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Peak groundwater depletion in the High Plains Aquifer, projections from 1930 to 2110

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ABSTRACT

Peak groundwater depletion from overtapping aquifers beyond recharge rates occurs as the depletion rate increases until a peak occurs followed by a decreasing trend as pumping equilibrates towards available recharge. The logistic equation of Hubbert's study of peak oil is used to project measurements at a set of observation wells, which provide estimates of saturated thickness and changes in groundwater storage from 1930 to 2110. The annual rate of depletion in High Plains Aquifer of the central USA is estimated to have peaked at 8.25×10^9 m^3/yr in 2006 followed by projected decreases to 4.0×10^9 m^3/yr in 2110. The timing of peaks follows a south–north progression, with peaks occurs in 1999 for Texas, 2002 for New Mexico, 2010 for Kansas, 2012 for Oklahoma and 2023 for Colorado; peaks do not occur before 2110 for Nebraska, South Dakota and Wyoming. The manifestation of peak groundwater depletion contributes towards the more comprehensive understanding necessary to assess potential vulnerabilities in the water-food nexus posed by aquifer depletion.

1. Introduction

Groundwater provides a freshwater source that is relatively reliable and stable, and has contributed towards the security and sustainability of irrigated agriculture. Groundwater is studied in the High Plains Aquifer, where depletion of these resources has contributed towards net agricultural productivity (Ripl, 2003) and supports one of the most important food production regions of the world (Steward et al., 2013). Water scarcity is a growing concern for global food supplies (Rosegrant and Cline, 2003), particularly as depletion threatens groundwater stores (Lilienfeld and Asmild, 2007). This study addresses the sustainability concerns that exist for the High Plains Aquifer (Scanlon et al., 2012), and the need for a more complete understanding of depletion to support groundwater management (Alley et al., 2002).

Projections of change in groundwater storage are developed for each State overlying the High Plains Aquifer from 1930 to 2110. These curves identify the occurrence of peak groundwater depletion, where the rates of depletion increase in intensity towards this peak and then decrease at later times. Methods use theory from the Hubbert curves of peak oil (Hubbert, 1956) and extend application of logistic equations from Kansas (Steward et al., 2013) to the High Plains Aquifer. Projections and analysis of peak groundwater depletion and its variations across the hydrologic and geologic setting provide a context to understand the evolution of groundwater stores over time. Identification and quantification of the peak limitations in natural resources (Seppelt et al., 2014) contributes towards understanding vulnerabilities and sustainability challenges within the water-food nexus (Gerbens-Leenes et al., 2009).

2. Methods: approximating a well hydrograph

2.1. The High Plains/Ogallala Aquifer

The High Plains Aquifer study region is situated in the central plains of the USA and its location is identified in Fig. 1. This temperate semi-arid grassland, with limited surface water supplies in ephemeral streams and playa lakes and few perennial rivers and lakes amongst the Sand Dunes in northern Nebraska, was once known as the "Great American Desert". Irrigation using surface water began in the late 1800s by diverting streamflow into irrigation canals, and storage reservoirs were constructed later to supplement streamflow during dry years

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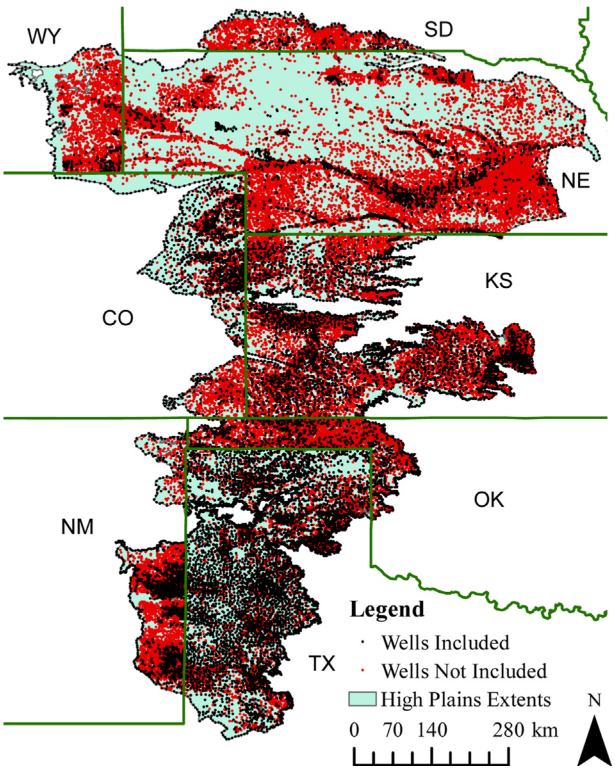


Fig. 1. The High Plains/Ogallala Aquifer study region in Colorado (CO), Kansas (KS), Nebraska (NE), New Mexico (NM), Oklahoma (OK), South Dakota (SD), Texas (TX), and Wyoming (WY) is situated in the central plains of the USA, and contains a northern basin above the Smoky Hills River in central Kansas, a central basin, and a southern basin in Texas and New Mexico below the Red River (Gutentag et al., 1984). The observation wells from USGS (2015c) identify those used in this study, and GIS images are projected in the Geographic Coordinate System North American Datum of 1983 following Cederstrand and Becker (1999).

(Gutentag et al., 1984, p. 40). More reliable groundwater sources were tapped by well diggers starting in the 1880s; however, the dust bowl of the 1930s along with the development of modern pump hydraulics and improvements in irrigation technology stimulated development of groundwater reservoirs (Opie, 2000). Groundwater pumping in the High Plains Aquifer currently supports 30% of the irrigated agriculture in the USA (USGS, 2015a) and has transformed the region into the "Breadbasket of the World", yet concerns exist about its long-term prospects (Ziolkowska, 2015).

The High Plains Aquifer is "saturated, generally unconsolidated deposits" of mostly "near-surface sand and gravel deposits" overlying a base formed by erosional surface cuts in bedrock from the Permian to Tertiary periods (Dugan et al., 1994; Gutentag et al., 1984). The Ogallala formation is the predominant component of the High Plains Aquifer, and was formed from erosion of the Rocky Mountains west of the region, and the transport and subsequent deposition of this material to lower eastward elevations (Sophocleous, 2012). The geologic units are isolated from adjacent units along the western edge with little groundwater entering along this boundary (Gutentag et al., 1984, p. 20). Regional flow is driven by recharge across the west-to-east hydraulic gradient in the sloping aquifer, and streams that are ephemeral in their western upstream reaches may become perennial if their eastern downstream reaches incise the groundwater table (Gutentag et al., 1984, p. 28). Regionally, the High Plains Aquifer behaves like a water-table aquifer (Gutentag et al., 1984, p. 1, 57) and the upper boundary is the groundwater table (McGuire et al., 2003, p. 26).

The observation wells used to study groundwater depletion are shown in Fig. 1, and identify areas where groundwater levels, and thus changes in groundwater storage, have been measured. The data associated with each well may be visualized as a hydrograph containing a set of M measurements of groundwater head h_m at time t_m, where m varies between 1 to M. For example, hydrographs are shown in Fig. 2 for representative wells in the state overlying the High Plains Aquifer. These measurements are above the elevation of the base of the aquifer B for each well and below a maximum groundwater elevation denoted $h_{\max}$. This figure also illustrates the logistic equation, which passes through the groundwater measurements and provides an approximate function to evaluate changes in groundwater head (water table elevation at a well) over time. This equation and its application to observation wells throughout the High Plains Aquifer is described next.

2.2. Logistic equation and regression

The logistic equation is an S-shaped curve that was developed by Verhulst (1838) to study population growth bounded by an upper limit imposed by carrying capacity. This equation has been applied extensively throughout the sciences to study processes that asymptotically progress from one value to another over time. Within the geosciences, it has been used to study natural resources depletion following the seminal work of Hubbert (1956) for fossil fuels. Recently, Gleick and Palaniappan (2010, Fig. 4) proposed use of the logistic equation to study the production of renewable freshwater supplies and the manifestation of this theoretical curve in unsustainable groundwater extraction (Gleick and Palaniappan, 2010, Fig. 6). Their conceptualization of "peak nonrenewable water" with the logistic equation provides a foundation for this study of peak groundwater depletion in the High Plains/Ogallala Aquifer.

The logistic function was previously applied to study groundwater depletion in the Kansas portion of the High Plains Aquifer by Steward et al. (2013). The saturated thickness (head minus base) was approximated by

$$\hat{h}(t) - B = \frac{h_{\max} - B}{1 + e^{(a_0 + a_1 t)}} \tag{1}$$

where a_0 and a_1 are coefficients adjusted to match the measurements in the hydrograph of a well. It will be convenient later to use the dimensionless form:

$$\mathcal{H} = \frac{1}{1 + e^{\mathcal{T}}} \quad , \quad \mathcal{H} = \frac{h - B}{h_{\max} - B} \quad , \quad \mathcal{T} = a_0 + a_1 t \tag{2}$$

where the dimensionless saturated thickness $\mathcal{H}$ varies from 0 (empty aquifer) to 1 (full aquifer), and the dimensionless time $\mathcal{T}$ is equal to 0 when the aquifer is half-full.

The following method is used to obtain the coefficients a_0 and a_1 from the data for each well. Regression minimizes the objective function equal to the sum of the squares of the residual error of groundwater approximation at times t_m minus the measured head:

$$\mathcal{F} = \frac{1}{M} \sum_{m=1}^{M} \left[\hat{h}(t_m; a_0, a_1) - h_m \right]^2 \tag{3}$$

The Levenberg–Marquardt method (Levenberg, 1944; Marquardt, 1963) provides an algorithm to iteratively solve the system of two equations with the two unknown coefficients

$$\left(\mathbf{J}^{\mathrm{T}} \mathbf{J}|_q + \lambda \mathbf{I} \right) \left(\begin{bmatrix} a_0 \\ a_1 \end{bmatrix}_{q+1} - \begin{bmatrix} a_0 \\ a_1 \end{bmatrix}_q \right) = -\mathbf{J}^{\mathrm{T}} \mathbf{f}|_q \tag{4}$$

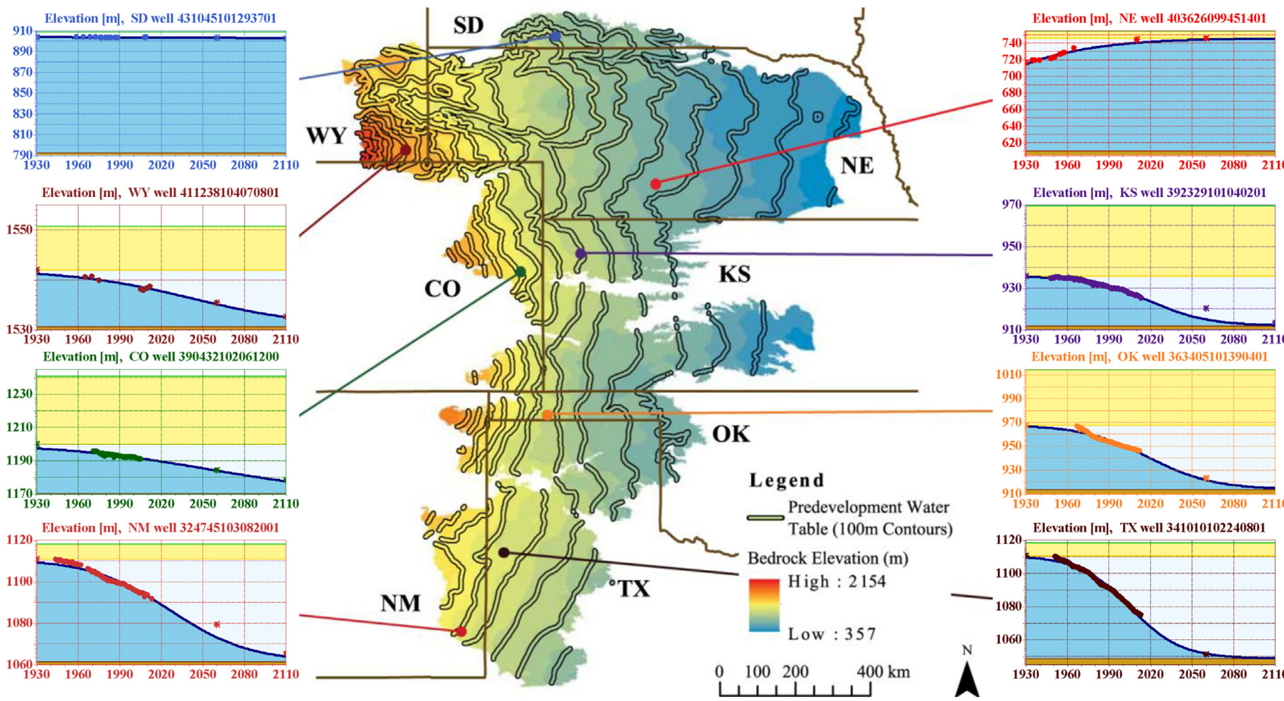


Fig. 2. Use of the logistic functions to project groundwater levels is illustrated by hydrographs for the representative observation wells in each state previously identified by Dugan et al. (1994, pp. 39–53) and McGuire (2011, p. 9). Within each hydrograph, groundwater stores lie in the blue saturated region above bedrock base elevation and below the groundwater level at the well. A vadose zone exists in the white region that is either dewatered (depleting conditions) or filled (increasing head over time), and in the upper yellow region that is above maximum groundwater elevation and below the land surface. The markers in each hydrograph show head measurements and extrapolation points (in 1930, 2060 and 2110) added using a linear trend. The elevations of bedrock and predevelopment groundwater head are shown across the region.

where **I** is the identity matrix, **J** is the Jacobian,

$$\mathbf{J} = \begin{bmatrix} \dfrac{\partial \hat{h}(t_1; a_0, a_1)}{\partial a_0} & \dfrac{\partial \hat{h}(t_1; a_0, a_1)}{\partial a_1} \\ \dfrac{\partial \hat{h}(t_2; a_0, a_1)}{\partial a_0} & \dfrac{\partial \hat{h}(t_2; a_0, a_1)}{\partial a_1} \\ \vdots & \vdots \\ \dfrac{\partial \hat{h}(t_M; a_0, a_1)}{\partial a_0} & \dfrac{\partial \hat{h}(t_M; a_0, a_1)}{\partial a_1} \end{bmatrix} \tag{5}$$

and **f** contains the residual errors.

$$\mathbf{f} = \begin{bmatrix} \hat{h}(t_1; a_0, a_1) - h_1 \\ \hat{h}(t_2; a_0, a_1) - h_2 \\ \vdots \\ \hat{h}(t_M; a_0, a_1) - h_M \end{bmatrix} \tag{6}$$

These matrices are evaluated using a_0 and a_1 in the qth iterate to solve for these coefficients in the next iterate, and λ is adjusted for each iterate following Marquardt (1963) until convergence is achieved.

This non-linear regression process was applied to the 14,016 study wells identified in Fig. 1. These wells contain 324,290 measurements of groundwater level that are shown on a single dimensionless plot in Fig. 3 by computing $\mathcal{H}$ and $\mathcal{T}$ using (2). This figure also contains the logistic equation, and identifies the measurements associated with the 8 reference wells in Fig. 2. The goodness of fit is quantified in Table 1 where the residual error (difference between measurements and the logistic equation approximation) is tabulated for the wells in each state. Our approximation closely matches observations with a root mean square error

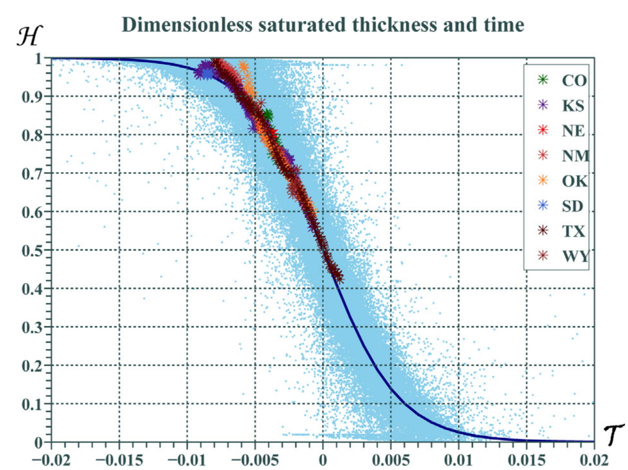


Fig. 3. The logistic curve is plotted on common axes of dimensionless saturated thickness $\mathcal{H}$ vs. dimensionless time $\mathcal{T}$. The observations for each well are individually fit to this curve and the measurements for the representative wells in Fig. 2 are identified, along with the 324,290 measurements of groundwater elevation at observation wells used in this study. Note that extrapolation points at 1930, 2060 and 2110 are not shown.

of 1.73 m, which is minimized in the objective function (3), and an average absolute error of 1.07 m for the entire High Plains Aquifer.

2.3. Rectifying data sources

2.3.1. Data and projections

A variety of sources provide groundwater data at the observation wells. The base elevation and groundwater elevation during predevelopment conditions before large-scale pumping were presented as contour maps in the USGS Regional Aquifer-System Analysis (RASA) of the High Plains Aquifer (Gutentag et al., 1984).

Table 1
The total number of observation wells are reported for each State and the High Plains Aquifer, along with the number of wells dropped from computations due to too few observed data points and too small a time interval when measurements exist. Subtracting these gives the number of wells and their measurements used in this study. Estimates of groundwater elevation for 2010 were added from a kriged surface to wells with all measurements before 1990, and 2110 estimates from linear regression were added for wells where this value lies between the aquifer base and maximum head. The residual difference between observed measurements and the functional approximation is presented as the root mean square error and the mean absolute error.

	CO	KS	NE	NM	OK	SD	TX	WY	High Plains
Well total	4489	14,499	19,980	6039	5014	2992	11,506	3942	68,461
Well dropped ($M < 4$)	3252	10,687	17,467	4611	4230	2874	5615	3554	52,290
Well dropped (time < 10 yr)	55	554	647	100	124	2	538	135	2155
Well used in study	1182	3258	1866	1328	660	116	5353	253	14,016
Number of observations	22,570	101,082	54,227	21,155	13,223	1119	107,268	3646	324,290
Added 2010 estimates	443	912	1185	421	184	2	1565	194	4906
Added 2110 linear trend	649	1748	1172	378	317	67	1843	158	6332
Residual: rmse [m]	1.376	1.689	1.022	1.346	1.972	0.623	2.121	1.659	1.730
Residual: mean absolute error [m]	0.890	1.013	0.590	0.898	1.160	0.442	1.435	1.012	1.070

Digitized versions of the contour lines (Cederstrand and Becker, 1998, 1999) were used with the Topo to Raster tool of ArcGIS to generate raster cells across the region. The land surface elevation is also available as raster cells at the National Elevation Dataset (USGS, 2015b) in a Digital Elevation Model with 1 arc-second resolution (approximately 30 m). The values of base, predevelopment groundwater elevation and land surface elevations at observation wells are obtained by extracting their value from these cells using ArcGIS. The groundwater level in observation wells are measured by many agencies across the High Plains, and these data have been coalesced into a consistent set of head measurements by the USGS (2015c) that are organized as a time series for each well.

These disparate data sets provide complementary information for projection of groundwater level in wells. Steward et al. (2013) developed procedures to rectify the data across sources for projections in Kansas that are extended here to work across all wells in all states. An overarching goal was to retain as many observation wells and measurements as possible using an objective procedure that consistently eliminated all wells from all states with data that gave unreasonable projections. While for the long time interval 1930–2110, the logistic function provides a mathematical form with continuous variation, for a short time interval at individual wells it does have the capacity to dewater or fill the aquifer too quickly. Projections are considered unreasonable when the projected annual declines over a short period of time are not consistent with those of neighboring wells. Unreasonable projections were identified by analyzing the maximum annual changes in head two ways, first by analyzing hydrographs of individual wells, and secondly by developing maps of the annual change in saturated thickness for the entire study region to identify points where individual wells deviated from their neighbors.

2.3.2. Groundwater level measurements during the recovery period

The groundwater level near irrigation wells is impacted by the annual pumping cycle where the groundwater level declines during the pumping period, it increases after pumping stops and before the next pumping cycle begins (Theis, 1935), and a residual drawdown occurs in depleting conditions. This may result in significant annual fluctuations of groundwater level in observation wells (to tens of meters) particularly during high water stress periods such as the 2011 drought conditions in the Southern High Plains (Mullican, 2012). Consequently, observation measurements are usually made in the winter and early spring although field conditions in the Northern High Plains may require late fall measurements (Dugan et al., 1994, p. 23). Groundwater levels are also impacted by evolving irrigation practices, where the pumping period in the southern High Plains has begun earlier due to falling water tables and decreasing pumping rates and off-season irrigation before planting is commonly used to replenish soil water (Stone et al., 2008). This study only uses data for water level measurements from December 1 to January 31 so that measurements are later in the recovery period but before the next pumping period, following the criteria used in Steward et al. (2009a, 2013). Note that some wells identify measurements taken while wells are pumped, and these data were also discarded.

2.3.3. Insufficient number of data points

The next step in preparing data to match the logistic curve is to discard observation wells where the “fitting process is fundamentally flawed” by “insufficient data to make a meaningful fit” (Brandt, 2007). We chose to not include observation wells where the number of measurements is $M < 4$, thus avoiding an overspecification ratio (Janković and Barnes, 1999) of less than 2 times the number of coefficients (a_0 and a_1). While this criterion removed 52,290 of the 68,461 observation wells from consideration in the High Plains as shown in Table 1, most of these wells (40,167) were discarded because they did not have any measurements during the recovery period. Note that the wells included and excluded from this study are shown in Fig. 1.

2.3.4. Recent observations for wells with early data

The data for groundwater level available from the USGS (2015c) contained values beginning in 1930 (1935 for Oklahoma) and ending in 2010, 2011 or 2012 depending upon the state. Many wells only have measurements in early years, and the water level from neighboring wells was used to estimate the recent groundwater level in these wells as follows. A surface of groundwater level in 2010, the latest year for which all states had data, was developed using universal kriging in ArcGIS for all wells with data during the 2010 recovery period. The value was extracted from this surface at each observation well, and this 2010 estimate was added to the data for measured groundwater elevation for wells with no measurements since 1990. This criterion was established by analyzing the hydrographs of a large sample of wells in each state where the difference between the 2010 estimate and the logistic equation was significant – larger than $0.1(h_{max} - B)$ or 10 m. For wells with recent data, the 2010 estimates were dropped because the measured data are more accurate. However, wells with measurements only before 1990 benefited by enabling the logistic equation to approximate the recent 2010 estimates and establish a longer-term trend that prevented many of the wells from dewatering much more quickly than their neighboring wells. Note that the surface obtained by kriging wells with measurements in 2010 generated estimates at some of the wells without 2010 data that were outside the range of B to $h_{\max}$ (particularly in regions where the aquifer is very thin), and so these estimates were not used for those wells. The number of wells where 2010 estimates were added is tabulated in Table 1.

2.3.5. Insufficient time interval

Many of the observation wells had a very limited time interval over which measurements were made and a meaningful fit for the logistic equation could not be achieved. It was decided to drop all wells with less than 10 years of data, including the 2010 estimate if it was added. This criterion was developed by viewing the hydrographs of a large sample of wells with measurements over time intervals of different lengths. It was found that, while the logistic equation provided reasonable estimates for most wells with 9 years of data, there were some wells where the maximum annual change was larger than expected. This did not occur for wells with 10 and 11 years of measurements; the rate of decline (or increase) was reasonable across the 1930–2110 period. The number of wells dropped due to an insufficient time interval is also tabulated in Table 1.

2.3.6. Reconciling base and maximum head elevations

The elevations of the base of the aquifer, B, and the maximum head, h_{max}, provide bounds for the saturated groundwater zone. It is assumed that data sources progress from most to least accurate elevations as:

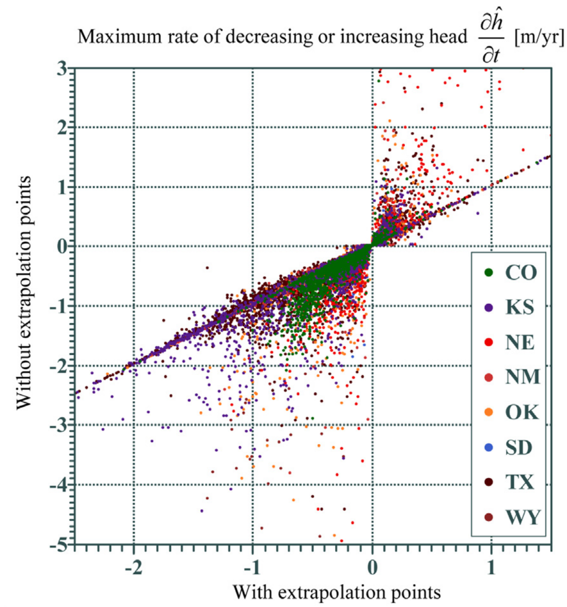
1. Observation well measurements (land surface minus measured depth to water).
2. DEM (Digital Elevation Model) land elevation (less accurate than surveyed wells).
3. Bedrock and predevelopment water level (interpolated between contour lines).

The base elevation B is set equal to the bedrock level from the interpolated contours and lower than all groundwater measurements. These calculations use the lowest observed measurement for all times at each well minus a small tolerance of $\delta = 0.02 \times$ (highest head − lowest head measured in well) to ensure that all head measurements are above B. The maximum head h_{max} is set equal to the predevelopment water level from the interpolated contours, lower than land elevation since the aquifer is unconfined (Gutentag et al., 1984), and higher than the highest observed measurement at each well plus the tolerance δ so all head measurements are below h_{max}.

2.3.7. Establishing consistent trends

The logistic equation, (1), provides a groundwater level that asymptotically approaches the base of the aquifer B for depleting conditions or the maximum head h_{max} for a rising water table with passing time. However, this function has the capacity to deplete or fill the aquifer too quickly at rates inconsistent with neighboring wells. This behavior is controlled by adding points obtained by extrapolating along a linear trend in data as per Steward et al. (2013). Extrapolation points are calculated for all wells using linear regression through the observation data (including 2010 estimated points) and adding the value along this linear trend in 1930 and 2060, while making sure these points are in the saturated zone – between $B + .001(h_{max} - B)$ and $h_{max} - .001(h_{max} - B)$. Another extrapolation point is added from the linear trend at 2110, but only if this point lies between B and h_{max}. (Table 1 identifies the number of observation wells in each state where the linear trend extended through 2110 and this extrapolation point was added.) These points are illustrated for the representative wells in Fig. 2.

Addition of the extrapolation points did not significantly change the projections of most wells, but they were needed for some to achieve reasonable projections consistent with neighboring wells. The sensitivity of results to adding these extrapolation points to



Absolute change in max rate from extrapolation points					
change [m/yr]	0–0.1	.1–.25	.25–.5	.5–1	>1
count	10458	1567	919	641	431
percent	75%	11%	7%	4%	3%

Fig. 4. The sensitivity of adding extrapolation points in 1930, 2060 and 2110 is quantified by plotting the maximum rate of decline occurring with and without this criteria (with 13,911 of 14,016 wells inside the visible data range).

the measured data is quantified in Fig. 4. The maximum annual change in groundwater level occurring during the study period 1930–2110 is plotted for each well along the x-axis with inclusion of the extrapolation points and along the y-axis when they are not included. This illustrates that most of the wells are not significantly impacted by the inclusion of extrapolation points (75% change by less than ±0.1 m/yr or one tick mark on the axis), and that water level continues to decline or increase regardless of whether points are included. The extrapolation points help the maximum rate of annual change in groundwater elevation to align with realistic values. For example, while Texas experienced localized areas with declines exceeding 3 m/yr in 2011, only a very few wells consistently declined by 1.5 m/yr or more across a 10 year period (Mullican, 2012).

The good quality of fit of the logistic equation to the observation measurements is quantified in Table 1, using the average absolute value of the difference between the measurements and the approximate function and the root mean square error [square root of (3)]. While these methods largely follow those established by Steward et al. (2013), they have been adapted to work for the data across the High Plains Aquifer region. For example, the assumption in Steward et al. (2013) of a declining water table was dropped (along with criterion to drop wells if the predevelopment groundwater level from the logistic equation was too low) since many regions of the High Plains region experienced increasing groundwater levels. Many of the observation wells with rising groundwater level, where $a_1 < 0$ in (1), occur in recharge zones beneath ditch irrigation, and are due to early recovery from the low groundwater levels of the dust bowl (Gutentag et al., 1984, p. 45). Due to changed criteria and regression method and addition of new wells (3258 Kansas wells in Table 1 with inclusion of wells in central Kansas vs. 1601 wells in the subset of Kansas studied in Steward et al. (2013)), the absolute difference dropped from 1.522 m (Steward et al., 2013, Eq. (3)) to 1.013 m for Kansas in Table 1.

3. Application: tapping the High Plains Aquifer

The methods to analyze well hydrographs are applied to study groundwater depletion due to tapping groundwater in the High Plains Aquifer beyond the recharge rate. The changes in groundwater stores over time may be evaluated using the observation wells as follows. The saturated thickness of the unconfined aquifer at each well is equal to the groundwater level minus the aquifer base, $h - B$, and the approximate value may be computed for any time by evaluating the logistic equation, (1). This saturated thickness is multiplied by the specific yield, S_y, and integrated over the surface area of the aquifer to give the volume of groundwater in storage:

$$S(t) = \iint\limits_{\text{aquifer}} \left[\hat{h}(t) - B\right] S_y \quad \mathrm{d}A \tag{7}$$

where the storage S varies over time with changes in the saturated thickness. GIS data exists for the High Plains region for both the specific yield (Cederstrand and Becker, 1998) and the lateral extent of the aquifer (Qi, 2010).

Computation of storage was implemented as follows. First, the groundwater storage was computed for predevelopment conditions before wells began extracting significant quantities of groundwater by evaluating the logistic function for each well at time $t_0 = 1930$:

$$S_0 = S(1930) = \iint\limits_{\text{aquifer}} \left[\hat{h}(1930) - B\right] S_y \quad \mathrm{d}A \tag{8}$$

This was calculated in ArcGIS using universal kriging with second-order trend removal to obtain a raster of saturated thickness from the observation wells, multiplying this by a raster of specific yield, and summing for cells overlying the High Plains Aquifer. The computed values of predevelopment storage are tabulated in the first row of Table 2 for each State and the High Plains.

The annual change in storage is obtained from the change in water level in (7) over the time interval $\Delta t = 1$ yr:

$$\Delta S(t) = \iint\limits_{\text{aquifer}} \left[\hat{h}(t) - \hat{h}(t - \Delta t)\right] S_y \quad \mathrm{d}A \tag{9}$$

This was calculated for all years from 1931 to 2110 by computing the difference in saturated thickness for all wells, kriging across the High Plains to develop a set of rasters for annual change, and evaluating the integral by summing cells. The volume of water in storage at year t is then computed by adding these changes to the predevelopment storage:

$$S(t) = S_0 + \sum_{\tau=1931}^{t} \Delta S(\tau) \tag{10}$$

Table 2 shows results for the volume of groundwater in storage at 10 year increments. Note that these results are aggregated to the State level to aid in comparison with previous studies, however, they are obtained from integration of detailed projections of depletion across the region.

These spatial patterns of saturated thickness and its evolution over time are illustrated in Fig. 5. The image of predevelopment storage shows the surface obtained by geospatially interpolating the values from the logistic equation in 1930 at the observation wells. The subsequent images at 30-year intervals were obtained by successively adding the rasters of annual water level change, and the raster cells are visualized using the same symbology as Steward et al. (2013). These spatial patterns help interpret the values in Table 2. The largest groundwater stores exist in Nebraska and the northern basin of the High Plains Aquifer (McGuire, 2011). And significant depletion occurs, particularly within the central and southern basins, as well as the southern portion of the northern basin.

Two procedures were implemented to enable integration of storage to match geological constraints. First, the lateral boundaries of the High Plains Aquifer occur at the physical limits of the geologic units (Gutentag et al., 1984, p. 20) where "little or no saturated thickness" occurs (Dugan et al., 1994, p. 22). This boundary was established by locating boundary points along the outline of the High Plains Aquifer with zero saturated thickness and with no annual change in saturated thickness. These boundary points were identified at approximately 5 km increments along the aquifer boundary (Qi, 2010) by writing a python script in ArcGIS. They were then used with the measurements at observation wells to establish the surfaces of groundwater elevation in Fig. 5 and to compute storage with the integrals (7)–(9). Note that the summation of predevelopment plus every annual estimate of change in storage through 2110 is listed in the second to last line of Table 2, and this accurately matches the storage listed on the last line of this Table obtained directly from (7) using groundwater elevation at $t = 2110$. This procedure to force saturated thickness to zero along the boundary contributed towards the accuracy of this summation, since groundwater is neither gained nor lost due to incorrect kriging near the lateral edges of the aquifer where wells do not exist.

A second geological constraint is imposed by the Wheatland and Whelan faults in Wyoming were a vertical displacement of about 300m occurred, and the bedrock elevation and fill material have different properties on each side (Gutentag et al., 1984, pp. 16–17). Consequently, spatial interpolation was performed independently on each side of the fault. This was accomplished by selecting wells on one side of the faults, kriging them and masking out the aquifer on the other side of the faults; then switching the selection, kriging only those wells and masking again; and then adding the two rasters. In both cases, wells along the boundary were included on the correct side of the fault, and this process was repeated for all years to develop the rasters of annual groundwater level change. The choice to subdivide the domain at the fault follows (Luckey et al., 1986, p. 36) who did not model the area NW of the fault, but instead applied a constant head Dirichlet condition along the SE boundary of the fault for the High Plains. The difference in groundwater level across the fault line in Wyoming is evident in Fig. 5.

The spatial distribution of change in groundwater elevation in Fig. 6 illustrates the summation of annual change over 30 year intervals, and matches the patterns observed in retrospective studies. Within the 1930–1960 period, the groundwater level rose in areas with strong surface water interactions in response to recovery from the drought of the 1930s and land-use change (Gutentag et al., 1984, p. 45), as well as ditch irrigation particularly along the Platte River in Nebraska (Luckey et al., 1981). Regions with high groundwater depletion formed as high-capacity irrigation wells were developed following a south-north trend, with declines in the groundwater table becoming evident by 1940 in the southern basin, by 1950 in the central basin and by 1960 in the northern basin (Dugan et al., 1994, p. 19). The groundwater elevation has been stable in regions of the High Plains where groundwater was not tapped due to shallow saturated thickness incapable of supporting large capacity wells or land-use constraints imposed by soils and topography (Dugan et al., 1994, p. 19).

Analysis of the changes in storage helped to establish and refine procedures. The procedures and criteria to exclude observation wells with too few data points or too short a time interval were evaluated by identifying wells with annual changes that dewatered or filled the aquifer too quickly. This was accomplished by visually inspecting GIS images of annual change to identify isolated points where a well has a vastly different annual change than the surrounding wells. Note that the blue region of central

Table 2
The projected groundwater storage for each State and the High Plains/Ogallala Aquifer is reported in units of km^3 (10^9m^3), and as a fraction of predevelopment 1930 storage.

Year	CO	KS	NE	NM	OK	SD	TX	WY	High Plains
Storage obtained by cumulatively subtracting annual change in storage from 1930 storage									
1930	133(0%)	420(0%)	2599(0%)	58(0%)	145(0%)	71(0%)	578(0%)	89(0%)	4093(0%)
1940	133(0%)	417(1%)	2606(−0%)	57(1%)	146(−0%)	71(−0%)	571(1%)	89(0%)	4089(0%)
1950	132(1%)	413(2%)	2611(−0%)	56(2%)	146(−0%)	71(−1%)	560(3%)	88(0%)	4077(0%)
1960	130(2%)	407(3%)	2613(−1%)	55(5%)	145(−0%)	72(−1%)	543(6%)	88(1%)	4053(1%)
1970	127(5%)	397(5%)	2614(−1%)	53(8%)	144(1%)	72(−1%)	521(10%)	87(2%)	4014(2%)
1980	123(8%)	383(9%)	2612(−0%)	51(12%)	143(2%)	72(−1%)	493(15%)	86(3%)	3962(3%)
1990	118(11%)	365(13%)	2608(−0%)	47(18%)	140(4%)	72(−2%)	460(20%)	85(4%)	3897(5%)
2000	113(15%)	345(18%)	2602(−0%)	43(25%)	136(6%)	72(−2%)	423(27%)	84(5%)	3820(7%)
2010	107(20%)	322(23%)	2594(0%)	40(31%)	132(9%)	72(−2%)	387(33%)	83(6%)	3738(9%)
2020	100(25%)	300(28%)	2585(1%)	37(36%)	127(12%)	73(−2%)	355(39%)	82(7%)	3659(11%)
2030	94(30%)	279(33%)	2574(1%)	35(39%)	123(15%)	73(−2%)	328(43%)	81(8%)	3587(12%)
2040	87(34%)	261(38%)	2563(1%)	34(41%)	119(18%)	73(−2%)	307(47%)	80(9%)	3524(14%)
2050	81(39%)	245(42%)	2551(2%)	33(43%)	116(20%)	73(−2%)	289(50%)	79(11%)	3466(15%)
2060	75(43%)	230(45%)	2538(2%)	32(44%)	112(23%)	73(−2%)	274(53%)	78(12%)	3413(17%)
2070	70(47%)	218(48%)	2524(3%)	31(46%)	109(25%)	73(-2%)	262(55%)	77(13%)	3364(18%)
2080	65(51%)	206(51%)	2509(3%)	31(47%)	106(27%)	73(−2%)	251(57%)	75(15%)	3317(19%)
2090	61(54%)	196(53%)	2493(4%)	30(47%)	104(29%)	73(−2%)	242(58%)	74(17%)	3273(20%)
2100	57(57%)	187(55%)	2477(5%)	30(48%)	101(30%)	73(−2%)	234(60%)	73(18%)	3231(21%)
2110	53(60%)	179(57%)	2460(5%)	29(49%)	98(32%)	73(−2%)	227(61%)	71(20%)	3190(22%)
Storage obtained from saturated thickness at 2110									
2110	54	179	2475	28	99	71	227	74	3207

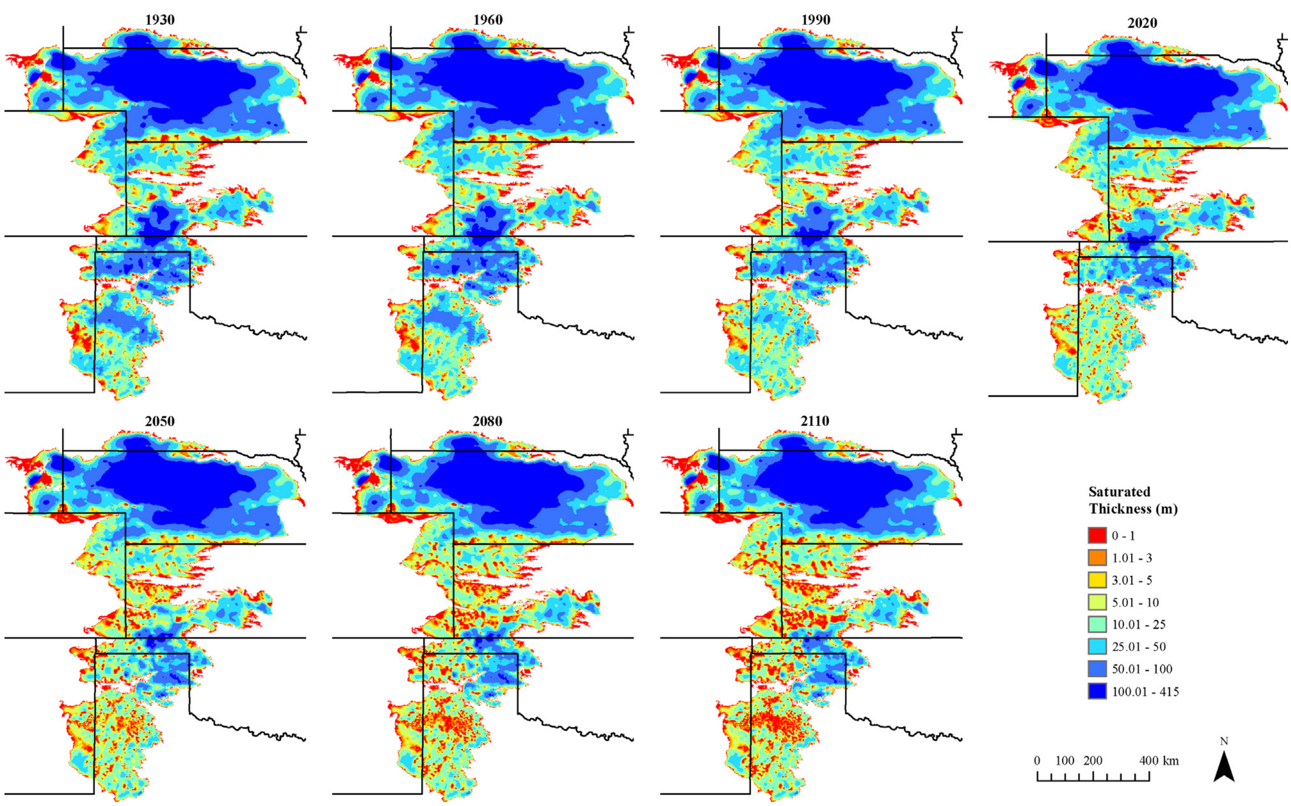


Fig. 5. The projected saturated thickness is shown at 30 year intervals from 1930 to 2110.

Nebraska in Fig. 6 for later times is an artifact of kriging an area with no observation wells; it is not due to wells filling the aquifer too quickly. Visual inspection of annual change augmented analysis of the hydrographs of sample wells and provided interpretation within the regional context of neighboring wells. It also identified a few wells with isolated measurements that were inconsistent, where the well behaved differently than surrounding wells (e.g, one well had one measurement of depth to water of 24 instead of 240). The groundwater elevation is also stable along the boundary of the High Plains, where change in water level was forced to zero by introduction of the boundary points. By adding these points, the projections of saturated thickness in Fig. 5 along the boundary are consistent with those in previous studies (Luckey et al., 1986; Dugan et al., 1994).

Our results for groundwater storage closely match previous studies as shown in Table 3, and comparisons across methods are briefly described. First, storage results are compared to the USGS RASA study for predevelopment storage in 1930 and for 1980. Our predevelopment storage is lower in Texas because the observation wells show lower groundwater elevation along the intersection of

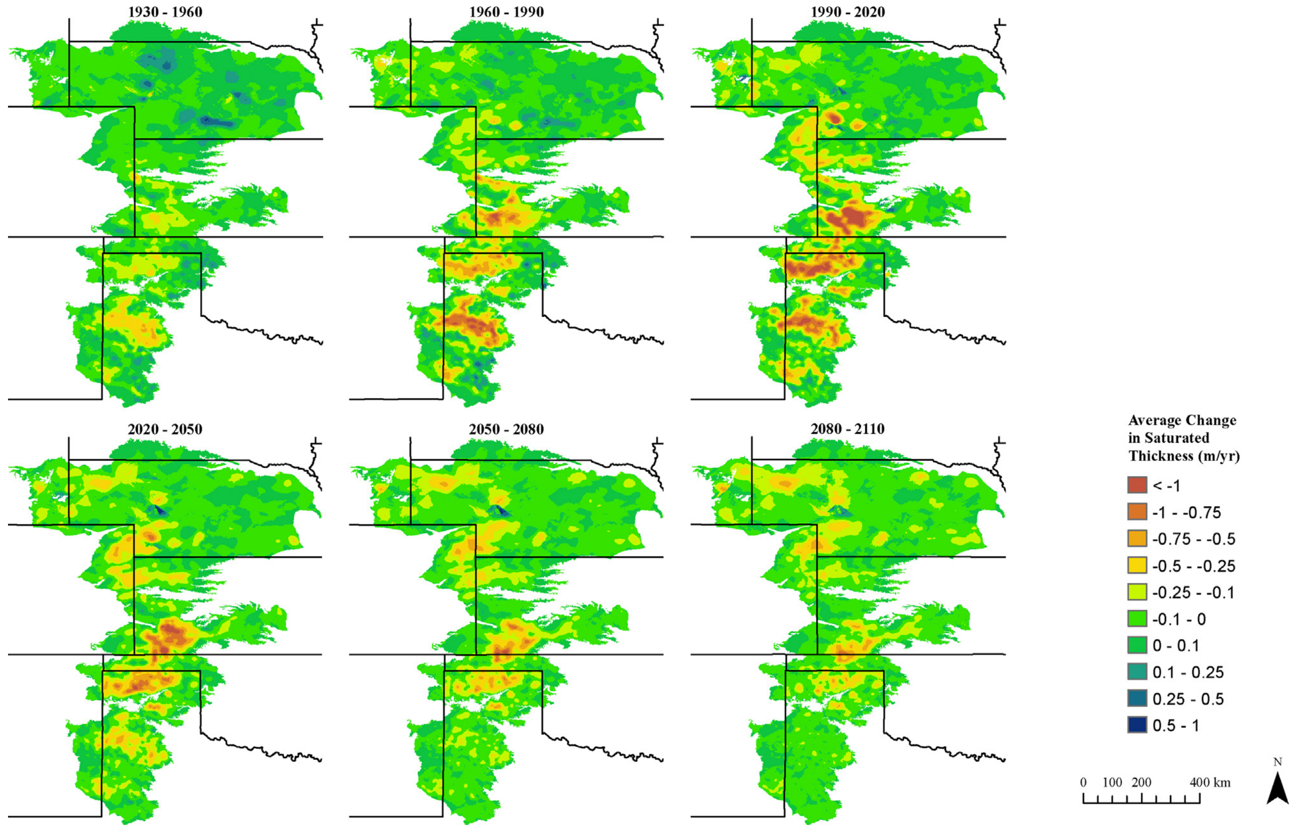


Fig. 6. The projected annual change in saturated thickness averaged over 30 year intervals is shown.

the Canadian River with the Ogallala and also along the eastern escarpment than Luckey et al. (1981) and Gutentag et al. (1984). Other state results agree very well, although our model has slightly smaller storage due to the lower saturated thickness of observation wells in Colorado (north of the Platte and south of the Arkansas River) and in New Mexico (in the southern lobe). The next results in Table 3 show a different estimate of predevelopment storage that was found by McGuire et al. (2003, p. 18) using the earliest available groundwater level measurement in over 20,000 wells. This estimate is specified along with the median measurement year for each State and the storage for these years as in Table 2. Differences exist between these results for a variety of reasons, such as the sets of observation wells and timing of measurements, and the methods of evaluating the storage integrals (Dugan et al., 1994, p. 26) used to estimate water-level change and storativity in the High Plains Aquifer (Konikow, 2013). Differences also exist due to the time range of observations, which were limited to the recovery period in our study, whereas Dugan et al. (1994, p. 23) studied consistency across year-to-year of fall measurements in the Northern High Plains. And so, the rest of the results in Table 3 compare the changes in storage by subtracting our results from our predevelopment storage and other studies change from their predevelopment storage. Our model results are listed for changes both since 1930 and since the median measurement year when compared to studies that utilized the newer predevelopment storage estimates, to aid in comparison of results obtained using a changing set of recent observation wells (Steward et al., 2013).

The robustness of our methods is illustrated by the good match found between our results and the retrospective studies across the states and the High Plains Aquifer in Table 3. While methodology is founded in a previous study of western Kansas by Steward et al. (2013), adaptation of those methods to become useful across the High Plains resulted in adding wells to the Ogallala portion of Kansas where declining groundwater levels occur. Wells with increasing water levels are now included, such as the Nebraska well hydrograph in Fig. 2. Such wells with increasing and near-steady levels exist in the Equus Beds and Great Bend Prairie Aquifers of central Kansas, which are managed for sustainable development, and the regions of shallow depth to water along the edges of the High Plains Aquifer Stanton et al., 2011, Fig. 9), all of which were outside the Ogallala study region in Steward et al. (2013). The depletion of the Ogallala is 12, 119, 271 and 341 km^3 by 1960, 2010, 2060 and 2110 (Steward et al., 2013, Fig. 1), while inclusion of the increasing areas gives 13, 98, 190 and 241 km^3 in Table 2. Examination of the differences between forecasts and the reasons for these differences (Lynch, 2002) illustrate that the eastern extents and regions of shallow depth to water of Kansas, where groundwater levels are largely increasing or stable in Fig. 5, contribute towards a persistent, long-term store.

The curves illustrating peak groundwater depletion in Fig. 7 illustrate the annual change of storage for each State, which is calculated using (9). The timing of depletion is identified at intervals of 5% reduction in predevelopment storage, and correspond to the values in Table 2. These curves illustrate the impact of tapping groundwater beyond the rate of recharge. Increasing groundwater extraction occurs through the period of well development until a peak occurs and then extractions from storage tails downward towards the remaining long-term stores of groundwater. The recharge occurring into the High Plains Aquifer was also computed in ArcGIS by integrating recharge from Dugan and Zelt (2000) across each state, and the values are shown in Fig. 7. Peak groundwater depletion in the High Plains is illustrated in Fig. 8 by summing each state's contribution to depletion from storage. Peak groundwater depletion occurred at 2006 for the High Plains, and the annual volume extracted from storage in 2110 following existing trends is forecast to be approximately 50% of the peak rate. The volume

Table 3
The groundwater storage and change in storage obtained using the logistic function approximation is compared with previous retrospective studies in units of km^3 (10^9 m^3).

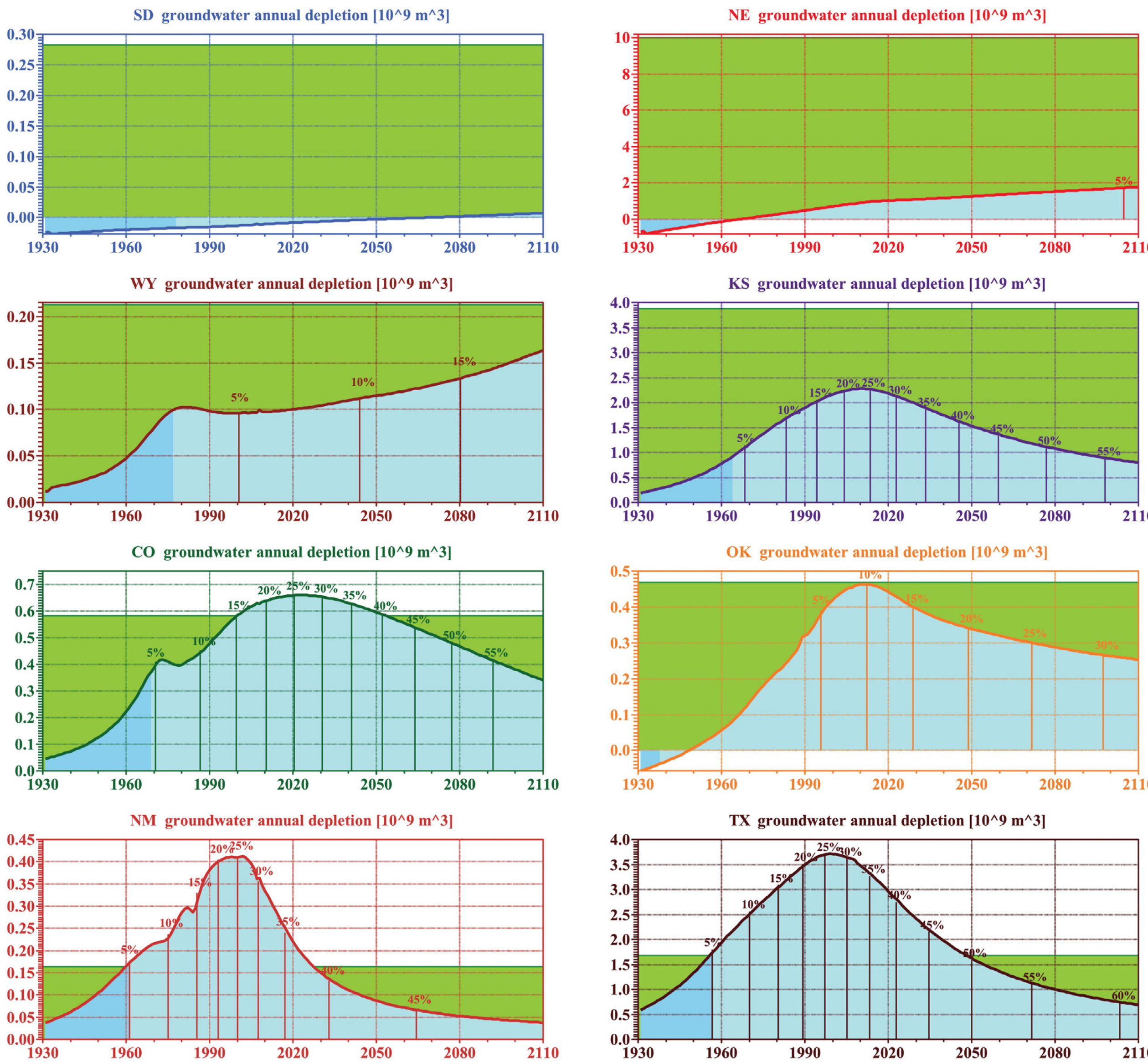
CO	KS	NE	NM	OK	SD	TX	WY	High Plains	study
Storage in predevelopment before pumping									
133	420	2599	58	145	71	578	89	4093	Our study (1930)
155	430	2627	73	146	74	622	86	4213	Gutentag et al. (1984, Tables 8, 11)
123	383	2612	51	143	72	493	86	3963	Our study (1980)
148	395	2627	62	136	74	481	86	4009	Gutentag et al. (1984, Table 8) and Dugan et al. (1994, Table 1)
Storage in predevelopment before observation wells (median 1957)									
(1969)	(1964)	(1952)	(1961)	(1938)	(1978)	(1957)	(1977)	(1957)	(year) (McGuire et al., 2003, Table 1)
127	403	2612	55	146	72	549	87	4051	Our study (median 1957)
117	396	2464	57	144	73	587	75	3913	Stanton et al. (2011, p. 58)
Change in storage, predevelopment to 1980									
10	37	−13	7	3	−1	86	2	131	Our study (1930–1980)
7	36	0	11	10	0	141	0	205	Gutentag et al. (1984, Table 11) and Dugan et al. (1994, Table 5)
7	36	0	12	10	0	140	0	205	Luckey et al. (1981, Sheet 1) with Gutentag et al. (1984, Table 6)
Change in storage, predevelopment to 1992									
16	58	−8	11	6	−1	126	4	212	Our study (1930–1992)
13	63	2	13	12	0	151	−1	253	Dugan et al. (1994, Table 5)
Change in storage, predevelopment to 1999									
20	73	−4	14	9	−1	151	4	266	Our study (1930–1999)
14	56	9	11	9	0	122	2	223	Our study (median 1957–1999)
19	71	−20	16	17	−1	186	2	290	McGuire (2001, Table 2) with Gutentag et al. (1984, Tables 2, 6)
Change in storage, 1980 to 1999									
9	36	9	7	6	0	66	2	135	Our study (1980–1999)
11	33	−20	6	5	−1	39	2	75	McGuire (2001, Table 2) with Gutentag et al. (1984, Tables 2, 6)
Change in storage, predevelopment to 2000									
20	75	−3	14	9	−1	155	4	273	Our study (1930–2000)
14	58	9	11	9	0	126	2	229	Our study (median 1957–2000)
16	71	−5	16	14	0	151	0	263	McGuire et al. (2003, Table 4) with Gutentag et al. (1984, Tables 2, 6)
14	58	−5	10	14	0	153	0	244	McGuire et al. (2003, Table 5)
14	60	−1	11	14	0	160	0	259	Konikow (2013, Table 1)
Change in storage, predevelopment to 2007									
25	90	2	17	12	−1	181	5	331	Our study (1930–2007)
18	74	15	14	12	0	151	3	287	Our study (median 1957–2007)
21	78	26	12	15	1	173	3	329	Stanton et al. (2011, p. 58)
Change in storage, 1900 to 2008									
24	80	20	14	16	1	182	3	341	Konikow (2013, Table 1)
Change in storage, predevelopment to 2009									
26	95	4	18	13	−1	188	5	348	Our study (1930–2009)
20	78	16	15	13	0	158	3	303	Our study (median 1957–2007)
24	80	20	14	16	1	178	3	336	McGuire (2011, Table 3)

of groundwater in storage is also presented. While large shares of groundwater have already been depleted in Colorado, Kansas, New Mexico and Texas; large stores still remain, particularly in Nebraska. The third set of curves show annual decline in groundwater stores plus average recharge. While groundwater pumping have significantly tapped stores in many localities, the available recharge provides future supplies.

4. Discussion: peak groundwater depletion

Groundwater depletion caused by overtapping an aquifer beyond the rate of available recharge is occurring throughout much of the High Plains Aquifer region. The timing of peak groundwater depletion in Fig. 7 follows a south-to-north trend with peak depletion occurring earlier in the southern states, where "parts of it could be depleted within 30 years" (Richey et al., 2015). This process depends on well development and extraction rates, and recharge events. A rising groundwater table is observed in the peak groundwater curves of Nebraska, Oklahoma and South Dakota during recovery after the Dust Bowl of the 1930s and, in Nebraska, due to enhanced recharge from ditch irrigation where surface water became transported to the High Plains Aquifer (Gutentag et al., 1984, p. 45). While pumping groundwater in the High Plains Aquifer began in late 1800s (Opie, 2000), it wasn't until the 1930s that development began in New Mexico and Texas within the southern basin (Luckey et al., 1986, p. 10). The start of the development period moved northward: to Colorado, Kansas and Oklahoma in the central basin in the 1940s and 1950s (Luckey et al., 1986, p. 9, 19), to Nebraska and Wyoming in the northern basin in the 1950s and 1960s (Luckey et al., 1986, p. 33). Significant development of South Dakota had not yet occurred by 1980 (Luckey et al., 1981). As well development and extraction rates increased, declines in groundwater elevation became evident by 1940 in the southern basin, by 1950 in the central basin, and by 1960 in the northern basin, although the water table in areas unsuitable for extensive irrigation remained stable (Dugan et al., 1994, p. 19).

The peak groundwater curves of the High Plains Aquifer follow a functional form similar to the Hubbert curve used to study peak oil (Hubbert, 1956), with increasing production and growth during the development phase until a resource-limited peak occurs followed by a long trend of decreasing production (Bentley and Boyle, 2008). The occurrence of peak groundwater depletion is quantified using the logistic function in (1) and (2) to project groundwater elevation at wells. The logistic function was used by Hubbert (1956) to fit data in the study of peak resource depletion of coal, oil, gas and uranium resources. It was also suggested by Gleick and Palaniappan (2010) for use in studying the depletion of renewable and non-renewable



state	CO	KS	NE	NM	OK	SD	TX	WY	High Plains
peak groundwater year	2023	2010	–	2002	2012	–	1999	–	2006
recharge $[10^9 m^3/yr]$	0.58	3.88	10.01	0.16	0.47	0.28	1.68	0.21	17.27

Fig. 7. Peak groundwater depletion is illustrated in plots of annual change in groundwater storage over time for each State. The change in storage is shown with the blue shaded area represents the depleted or filled volume (darker blue before median year from McGuire et al. (2003) in Table 3), and the fraction of depletion in storage since 1930 is shown at 5% intervals. Mean annual recharge to the High Plains Aquifer is shown within the green shaded area.

water supplies. Other similar functional forms have been proposed (Bartlett, 1999), for example to account for the occurrence of multiple peaks and asymmetry occurring in some production profiles (Brandt, 2010). And yet the results from Hubbert's analysis were eventually confirmed by the National Academy of Sciences (http://www.hubbertpeak.com/hubbert/tribute.htm). The Hubbert curve provided evidence of the depletion of conventional oil and and promoted attention towards mitigation of the inevitable consequences of depletion (Brandt, 2007).

Hubbert applied a single equation equal to the derivative of the logistic equation across a petroleum reservoir. To account for spatial asymmetry in the timing of depletion rate across an aquifer, we integrate the difference of the logistic equation using (9) at the set of observation wells across the “geological characteristics and exploration history of the region” (Sorrell and Speirs, 2010, p. 227). While the peak groundwater curves do not fit the idealized Hubbert form for the entire aquifer, they reflect differences in the timing of depletion occurring across the High Plains Aquifer. These differences lead to asymmetry in peak groundwater depletion curves, which also occurs in some of the world's peak oil curves (Brandt, 2007).

Extensions to the mathematical form of the logistic equation could be envisioned to account for a more detailed analysis of the depletion process at a well. For example, Steward et al. (2009b, Eq. (29)) developed a closed-form solution for dewatering or filling a sloping aquifer (our problem) as a logistic equation, which was applied to aquifer tapping by Bulatewicz et al. (2014). In general, this depletion process may be influenced by the rate of change at the point of water use ($\mathcal{H}_0$ in Eqs. (17a) and (23)) by the different aquifer responses occurring upgradient and downgradient

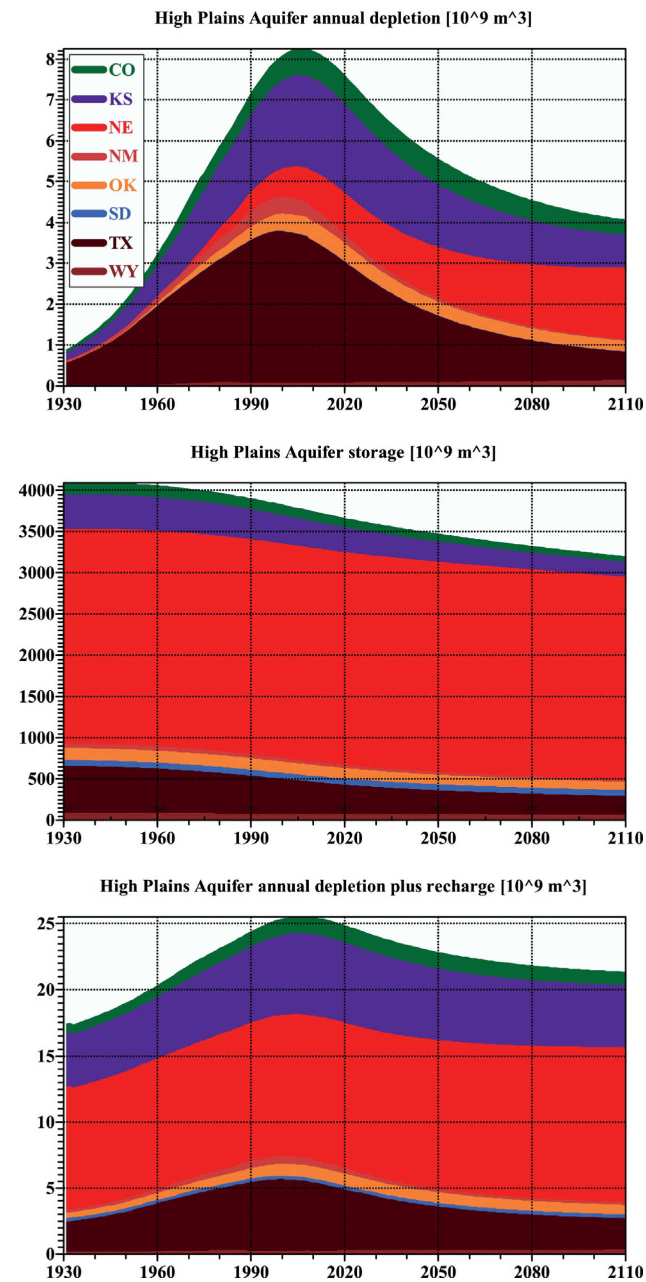


Fig. 8. Peak groundwater depletion for the High Plains Aquifer illustrates the aggregate annual tapping of groundwater from storage.

from a point of extraction ($\mathcal{H}^{\infty}$ in denominator on left side of Eq. (28)) and by the dispersion of an advancing depletion zone about the front of the kinematic wave (exponential term in numerator of right side of eq.28); all equations and variables are from Steward et al. (2009b). Furthermore, the depletion process is influenced by the convolution of water-use extractions and recharge at different locations and times Steward et al. (2009b, Eq. (40)). Differences in aquifer properties (e.g., changes in specific yield with depth) pose additional complexities.

In spite of the complexity of the process of aquifer depletion, the logistic function provides meaningful results. It is clear from Fig. 3 that the observed measurements of groundwater depletion span the range of a full to empty aquifer. This progression from full to empty has already occurred at some locations within Figs. 5 and 6. Across these historical measurements, the errors associated with approximating groundwater elevation with the logistic function are small in Table 1. Spatially integrating groundwater depletion using the logistic function estimates across these wells also matches previous studies very well in Table 3.

Future projections in the peak groundwater curves reflect differences between tapping groundwater and petroleum. Hubbert (1973) proposed a set of future possibilities after the period of exponential growth that could stabilize at a maximum or sustainable level, or decline to zero. Peak groundwater curves are limited by the recoverable groundwater available in storage, where the Ultimately Recoverable Reserves (URR) is equal to the integral of the depletion rate over all time (Hubbert, 1956). A major difference between fossil fuels and groundwater is that aquifer sources are renewable, although replenishment rates may be very slow (Steward et al., 2013), and groundwater may continue to be pumped at or below the rate of recharge when tapping groundwater from storage is not feasible. Thus, pumping may approach a long-term limit equal to recharge (Gleick and Palaniappan, 2010). The URR in petroleum has evolved over time as technologies such as hydraulic fracturing adapt to enable larger extraction rates. The peak groundwater depletion curves would be influenced by adoption of future strategies to tap new groundwater supplies as well as those technologies that would augment existing stores, such as the proposed Kansas aqueduct project.

The occurrence of peak groundwater depletion is useful in assessing the groundwater stores available into the future to support agricultural production. Long-term supplies will eventually transition to at most the available portion of the recharge rate, shown in Fig. 8 from Dugan and Zelt (2000), that are not captured by the ecological needs of the region (Gleick and Palaniappan, 2010). Estimates of peak groundwater depletion are influenced by the spatial extent of a study region as depleting regions become summed with regions with steady or increasing water tables. For example, the results from Steward et al. (2013) in the western Ogallala portion of Kansas illustrate a more rapid depletion process than when the south central portion of Kansas is added in Fig. 7 where larger recharge rates provide storage with a longer useful lifetime (Sophocleous, 2012). Aggregation of the peak curves for each State in Fig. 7 cumulatively provide the peak for the High Plains Aquifer in Fig. 8 reflects dewatering across this larger region. Clearly, other factors will influence the needs for future groundwater extractions, such as changing climatic conditions (Scanlon et al., 2012), adoption of optimal irrigation allocations (Hassan-Esfahani et al., 2015), and future water-use efficiencies in crops (Steward et al., 2013). This study projects the peak groundwater depletion occurring due to existing trends in water-use.

5. Conclusions

The concept of peak groundwater depletion is introduced and quantified for the High Plains Aquifer. Methods are presented to fit the hydrographs (Fig. 2) of a set of observation wells (Fig. 1) to the logistic equation (1). This provides estimates of saturated thickness (Fig. 5) and the average annual rate of depletion (Fig. 6) that were integrated over the aquifer to project existing trends back to predevelopment 1930 conditions and forward into the future (Table 2). Spatial integration provide estimates of the volume of groundwater depletion at the state level in Fig. 7 and at the High Plains Aquifer level in Fig. 8. These patterns reveal the occurrences of peak groundwater depletion as the rate of tapping aquifer stores increases towards the peak, and then decline as the limits become reached in the availability of extractable groundwater.

Peak groundwater depletion has already occurred for the High Plains (Fig. 8). The depletion process is spatially variable with peak groundwater for each State projected to occur in 2023 for Colorado, 2010 for Kansas, 2002 for New Mexico, 2012 for Oklahoma and 1999 for Texas. Peaks have not yet occurred for Nebraska, South Dakota and Wyoming (Fig. 7). Analysis includes wells with increasing or near-steady water tables that contribute to these longer-term supplies, and groundwater will continue to exist at some locations for the foreseeable future, even in states that have already experienced large declines in storage.

The groundwater resources of the High Plains Aquifer are vulnerable, and society has the opportunity now to better understand the tapping processes and to plan for a more resilient future. Lessons in natural resources management over the past several decades since peak oil was first forecast reveal a pattern of early public attention that waned over decades as global peaking became "an epochal non-event" (Bardi, 2009). Recently, attention has become refocused on the eminently dwindling global supply (Bentley and Boyle, 2008) and efforts are focusing on transition towards inevitable substitutes (Brandt, 2010). The energy sector is learning to utilize the remaining resources to put long-term solutions into place, rather than fighting for the remaining resources. Similarly, society has time now to develop and implement strategies to deal with future groundwater shortages, and to plan for pumping of groundwater in regions that currently are not experiencing depletion. Water resources, unlike many important commodities, do not have a substitute (Postel et al., 1996). Once it has been consumed, it will be gone for the foreseeable future in many areas. Managing the risk of vulnerable groundwater supplies must balance the needs of today with future possibilities to enhance the resiliency of this most important resource.

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Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at http://dx.doi.org/10.1016/j.agwat.2015.10.003.

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