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Analytical modeling of irrigation and land use effects on streamflow in semi-arid conditions

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SUMMARY

Availability and uncertainty in input data are the primary constraints of groundwater modeling. Analytical models assimilate the key and important data, but capture the major traits of the watershed. We study a baseflow-dominated stream, Frenchman Creek in southwestern Nebraska, USA, which has experienced large streamflow reductions since the 1960s and is a subject of various actions on water rights appropriation. The new element of the model is simultaneous analytical consideration of groundwater pumping and land use change effects. Analytical stream depletion rate calculations by various methods show that pumping from the 462 irrigation wells in the basin consumed a large amount of baseflow. The simulated streamflow at the outlet of Frenchman Creek with minimal calibration compares favorably with observed streamflow and indicates the viability of an analytical approach to watersheds with limited hydrogeologic data.

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1. Introduction

Modeling of groundwater (GW)–surface water (SW) interactions to gain a better understanding of the hydrologic system is important in the Great Plains region of the United States where groundwater pumping for irrigation is widespread. This area has undergone tremendous land use changes from native rangeland to cropland. This proliferation of irrigation and land use change has led to streamflow and GW level declines throughout the Great Plains during the past century (McGuire, 2011). In order to better understand the effects of irrigation and land use changes on the water budget and streamflow and to predict and mitigate future declines, water resources management in the western USA rely on modeling the GW–SW interactions predominately using numerical models (e.g., Rossman and Zlotnik, 2013). Usually, natural conditions are highly complex and the model's ability to produce reliable outputs rests on the quality and quantity of input data. As well, model development is labor-intensive. Therefore, analytical models may be a good supplement or alternative for stream water budget assessment because they focus on key processes and are easy to implement and provide a water management tool for understanding the consequences of water management policies.

Analytical studies by Theis (1941), Glover and Balmer (1954), Hantush (1965), Jenkins (1968), Hunt (1999); Zlotnik et al. (1999); Butler et al. (2001) and others, and well summarized by Barlow and Leake (2012), focused largely on stream depletion by irrigation wells in various hydrogeological conditions. Jenkins' (1968) analysis became the standard analytical approach for stream depletion rate (SDR) assessment for water management in the Mid-West and other regions of the USA, but newer methods have not been compared with this technique in hydrological applications. Areas where SDR were assessed using analytical techniques may vary in magnitude from km^2 (e.g., Hunt et al. (2001), Kollet and Zlotnik (2003), Langstaff (2006), Fox (2004) and Fox et al. (2011)) to hundreds of km^2 (e.g., Foglia et al., 2013).

Abbreviations: FC, Frenchman Creek; GIS, geographic information system; GW, groundwater; NDNR, Nebraska Department of Natural Resources; RRGWM, Republican River Groundwater Model; SDR, stream depletion rate; SRR, stream recharge rate; SW, surface water.

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The influence of land use change on recharge and streamflow is also a concern (Dugan and Zelt, 2000; Sophocleous, 2005; Oudin et al., 2008; Wilcox et al., 2008; Zheng et al., 2009; Stonestrom et al., 2009; McMahon et al., 2011; Perez et al., 2011; Zeng and Cai, 2014), but the application of analytical techniques to this problem have not been fully explored yet. Knight et al. (2005) addressed the influence of land use practices; they evaluated the effect of GW recharge changes on baseflow to a stream in South Australia, but GW withdrawals for irrigation were not considered. Foglia et al. (2013) accounted for pumping and land use aspects, but utilized a more complex hydrological model of the watershed.

The goal of this study is to combine the analytical methods of the stream depletion and stream recharge analyses, considering the effects of GW pumping for irrigation and land use changes on streamflow and apply it jointly to Frenchman Creek (FC). Another goal of this study is to investigate conditions for application of more recent approaches in SDR evaluations by comparing Jenkins' and Hunt's methods.

2. Study area

Our study area is the salient example of declining streamflow and large land use changes over the last century like in many countries around the world. FC of southwestern Nebraska, USA, is a tributary of the Republican River (Fig. 1) and has been the subject of a number of hydrological studies and numerical models of various domains, scales, and complexity (Condra, 1907; Cardwell et al., 1963; Lappala, 1978; Peckenpaugh et al., 1995; Szilagyi, 1999, 2001; Burt et al., 2002; Republican River GW Modeling Committee, 2003; Zeng and Cai, 2014; Demissie et al., 2014).

Our study encompasses both the SW and GW basins of FC between the Imperial and Culbertson stream gages (Fig. 1). The SW basin area is 985 km^2 and the GW basin area is 1308 km^2. The semi-arid area receives an average of 500 mm of precipitation per year with 75% of that precipitation occurring in the growing season from April to September (National Climatic Data Center (NCDC), 2011, http://www.ncdc.noaa.gov). Land use in 2009 includes 50% rangeland, 15% dry cropland, 16% terraced land, and 16% irrigated land with a total of 462 irrigation wells. The remaining 3% of land is open water, riparian vegetation, roads, and municipalities (provided in table format by T. Tietjen, personal communication, 2011). From 1928, or predevelopment period, more than 450 irrigation wells have been installed (NDNR Wells, 2011, http://dnrdata.dnr.ne.gov).

The topography influences the land use. Along the flat FC valley, the dominant land use is irrigated crops. The valley is surrounded by rolling hills and steep canyons covered by native rangeland and terraced dry cropland. The western parts of the region are relatively flat or gently sloping and have a mixture of dry cropland and irrigated land.

The principal aquifer in the study area is the High Plains Aquifer, where GW generally flows from west to east until it discharges to FC as baseflow. FC has a flow-through regime above Enders Reservoir and is naturally a gaining stream below Enders Reservoir; receiving all GW that flows into the basin. There is no GW flow out of the basin (Fig. 2).

FC flows from west to east and is the central water body in the SW network between Imperial and Culbertson. Stinking Water Creek, the only significant tributary of FC, flows south and empties into FC at Palisade, Nebraska. Enders Reservoir, located 5 km downstream from the Imperial stream gage on FC, was constructed

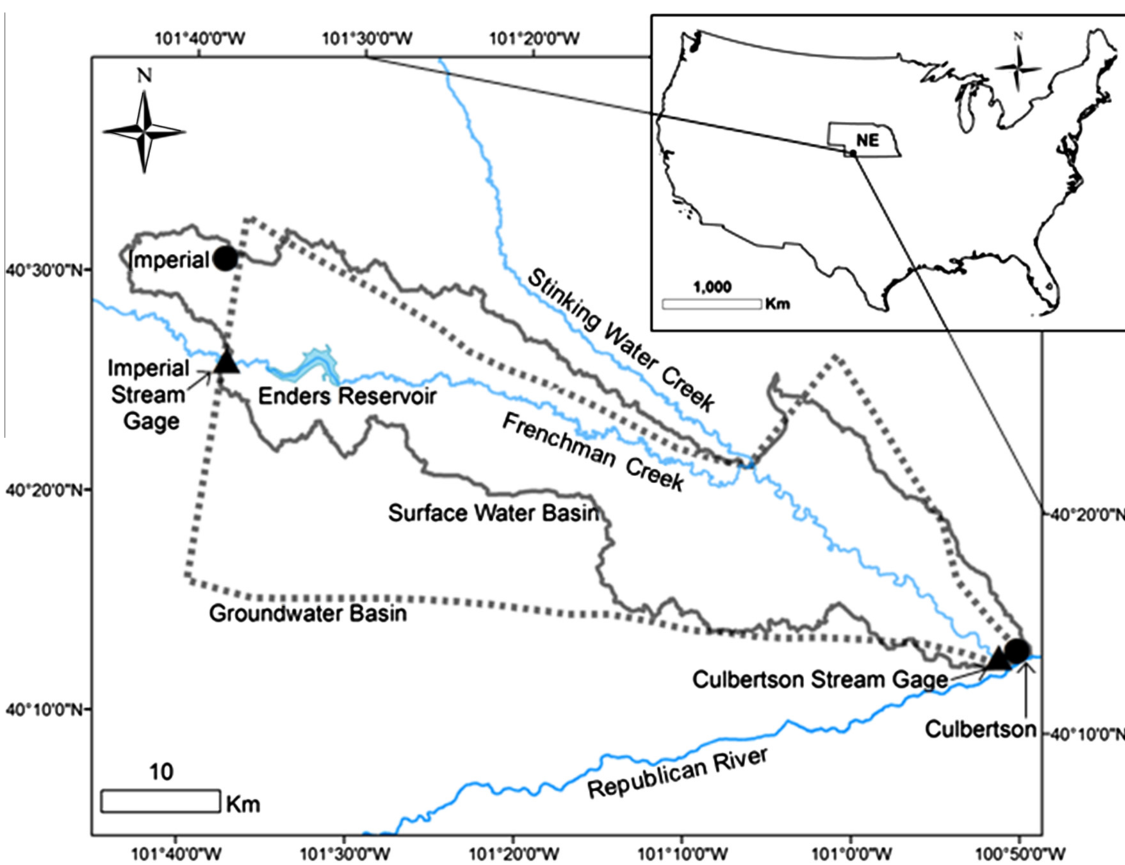

Fig. 1. The GW and SW basins of Frenchman Creek between the Imperial and Culbertson stream gages, Nebraska, USA.

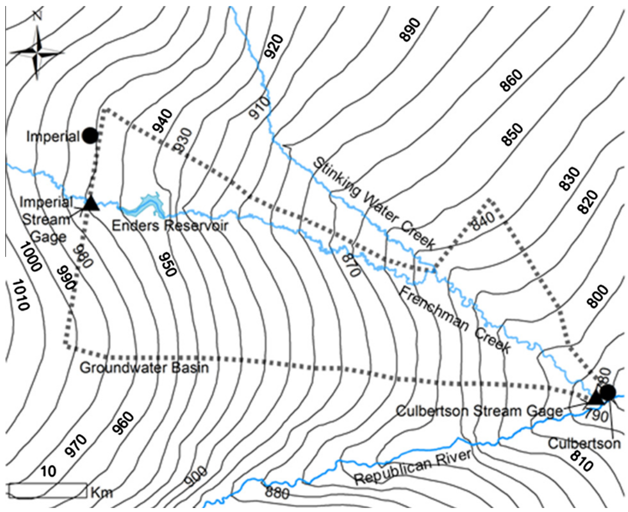

Fig. 2. The September 2008 equipotential lines (hydraulic head in meters) delineating the GW basin of Frenchman Creek between the Imperial and Culbertson stream gages.

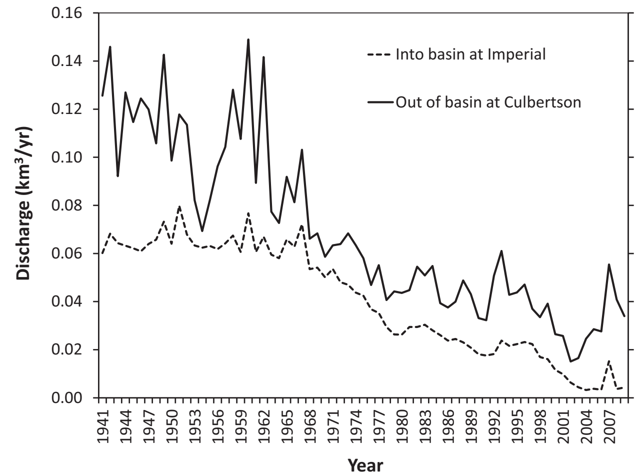

Fig. 4. Mean annual stream discharge at the Imperial and Culbertson stream gages for Frenchman Creek.

between 1947 and 1951 with a maximum capacity of 0.043 km^3 (Figs. 2 and 3). FC has experienced significant streamflow declines since the mid-1960s (Fig. 4) which have resulted in shortening of the perennial stream length over time, but records indicate perennial surface flow between Imperial and Culbertson stream gauges. Frenchman Creek has the largest virgin water supply allocation of any other Republican River tributary. These observed streamflow declines and the proliferation of GW development for irrigation inhibit Nebraska's compliance with the Republican River Compact allocations of 1943 (Republican River Compact, 1943). This is still an area under litigation between Kansas and Nebraska in the U.S. Supreme Court as of 2012 (Republican River case back in court, 2012).

3. Conceptual model

3.1. Boundaries and boundary conditions

The model domain is located between the Imperial and Culbertson stream gages (Fig. 1). The water table is relatively stable annually as apparent by comparison of the 1995 and 2008 water table maps (Summerside et al., 2001 and RRGWM Committee, 2003), although some minor to moderate changes of 0–15 m saturated thickness have been observed since pre-development along the western edge of the GW basin boundary (McGuire, 2011). The GW basin boundaries (Fig. 2), delineated by tracing streamlines perpendicular to the September 2008 equipotentials, are used for description of any process that affects GW flow, such as stream depletion, GW recharge, and baseflow. Any segment of the GW basin that receives GW inflow is treated as a constant head boundary for all modeled years (1941–2009). Other segments are treated as no-flow boundaries for all modeled years (1941–2009) and FC is treated as a constant head boundary. The SW basin boundaries were delineated using ArcGIS tools and a Digital Elevation Model obtained from the NDNR Databank (NDNR, 2011a, http://dnr.ne.gov/databank/dem.html).

3.2. Water budget

We consider the annual stream water budget between the Imperial and Culbertson gages at any i-th year between 1941 and 2009, using units $[L^3/T]$. The inflow includes streamflow into the basin from FC at the Imperial stream gage (q_{in}^{FC}), streamflow into the basin from Stinking Water Creek at Palisade (q_{in}^{SWC}), GW inflow to the basin from the unconfined aquifer in the west that eventually discharges to FC as baseflow (Q_{in}), contributions from overland

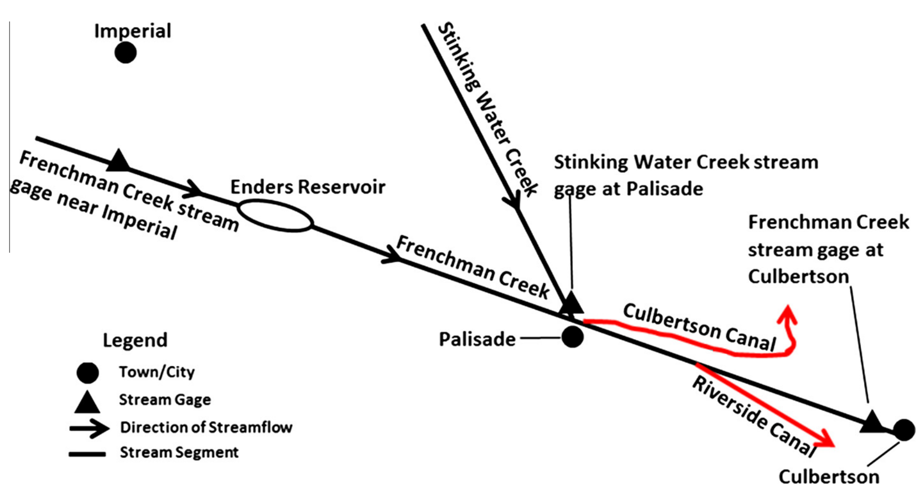

Fig. 3. Plan view schematic of the Frenchman Creek surface water network between the Imperial and Culbertson stream gages.

flow (q) associated with different land uses, and stream recharge rate (SRR) originating from GW recharge associated with different land uses. It is important to notice that GW inflow to the basin Q_{in} can be a good representation of groundwater flow entering the stream, considering the multi-decadal time frame of our study and relatively stable configuration of the water table on the boundary. Outflow includes stream depletion rate (SDR) caused by GW pumping for crop irrigation, canal diversions (q^C), and losses from FC evaporation (E^{FC}). Finally, q^{FC}_{out} is the modeled streamflow rate out of the basin at Culbertson on an annual basis:

$$q^{FC}_{in}(i) + q^{SWC}_{in}(i) + Q_{in}(i) + q(i) + SRR(i) - \mathrm{SDR}(i)\ldots \\ - q^{C}(i) - \Delta S^{R}(i) - E^{FC}(i) = q^{FC}_{out}(i) \quad (1)$$

The term ΔS^R accounts for changes in storage of Enders Reservoir which includes FC inflow and outflow and reservoir evaporation, which altered streamflow largely during its construction and filling period from 1951 to 1957.

The flow out of the basin for each i-th year is the subject of study; therefore each component must be assessed for each i-th year.

4. Evaluation of data and components in the SW budget

4.1. Input data

An extensive database has been compiled from various sources with the most important characteristics given in Supplementary Material 1. The streamflow data for q^{FC}_{in} and q^{SWC}_{in} were obtained from the U.S. Geological Survey's National Water Information System on-line database (USGS, 2011, http://waterdata.usgs.gov/nwis) and the Nebraska Department of Natural Resources (NDNR) on-line stream gaging database (NDNR, 2011b, http://dnr.ne.gov/docs/hydrologic2013.html). The streamflow data measured by the USGS achieved mostly a "Fair to Good" rating (95% of daily discharges within 10% of true value). The canal diversions data for q^C were obtained from the NDNR on-line stream gaging database (NDNR, 2011b, http://dnr.ne.gov/docs/hydrologic2013.html). The reservoir data for storage ΔS^R were obtained from the Bureau of Reclamation (BOR) (USBOR, 2013, http://www.usbr.gov/gp-bin/arcweb_edne.pl) and given in Supplementary Material 1. The NDNR and BOR do not provide accuracy or uncertainty ranges for their data. Based on local practices, an estimated accuracy of 10% is appropriate.

Irrigation well locations and SW network data were obtained from the NDNR GIS Processing Site (NDNR GIS, 2011, http://dnr.ne.gov/databank/DataTypeList.html). Hydraulic head levels and aquifer properties used in this study including saturated thickness, hydraulic conductivity, and storativity are Republican River GW Model (RRGWM) calibrated parameters and post-calibration simulated head levels, obtained from the NDNR (personal communication, 2011). The RRGWM Committee did not provide a sensitivity or uncertainty analysis for the calibrated values. However, calibrated values were consistent with hydrostratigraphy of the area discussed by Cardwell et al. (1963) and within ±10%. Irrigation well pumping data was obtained from the NDNR (personal communication, 2011). Land use data was obtained from T. Tietjen (personal communication, 2011).

In this study area, land use data were limited in some years. Spatial rangeland data was only available for 2005 from the Center for Advanced Land Management Information Technologies (CALMIT), University of Nebraska-Lincoln (Dappen et al., 2007). The area and the center of the rangeland area were found for each year, and distance from the center to the stream was used in calculations for each year. Individual dry cropland fields were randomly distributed over the basin during the simulation period. Therefore, the center of the GW basin was used to determine the distance from the cropland fields to the stream. For terraced fields, distance to the stream was calculated for each individual field. Finally, distance from the stream to each irrigated field was taken equal to the distance from the stream to the appropriate well.

Terms Q_{in}, q, SDR, and SRR must be calculated using parameters of the aquifer, individual wells, and characteristics of land use in order to obtain the streamflow rate out of the basin, q^{FC}_{out}. We will discuss each of these factors and focus on computational aspects, but it is important to note that the effect of these parameters on streamflow in a given year may depend on water management far into the past.

4.2. Calculation of Q_{in}

Regional data indicate that the water table configuration has changed little over the period of most intensive pumping along the north, south, and eastern boundaries between two published surveys (Summerside et al., 2001; McGuire, 2011), which means that the regional GW flow system is in a steady state for over more than decade. Parts of the western edge of the GW basin declined 3–15 m since pre-development (McGuire, 2011). Therefore, GW discharge from the adjacent unconfined aquifer, crossing the western boundary of the study area in the west (Fig. 2) after a certain time lag seeps into the stream and becomes approximately $Q_{in}(i)$ [L^3/T]. This time lag can differ between various segments of the western boundary and FC locations, but total discharge across the western boundary (Q) is equal to $Q_{in}(i)$ and can be calculated from the Darcy equation as follows:

$$Q = K \cdot A \cdot \frac{dh}{dl} \approx Q_{in}(i) \quad (2)$$

where K [L/T] is saturated hydraulic conductivity, A [L^2] is the cross-sectional area of the aquifer and dh/dl [–] is the hydraulic gradient, taken at the western edges of the GW basin. Using RRGWM calibrated $K = 10$ m/d, $A = 2.66 \times 10^6$ m^2, based on the width of the western borders of the GW basin (42,270 m) and aquifer saturated thickness $b = 63$ m, and $dh/dl = 0.0032$, we estimated $Q_{in}(i) = Q = 0.031$ km^3/yr. The major uncertainty stems from the saturated thickness b, while relative uncertainty of K along the western edge of the model domain is ±10%, and much higher accuracy is apparent from regional slope estimates (5%). Long-term observations of water-level changes (McGuire, 2011, Figs. 2 and 3C) indicate reduction of saturation thickness at some locations at the western boundary on the order ~10 m. Therefore, $Q_{in}(i)$ may be biased toward higher values as much as 15%. Although currently the stream gauge at Imperial shows continuous flow, the upstream changes especially extensive in Colorado, may upset this quasi-steady state and result in dry channel conditions.

4.3. Stream depletion rate (SDR)

For the SDR evaluation, two analytical solutions, Jenkins (1968) and Hunt (1999) are the most common. SDR, induced by GW pumping from a well is determined by a time scale (t_a), defined by aquifer transmissivity T [L^2/T], storativity S [–], and distance l [L] from a given stream (Fig. 6). This time scale for GW flow between the stream and aquifer is termed "stream depletion factor (sdf)" after Jenkins (1968):

$$t_a = \frac{Sl^2}{T} = sdf \quad (3)$$

(A similar time scale is applicable for analyses of propagation of any local GW perturbations over the distance l). Needed parameters on a grid with 1 km × 1 km cells were provided by RRGWM, where T ranged from 34.6 m^2/day to 1115.7 m^2/day, with average 454.1 m^2/day, and S ranged from 0.174 to 0.225, with an average of 0.187. This applies to various sources of drawdown changes.

Then, SDR induced by an individual well using the Glover–Balmer–Jenkins approach (a stream that fully penetrates the aquifer and a perfect stream-aquifer connection (Fig. 6a)) can be rewritten for convenience as follows (Zlotnik, 2004):

$$SDR(t) = Q_W erfc\left(\frac{1}{2\sqrt{t/t_a}}\right) \quad (4)$$

where t is well operation time, and Q_W is constant pumping rate. This rate is calculated by spreading pumped water volume from the irrigation season to a full calendar year; changes from year to year are accounted by convolution, and effects of multiple wells are accounted by superposition (Wallace et al., 1990; Barlow and Leake, 2012).

To estimate the time scale for SDR, one obtains t_a = -sdf = 18 years taking average T and S values, and l = 4000 m that reflects average well-stream distance.

A more detailed SDR representation of field conditions by Hunt (1999), considering streambed properties and partial penetration of the stream in the aquifer (Fig. 5b) explicitly can be written using a function of two arguments, $D_H(u, v)$ as follows (Zlotnik, 2004):

$$SDR(t) = Q_W \cdot D_H\left(\frac{t}{t_a}, \frac{l}{B_S}\right), \; B_S = \frac{2T}{\lambda}, \quad \lambda \approx WK'/b' \quad (5)$$

$$D_H(u, v) = erfc\left(\frac{1}{2\sqrt{u}}\right) - e^{v^2u+v} erfc\left(\frac{1}{2\sqrt{u}} + v\sqrt{u}\right) \quad (6)$$

where the streambed leakage coefficient λ [L/T] is defined by a stream width W, streambed hydraulic conductivity K', and a streambed thickness b' that are determined by sedimentological processes. In spite of ubiquitous studies of K' (e.g., Calver, 2001; Chen, 2004), values of parameter b' (and consequently, λ) are highly uncertain. In our case, data have been obtained by aquifer tests from the stream-analogue, Prairie Creek near Central City, Nebraska (Kollet and Zlotnik, 2007): $\lambda = 20$ m/d. Although Reeves (2008) and Reeves et al. (2009) suggested deducing these parameters from streambed-well geometry, this method has not been verified yet. It is important to note that the effect of pumping from wells on SDR must account for returnflow resulting from irrigation. This returnflow partially dampens the effect of pumping. Therefore we will also use the term "net SDR" that takes into account this effect.

4.4. Stream recharge rate (SRR)

There are two sources of SRR [L^3/T]. The first source of SRR is precipitation; a fraction of precipitation that goes to SRR from each land use is given by the dimensionless coefficient γ, which is specific for each land use (rangeland, dry cropland, terraced land, and irrigated land) and certain development periods. The second source is returnflow from irrigation, and a dimensionless coefficient α denotes some fraction of the pumping rate Q_W that becomes irrigation returnflow and contributes to SRR.

The stream recharge process is affected by two important time scales. Firstly, a fraction of precipitation or irrigation water traverses the vadose zone and arrives as GW recharge to the water table with some time lag, t_{lag} after the land use change; secondly, this GW recharge travels to the stream as a GW flow and becomes baseflow.

Assessment of the GW flow component is a relatively straightforward procedure. Consider a rectangular field with recharge rate corresponding to a given land use, R^{GW} [L/T]. The field has length Y [L] (aligned along the y-axis and parallel to the stream), width X [L] (aligned along the x-axis and perpendicular to the stream), and is centered at distance l [L] from the stream (Fig. 6).

Using the Polubarinova-Kochina (1962) solution for GW flow under the rectangular recharge source and the superposition principle, one obtains a solution as follows:

$$SRR(t) = 2R^{GW}Y \cdot l\left(\frac{t}{t_a}\right)\left[ierfc\left(\frac{1 - X/(2l)}{2\sqrt{t/t_a}}\right) - ierfc\left(\frac{1 + X/(2l)}{2\sqrt{t/t_a}}\right)\right] \quad (7)$$

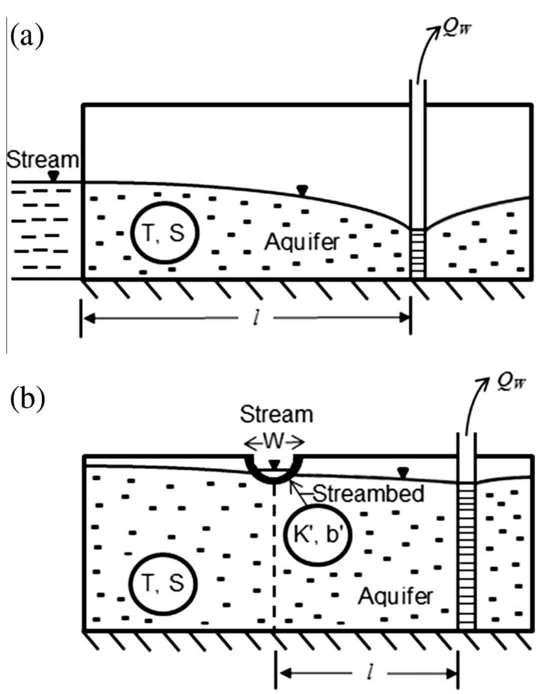

Fig. 5. Schematic cross section showing (a) Jenkins (1968) and (b) Hunt (1999) model characteristics.

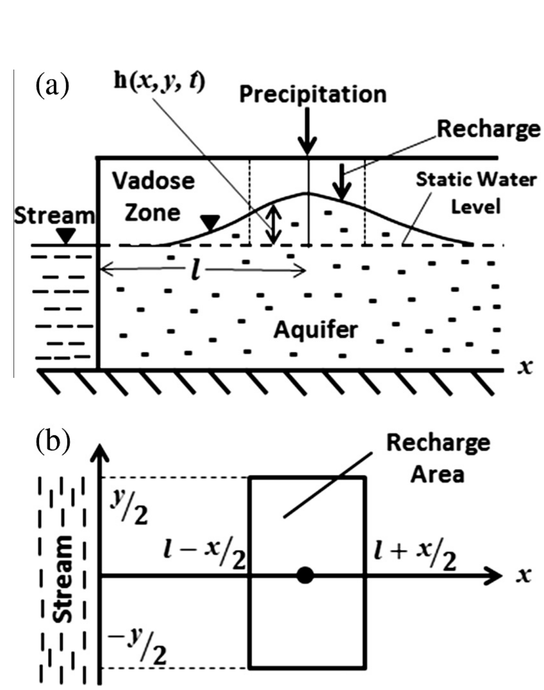

Fig. 6. Schematic diagrams showing parameters of the stream recharge equation; (a) Cross section, (b) Plan view.

where t is the time since the moment when recharge through the vadose zone arrived to the water table, R^{GW} is any land-use related GW recharge rate [L/T]. A similar equation in a different notation was obtained by Knight et al. (2005). The function $ierfc(x)$ is defined as follows:

$$ierfc(x) = \frac{1}{\sqrt{\pi}} e^{-x^2} - x \cdot erfc(x) \quad (8)$$

Eq. (7) accounts only for the time lag associated with the GW flow toward the stream from an area with changing land use (e.g., Sophocleous, 2012). Determination of distance l for each land use was described in Section 4.1.

To apply Eq. (7) to each specific land use, the recharge rate for each source is introduced. For example, recharge due to precipitation only, $R^{GW} = R_P^{GW}$ is calculated as a dimensionless fraction of annual precipitation rate P [L/T], denoted γ^{LU}:

$$R_P^{GW}(i) = P(i - t_{lag}) \cdot \gamma^{LU}(i - t_{lag}) \quad (9)$$

Here, t_{lag} [T] is time lag expressed in years which accounts for the period that soil moisture changes traversed the vadose zone. To estimate t_{lag}, various approaches based on Richards' equation are advocated (Cook et al., 2003; Niswonger et al., 2006; Sophocleous 2012). However, these techniques require accurate knowledge of subtle differences between the pre- and post-development moisture conditions and unsaturated zone properties, which are unavailable at the watershed scale due to heterogeneity. Therefore, we selected a range of the time lags from values t_{lag} = 0, 2, 5, 10, and 20 years and used calibration to observed stream discharge.

A similar approach was used to obtain $R^{GW} = R_Q^{GW}$ for irrigation returnflow, where the term

$$R_Q^{GW}(i) = Q_W(i - t_{lag}) \cdot \alpha(i - t_{lag}) \quad (10)$$

is substituted into Eq. (7). γ^{LU} and α are not constant for all times, they change to account for different agricultural practices (periods) in time.

4.5. Calculation of overland flow rates

A contribution of a specific land use to streamflow is calculated by a dimensionless fraction β^{LU} of annual precipitation rate P [L/T] in year i. This contribution from the area A_{SW}^{LU} [L^2] within the SW basin arrives to the stream via overland flow within the same year:

$$q(i) = A_{SW}^{LU}(i) \cdot P(i) \cdot \beta^{LU}(i) \quad (11)$$

Overland flow is assumed to discharge to FC within the same year as the precipitation event. β^{LU} is not constant for all times; it changes to account for different agricultural practices.

4.6. Calculation of changes in reservoir storage

Enders Reservoir's only source of water is FC and its only outlet is to FC. Therefore, the reservoir storage changes ΔS^R [L^3/y] were calculated as the difference between the inflow and outflow of FC inflow at Enders Reservoir and reservoir evaporation. It is important to note that the reservoir began to fill in 1950.

5. Model calibration

5.1. Target function

The objective of model calibration was to minimize the average absolute relative error $\overline{RE}$ over all modeled years:

$$\overline{RE} = \frac{1}{69} \sum_{i=1941}^{i=2009} |RE(i)| \quad (12)$$

The relative error $RE(i)$ for i-th year is defined as the difference between modeled streamflow q_{out}^{FC} and observed streamflow, normalized by observed streamflow.

5.2. Parameters

Land use change alters the land surface and soil properties, causing changes in overland flow, infiltration, and stream recharge rates. The model was calibrated by adjustment of parameters that control the returnflow (α), overland flow (β), stream recharge (γ), and t_{lag}. Currently, watersheds with developed irrigation and stream discharge records have appropriate estimates, and strong constraints on these parameters are available from regional and national data sources. In our case, α is based on a range of values from RRGWM Committee (2003), Kendy and Bredehoeft (2006), and Dewandel et al. (2008), and regional β and γ values were obtained by Dugan and Zelt (2000), McMahon et al. (2006), Scanlon et al. (2007), Twombly (2008). Manual trial-and-error adjustment of the parameters with these constraints resulted in values that account for climate, hydrologic conditions, and land use of the region.

The $\alpha, \beta,$ and γ values for each land use are listed in Tables 1–3. Superscripts denote land use (i.e. R = rangeland, DC = dry cropland, T = terraced land, ir = irrigation water applied to a field, and P = precipitation that falls on irrigated land). The time ranges listed in these tables reflect different stages of agricultural developments. For example, the three α time ranges correspond to the dominance of surface irrigation (1941–1955), the emergence of center-pivot irrigation (1956–1985), and further improvement of irrigation efficiency (1986–2009).

The model was calibrated using t_{lag} = 0, 2, 5, 10, and 20 years while keeping base parameters $\alpha, \beta,$ and γ constant for each t_{lag} value and also small adjustments to α, β, and γ for each t_{lag} value that required several manual iterations. The absolute relative errors corresponding to these t_{lag} values were 4.9%, 4.6%, 4.3%, 4.4%, and 6.2%, respectively. Note that the lowest $\overline{RE} = 4.3\%$ yields t_{lag} = 5 years is consistent with regional values obtained by Rossman et al. (2014). Annual relative error is under 10% for 61 out of the 69 modeled years (demonstrated in Supplementary Material 2, Fig. S2-2).

5.3. Sensitivity analysis

Results of the sensitivity analyses to selected parameter values are shown in Table 4. Surface water budget components that were obtained from direct measurements and two data sources ($q_{in}^{FC}, q_{in}^{SWC}$, and q^C) and land use characteristics (indexed parameters α, β, and γ, denoted in Section 5.2) were constrained by studies of irrigation as explained above. Change of any parameter by ±10% results in an increase of the absolute relative error in excess of the base value 4.3%. Among all parameters, increases of these parameter magnitudes are most notable for Frenchmen Creek inflow (measured) and irrigation practices (returnflow parameter α).

6. Results

6.1. Dynamics of SDR

The SDR estimates using the Jenkins (1968) and Hunt (1999) solutions yielded very similar results (Fig. 7). The first irrigation

Table 1
Irrigation returnflow coefficient α values.

Time periods	1941–1955	1956–1985	1986–2009
α	0.25	0.1	0.03

wells installed in 1928 had a total SDR of less than 0.0010 km^3/yr and differences between the Jenkins and Hunt solutions were 2.5%. By 1941, there were 15 irrigation wells, and the two solutions had a difference of 1.3%. By 1962, 82 operational irrigation wells increased SDR to 0.0045 km^3/yr, and the two solutions had a difference of 0.28% that remained under 1% thereafter. By 1978, 332 operational irrigation wells increased SDR to 0.039 km^3/yr. In 2002, there were 461 irrigation wells and SDR peaked at 0.056 km^3/yr. In 2009, there were 462 irrigation wells and SDR dropped to 0.0360 km^3/yr, and the two solutions had a difference of 0.010%.

6.2. Overland flow rates

Total overland flow, characterized by annual fluctuations and a decreasing trend, declined from 0.02 km^3/yr in 1941 to 0.012 km^3/yr in 2009 (Fig. 8). Dry cropland provided the most overland flow to FC. In 1941, overland flow from dry cropland constituted 72% of total overland flow. By 2009, dry cropland constituted 46% of total overland flow. The next most substantial amount of overland flow came from rangeland, which was fairly consistent throughout the modeled period. In 1941, rangeland constituted 28% of total overland flow and in 2009 rangeland constituted 49% of total overland flow. Terraced land constituted less than 5% of total overland flow from 1941 to 2009. Overland flow from irrigated land was between 0% and 1% of total overland flow for all modeled years (see Fig. 8).

6.3. Stream recharge rates

The total stream recharge rate increased from 0.0004 km^3/yr in 1941 to 0.029 km^3/yr in 2009 (Fig. 9). Two time periods are marked by different land uses. Prior to 1961, dry cropland and rangeland constituted greater than 50% of total stream recharge. After 1961, returnflow constituted greater than 50% of total stream recharge. In 1941, stream recharge from dry cropland constituted 33% of total stream recharge, but by 2009, it decreased to 2%. From 1941 to 1958, stream recharge from rangeland constituted 51% to 67% of total stream recharge. After 1958, stream recharge from rangeland decreased to 6% of total stream recharge. In 1941, stream recharge from returnflow constituted less than 1% of total stream recharge. By 2009, returnflow constituted 86% of total stream recharge. Precipitation on irrigated land, a minor contributor to stream recharge, increased from less than 1% of total stream recharge in 1941 to 3% in 2009. Terraced land constituted less than 1% of total stream recharge prior to 1960, but by 2009 it constituted 3% of total stream recharge.

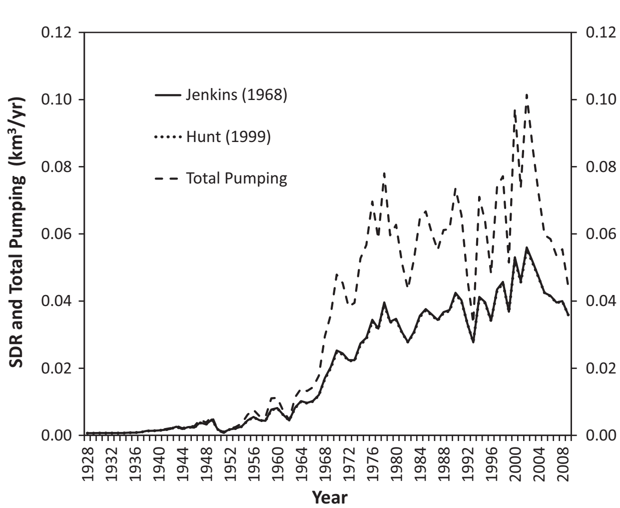

Fig. 7. Estimated dynamics of stream depletion rate (SDR) and total irrigation pumping in the study area. Note that the Jenkins (1968) and Hunt (1999) models are indistinguishable.

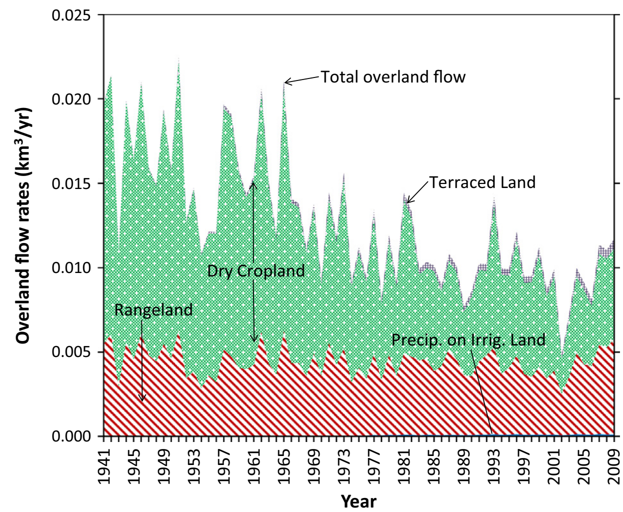

Fig. 8. Composition of overland flow rate estimates (km^3/yr) for each land use.

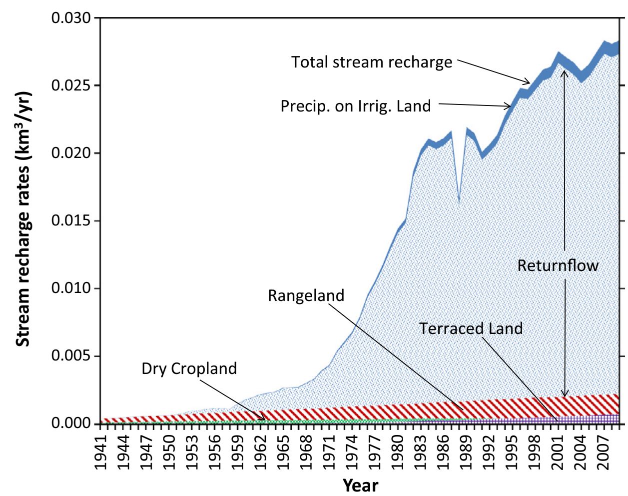

Fig. 9. Composition of stream recharge rate (SRR) estimates (km^3/yr) for each land use.

6.4. Streamflow at Culbertson and water budget

Eight major components of streamflow (SW inflow into the basin at Imperial and inflow from the tributary, Stinking Water Creek at Palisade, GW inflow to the basin, stream depletion due to irrigation including irrigation returnflow, land use change contributions via changes in *SRR* and overland flow but excluding recharge from irrigation returnflow, canal diversions, reservoir storage changes, and evaporation from FC and Enders reservoir are plotted in Fig. 10. The summation of inflow and outflow budget components in (Fig. S2-1) helps to compare contributions from all components to stream discharge at Culbertson each year from 1941 to 2009 (see Supplementary Material 2 for Fig. S2-1).

The modeled streamflow at Culbertson replicates all the trends visible in the observed streamflow record with the average absolute relative error, $\overline{RE}$ at 4.3% (Fig. 11). From 1941 to 1951,

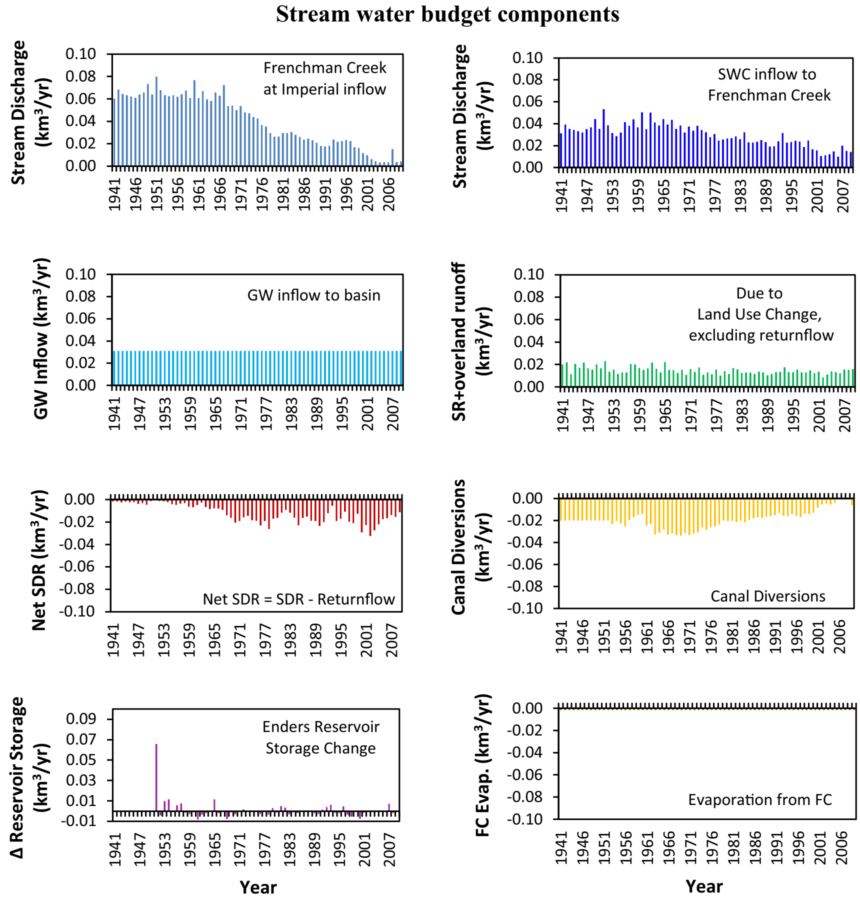

Fig. 10. Comparative analysis of stream water budget components.

streamflow is fairly consistent with some fluctuations. From 1951 to 1954, streamflow declined rapidly as did modeled streamflow and increased rapidly from 1955 to 1958. After a period with relatively steady streamflow from 1959 to 1964, large declines in streamflow rate began in 1965 and continued to 2002 after which observed streamflow increased temporarily.

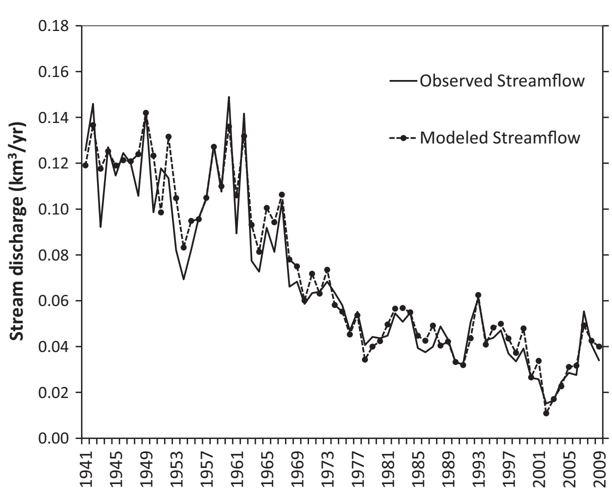

Fig. 11. The modeled and observed discharge of Frenchman Creek at the Culbertson gage from 1941 to 2009.

7. Discussion of streamflow declines

7.1. SDR evaluation

Both the Jenkins (1968) and Hunt (1999) SDR estimates exhibit yearly fluctuations due to variations in the amount of GW pumping (Fig. 7). For example, increased precipitation in 2005–2009 led to less pumping and a decrease in SDR; in dry years like 2002, increased pumping resulted in an increase in SDR.

SDR estimates by Jenkins (1968) and Hunt (1999) are very similar for two reasons. Firstly, with increased well-stream distances, these solutions converge. Secondly, the assigned value $\lambda = 20$ m/d was obtained by analogy with the very similar Prairie Creek, also in Nebraska (Kollet and Zlotnik, 2003, 2007), which does not significantly delay SDR even in Hunt's (1999) model. This analysis shows that for our study area, the Jenkins (1968) approach is valid and preferred over the Hunt (1999) approach despite the more detailed model of the latter. Discussion of secondary reasons is relegated to Supplementary Material 3.

7.2. SDR impacts on baseflow

SDR due to GW pumping for irrigation is a major concern for water management because it directly affects baseflow. The FC baseflow is a sum of GW flow Q_{in} into the basin and *SRR* (Fig. 12). The relationships between modeled baseflow ($Baseflow_{model}$) and SDR indicate times when wells captured almost all the baseflow to the stream (Fig. 12). This occurred during the most intense drought years with minimal overland flow in 1978 and

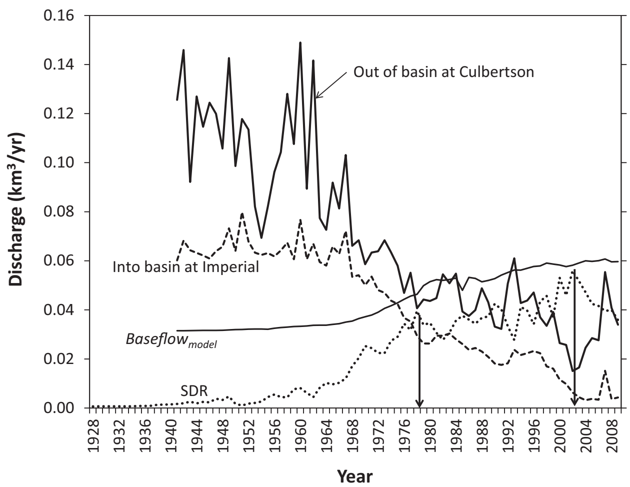

Fig. 12. Comparison of *Baseflow$_{model}$* with SDR between gages at Imperial and Culbertson. Arrows indicate years 1978 and 2002 when SDR captured almost all baseflow.

2002 (arrows in Fig. 12); this means the stream remained flowing primarily due to SW inflow into the basin from Imperial and from Stinking Water Creek, and overland flow within the SW basin. In both cases, there was a substantial increase in pumping, especially in 1978 prior to more stringent regulations. This is further discussed in Supplementary Material 3.

7.3. Land use change

The effect of land use change, excluding irrigation, is a significant contributor to streamflow but is not a significant cause of streamflow declines (Fig. 10). Although it is significant, it is fairly constant over time and does not affect stream discharge significantly. A more detailed explanation and relation to hydrology, agricultural practices, and published data can be found in Supplementary Material 4. Parameter α varied over time with the transition from flood irrigation to the more efficient center-pivot irrigation (Table 1), and this transition is nearly complete today. Parameters β and γ are conservative (Tables 2 and 3) and their significant change is not expected, unless there will be a climate shift within a decade.

Table 2
Overland flow coefficient β values related to land use.

Time periods	1941–1970	1971–2009
β^R	0.02	0.02
β^{DC}	0.045	0.04
β^T	0.005	0.005
β^{ir}	0	0
β^P	0.001	0.001

Table 4
Sensitivity analysis showing the average absolute relative error when there is a ±10% change in model parameters relative to the best calibrated parameter values.

Parameter	Increase 10%	Decrease 10%
q^{FC} inflow	5.7	4.4
q^{SWC} inflow	5.2	4.3
q^C	4.1	4.9
α	5.0	4.3
β^R	4.7	4.4
β^{DC}	4.4	4.4
β^T	4.6	4.6
β^P	4.6	4.6
γ^R	4.6	4.5
γ^{DC}	4.6	4.5
γ^T	4.6	4.5
γ^P	4.6	4.5

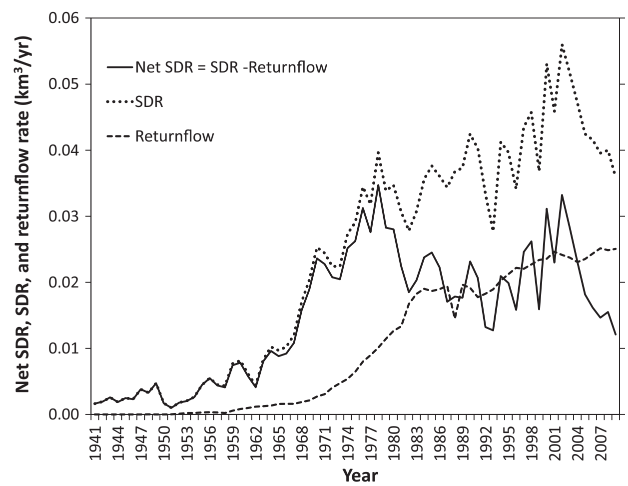

Fig. 13. The net effect of SDR when considering returnflow.

7.3.1. Returnflow impacts on SDR

An increase in modeled baseflow from 1972 to 2007 (Fig. 12) is attributed to two mechanisms of recharge. The first one is irrigation returnflow, resulting from inefficient irrigation practices in the 1940s and 1950s and eventually reaching FC through the aquifer. Another one is the overall increase in irrigated land area and corresponding recharge from precipitation compared to previous land use. This combination of inefficient surface irrigation and the overall expansion of irrigated area from 1941 to 1978 provided a buffer of increased recharge and then baseflow from 1972 to 2007 that dampened the effects of SDR. If from 1941 to 1978 irrigation efficiency was higher (like today) then irrigation returnflow and ultimately baseflow would be less than in modern conditions.

The real effect of irrigation wells is net SDR, or SDR reduced by the fraction of pumped GW that returns to the aquifer with

Table 3
Recharge coefficient γ values related to land use.

Time periods	1910–1940	1941–1970	1971–2009
γ^R	0.04	0.03	0.04

Time periods	1910–1930	1931–1940	1941–1960	1961–1980	1981–2009
γ^{DC}	0.02	0.015	0.025	0.03	0.02

Time periods	1941–1960	1961–1980	1981–2009
γ^T	0.17	0.16	0.15
Time periods	1941–1960	1961–1980	1981–2009
γ^P	0.09	0.08	0.07

irrigation application and constitutes returnflow. Fig. 13 shows the dampening effect of irrigation returnflow on SDR. In fact, net SDR has stabilized since 1980. According to new trends, the fraction of SDR that became returnflow (α) decreased from 25% for 1941 to 1955 to 3% for 1986 to 2009 due to the conversion from flood irrigation to center-pivot irrigation and reduced infiltration. This will have a tremendous impact on baseflow in the near future as less irrigation water is available for stream recharge. The baseflow will likely decrease further, maybe to 1986 levels, if not to 1940s and 1950s levels. This will bring SDR and baseflow curves in Fig. 12 much closer than they are today and cause more stress on the stream, because decreased baseflow will cause streamflow declines.

7.4. Reservoir effects

Reservoir storage changes have an important effect on the stream budget and subsequent model accuracy (see Fig. S2-2 in Supplementary Material 2). Without consideration of ΔS^R, the RE is significantly higher during reservoir construction and filling from 1951 to 1954 (20–35%) compared to RE with the ΔS^R consideration (10–15%). RE is much smaller and random when ΔS^R is considered.

7.5. Canal diversions

Unlike SDR's lag time, canal diversions resulted in a direct and immediate reduction of streamflow (Fig. 10). From 1941 to mid-1980s canal diversions decreased streamflow more than net SDR and became the primary component of Eq. (1). After 1980, annual canal diversions steadily declined and from 2001 to 2009 diversions were almost negligible or zero in Nebraska to meet the Republican River Compact allocations. Also, 2001–2009 canal diversions were much smaller than previous decades because of the severe early 2000s drought. This trend is likely to persist in the future as Nebraska attempts to meet the Republican River Compact allocations in the watershed with extensive GW-based irrigation. As of 2009, canal diversions are another factor causing the streamflow declines at Culbertson, in addition to SDR by irrigation wells, but they have opposite temporal trends of influence on streamflow.

7.6. Evapotranspiration

To corroborate our results with independently obtained estimates, annual evapotranspiration ET [km^3/yr] was calculated from the model results by subtracting total overland flow q and total SRR, excluding returnflow, from annual precipitation rate P within the basin as follows: $ET(i) = P(i) - q(i) - SRR(i)$.

The calculated average for ET for the decade of 2000 to 2009 is $ET_{ave} = 0.52–0.0093–0.004 = 0.51$ km^3/yr using average precipitation and modeled overland flow and SRR values for the 2000s. Szilagyi et al. (2003) estimated long term evapotranspiration in Nebraska, and estimates for the FC study area indicate a rate of 0.48 m/yr, or 0.48 km^3/yr. The difference is only 6% indicating the high accuracy of the simulated water budget in this study.

7.7. GW inflow

GW inflow into the stream in a given year differs from GW inflow to the basin. Two factors may affect this difference: (a) regional decline in water level on the watershed boundary resulting in saturated thickness (McGuire, 2011) and corresponding transmissivity changes from 0 up to 25% with an average on the order of 10% at various boundary segments over time and (b) a lag time between these changes at the boundary and at the stream that can be characterized by Eq. (3).

The first factor is due to the decrease of quantity of water delivered to the watershed boundary, while the second factor delays (mitigates) the arrival of these changes from this boundary to the stream reaches and ensuing decrease in Q_{in}. If instantaneously arrived from the boundary to the stream reaches, changes in saturated thickness would result in watershed and stream budget error of $\sim$10%. The second factor, assumption of steady (not reduced) transmissivity commonly used in analytical models (e.g., Foglia et al., 2013) leads to an overestimation of SDR and more conservative assessment of remaining SW resources.

Conservative estimates of the lag times between changes on the watershed boundary and changes on the stream by Eq. (3) are on the order of a century or more for the stream reach based on distance to the boundary (average $\sim$20 km), transmissivity, and the storage coefficient. Therefore, such error will be less than 10% considering the available time series (69 years). And indeed, inspection of the simulated stream budget and discharge q^{FC}_{out} in (Figs. 10 and 11) exhibits a small systematic bias on the order of +5% (also see discussion of relative error and Fig. S2-2 in Supplementary Material 2), as expected from the analyses above. This accuracy is quite adequate for the methodology.

8. Conclusions

The main goal of the study was to develop a simple model for simulating stream discharge under the influence of GW withdrawals and land use changes at a heavily irrigated watershed. This approach uses the stream water budget with an emphasis on a simplified description of various processes on the watershed. Modeling involves an analytical estimation of stream depletion rates and stream recharge rates with simplified treatment of overland flow. The role of the vadose zone is described by one parameter, a time lag between precipitation or irrigation applications and groundwater recharge to the water table. Land use parameters were obtained from the published and regional data with minor adjustments and sensitivity analyses. The model is applicable to other watersheds where streamflow data are available at the end points of a stream reach.

Application of this approach at the Culbertson gage on Frenchman Creek, Nebraska to the period from year 1941 to 2009 provides accurate estimates of stream discharge: relative error less than 10% for 88% of the modeled years at the stream gage (61 years of the total 69) and a 4.3% average relative error. Modeling of the pumping and land use change effects is consistent with field observations for low flow events in 1978 and 2002. Model-estimated evapotranspiration rates compare well with previously published data. The simplicity of the model also facilitates identification of causes and magnitude of possible errors in input data by sensitivity analyses and providing a measure of uncertainty due to various assumptions and imprecise data.

Another goal of the study, comparison of the applications of the Jenkins (1968) and Hunt (1999) SDR equations, gave insight into the differences between the two solutions with practical ramifications. Both equations produce practically identical results in our case. It is expected that when the stream is hydraulically well connected to the aquifer, multiple wells are relatively uniformly distributed over the domain, and distances to the stream far exceed stream width, the Jenkins (1968) solution should be used. The more parameter-demanding Hunt (1999) solution may be useful for more local studies with wells near the stream, when streambeds are of low permeability.

Analysis of the stream recharge processes aided in understanding the land use change effects on streamflow and the SW water budget. Using simple partitioning and tracing of annual precipitation into stream recharge and overland flow, it was found that the conversion of rangeland to dry cropland and terraced land has minor effect on stream recharge rates, overland flow rates, and streamflow on the Frenchman Creek watershed in comparison with irrigation withdrawals. The watershed lag time, which accounts for the period that soil moisture changes traversed the vadose zone was determined at 5 years, consistent with analyses by Rossman et al. (2014). However, the total effect of precipitation that becomes overland flow and stream recharge is substantial and roughly equal to the net effect of wells. Irrigated land is the most significant contributor to streamflow via returnflow. Returnflow from inefficient surface-irrigation techniques that were common prior to the 1960s contributes a significant amount of baseflow to the stream and dampens the effects of SDR.

The model shows that stream depletion due to GW pumping and canal diversions are the main causes of the observed streamflow declines of Frenchman Creek. However, they have opposite temporal trends over the modeling period (1941–2009); stream depletion from pumping results in a greater impact on Frenchman Creek over time than diversions. Land use change (conversion of rangeland to dry cropland and terraced land) and reservoir evaporation caused streamflow declines, but not to the same extent as GW pumping and canal diversions. Analysis also shows that inefficient water use on irrigated land in the past provided some enhancement of baseflow that is being exhausted by continuous pumping.

Conditions of a perennial stream are important for applications of our approach, because data from both the upstream and downstream gauges are needed. For example, this method could not be extended upstream from Imperial, because the stream gauge at Champion, 12 km upstream did not have adequate data for the time period. In this particular study, the investigated section of Frenchman Creek within the study area was never dry. Hopefully, more observations will become available to assess the model in the future (Konikow and Bredehoeft, 1992).

This approach provides a relatively expeditious and accurate assessment of a stream water budget compared to more complex models. This tool can be used to evaluate future possible changes to streamflow under past and future water management decisions.

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Appendix A. Supplementary material

Supplementary data associated with this article can be found, in the online version, at http://dx.doi.org/10.1016/j.jhydrol.2015.12.006.

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