant
P9	Insignificant	T18	Insignificant	GW-T4	Significant

where the water table does rise above the streambeds, can be affected because groundwater not only sustains local streams but also regional streams, particularly in arid environments (Schaller and Fan, 2009).

Additionally, we assessed the regional significance of trends in annual mean streamflow using the Regional Kendall's S test for the period of most intensive irrigation development (1941–1980). The study area was divided into two main regions as ''Region 1 (North)'' and ''Region 2 (South)'' based on the observed patterns in groundwater–surface water connection. That is, the first region included streams in Nebraska, Colorado, and Kansas (the first 34 gauges from N1 to K13) which were revealed to be predominantly influenced by groundwater, while the second region contained the remaining 30 gauges in Oklahoma and Texas (from O1 to T22) that were mostly surface runoff-fed. Results indicated that identified trends at individual sites in Region 1 are field significant at the 5% level, confirming that there is a regional decreasing trend in annual streamflow in the north of the study area in response to pumping. On the other hand, the observed annual decreases in streams in Region 2 were not field significant, and, thus, streamflow depletion is not regionally consistent. Nevertheless, it should be noted that substantial dissimilarities in record periods of stream gauges in Region 2 most likely have affected the analysis results.

In summary, all trend, step change and regional analysis of mean annual streamflow reveal a significant flow reduction in the North and less so in the South. This is consistent with the regional patterns emerged from the earlier analysis of streamflow–groundwater connection, that is the effect of irrigational pumping is more prominent on the rivers in the Northern High Plains with a gradual decrease towards the Southern High Plains. Also, we note that the results of step and trend changes are not affected by data gaps in the time series since both the Mann–Kendall and Student's t-test are insensitive to missing data (Kundzewicz and Robson, 2000).

4.2.2. Changes in dry-season streamflow

In the High Plains, irrigation is applied most intensively from late June through August due to low precipitation and high crop water demand (Moore and Rojstaczer, 2001). Therefore, the effect of pumping is likely to be more clearly observed on July and August streamflow. For this reason, mean annual July and August, referred to as ''dry-season'' hereafter, streamflow time series of the same 64 gauging stations are analyzed using the Mann–Kendall trend test. The resulting trends are shown in the 5th column of Table 6 and Fig. 7b shows their spatial distribution. Surprisingly, the number of stations with significant downward trends decreased from 36 in mean annual streamflow to 24 in dry-season streamflow. Of the 24 stream gauges with decreasing trends, 13 (54%) are in Nebraska, 7 (29%) in Kansas, 3 (13%) in Oklahoma, and 1 (4%) in Texas. No stream gauges in Colorado had significant trends.

Among the 12 stream gauges that went from decreasing trend in the mean annual flow to no-significance in the dry-season flow, four (N2, N13, C3, and T15) are under the influence of dams. (The regulated stream gauges over the study area are emphasized in bold in Table 6.) Hence, it is possible that summer discharge rates measured at these gauges have been affected by flow regulations which tend to dampen seasonal variability and increase dry-season flow (Haddeland et al., 2006b). As for the other gauges, the high natural variability of streamflow during the summer months might be hampering the detection of trends by relatively simple statistical methods (Miller and Piechota, 2008). Widespread thunderstorms and extreme weather events across the region from April to September lead to large variations in rainfall as well as runoff, especially in the Southern High Plains where streamflow is maintained mainly by rainfall-generated surface runoff. This might be the reason why Kansas is the most affected state with noticeably fewer number of trends in dry-season as compared to the number of annual trends; further south, summer thunderstorms dominate both annual and summer streamflow. Fig. 8 shows the mean July–August time series of those gauges that fail to show significant trends in dry-season flow but have decreasing trends in annual flow. The time series of each of these gauges clearly show a decreasing trend, however the decrease is not statistically significant. Although most of these gauges have missing or relatively shorter period of records, this can not be the main reason of insignificant dry-season trends, since there are gauges with similar record periods that show significant decreasing trends both in annual and dry-season flow.

One other possible explanation for the decrease in the number of dry-season trends might be the lag between groundwater pumping and streamflow reduction. That is, summer pumping may lead to a fall and winter streamflow depletion; hence the pumping signal is stronger in the annual flow and can not be detected in dry-season flow. It should be noted that this is the case if the water table is lowered over large regional scales and the groundwater is feeding the downstream rivers.

The difference in the size of drainage area among the gauges could be another factor, because the larger the river basin, the longer are the flow paths, and hence the longer the response time between the groundwater and the river signals. However the plot of Mann–Kendall Z-scores against the drainage basin area indicates no such relationship (not shown).

Hantush (1964) recognized that there are two components leading to total streamflow depletion: reduced baseflow and induced streamflow infiltration (or seepage to the groundwater below). Earlier studies argue that although both components are caused by seasonally-pumped wells, the impacts of the former continue during the non-pumping period, while the residual effects of the latter disappear as the pumping stops (Chen and Yin, 2001; Chen and Shu, 2002). Chen and Yin (2001) show that as the hydraulic head difference between the stream and the aquifer increases, i.e., as the water levels continue to decline, the rate of baseflow reduction also increases, but the streamflow infiltration does not occur until a reversed hydraulic gradient is established between the two. Hence, it is reasonable to assume that the rivers that fail to show a significant trend in the dry-season, but significantly decrease annually, are affected only by the first component of total streamflow depletion which is baseflow reduction. The pumping-induced stream infiltration does not happen in these rivers; most probably because a reverse hydraulic gradient is not established due to the high rate of summer pumping which lowers the water table so quickly that the connection between the river and the aquifer is lost. After summer, when the pumping stops, the rivers re-connect with the aquifer as the water levels start to recover; nevertheless streamflow continues to be depleted during the non-pumping period as a result of ongoing baseflow reduction. Since the water levels cannot recover fully back to the previous conditions before the beginning of the next pumping season, total depletion will tend to increase after each pumping season. Additionally, pumping effect of the wells farther away from the rivers also kicks in during the post-pumping period further reducing the annual streamflow (Chen and Yin, 2001).

Table 9 shows the step-change analysis statistics and *t*-test results for the monthly dry-season flow and Fig. 7e marks the percent change at each gauge from period 1 (1941–1950) to period 2 (1961–1980). The results are similar to the trend results in that less number of sites with significant changes is detected than the annual mean step changes. Between the two periods, the rate of streamflow change varied from 48% more flow at gauge T3 to 90% less flow at gauge O8; larger than the observed annual step changes for the same gauges in both directions. The increase in flow at gauge T3 during the dry-season is twice of the annual flow increase indicating that the river is mostly recharged in summer. Significant step changes between pre- and post-irrigation periods are observed only at gauges N12 and O8, which are unlikely related to changes in precipitation since corresponding data do not reveal any significant step-changes. However, it is also not certain if the observed streamflow depletion is caused by groundwater pumping due to insufficient summer records (Table 10). All Kansas gauges (K2, K8, and K10) with significant annual step-changes fail to do so in dry-season flow consistent with the trend results. Still, the flow rate at all three gauges has decreased at least more than 50% from the earlier period to the later. It is particularly interesting that despite significant decreases in both precipitation and groundwater levels, no significant changes could be detected at gauge K2, likely because of the limited summer records during the first period.

Again, the regional significance of trends in annual dry-season streamflow was assessed by the Regional Kendall's *S* test for two regions (Region 1 and Region 2) over the period of 1941–1980. Results showed that trends in dry-season were not field significant in both of the regions implying that the individually detected decreasing trends might have occurred by chance.

4.2.3. Changes in the number of low-flow days

The third and last hydrologic variable analyzed for streamflow reduction is the annual number of low-flow days in the discharge records. To establish a statistically significant low-flow value for the streamflow time series, a typical 7-day 10-year (7Q10) low flow index is used which is computed by finding the lowest average discharge that occurs over any 7-consecutive days at a recurrence interval of 10 years (Gupta, 1995; Smakhtin, 2001; Risley et al., 2008). The number of days with a flow record equal to or less than the 7Q10 statistic in each year is counted within the time series data and the total number is subjected to the trend and step-change analysis. Since a reliable 7Q10 value could not be determined for non-daily time series, the stream gauges without daily records are discarded from the analysis reducing the total number

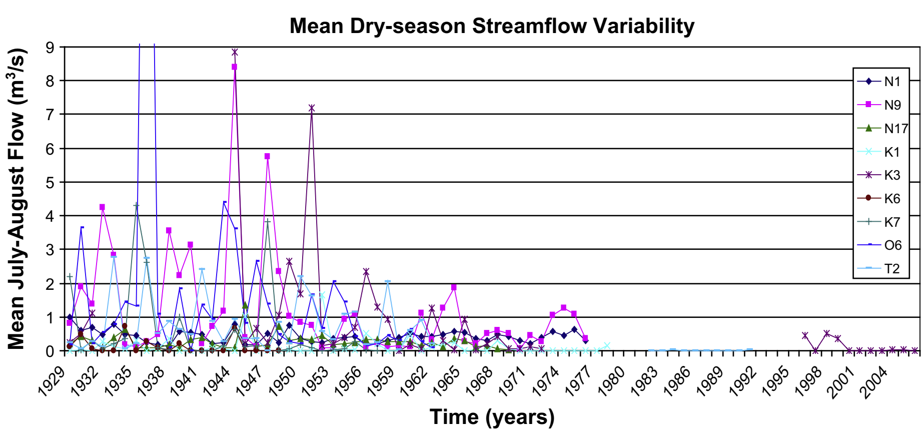

Fig. 8. Time series of mean July–August flow at the gauges that fail to show significant trends in dry-season flow but have decreasing trends in the mean annual flow.

Table 9
Step change test results of monthly dry-season (mean July–August) streamflow.

Stream sites	Period 1 (1941–1950)			Period 2 (1961–1980)			t Statistic	Degrees of freedom	p-Value	Two-tail test Trend (5%)	Change in means (%)
	Mean	Variance	No. of observations	Mean	Variance	No. of Observations					
N4	0.749	0.370	20	0.441	0.296	40	1.911	35	0.0642	Insignificant	−41.1
N12	1.754	0.030	20	1.390	0.279	40	3.956	53	0.0002	Significant	−20.8
K2	1.107	0.566	10	0.516	0.976	40	2.076	18	0.0525	Insignificant	−53.4
K8	2.719	25.929	20	0.611	2.5056	40	1.809	21	0.0848	Insignificant	−77.5
K10	2.285	15.561	16	0.729	1.392	40	1.550	16	0.1406	Insignificant	−68.1
T2	0.692	1.222	10	0.731	0.645	38	−0.105	12	0.9181	Insignificant	5.7
O8	1.099	1.177	20	0.109	0.028	40	4.058	19	0.0007	Significant	−90.1
T3	0.050	0.019	20	0.073	0.032	26	−0.505	44	0.6161	Insignificant	47.6
T18	1.169	11.708	20	0.195	1.273	40	1.240	21	0.2287	Insignificant	−83.3

Table 10
Summarized step-change test results of monthly mean dry-season streamflow, precipitation, and water table elevation.

Precipitation sites	Dry-season trend results (5%)	Stream sites	Dry-season trend results (5%)	Groundwater sites	Dry-season trend results (5%)
P1	Insignificant	N4	Insignificant	GW-N5	NaN
P2	Insignificant	N12	Significant	GW-N6	Insignificant
P3	Significant	K2	Insignificant	GW-K1	Significant
P4	Insignificant	K8	Insignificant	GW-K2	Significant
P5	Insignificant	K10	Insignificant	GW-K5	NaN
P6	Insignificant	T2	Insignificant	GW-T2	NaN
P7	Insignificant	O8	Significant	–	
P8	Insignificant	T3	Insignificant	GW-T3	NaN
P9	Insignificant	T18	Insignificant	GW-T4	NaN

of stations from 64 to 53. The 7Q10 values at 38 of these stations are equal to zero.

The Mann–Kendall test results of the number of low-flow days are shown in the last column of Table 6 and the spatial distribution of trends are depicted in Fig. 7c. There are 10 (19%) stream gauges with decreasing, 19 (36%) with increasing, and 24 (45%) with insignificant trends. The number of increasing trends is nearly twice the number of decreasing trends. Of the 19 gauges with significantly increasing trends, 8 (44%) are in Nebraska, 1 (33%) in Colorado, 6 (46%) in Kansas, 1 (13%) in Oklahoma, and 3 (27%) in Texas. Almost half of the stations in Nebraska and Kansas exhibit increasing number of low-flow days indicative of rivers with less flow for longer periods. The majority of stations with significantly increasing trends is grouped in and around the Republican River basin, where significant decreasing trends in annual and/or dry-season streamflow are also observed earlier. Among the stations with significantly increasing number of low-flow days, there are only three gauges (C2, K5, and T3) without any significant trends in either annual or dry-season flow. From our earlier findings, Colorado and Kansas are already recognized as transition zones where local rivers are fed by both surface runoff and groundwater, hence, the observed increases in the number of low-flow days at these gauges have probably resulted from the decreasing summer precipitation detected at the nearby rainfall station P3 (Fig. 9b). However, the precipitation data associated with gauge T3 shows no such trend, therefore, the increase detected at this station might be related to an increase in temperature or a decrease in the number of heavy rain events since streamflow in Texas is known to be dominated by summer thunderstorms. The small drainage area of T3 might be an additional factor in shortening the response time to the changes in climate.

The greatest percentage of insignificant trends (67%) is observed in Colorado, followed by Texas (64%), Kansas (46%), Oklahoma (38%), and Nebraska (33%). Excluding Colorado, which has only three stations with relatively short periods of record, it is noted that the number of trends that could be detected significantly are lowest in the South with a gradual increase towards the North. This is also in agreement with our earlier results of streamflow–groundwater connection degree, that is, the Northern High Plains rivers are primarily fed by groundwater whereas the Southern rivers rely more on surface runoff. Of the 24 stations with insignificant trends, 11 have no significant trends in neither annual nor dry-season flow and are located in Colorado, Kansas, and, mostly, in Texas. The fact that Texas is the state with the greatest number of insignificant trends in number of low-flow days, as well as in annual and dry-season flow, is further indicative of the weak groundwater–streamflow connection in this region.

Out of 10 stream gauges with significantly decreasing number of low-flow days, 4 are in Oklahoma (50%), 4 in Nebraska (22%), 1 in Kansas (8%), and 1 in Texas (9%). Most of these gauges are located away from the areas of significant groundwater decline and three of them (N2, N11, and T12) are regulated. Hence, the observed decreases in low-flow days at these three stations are probably results of flow regulations. The decline in low-flow days at the Nebraska gauge N1 despite the significant decreases in annual streamflow and annual precipitation (Fig. 9a) indicates that the river is sustained by groundwater throughout the year. Because, even the total volume of flow decreases over the period of record, the days in which the flow rate drops below the 7Q10 value are not reduced. Unlike the other gauges in the Republican River basin, N17 shows a decreasing trend, likely because of the missing data after the 1980s (1946–1986). Low-flow rates generally appear after the 1980s in the records of most stations in the Republican River basin even though the annual groundwater pumpage did not increase much between 1974 and 1995 (see the insert in Fig. 2c) and the annual precipitation shows no significant trend (Fig. 9a). The reason of this might be the increased sensitivity of streamflow to depletion resulting from the continuous groundwater exploitation year after year (Chen and Yin, 2001) or the more significant use of surface water for irrigation in Nebraska as mentioned earlier. On the other hand, the gauges in the Oklahoma panhandle (O1, O6, O7, O8, and K10) that exhi-

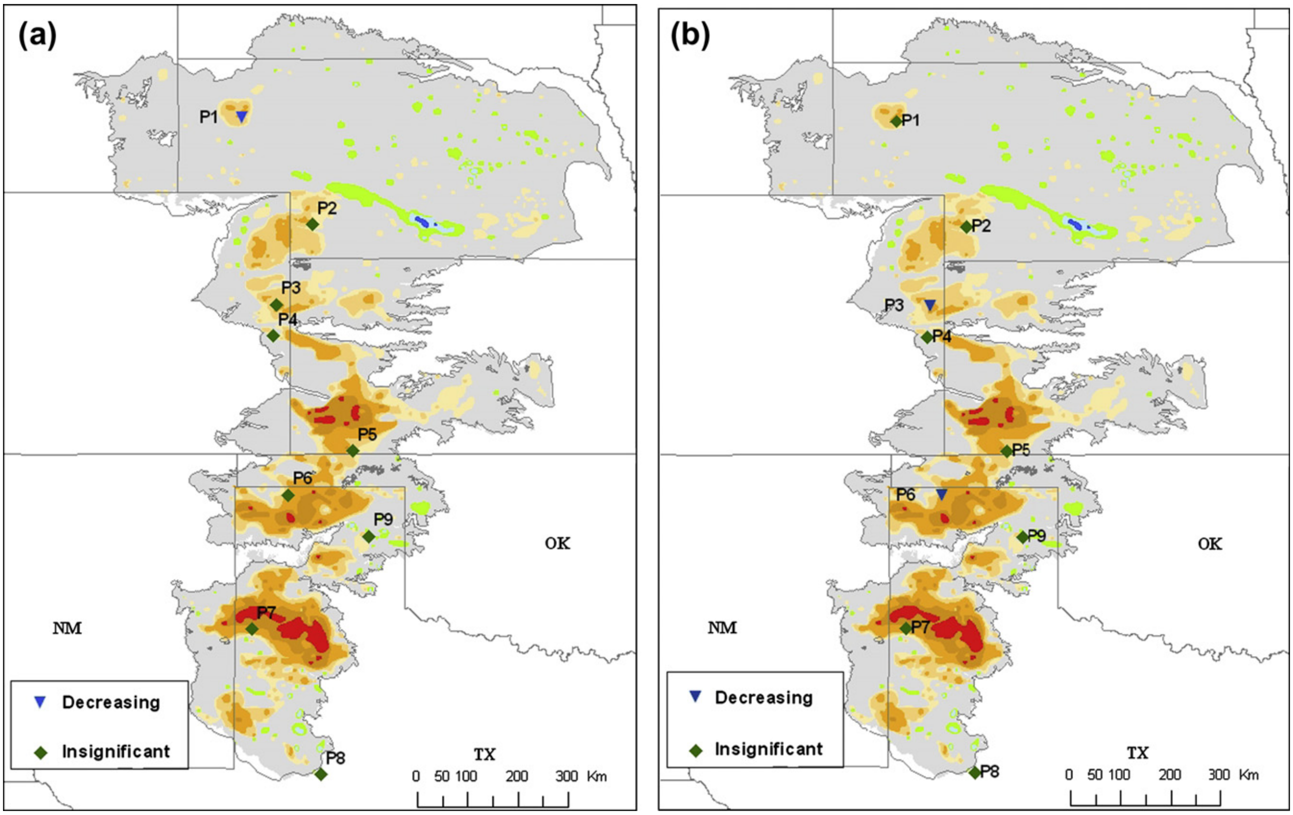

Fig. 9. Spatial distribution of trend analysis based on: (a) total annual precipitation, (b) total dry-season (mean July and August) precipitation (▼: precipitation station with decreasing trend, and ♦: precipitation station with no trend).

bit decreasing trends in the number of low-flow days are located in areas where small declines in groundwater levels (<3 m) are observed. Hence, any changes in streamflow have probably been minor.

Step-change analysis statistics and *t*-test results of the number of low-flow days are shown in Table 11 and the percent change at each gauge from the first period (1941–1950) to the second (1961–1980) are displayed in Fig. 7f. The results show that the number of low-flow days increased between the two periods at almost all stream gauges, but the increase is significant at only four (K2, T2, T3, and T18) of them. The only gauge that shows a decrease in the number of low-flow days from the first period to the second is K8 (−17.3%), but this gauge is regulated; again low-flow rates likely have been altered by flow regulations. The percent of increase is greatest at gauges K2 and T2 (100%), followed by the gauges T3 (86.3%), T18 (60.7%), O8 (59.3%), and K10 (18.5%). It is remarkable that all gauges in Texas exhibit significant increases in the number of low-flow days despite no significant step-changes could be detected at any of them in annual and dry-season flow as well as in the precipitation data. In fact, this further indicates that rivers in Texas are sustained by surface runoff since, although the total volume of flow has not changed, the low-flow frequency has increased. If these rivers were also sustained by groundwater, then they would show decreases in annual and/or dry-season streamflow as well. It has been already recognized that summer thunderstorms dominate streamflow in the Southern High Plains. Therefore, the increase in low-flow days at these gauges is most likely related to the changes in the number of extreme rainfall events.

Although precipitation and water table data could not be examined for such a step-change, earlier results of the corresponding annual step-changes in precipitation and groundwater can be used as

Table 11
Step change test results of annual number of low-flow days.

Stream sites	Period 1 (1941–1950)			Period 2 (1961–1980)			*t* Statistic	Degrees of freedom	*p*-Value	Two-tail test Trend (5%)	Change in means (%)
	Mean	Variance	No. of observations	Mean	Variance	No. of observations					
N4	1.300	6.678	10	2.050	7.103	20	−0.742	19	0.4671	Insignificant	57.7
N12	0.000	0.000	10	0.150	0.239	20	−1.372	19	0.1860	Insignificant	–
K2	13.40	218.80	5	190.30	18546.75	20	−5.677	21	0.0000	Significant	100.0
K8	43.40	3264.93	10	35.90	1523.15	20	0.374	13	0.7144	Insignificant	−17.3
K10	17.00	647.00	9	20.15	381.29	20	−0.330	12	0.7471	Insignificant	18.5
T2	45.33	1211.47	6	125.89	10046.77	19	−2.980	23	0.0067	Significant	100.0
O8	70.20	2515.51	10	111.80	4982.06	20	−1.859	24	0.0753	Insignificant	59.3
T3	169.70	6801.34	10	316.08	470.91	13	−5.469	10	0.0003	Significant	86.3
T18	172.70	7944.46	10	277.50	7148.79	20	−3.088	17	0.0067	Significant	60.7

an analogy. Thus, the significant step-changes in the low-flow days at gauge K2 from pre-irrigation to the post-irrigation period has probably resulted from the significant decreases both in precipitation and groundwater levels between the two periods since it has already been shown that rivers in Kansas are sustained by both surface runoff and baseflow.

Finally, the regional significance of identified trends in the number of low-flow days were evaluated over 1940–1980 for Region 1 and Region 2 resulting in a lack of field significance for both regions. Hence, the possibility that they might have occurred by chance could not be eliminated.

5. Summary and conclusions

The High Plains aquifer, in the Great Plains of USA, has undergone substantial declines in groundwater levels since the onset of widespread irrigational pumping in the 1940s. This study examined the annual and seasonal impacts of this long-term, large-scale groundwater pumping on streamflow regimes in the High Plains at the regional scale. We analyzed trends and step-changes in annual streamflow, dry-season flow and in the number of low-flow days at 64 and 9 stream gauges, respectively, in conjunction with changes in precipitation and water table. Also, we assessed the field significance of trends in those variables using a regional average test statistic to evaluate the effect of spatial correlation among the stream gauges studied.

Several indicators revealed spatial differences in the degree of hydraulic connection between groundwater and streamflow based on the hydro-climatic gradients across the High Plains. There is a systematic decrease in the degree of groundwater–streamflow connection from the Northern to the Southern High Plains. The trend and step-change results in mean annual streamflow confirm this spatial tendency: streamflow depletion is more significant in the North, gradually becoming less apparent towards the South. However, fewer gauges are detected with significant trends and step-changes in dry-season (mean July–August) flow. Various factors could have contributed to this such as: (1) dam regulations might have affected the summer flow rates, (2) large variations in summer rainfall might have impeded the trend detection, particularly in Kansas and Texas, (3) rivers downstream from the irrigated area might reflect the pumping signal later in the year due to the lag between groundwater level and streamflow response, and (4) rivers in areas of large water decline become disconnected from the aquifers due to extensive summer pumpage, and re-connect after summer when the pumping stops and water levels start to recover. The spatial distribution of the dry-season trends is in agreement with that of the annual trends; the largest number of significant decreasing trends is in Nebraska, and the greatest number of stations with insignificant trends is in Texas while both decreasing and insignificant trends are detected in between. Namely, the Republican River basin, the Arkansas River basin, and the Oklahoma panhandle are the regions with the most significant declines in annual and dry-season streamflow. A different pattern emerges in the spatial distribution of trend and step-change results of the number of low-flow days; not only decreasing but also increasing trends are observed. Increasing trends are

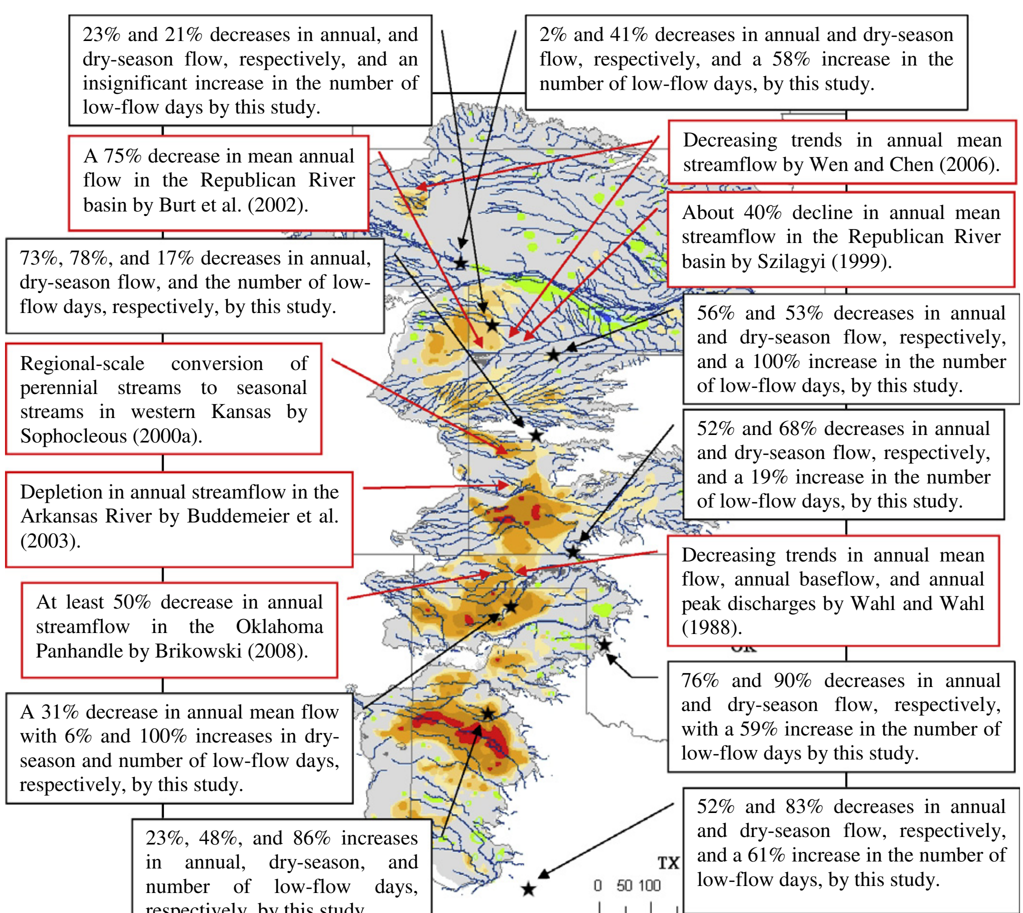

Fig. 10. Results of this study (in black boxes) together with the findings from earlier studies (in red boxes) related to the changes in streamflow variables over the High Plains aquifer.

mostly grouped in the Republican River basin and a few are observed in the Arkansas River basin; however the Oklahoma Panhandle is dominated by decreasing trends. More stream gauges with significantly increasing number of low-flow days are detected in Texas, likely resulting from changes in the frequency of extreme weather events that, as the findings of this study indicate, sustain the local streams in Texas. The significant increases in the number of low-flow days at the Texas gauges, which fail to show any significant step-changes in annual and dry-season flow, from the pre-irrigation period to the post-irrigation further supports this argument.

The trend results in annual and dry-season streamflow provide observational evidence of decreased streamflow across the High Plains region consistent with the regional pattern of streamflow–groundwater hydraulic connection. The similarities in step-changes of streamflow and groundwater at select locations imply that the observed trends in streamflow variables are attributable to changes in groundwater levels. The disagreement between the precipitation and streamflow trends further supports this argument. Extensive irrigational pumping causes depletion, more severely, in the Northern High Plains streams, and to a lesser extent in the Southern streams. Recently, Krakauer and Fung (2008) reported that the trends in annual mean streamflow are well-correlated with the trends in precipitation over the United States for the period 1920–2007. However, of all regions in the US, they identified the Great Plains as the only region where streamflow was least sensitive to the variations in precipitation. Therefore, the observed decreases in streamflow, especially in Nebraska, can be confidently attributed to the pumping of groundwater as opposed to any change in precipitation. This is also supported by the results of regional analysis which revealed that identified trends in annual streamflow in Nebraska, Colorado, and Kansas (Region 1) were field significant at the 5% level for the period of irrigation development (1941–1980). However, we can not eliminate the possibility that trends in annual streamflow in Oklahoma and Texas (Region 2), and trends in dry-season flow and the number of low-flow days in Region 1 and Region 2 might have happened accidentally as they were not field significant at the 5% level.

The results of this study may have important implications regarding the extents of the impacts that human beings exert on the regional water resources. The findings point out to a more notable impact of groundwater pumping on regional streamflow than a corresponding impact of precipitation in the High Plains region. Fig. 10 summarizes the observed changes in streamflow variables over the High Plains by earlier studies together with new contributions from this study. The consistency of the streamflow depletion over such a large area indicates the regional characteristic of the streamflow trend. Despite the reported increase in precipitation over the Great Plains during the last two decades of the 20th century (Garbrecht and Rossel, 2002; Garbrecht et al., 2004), our results indicate that streamflow depletion persists in recent decades with a possibility of becoming worse in the subsequent years due to the increasing tendency of streams to deplete as a consequence of prolonged and excessive withdrawal of groundwater year after year.

The results presented here in general agree with the previous findings, and also fill the spatial gaps using as much information as possible and a consistent methodology throughout the region. Spatial differences in the occurrence and direction of trends reveal that a systematic analysis of trend detection for the entire aquifer is crucial to establish the regional significance of groundwater pumping on surface water resources. By focusing on regional patterns and end-members, this study serves as a synthesis of streamflow depletion induced by large-scale and long-term groundwater pumping over the High Plains aquifer.

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