GROUNDWATER STORAGE CHANGE IN NEW MEXICO AQUIFERS
PART 1. Method for Estimating Groundwater Storage Change in Variably Confined Aquifers in New Mexico
PART 2. Estimates for Groundwater Storage Change in the New Mexico Southern High Plains Aquifer

Compiled and edited by Alex J. Rinehart and Ethan Mamer

New Mexico Bureau of Geology and Mineral Resources, New Mexico Institute of Mining and Technology,
801 Leroy Place, Socorro, NM 87801. Email: alex.rinehart@nmt.edu, ethan.mamer@nmt.edu

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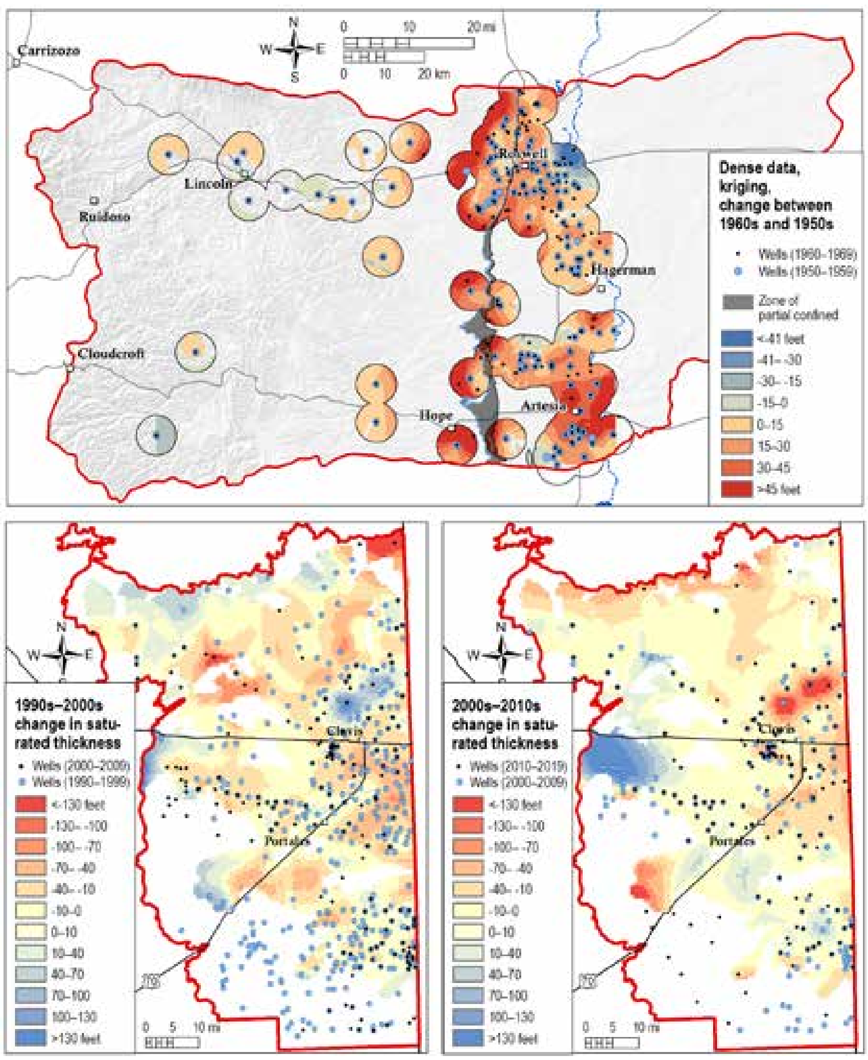

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New Mexico Bureau of Geology and Mineral Resources

A division of New Mexico Institute of Mining and Technology

Socorro, NM 87801
(575) 835 5490
Fax (575) 835 6333
geoinfo.nmt.edu

ABSTRACT

This work combined method development for systematically estimating groundwater storage changes in variably confined and stacked New Mexican aquifer systems, and groundwater storage change analysis in the Quay-Roosevelt-Curry County (QRC) and Lea County areas of the Southern High Plains (SHP). For both studies, the primary limitation on reliable estimates was having enough consistently located water level measurements through time. To develop and test the method for estimating storage changes in variably confined aquifers, part of the Pecos Slope and Roswell Artesian Basin, and the Jurassic aquifers of the San Juan basin were used. The Pecos Slope and Roswell Artesian Basin contained two aquifer systems: a variably confined Permian aquifer system; and an overlying shallow unconfined alluvial aquifer system. In order to estimate storage changes in variably confined aquifers, the aquifer boundaries had to be identified, generally through well log interpretation or literature review, and wells had to be correctly located and identified as being completed in an aquifer of interest. During the analysis, the boundary between confined and unconfined parts of the aquifer had to be tracked. This method appeared to reasonably reflect changing water management practices in the region, though the reliability of the estimates was limited by the changing well network. The Jurassic aquifer system of the San Juan Basin shows the lower bound of well density valid for the method. Due to sparse data, these estimates were problematic and only showed changes from the 1970s to the 1980s.

In the SHP aquifers, we improved on our earlier method to incorporate hydraulic and geologic information from other studies. We fitted the water elevation with a two dimensional polynomial, and then interpolated a surface of the residuals of the water elevation using ordinary kriging. The results of the study showed that the QRC area had dwindling zones of reliable saturated thickness and had lost at least 8 Maf of groundwater storage since the 1950s. The estimates in the Lea County area were more problematic because of changing well networks. However, while there had been substantial decreases in saturated thickness, the overall withdrawals were less than in the QRC area and the starting saturated thickness was greater. The aquifer-wide storage change rate was close to the withdrawal estimates of the Office of the State Engineer in the Lea County. In summary, we had extended and improved our previous method to find groundwater storage changes in variably confined and stacked aquifers, and in areas with extensive hydrogeologic data. We were still limited in our work by the density and extent of long-term groundwater level measurements.

Keywords: New Mexico, confined/unconfined aquifers, groundwater storage change, regional hydrology, Pecos Slope, Roswell Artesian Basin, San Juan Basin, Southern High Plains, Ogallala Formation

Unit Conversions

1 km^3 = 810715 acre-feet = 810.715 kaf = 0.810715 Maf

1 km = 0.621371 miles

1 m = 3.28084 ft = 39.3701 in

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SUMMARY

Groundwater is a primary source of freshwater in New Mexico and dominates surface water use in many places across the state (Longworth et al., 2013). This is particularly true outside of the Rio Grande valley (Longworth et al., 2013). This makes understanding changes in groundwater storage vital to the state—we need to know how much water is leaving storage and, where possible, how much is left. The current study extended the work done in previous years on unconfined basin-fill aquifers across the state.

This year, we had two prongs of work. First, we adapted the methods of Rinehart et al. (2015 and 2016) to estimating groundwater storage change in variably confined and multilevel (i.e., stacked) aquifer systems (Part I). To develop and test the method, we focused on the variably confined Permian and unconfined alluvial aquifers of part of the Pecos Slope and Roswell Artesian Basin, and on the Jurassic aquifer system in the San Juan Basin (Fig. 1). Second, we adapted the method of Rinehart et al. (2015 and 2016) to estimate changes in water level and saturated thickness, and changes in overall groundwater storage, in the Southern High Plains (SHP) aquifer in central-east and southeastern New Mexico (Fig. 1; Part II). The primary aquifer in this region is in the Ogallala Formation. The SHP region was broken into a southern region of the Ogallala Formation in Lea County, and a region including the Ogallala Formation in Quay, Roosevelt and Curry Counties.

In Part I, we developed a method for estimating storage changes in variably confined aquifers. In these systems, the aquifer transitions from a confined, or pressurized, aquifer to an unconfined aquifer. When a aquifer was confined, storage changes occured due to elastic deformation of the water and mineral framework (Schwartz and Zhang, 2003). An increase in pressure forced the pores open and squeezed the water molecules together; a decrease in pressure did the opposite (Schwartz and Zhang, 2003). This means that (a) the entire thickness of the aquifer was involved in the storage change, not just the volume near the top of the aquifer, and (b) for confined aquifers of thicknesses seen in New Mexico, the volume of water released per unit drop in head (the storativity) was 10 to 1000 times less than in unconfined aquifers (Schwartz and Zhang, 2003). In unconfined aquifers, storage changes occured by draining or filling pore space. Here, the storage coefficient was near to the effective porosity (Schwartz and Zhang, 2003).

In large part, the estimates in variably confined aquifers were made using the same method used in Rinehart et al. (2015 and 2016): data review and quality assurance using U.S. Geological Survey and other trusted sources of data; finding median decadal water levels for each well; interpolating the water levels using inverse distance weighting and ordinary kriging methods; finding the differences in the water levels from decade to decade or as compared against a base decade; and using an appropriate storage coefficients to calculate the total storage changes in the region. The major additions to the process were accurately determining the top elevation of the aquifer system throughout the region of interest, keeping track of where the aquifer is confined vs. unconfined, and using the appropriate storage coefficient (storativity vs. specific yield) to calculate the storage change.

Using the correct storage coefficient involved not just mapping where the system was confined vs. unconfined. It also involved calculating the storage change in regions that change from confined to unconfined through time. To do this, a confined storage change was calculated for the change in water level down to the top of unit, and was then added to an unconfined change for water level changes below the top of the aquifer unit.

The Pecos Slope and Roswell Artesian Basin was one of best characterized aquifer systems in New Mexico, with well measurements going back to the early 1900s when the first wells were drilled into the confined Permian aquifer system and overlying unconfined alluvial aquifer system. We choose this system because there was a dense network of water level measurements in time and in space, and there was a well-exposed geologic contact between the confining unit (the Artesia Group) and the Permian

Figure 1. Study regions. Pecos Slope and San Juan Basin are analyzed in Part I, and the Southern High Plains is analyzed in Part II.

aquifer system (from bottom to top, the Yeso Formation, Glorieta Sandstone and San Andres Limestone; Land and Newton, 2007, and Newton et al., 2012). Through a literature review, review of well logs and correlation of logs with cuttings, the top of the San Andres Limestone was mapped out at individual wells. These points were interpolated using a nearest-neighbors algorithm to define the elevation of the aquifer top. Water level measurements were taken from water wells that were completed in the aquifer formations based on the U.S. Geological Survey (USGS) formation code assigned, on reports from the NMBGMR, and on the reports of wells from the Pecos Valley Artesian Conservancy District. All water levels were converted from depth-to-water to elevation to be consistent with the geologic surface interpolation.

We were able to perform a reasonable analysis of groundwater storage changes in the Pecos Slope and Roswell Artesian Basin aquifers. Decade-by-decade differences were reasonable from the 1900s through the 2010s, with an increase in groundwater storage corresponding changes in groundwater management practices. However, there had been significant variation in the well network through time, leading to overestimates of gains, which was partially exaggerated by the high impact of adding or removing wells from the unconfined upland regions of the aquifer.

We attempted to perform this analysis in the San Juan Basin based on the geologic framework of Kelley et al. (2013). After reviewing the water level data in the region, only one pair of decade for one aquifer system was found to have acceptable water level measurement density. This was the 1970s to the 1980s in the Jurassic aquifer system that was made up of the members of the Morrison Formation, the Wanakah Formation, the Cow Springs Formation, the Bluff Sandstone and the Entrada Sandstone (Kelley et al., 2013). Most of the aquifer that was well covered was along the Chaco Slope on the southern margin of the San Juan Basin, or near the Chuska Mountains near the western edge of the San Juan Basin. There was little data in the center of the San Juan Basin. For this 20 year interval, we were able to generate reasonable estimates, but these estimates were complicated by the lack of spatial coverage. Fundamentally, this region showed the lower limit for the well coverage needed to perform our analysis, or, indeed, might be too sparse. This combined with the results of Rinehart et al. (2016) implied that much of the state will not be able to be analyzed in the future. There simply was not enough data.

In Part 2, we performed storage change calculations in the Southern High Plains building from the work of Rawling and Rinehart (in review) and Rinehart et al. (2015 and 2016). Because of the large body of literature in the region and in particular using the results of Hart and McAda (1985), we were able to calculate saturated thicknesses and to use a spatially varying specific yield. We also used a more sophisticated method for estimating the water level surfaces, where we fitted a two-dimensional (northing and easting) third-order polynomial to the estimate mean water elevation, and then interpolated the residual water levels left after detrending with the mean fitted surface. The fitted polynomial plus the residual interpolation surface gave the estimate of a median decade water elevation at all points in the study area. The additional steps also allowed us to better visualize the uncertainty caused by changing well networks through the leave-one-out (LOO) cross-validation of the kriging interpolation. We divided the study area into the QRC area and the Lea County area based on hydrologic boundaries (Hart and McAda, 1985). The QRC area showed large declines since the 1950s, with a total volume of 8 Maf removed compared to a decrease of 3 Maf in the Lea County area. We found that much of the QRC area has either gone dry (water table dropped below the bottom of the aquifer) or has less than 30 ft remaining in saturated thickness. Thirty feet of saturated thickness was roughly what is required for high volume pumps to run (Hecox et al., 2002). The Lea County area showed milder decreases in saturated thickness. The original thickness of the aquifer was greater in Lea County area than in the QRC area (Hart and McAda, 1985), and the groundwater withdrawals reported by the NMOSE have been less in the Lea County area than in the QRC area (Longworth et al., 2008 and 2013). It was important to note that these estimates were significantly lower than predicted by the NMOSE (Longworth et al., 2008 and 2013) or by the USGS (McGuire, 2011). We believed that this is in part because of the areal restrictions of our change estimates as compared to the assumption-rich estimates of the NMOSE and sparse well networks necessarily used by the USGS that were interpolated across the entire region. However, these underestimates were also because of unpredictable changes in well measurement networks, especially in Lea County.

In the end, we feel that we proved that our new method is valid for the variably confined aquifers given enough data, and that we found the well density limit that was needed to perform a reasonable analysis. We also found spatially detailed estimates of groundwater storage change in the Southern High Plains aquifer system that appeared reasonable, especially when viewed as maps of saturated thickness changes.

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PART I.
Storage change in variably confined and stacked aquifer systems: Examples from the Pecos Slope and Roswell Artesian Basin, and the Jurassic aquifers of the San Juan Basin

Ethan Mamer, Alex J. Rinehart, Ron F. Broadhead, Trevor Kludt,
Brigitte Felix, and Cathryn Pokorny

I.1. Introduction

In New Mexico, a dominant source of water for all uses has been groundwater—in 2010, just over 2 Maf of surface water was withdrawn for use and 1.77 Maf of groundwater was withdrawn (Longworth et al., 2013). The vast majority of groundwater withdrawals were for irrigated agriculture (Longworth et al., 2013). The difference was made up in drinking water, industrial use, livestock agriculture, commercial use, mining and power production (Longworth et al., 2013). This means that, for New Mexico, aquifers were vital for the well-being of the state both in terms of economics and in terms of health. This dependence had been exacerbated during times of drought, when surface water supplies ran short and surface water users had to supplement with groundwater.

This study presented an adaptation of the previous method of Rinehart et al. (2015 and 2016) to estimate groundwater storage changes in confined, variably confined, and stacked aquifer systems. The following introduction to hydrogeology was adapted from Schwartz and Zhang (2003). An unconfined aquifer is a layer of saturated rock or other geologic material where the water table is open to the atmosphere. A confined aquifer is a layer of saturated rock where the water table (or piezometric surface) is not open to the atmosphere. Rather, the top of the aquifer is defined by a confining layer, or aquitard, that restricts upward flow. If the confining layer was breached by a well, then water rose through the breach until the pressure head in the aquifer is matched by the hydrostatic pressure of the column of water in the well. The elevation that the water rose to in the well casing was considered the 'potentiometric surface'. If the land surface was below the elevation needed to match pressures, then the well flows freely until the pressure of the aquifer was relieved. This was illustrated in Figure 2a.

To find groundwater storage changes in regions with variably confined and/or stacked aquifers, new steps were required beyond the method of Rinehart et al. (2015 and 2016) to define the aquifer vertical and plan-view extent, identify where it was confined, and whether different layers are in hydrologic communication. The previous work focused entirely on unconfined and alluvial or basin-fill aquifers (Rinehart et al., 2015 and 2016). The previous work did this for two reasons: (1) most of the population and industry in New Mexico used groundwater from unconfined basin-fill aquifers (Longworth et al., 2013); and (2) hydrogeologically, storage changes in unconfined basin-fill aquifers could be estimated directly from calculating water level change multiplied by the specific yield (Schwartz and Zhang, 2003). Variably confined aquifers could have up to four orders of magnitude variations in storage coefficients (specific storage vs. specific yield) depending if the aquifer was confined or unconfined (Scwartz and Zhang, 2003). Bedrock aquifers, such as limestone aquifers, were often also variable in the storage coefficients as a function of fracture density or cementation, causing greater spatial variability than that seen in basin-fill aquifers (Schwartz and Zhang, 2003). Finally, stacked aquifer systems were often capped by an alluvial unconfined aquifer, with aquitards, or low hydraulic conductivity layers interbedded between aquifer units. These patterns and challenges were summarized conceptually in Figure 2.

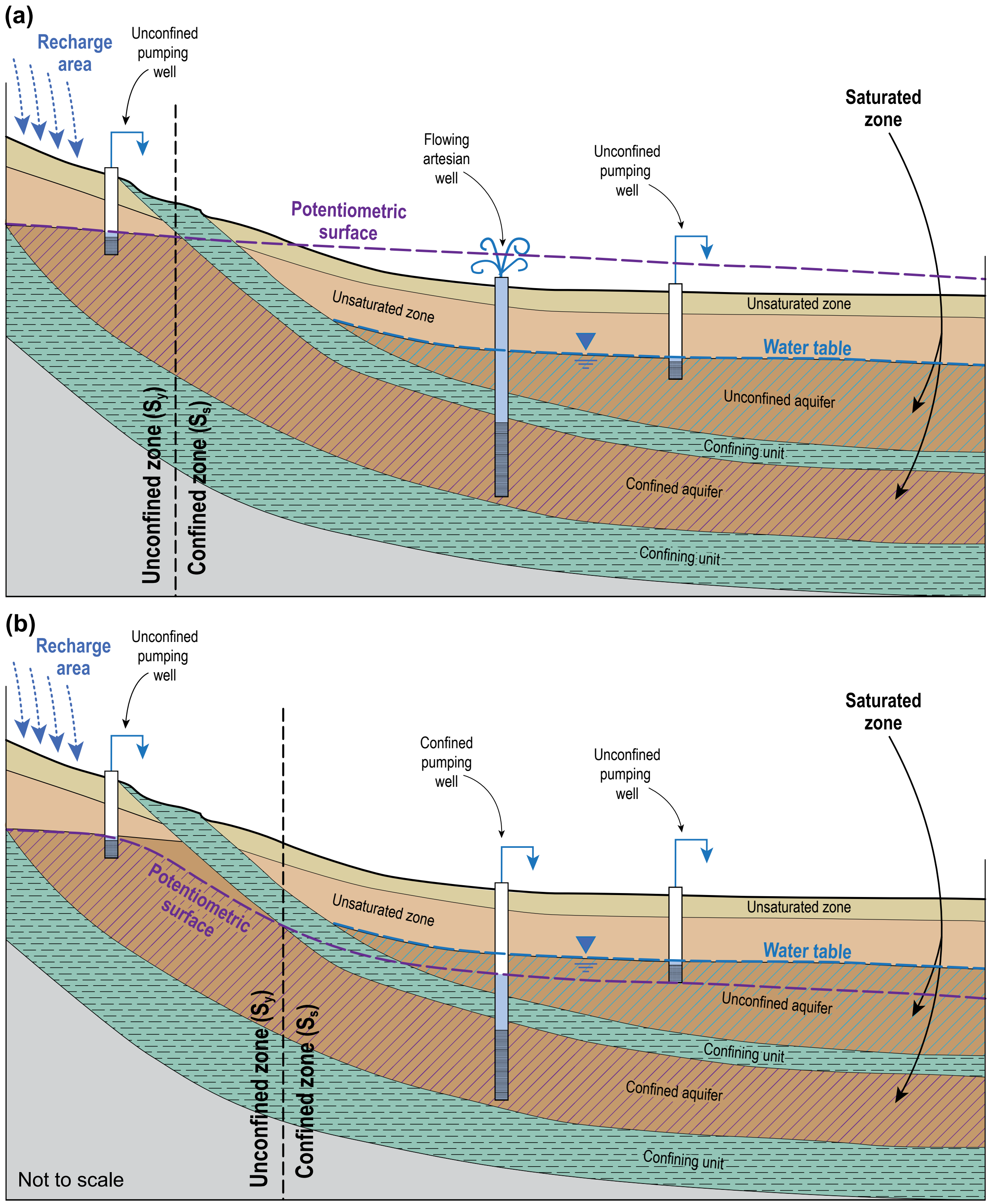


Figure 2. Conceptual diagrams demonstrating major concepts of confined and unconfined aquifers. (a) Cross-section of a rechargeable, variably confined, artesian aquifer system. Recharge enters the aquifer at higher elevations, before the aquifer dips under land surface and is confined by impermeable units. (b) The effects over pumping can have on an artesian system. As the potentiometric surface declines with pumping the confined portion of the aquifer shrinks. Where the aquifer is still confined, the potentiometric surface is now below land surface.

I.2. Background

I.2.1. Hydraulic background

While the majority of the method described in this report focuses on quantifying changes in water level, volumetric storage changes were defined by changes in water level and an appropriate storage coefficient that converted a change in head to a change in volume. The volume of water (V) removed from the ground could be calculated by multiplying the change in hydraulic head (Δh) per unit area (A) of the aquifer, by the appropriate storage coefficient (S) (Schwartz and Zang, 2003) (Equation 1).

$$V = \Delta h * A * S \tag{1}$$

There were two storage coefficients, specific storage and specific yield, that applied, respectively, to confined or unconfined aquifers (Fig. 2). The following description of the properties of confined and unconfined aquifers was summarized from Schwartz and Zhang (2003) and Batu (1998). A confined aquifer was one that was separated from atmospheric pressure by a relatively impermeable confining unit. Water would rise above confining unit when a confined aquifer was penetrated by a well. The elevation that the water rose to in the well casing was considered the 'potentiometric surface.' When a confined aquifer was pumped there was no dewatering of the saturated zone; instead, the aquifer was depressurized. The specific storage coefficient for the aquifer was used to determine the volume of water removed per unit head decline in a confined aquifer per unit area per unit thickness of the aquifer. The ability of the confined aquifer to support the load of rock above the aquifer was reduced by an amount proportional to the reduction in artesian pressure; as a result, the mineral grains and water in the aquifer was expanded, opening pore space and increasing the stored volume of water throughout the aquifer layer. Similarly, when the pressure was relieved, (i.e., the aquifer was pumped or the head was otherwise lowered), the framework grains in the aquifer were under less pressure and expand, thereby lowering the stored volume of water even though the aquifer was still saturated (Batu, 1998). Because the specific storage was related to the compressibility of minerals and water, as well as an effective framework-fluid coupling factor, it generally had values between 10^{-6} and 10^{-4}/m. However, pressure changes within a confined aquifer occured throughout the layer, so that the entire thickness of the aquifer changes in storage. To find the volume of water per change in pressure head per unit area, the specific storage was multiplied by the thickness of the aquifer to find the storativity. In New Mexico, most confining aquifers were relatively thin (10s to 100s of feet), yielding storativity values of between 10^{-2} and 10^{-5}.

When the piezometric surface, or water table, of an unconfined aquifer was lowered, the porosity was drained. Once again, an unconfined aquifer was one where the water table was open to atmospheric pressure. The conversion between the change in water level to the change in volume of water per unit area was called the specific yield. It was roughly the drainable porosity, with general values between 0.05 and 0.3 (Scwartz and Zhang, 2003).

The boundary between the confined and unconfined portions of the aquifer was defined as where the potentiometric surface intersected the bottom of the confining layer. As a confined aquifer was pumped the potentiometric surface would lower. As a result, the boundary between the unconfined and confined zone, where the potentiometric surface met the confining unit, moved down gradient (Fig. 2b) (Schwartz and Zang, 2003), and part of the formerly confined aquifer became unconfined. This corresponded to a dramatic change in storage coefficients as storage in that part of the aquifer transitions from being characterized by specific storage to specific yield coefficients (Schwartz and Zang, 2003).

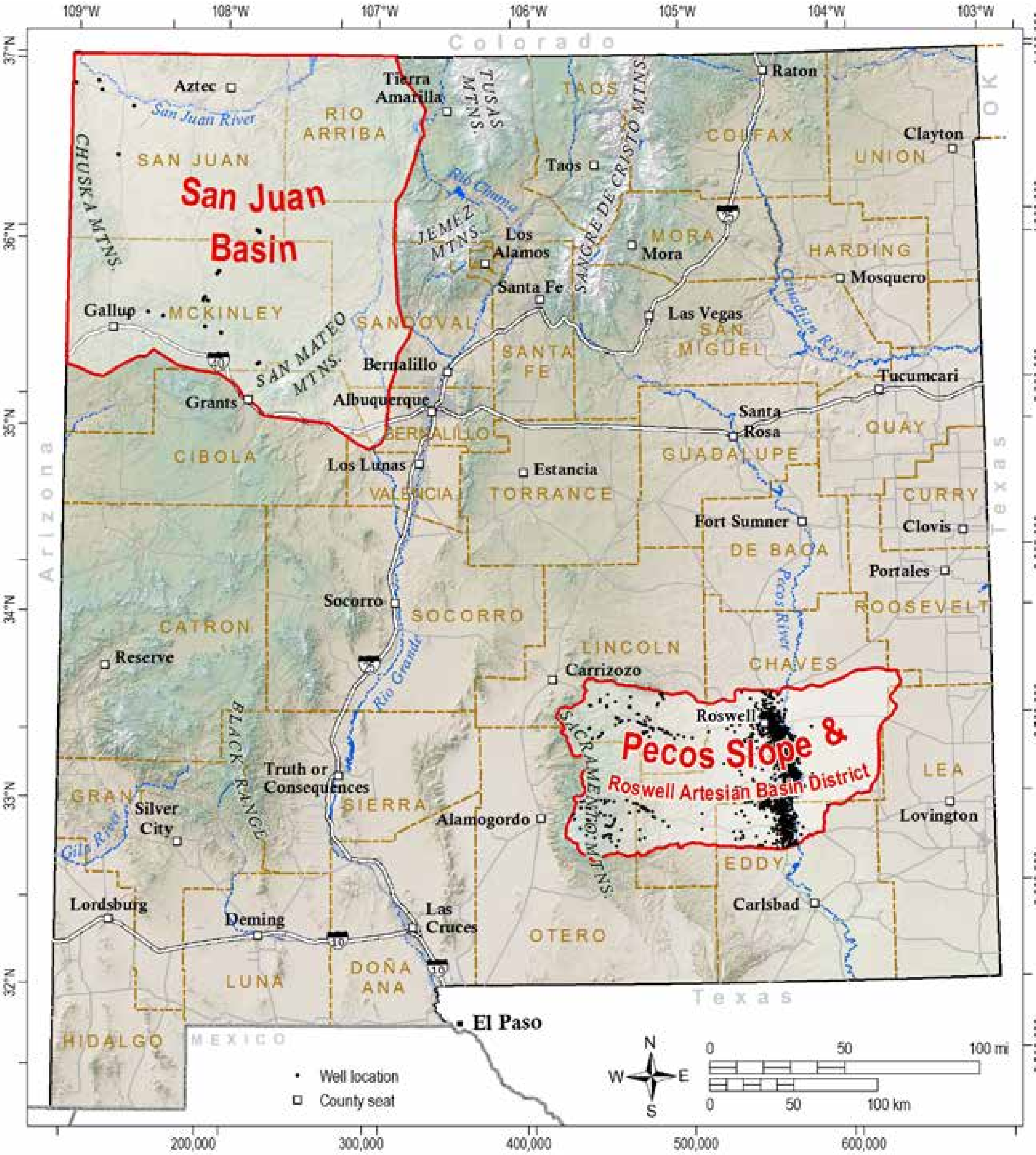


Figure 3. Regions in New Mexico where method for storage change calculations in variably confined or stacked aquifers was developed and tested. Dots represent available well sites used in analysis.

I.2.2. Study areas

In the previously completed reports that focused on the alluvial aquifers, study areas were separated by watershed boundaries as defined by the USGS HUC-8 delineation. This classification was prudent as alluvial aquifers were typically closely linked to surface water (Seaber et al., 1987). Using individual watersheds to delineate a study area is not as applicable when working with confined aquifers. For this project, the study area was instead defined by known variably confined aquifers and constrained by geologic controls.

Pecos Slope and Roswell Artesian Basin (PSRAB) region

The Pecos Slope (Fig. 3) is a geologically defined structural ramp to the west of Pecos River (Kelley, 1971). Its western extent and highest elevations are defined by the Mescalero arch in the north, the crest of the Sacramento Mountains (focus of this study), and the crest of Guadalupe Mountains in the south. From this high, the ramp descends to the east into the Delaware and Permian basins. The primary fresh-water-bearing rocks in this region are the Permian Yeso Formation, San Andres Formation and Glorieta Sandstone, which are capped by the impermeable Artesia Group along the lower part of the Pecos Slope. The Pecos Slope eventually runs under the Pecos River and eastward into the Delaware and Permian depositional basins.

We informally defined the Roswell Artesian Basin as the region in the Pecos valley that had artesian freshwater aquifers in the Permian strata mentioned above. Geologically, this is still part of the formal Pecos Slope; however, it had been historically managed separately and had distinct surface water-groundwater management issues.

We defined the our region of interest (the PSRAB region) as the part of the Pecos Slope stretching eastward from the crest of the Sacramento Mountains down slope to the lower Pecos Valley, which included the Roswell Artesian Basin. The hydrologic framework consisted of three parts: an unconfined Permian-aged aquifer along the eastern flank of the Sacramento Mountains, a confined Permian aquifer along the lower flank of the Sacramento Mountains and underlying the Pecos Valley (the Roswell Artesian Basin), and an unconfined alluvial and basin-fill aquifer overlying the confined Permian aquifer. The confined Permian aquifer was the most well-known confined aquifer in New Mexico (Land and Newton, 2007). The study area selected for this analysis consisted of a combination of four USGS HUC-8 basins in southeastern New Mexico; Upper Pecos-Long Arroyo, Rio Hondo, Rio Felix, and Rio Penasco (6,454 sq mi). The boundary of the study area to the west ran north-south along the crest of the Sacramento Mountains. The northern boundary of the study area started at Carrizo Peak in the northwest, ran east along the Capitan Mountains to the eastern boundary, roughly 50 miles east of Roswell. The southern boundary runs east-west, roughly 10 miles south of Artesia, extending west to where it reaches the ridge of the Sacramento Mountains. The south-flowing Pecos River is the primary surface water feature in the area. Several other smaller rivers flowed eastward from the mountains (Fig. 4a). The upper reaches of these rivers flow year-round. However, natural recharge to the unconfined aquifers, and irrigation along eastern reaches of the rivers siphoned off much of the natural flow and these rivers were dry by the time they reach the lower basin except after storms and during spring runoff. Elevation ranges from 11,981 ft in the mountains to the west, down to 3,300 ft on the valley floor along the Pecos River.

The primary population centers within the study area included Ruidoso, Roswell, and Artesia. The Roswell Artesian Basin in the Pecos Valley was one of the most intensively farmed areas in the state. The principal crops were alfalfa, cotton, sorghum, chiles and pecans (Land and Newton, 2007). The basin derived the majority of its irrigation water from groundwater stored in a shallow alluvial aquifer and from the underlying confined Permian aquifer. Because of the long history of agriculture and settlement in the region, this region had one of the best water level measurement networks in the state. The

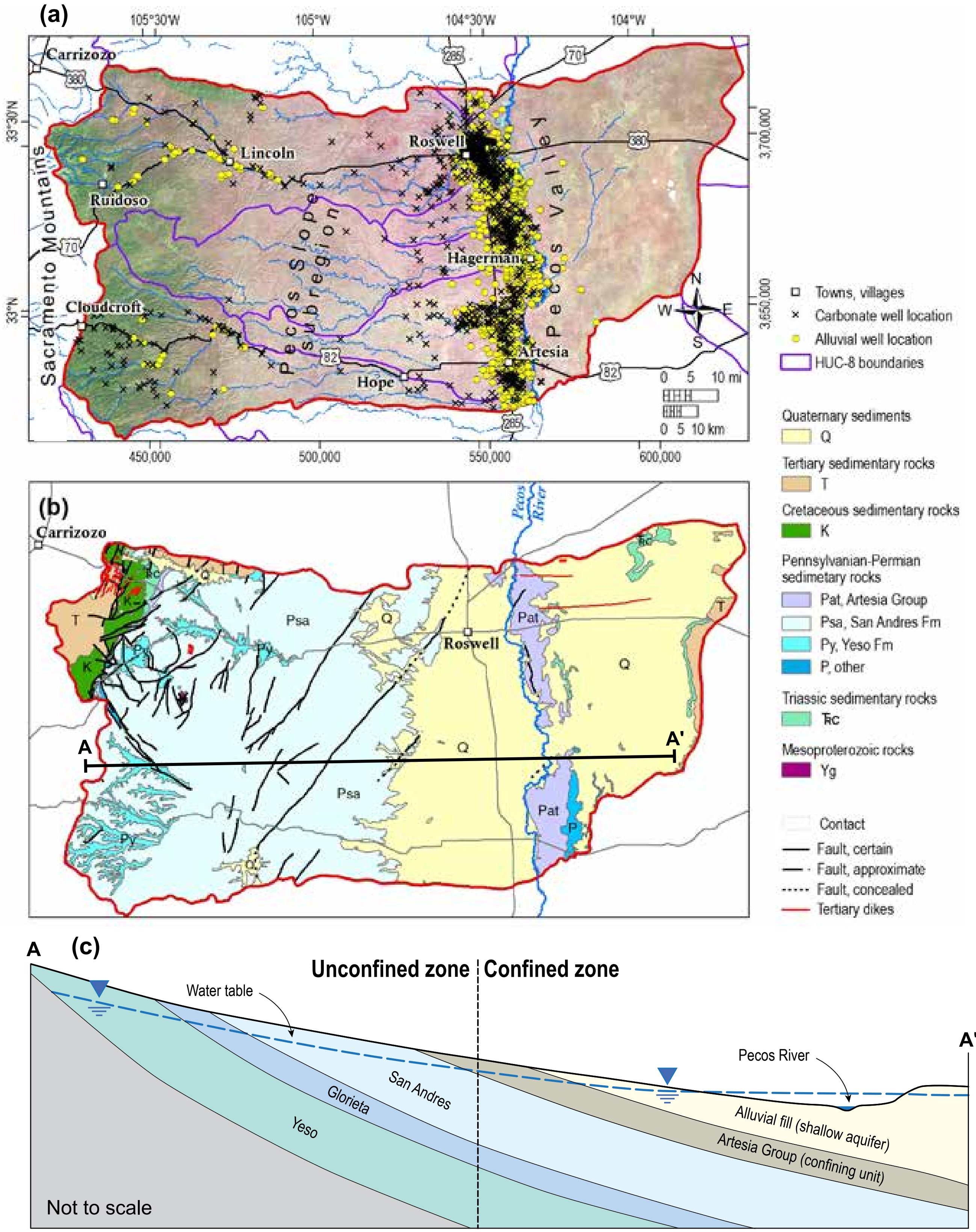


Figure 4. (a) Outline of the Pecos Slope and Roswell Artesian study area with HUC boundaries in blue, and well network used for analysis. (b) Geologic map (New Mexico Bureau of Geology and Mineral Resources, 2003). Units are grouped by geologic age, except for the Permian units that make up the aquifers studied here. (c) Conceptual geologic cross-section of the unit, going east to west (not to scale) depicting the shallow alluvial perched aquifer, as well as the partially confined Permian aquifer.

measurements were spatially extensive, repeated through time, and frequent in time. This system formed our best-case trial.

The primary geologic units found in the study area that form aquifers, from oldest to youngest, were the Permian Yeso Formation, the Glorieta Sandstone, the San Andres Formation, the Artesia Group, and the Quaternary alluvium and Gatuna Formation (Land and Newton, 2007). The Permian units were tilted, dipping roughly 1 degree to the east. There was very poor exposure of the Yeso Formation, as it was usually covered by the more resistant San Andres Formation and, where the Yeso Formation was exposed, it was prone to dissolution and weathering. The San Andres was exposed throughout much of the Sacramento Mountains, and across the Pecos Slope where it was overlain by the Artesia Group. The valley floor ass covered by the alluvium and older basin fill (Fig. 4b) (Land and Newton, 2007).

The Yeso Formation (~2,000 ft thick) was a heterogeneous formation composed of discontinuous limestone, siltstone, and sandstone, dipping gently to the east. Abundant fractures and karst features made it a productive aquifer. The Glorieta Sandstone overlies the Yeso Formation and was a relatively thin sandstone layer. Above the Glorieta Sandstone, the San Andres Formation (~1,000 ft thick) was made up of light-to-dark-gray and bluish-gray carbonate rocks (Newton et al., 2012). This group of carbonate and siliciclastic units made up the confined aquifer system. Above the San Andres was the low permeability Artesia Group, largely composed of siliciclastic and evaporite deposits. Inset against the Artesia Group, the silty-to-gravelly Gatuna Formation and younger alluvium (~350 ft thick) hosted the unconfined alluvial aquifer (D.B. Stevens, 1995).

To summarize, the Yeso Formation, Glorieta Sandstone and San Andres Limestone together formed an unconfined aquifer high on the slope of the Sacramento Mountains, and became a confined aquifer system lower down the slope where covered by the low-permeability Artesia Group (Fig. 4c). Above the Artesia Group, the Gatuna Formation and younger alluvium formed an unconfined aquifer.

San Juan Basin

The San Juan Basin was also used to test the method for analyzing water level changes in variably confined aquifers. Located in the northwest corner of the state of New Mexico on the Colorado Plateau (Fig. 5a), the San Juan Basin was a structural basin, in that the geologic units in the area all dip toward its center (Fig. 5c). The basin was made up of sedimentary units that range from Cenozoic through Permian in age. Interbedded shale units and other tightly packed formations resulted in numerous rechargeable variably confined aquifers (Fig. 5b.) (Kelley et al., 2014).

While there were numerous variably confined aquifers in the San Juan Basin, well measurement density determined which aquifers could be studied. Much of the San Juan Basin was sparsely populated, and as result, the well network was patchy. We found that in the 1970s and the 1980s the Jurassic aquifer had a large enough well network with which to test the method.

The portion of the Jurassic aquifer studied underlies four USGS HUC-8 basins, and emphasized the contrast between surface water drainages and groundwater aquifers. The HUC-8 basins overlying the aquifer drain into three distinct regions. The Rio San Jose HUC drained east into the Rio Grande basin. The Middle San Jose, and the Chaco HUCs drained northwest into the Upper Colorado basin, and the Upper Puerco drains west into the Lower Colorado basin. The wells screened in the Jurassic aquifer were primarily located on the Chaco Slope, northwest of the San Mateo Mountains. The well network was north of U.S. Interstate 40 between Grants and Gallup. There was another portion of the well network in the Shiprock area, northeast of the Chuska Mountains (Fig. 5a). The primary population centers in and around the well network were Gallup, Grants, and Shiprock.

The Jurassic units that were grouped together to form the aquifer studied here included the lower Morrison Formation, the Wanakah Formation, the Cow Springs Sandstone, the Bluff Sandstone, and Entrada Sandstone (Fig. 5d). The sandstones that made up the Jurassic aquifers were deposited in fluvial (Westwater Canyon Member of the Morrison Formation) and eolian (Entrada Sandstone,

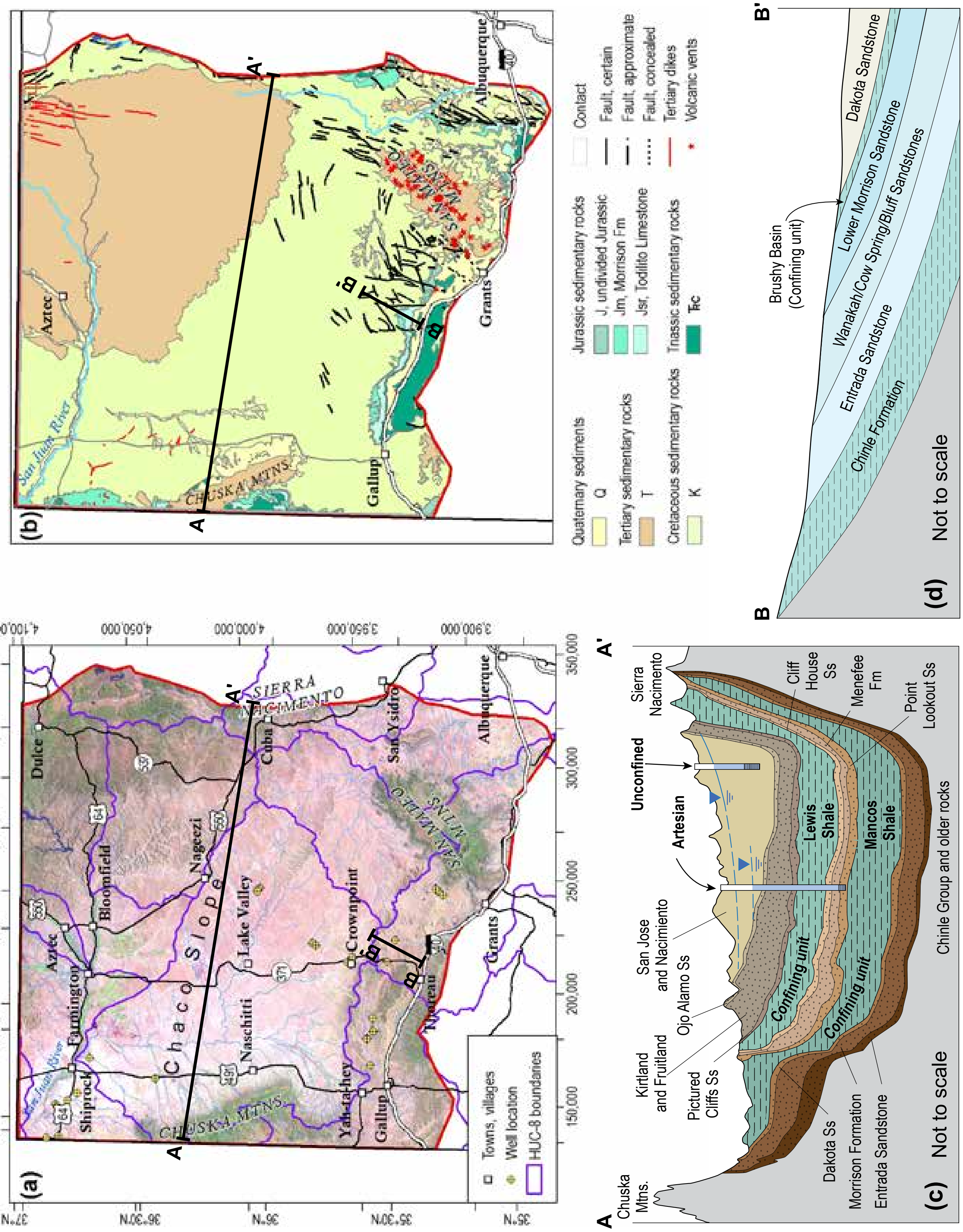


Figure 5. (a) Outline of the San Juan Basin with HUC boundaries in blue, and well network used for analysis. (b) Geologic map (New Mexico Bureau of Geology and Mineral Resources, 2003). Units are grouped by geologic age. (c) East-west conceptual cross-section of the entire San Juan Basin (not to scale). (d) A conceptual cross-section of the partially confined Jurassic aquifer units (not to scale).

Cow Springs Sandstone, and Bluff Sandstone) settings. These units were exposed on the margins of the basin in the foothills of Laramide uplifts where faulting and fracturing could enhance recharge (Kelley et al., 2014). Berry (1959) suggested treating all the Jurassic sandstones as a single hydrostratigraphic unit. He observed that the clays and siltstones that dominate the composition of the Brushy Basin Member, at the top of the Morrison Formation, acted as aquitards separating the Morrison, Bluff, Cow Springs, and Entrada sandstones from the Dakota Sandstone aquifer system above the Morrison (Berry, 1959). We treated the top of the Morrison Formation as the confining unit as the contact between the Morrison Formation and the Dakota Sandstone is well delineated (Kelly et al., 2014).

I.3. Storage change estimation method

Essentially, the new workflow for estimating groundwater storage change in variably confined aquifers added two additional steps to the method of Rinehart et al. (2015 and 2016). First, a literature review was undertaken for the region to decide which strata form the regional aquifer, which strata formed aquitards or seals, and whether or not the aquifers were confined. We defined tops of units in elevation. Then, the water level data was reviewed as in Rinehart et al. (2015), and interpolated as in Rinehart et al. (2015 and 2016). The water level surfaces were then compared to the aquifer tops to see where the aquifer system was confined or not. At this point, we began calculating changes in water levels and changes in groundwater storage; the confined and unconfined calculations were separated at this stage.

I.3.1. Definition of aquifers and confining units

A careful review of existing literature was required to group geologic units into aquifers and to determine which formations act as confining units. Once the aquifer units had been grouped and the confining unit had been determined, the top of the confined aquifer (bottom of the confining unit) was structurally contoured.

For the Permian aquifer in the Roswell Artesian Basin and parts of the Pecos Slope, we incorporated data from a database of formation elevation picks in the southeastern part of the state. As part of research activities at the New Mexico Bureau of Geology and Mineral Resources, this database was compiled including depths to the tops of major stratigraphic units (or stratigraphic "picks") within deep oil and natural gas exploratory wells that had been drilled in the Permian Basin of southeastern New Mexico. This database was used as the basis of aquifer top elevations in the Roswell Artesian Basin in this study, though additional well logs were used to refine the formation tops near the PSRAB area. The database contained stratigraphic picks for 424 wells in Eddy, Lea, Chaves, Roosevelt and De Baca Counties. The wells included in the database either penetrated Precambrian basement, or the oldest strata that were present in the region. Other criteria used for selecting wells for inclusion into the database were the availability of electrical resistivity wells logs. In most cases, the resistivity curves were accompanied by gamma ray or spontaneous potential curves. The electrical resistivity, gamma ray, and spontaneous potential logs were essential for accurate subsurface correlation of formation tops.

In addition, wells included in the database had sample logs on file at the NMBGMR. The sample logs were descriptions of drill cuttings plotted versus depth and provided a lithologic grounding to well log correlations, thereby increasing the accuracy and confidence of stratigraphic picks. In addition, for some wells the correlation of stratigraphic units within the Pennsylvanian and Permian sections was facilitated by reference to fusulinid determinations made on drill cuttings. Fusulinids were a type of marine microfossil that lived during the Pennsylvanian and Permian Periods that evolved rapidly and therefore were useful for age-related correlations of strata deposited within those two geologic Periods. Most of the fusulinid determinations on file at the New Mexico Bureau of Geology and Mineral Resources were made by the late R.V. Hollingsworth. In addition, there had been several projects undertaken and published at the Bureau where databases of more rigorously correlated formation tops were produced for selected stratigraphic units in the New Mexico part of the Permian Basin. These projects involved development of networks of well-log based cross sections that related the well log correlations to descriptions of drill cuttings and cores. This increased the geologic accuracy of correlations to a level greater than is possible by using well logs and sample descriptions without first establishing a network of cross sections. Stratigraphic units for which networks of well log cross sections were produced throughout the Permian Basin include the Abo and Bone Spring Formations (Permian), the Brushy Canyon Formation (Permian), all stratigraphic units within the Mississippian section, and the Woodford Shale (Devonian) and Wristen Formation (Silurian). Stratigraphic picks developed for these projects were also incorporated into the southeastern New Mexico stratigraphic database used. While these formations did not include the primary aquifers of the Roswell Artesian Basin, these detailed picks allow a better estimate of the elevations of overlying younger Permian aquifer materials.

In the San Juan Basin, the aquifer formation tops were identified from the compilation work of Kelley et al. (2014). This work focused on the sandy members of the Mancos shale, but also included estimates of formation tops for most of the stratigraphic section in the area.

In addition to well controls, the New Mexico 1:500,000 geology map (New Mexico Bureau of Geology and Mineral Resources, 2003) was used to add aquifer top elevation control points along the contact where the top of the aquifer unit dips below ground surface. The confining unit surface was contoured using a natural neighbor interpolation, using both well log picks and the surface contact elevation points. Additional control points were added in areas of poor data coverage to aid in the interpolation. Where the aquifer was exposed, as defined by the 1:500,000 geology map (New Mexico Bureau of Geology and Mineral Resources, 2003), a land surface DEM could be superimposed on the structural contour map to represent the top of the aquifer. In the case of the Pecos Slope and Roswell Artesian Basin, the top of the San Andres Limestone was contoured from the well logs and picks. In the San Juan Basin, the top of Morrison formation was previously contoured in Kelly et al. (2014), and the outcrop was defined by New Mexico Bureau of Geology and Mineral Resources (2003) and Kelley et al. (2014).

I.3.2. Well selection and data filtering

Wells from the Aquifer Mapping Database were first filtered by their reported completion code (New Mexico Bureau of Geology and Mineral Resources, 2003), as identified by the USGS unit description code. In the Pecos Slope and Roswell Artesian Basin, wells completed in the lower Permian units (the Yeso Formation, the Glorieta Sandstone and the San Andres Formation) were extracted for analysis of the variably confined aquifer. In the Pecos Slope, as there was sufficient data coverage in the alluvial aquifer, the storage change in the unconfined perched aquifer was also studied. Wells completed in Quaternary-Tertiary sedimentary units were analyzed separately. In the San Juan Basin, wells completed in the Jurassic siliciclastic aquifers were extracted: the Morrison, Wanakah, Summerville, Cow Springs, Bluff, and Entrada Formations.

The Roswell Artesian Basin well measurements were filtered similarly to the Rinehart et al. (2015) scheme. This method excluded measurements taken during the primary agricultural season (March 1 through October 31), and removed any measurement records flagged by the sampling agency as potentially compromised. The hydrograph for each well was then visually examined and spurious measurements were manually removed. Wells with fewer than three water level records were also excluded as there are not enough measurements to determine visually if the measurement was a stable representation of the depth-to-water (DTW).

The area studied in the San Juan Basin did not have the same well network density, or robust measurement history. As result, in order to obtain even a scarce well coverage, the filtering requirements were scaled back to preserve a broader well network and more water level measurements. This altered filtering method relied heavily on the manual analysis of the hydrographs. Each hydrograph was observed individually, and outlier measurements were removed from the records. Concerning wells with very few (1–3) water level measurements, the water level records in neighboring wells were taken into account to better understand how well the record fit in the data set.

Next, for each well, the median depth-to-water measurements were found for each decade. Once the measurements from each well were checked for data quality, we found a correlation length using variogram analysis (Cressie and Wilke, 2011), or the distance within which a water level at a well was statistically predictive, for the depth-to-water measurements in each aquifer system. A variogram was essentially a scaled covariance of measurements as a function of distance between well sites (Cressie and Wikle, 2011). At close distances, DTW measurements were well-correlated and the variance between well measurements within a short distance was low. With increasing distance, the variogram increased to the variance of the measurements (Cressie and Wikle, 2011). The distance at which the variogram reached an asymptote, or sill, was called the range, and was the distance at which measurements are no longer correlated (Cressie and Wikle, 2011).

I.3.3. Water level change analysis

Because of the complicated nature of the geostatistical analysis and the need to repeat the analysis on numerous time steps, it was necessary to automate the process by building an ArcGIS toolbox. The program began by reading a text file created by the user that details which decadal time steps would be processed together. The wells with water level measurements for a given decade were plotted and a buffer of the correlation length is constructed around the wells from the decades being compared. For each difference calculation (e.g., 1970s minus 1950s, or 1990s minus 1980s), we found the intersection of the buffered areas. Wells outside the buffer were excluded from the following interpolations. We had found using wells outside of the interpolation led to spurious changes on the edges of the depth-to-water interpolations.

Next, a surface contour map of the DTW was constructed from the average depth-to-water found in the remaining wells, for each decade. This surface was created using either ordinary kriging or inverse-distance-weighting (IDW) interpolations. At this point several additional steps were required for the water level change analysis of the variably confined aquifers.

To determine where the aquifer was confined and unconfined, the depth-to-water map was subtracted from a DEM to convert the surface to a potentiometric water elevation map. The water elevation map was then subtracted from the elevation of the confining unit. This produced a map of the height of water above and below the confining unit. Where the aquifer was confined the value was negative and where the value was positive the aquifer is unconfined. The map was then divided into the two unique maps (confined and unconfined) for separate analysis. For each time step, the older confined depth-to-water surface was subtracted from the younger confined depth-to-water surface to find the change in water level. The same was done for the isolated portion of the maps that was unconfined (Fig. 6).

As the water level fell, the boundary between the unconfined and confined zone moved down slope (Fig. 2, and Fig. 6). The location of this boundary fluctuated with the water table and was different each decade. The location of this boundary between the confined and unconfined zone played a significant role in the storage calculation. When finding water level change between two decades, the boundaries were combined and described as a ‘zone of confinement’ (Fig. 7).

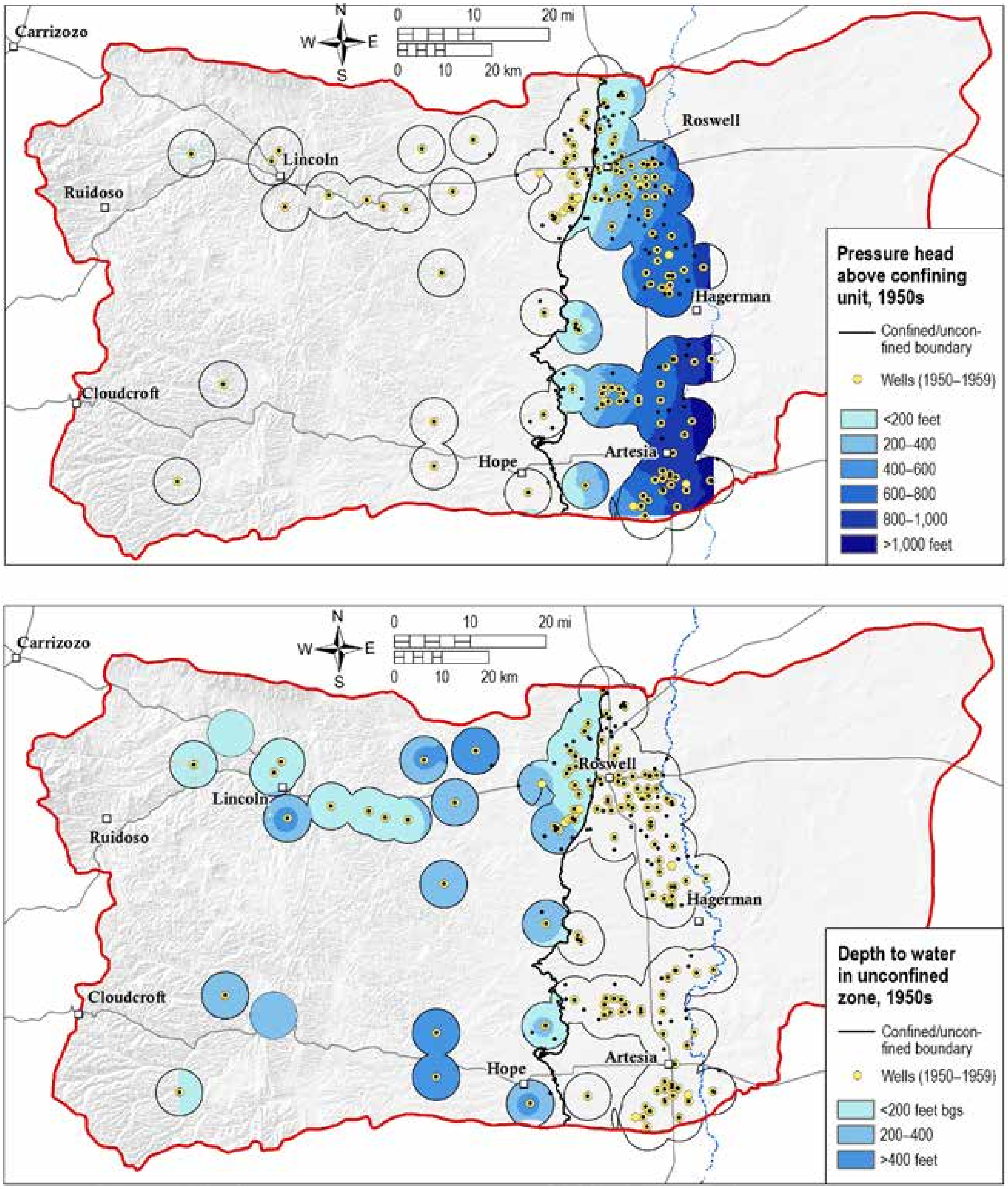


Figure 6. Example from the 1950s. Shows pressure head above confining unit, and depth to water below ground surface (bgs) in the unconfined zone.

I.3.4. Storage change calculation

Once the water level change maps were constructed, the total change in volume of the water storage, for both the confined and unconfined aquifers, could be estimated. To determine the change in volume we needed to know the storage properties of the aquifers in question: the specific yield of the unconfined portion, and the storativity for the confined portion. Storage coefficients were found in existing literature on the respective basins. For each decadal time-step comparison, the water level change in the confined and unconfined portion of the aquifer was multiplied separately by the applicable storage coefficient, multiplied by the grid cell area outputted by the interpolation as follows:

$$\textit{Unconfined Volume Change} = \textit{Cumulative unconfined WL change} \times \textit{specific yield} \times \textit{cell size} \tag{2}$$

$$\textit{Confined Volume Change} = \textit{Cumulative confined WL change} \times \textit{storativity} \times \textit{cell size} \tag{3}$$

and finally the confined and unconfined volume change estimate were added together, as in

$$\textit{Total Volume Change} = \textit{Unconfined Volume change} + \textit{Confined Volume change} \tag{4}$$

This returned an estimated change in total groundwater storage for the entire confined and unconfined aquifer system (Schwartz and Zhang, 2003).

In comparing the final calculated storage change in the Permian aquifer for the PSRAB area, we displayed calculated storage changes based on the difference in volume from the 1900s to 2010s, and from the 1950s to 2010s. The change in volume data was graphed in three distinct ways; the total difference, the moving difference, and the cumulative moving difference. The total difference compared the storage change for each decade against an initial measurement point. For example, on the graph showing the 1900s to 2010s, the storage change was compared against the starting point, the 1900s. The storage change in each subsequent decade was compared against the well network from the 1900s. The problem with comparing subsequent decades against the 1900s was that the well networks had changed drastically in succeeding decades. In the 1900s the buffered well network only covered 461 sq mi, and was focused entirely on the confined portion of the aquifer. The 1950s buffered well network covered 1,178 sq mi, and covered both areas that were confined or unconfined in the Permian aquifer system.

The moving difference was a measure of the change in storage from one decade to the next, graphing the volume change for each individual time step. For example, difference between each decade was calculated: the change in volume from the 1950s to the 1960s is reported, then the change in storage from the 1960s to the 1970s was reported, and so on. This series only found the change between decades, and was more sensitive to changes in the well monitoring network.

The cumulative moving difference found the cumulative, or moving, sum of the moving difference calculations. For example, it took the storage change reported for the 1960s and added it to the storage change reported in the 1950s. Next it took the volume change reported in the 1970s and added it to the cumulative volume change for the 1950s and 1960s, and so on.

In short, the total difference allowed us to examine storage changes keeping the region of interest as defined by the well networks static through time. The moving difference and cumulative moving difference allowed the region of interest to change as wells are added and removed from the monitoring network. The former gave a better sense of historical trends, while the latter gave a better sense of decade-to-decade changes.

I.4. Results

I.4.1. PSRAB region

Agricultural interests in the PSRAB region had relied on its groundwater resources for over 100 years (Hall, 2002). As a result, the water level records in the basin were amongst the most robust in the state. In the Permian aquifer, each decade from the 1900s to 2010s (except for the 1920s) had between 66 and 200 high-quality well measurements. The alluvial aquifer was more robust; each decade from the 1930s to 2010s had between 100 and 600 wells available in the monitoring network after filtering. The storage coefficients used to quantify storage change for the aquifers were found in D.B. Stevens (1995). The report compiled published storage coefficients from numerous studies conducted on the Permian aquifer system, as well as the alluvial system. We used the storage coefficients calibrated by groundwater flow model reported by D.B. Stevens (1995). The correlation length found for the alluvial aquifer was 6 km, and the specific yield was 0.17 (D.B. Stevens, 1995). The correlation length determined for the Permian aquifer was 5 km. In the unconfined portion of the Permian aquifer, 0.05 was used for the specific yield, and the storativity used for the confined portion was 0.0005 (D.B. Stevens, 1995).

From the 1900s through the 1950s, the monitoring network focused almost exclusively on the confined portion of the Permian aquifer. During this period, declines were relatively minor, on the order of 0.05 Maf. An example of this sparse well coverage could be seen in Figure 7a. In these figures, negative values indicated rises in the water level, and positive numbers indicated declines in water level. From the 1950s to 1960s, we saw the largest withdrawals, almost 0.5 Maf. During this time step the well coverage was greatly expanded as seen in Figure 7b. From the 1970s through the 1990s, groundwater storage appeared to recover before declining slightly in the 2010s (Fig. 8). While the total volume of water extracted from the Permian aquifer system had been relatively small, in the confined area between Roswell and Artesia there had been significant potentiometric declines in the potentiometric surface. In the Roswell area, the potentiometric surface had dropped by 50 to 60 ft over the 110 year measurement record. The Artesia area had experienced declines on the order of 150 ft over the entire record. This demonstrated how rapidly the potentiometric surface declined in confined areas due to extraction of small volumes of water.

Decreases in storage in the alluvial aquifer had been much more significant. Groundwater storage had steadily declined in the alluvial aquifer since the 1940s. In the 1980s, the rate of decline began to slow, though withdrawals remained significant. The 1990s saw a brief period of recovery before declining again. We estimated roughly 3.75 Maf of cumulative storage loss (Fig. 9, 10; Table 1).

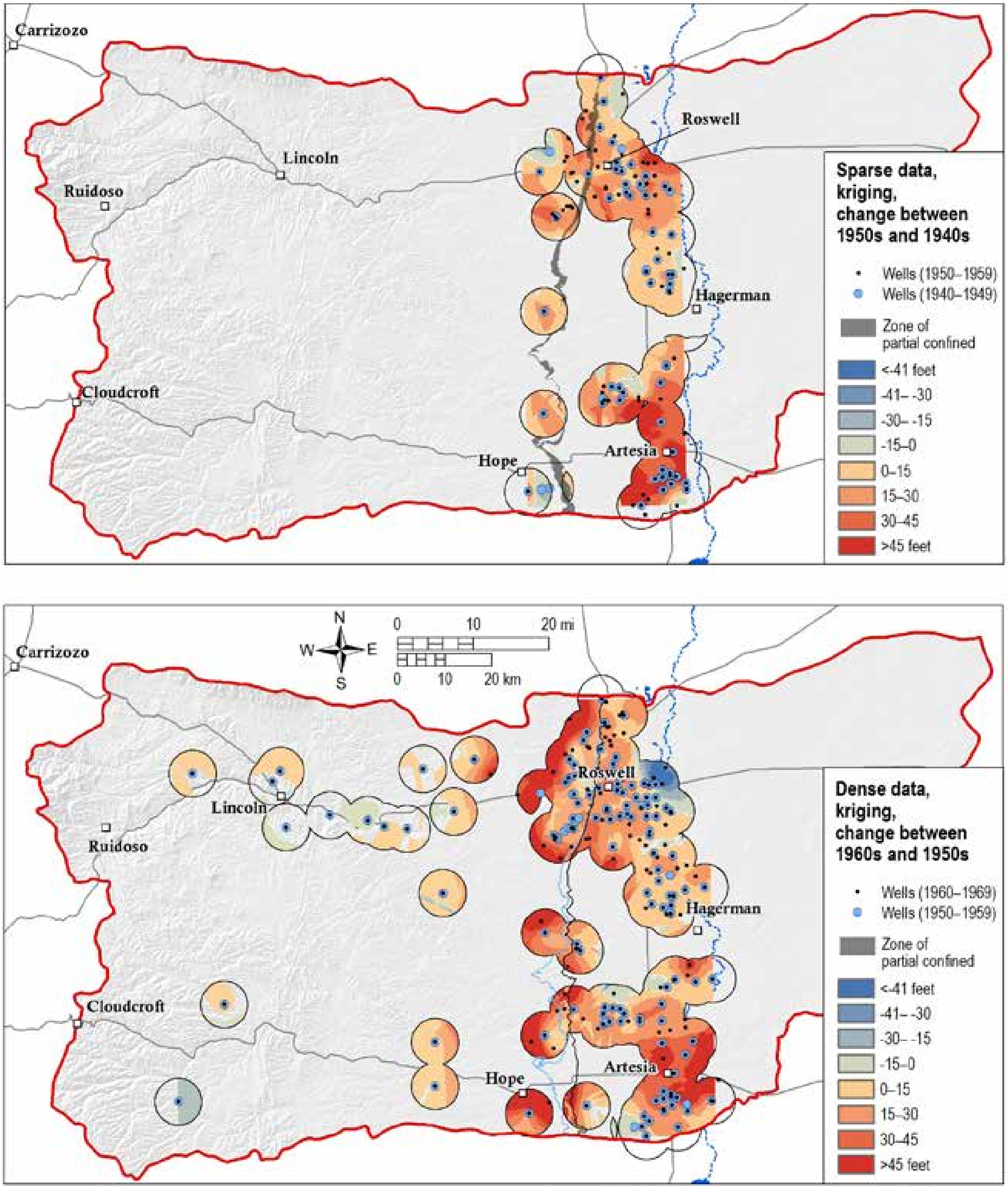


Figure 7. (a) Example of sparse data coverage of DTW change between the 1940s and 1950s in the Permian aquifer found in the Roswell Artesian Basin. (b) Example of dense data coverage of DTW change between 1950s and 1960s in the Permian aquifer.

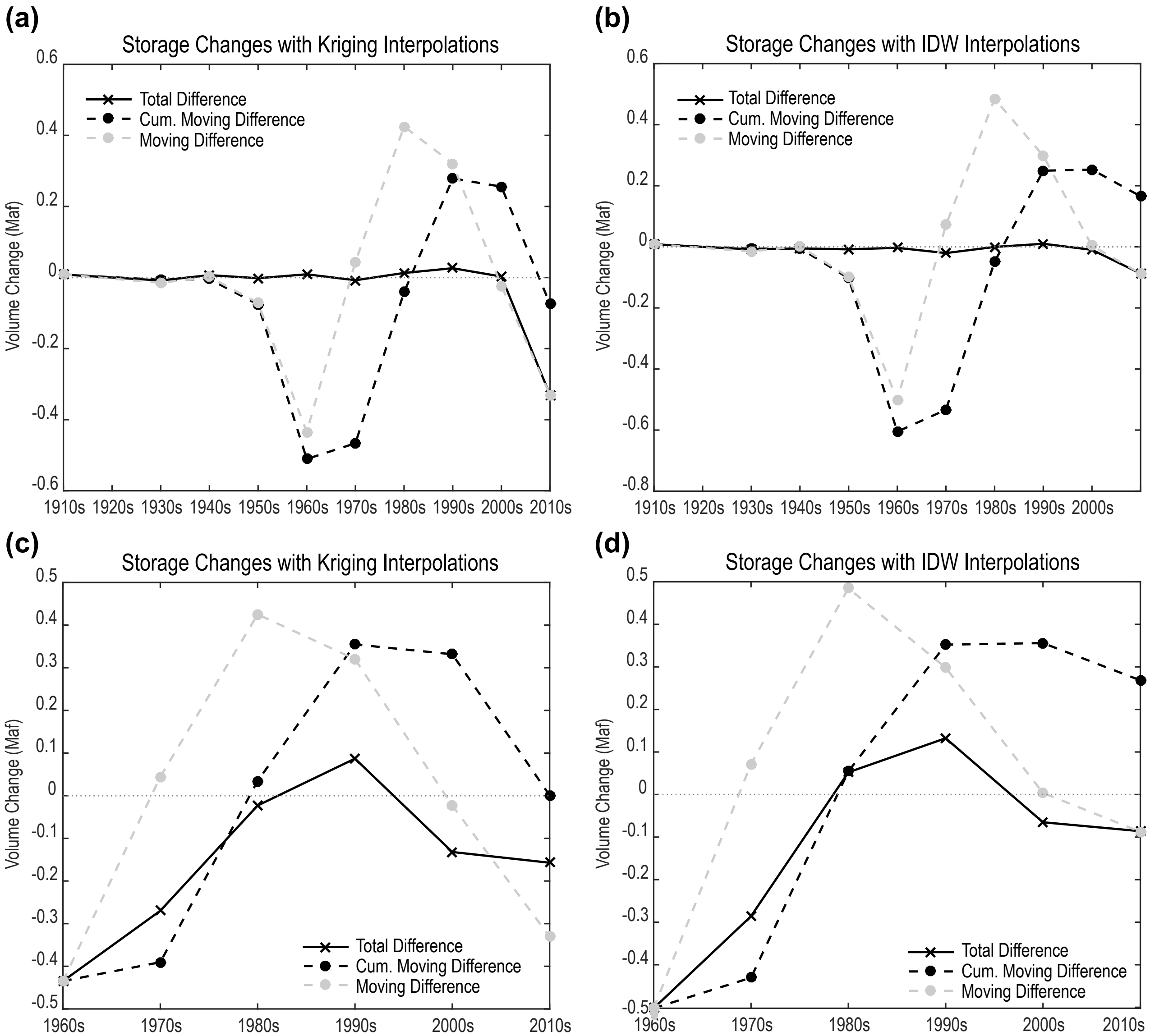
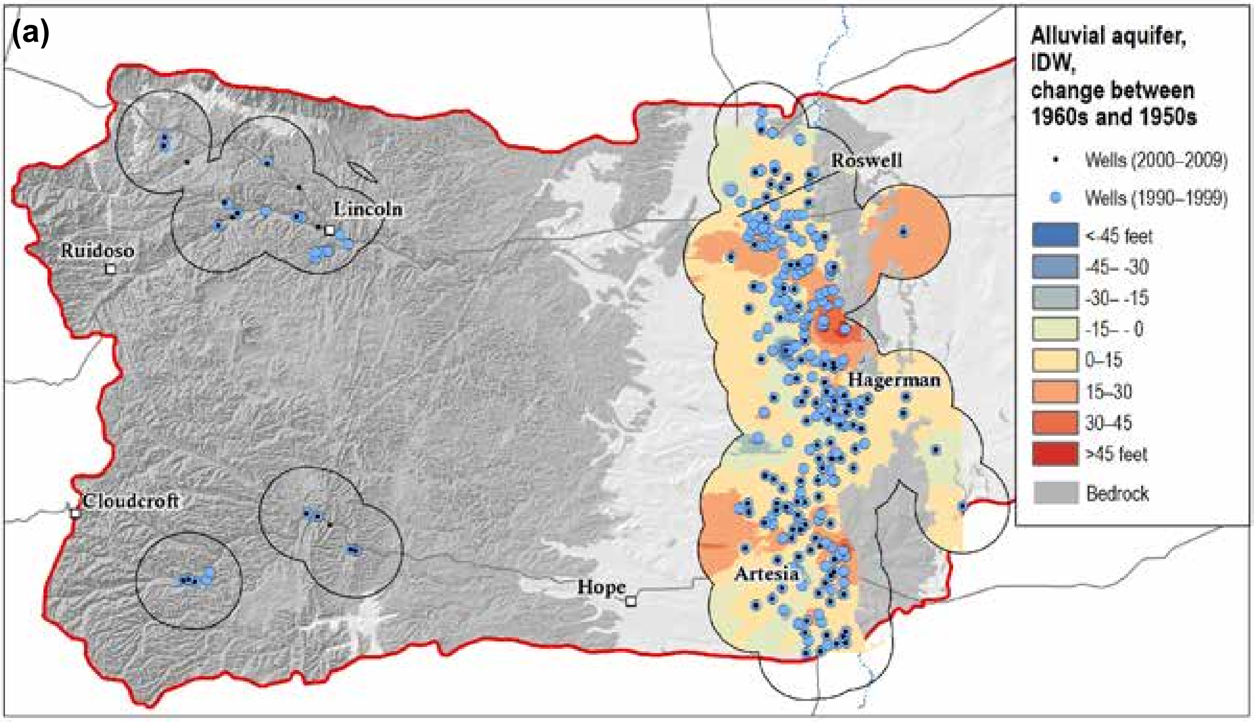


Figure 8. Estimated groundwater storage changes (million acre-feet) for each decade in the Permian Aquifer system in the Roswell Artesian Basin, with total changes (solid x), cumulative moving changes (black dashed dots), and moving changes (dark grey dashed dots) showing for (a, c) kriging interpolation and the (b,d) IDW interpolations in the Permian aquifer. (a, b) Storage change compared against the 1900s. (c, d) Storage change compared against the 1950s.



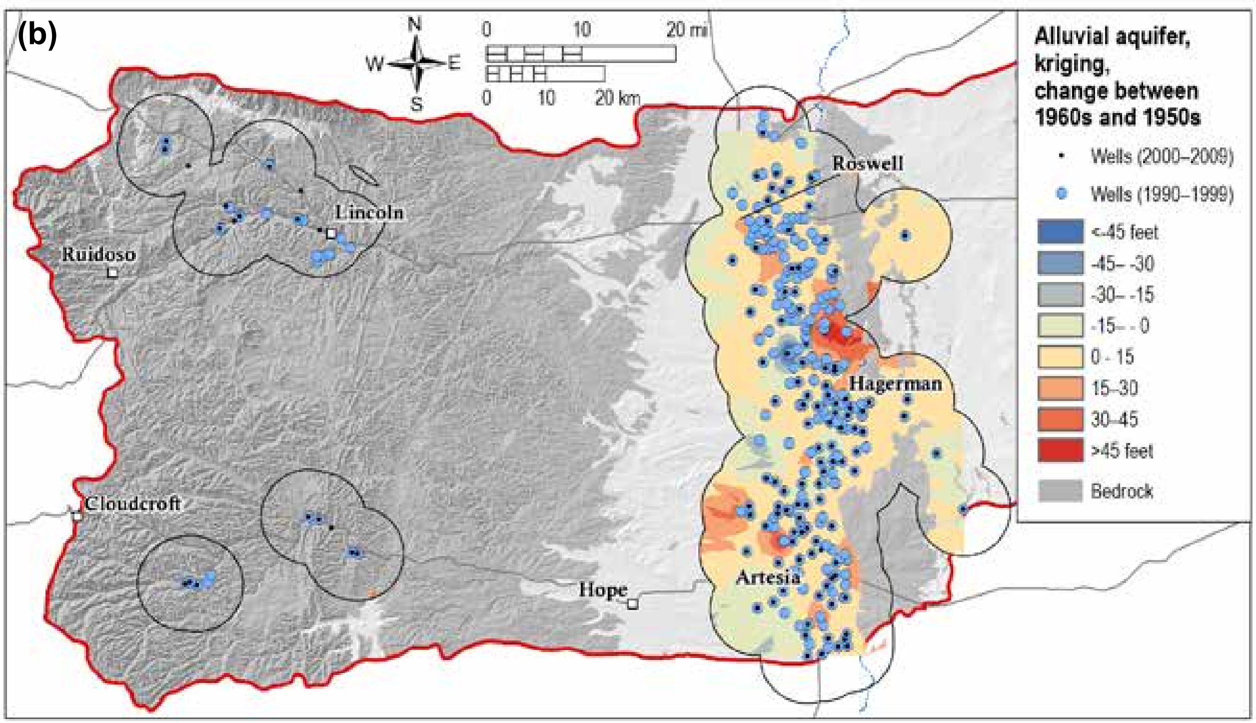


Figure 9. Water level changes between the 1950s and the 1960s in the perched alluvial aquifer using (a) IDW interpolation, and (b) Kriging interpolation.

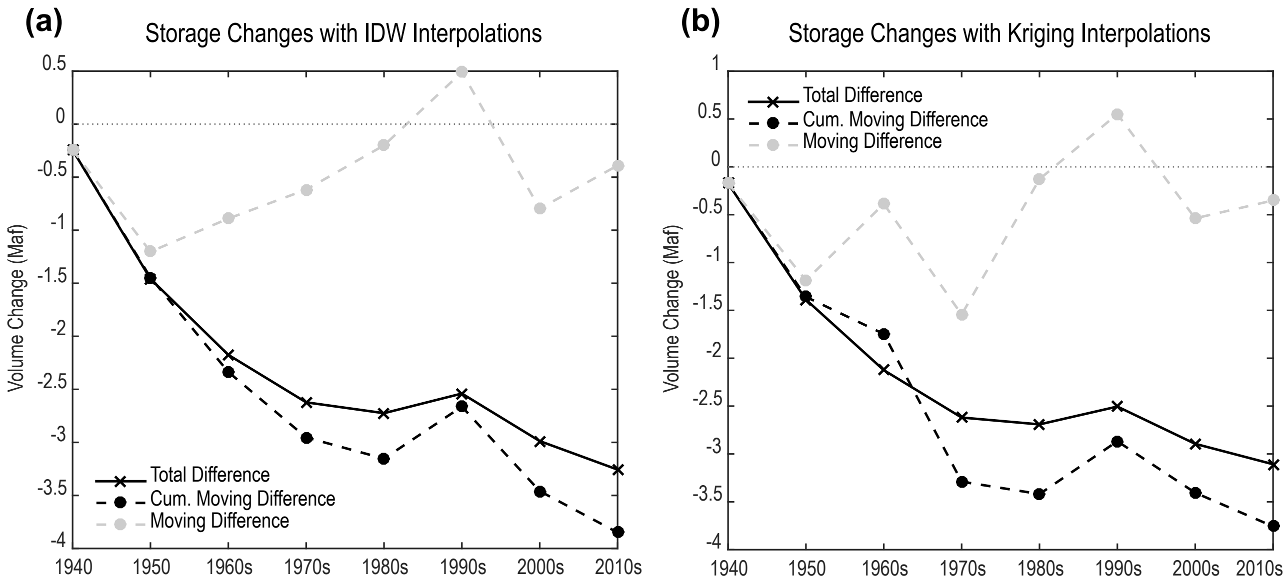


Figure 10. Estimated groundwater storage changes (million acre-feet) for each decade in the Perched alluvial aquifer, with total changes (solid x), cumulative moving changes (black dashed dots), and moving changes (dark grey dashed dots) showing for (a) IDW interpolation and the (b) Kriging interpolations.

Table 1. Groundwater storage changes for the Pecos Slope and Roswell Artesian Basin Permian and alluvial aquifers, and the Jurassic aquifers in the San Juan Basin.

STORAGE CHANGE (Maf)																						
	1910s		1920s		1930s		1940s		1950s		1960s		1970s		1980s		1990s		2000s		2010s	
	IDW	Krig.	IDW	Krig.	IDW	Krig.	IDW	Krig.	IDW	Krig.	IDW	Krig.	IDW	Krig.	IDW	Krig.	IDW	Krig.	IDW	Krig.	IDW	Krig.
PECOS SLOPE AND ROSWELL ARTESIAN BASIN																						
Permian Carbonate																						
Moving diff.	-0.006	-0.006	--	--	-0.02	-0.02	0.00	0.00	-0.10	-0.07	-0.51	-0.44	0.07	0.04	0.49	0.43	0.30	0.32	0.00	-0.02	-0.09	-0.33
Cumul diff.	-0.006	-0.006	--	--	-0.01	-0.01	-0.01	-0.01	-0.11	-0.08	-0.62	-0.52	-0.55	-0.48	-0.06	-0.05	0.24	0.27	0.24	0.24	0.15	-0.09
Total diff.	--	--	--	--	--	--	--	--	--	--	-0.51	-0.44	-0.29	-0.27	0.05	-0.03	0.13	0.08	-0.07	-0.13	-0.09	-0.16
Quaternary Alluvial																						
Moving diff.	--	--	--	--	--	--	-0.24	-0.17	-1.20	-1.19	-0.89	-0.39	-0.62	-1.54	-0.20	-0.13	0.49	0.55	-0.80	-0.54	-0.39	-0.35
Cumul diff.	--	--	--	--	--	--	-0.24	-0.17	-1.45	-1.36	-2.34	-1.75	-2.95	-3.29	-3.15	-3.42	-2.66	-2.87	-3.46	-3.41	-3.85	-3.76
Total diff.	--	--	--	--	--	--	-0.24	-0.17	-1.46	-1.39	-2.17	-2.12	-2.62	-2.62	-2.72	-2.69	-2.54	-2.50	-2.99	-2.89	-3.26	-3.11
SAN JUAN BASIN																						
Jurassic Siliciclastic																						
Moving diff.	--	--	--	--	--	--	--	--	--	--	--	--	--	--	-0.19	--	--	--	--	--	--	--

I.4.2. San Juan Basin

The groundwater level monitoring network in the San Juan Basin was not nearly as robust as the Roswell Artesian Basin. It was very difficult to find an aquifer with adequate well coverage, with a detailed structural contour map of the bottom of the confining unit. Even after adapting a modified filtering methodology, only the Jurassic aquifer had two consecutive decades with sufficient well coverage for analysis: the 1970s had 26 available well locations, and the 1980s had 58 locations (Fig. 11). Sparse well networks were found to introduce errors at the edges of the interpolated area. In Rinehart et al., (2016), IDW surface interpolation was found to reduce the effect compared to the kriging method. The correlation length

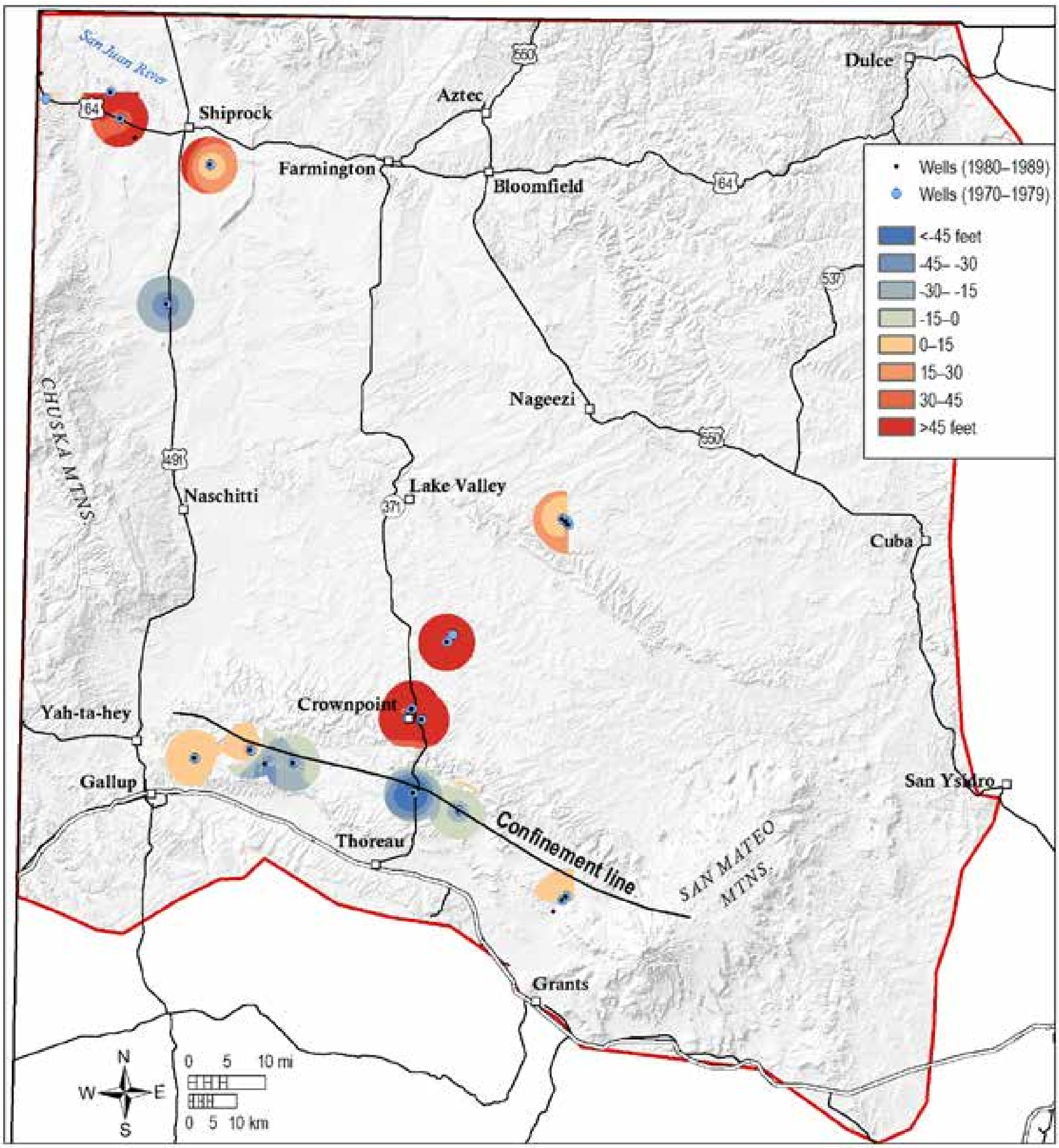


Figure 11. IDW interpolation of water level change in the Jurassic aquifers in the San Juan Basin between the 1970s and 1980s. The boundary where the aquifer switches from confined to unconfined is denoted by the line labeled 'Confinement line'.

found for the Jurassic aquifer was 6 km, the specific yield was 0.15, and the storativity value was 0.001 (Kelly et al., 2014).

While we only have two decades to compare, between the 1970s and 1980s, the areas with data near the edge of the confined/ unconfined zone showed a rise in water levels. As we moved further away from the edge into the confined zone, toward the center of the basin, there were declines (Fig. 12). The volumes calculated are found in Table 1.

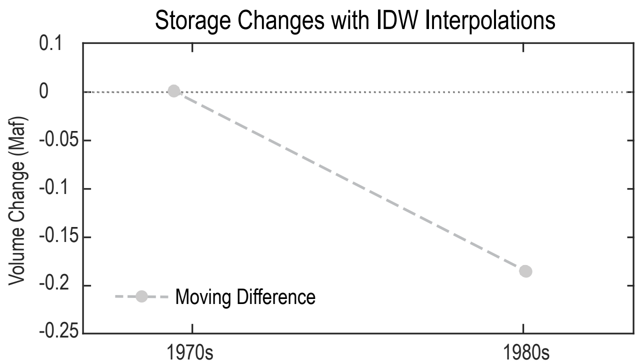


Figure 12. Estimated groundwater storage changes (Million acre-feet) from the 1970s to 1980s, using the IDW interpolations in the Jurassic aquifer in the San Juan Basin. Only the moving difference is shown.

I.5. Discussion and Limitations

I.5.1. Discussion

Our findings were an analog of the history of groundwater management in the Pecos Slope and Roswell Artesian Basin. The first well in the artesian system was drilled in 1892 (Shomaker, 2003). The discovery of free-flowing artesian water led to a boom in agricultural development in the area and by 1915 more than 1300 wells had been installed (Shomaker, 2003). In 1932, the Pecos Valley Artesian Conservancy District (PVACD) was established to help conserve the confined aquifer. This agency had been instrumental in maintaining and improving the valley's agricultural infrastructure, and implementing more efficient irrigation systems (Hall, 2002). The PVACD had purchased and retired almost 7,000 acres of irrigation rights. In 1956 the adjudication of water rights began and led to the retirement of an additional 12,000 'illegal' acres (Shomaker, 2003). In 1967, the PVACD required irrigation wells in the basin to be metered; this ensured that land owners were only accessing their allotted water rights (Barroll and Shomaker, 2003).

The groundwater storage change graphs that were produced for this study showed small declines in groundwater storage in the carbonate system in the early 1900s, becoming more pronounced in the 1950s and 1960s (Fig. 8). The groundwater policy implementations of the 1960s seemed to have had a significant impact on the storage change. Throughout the 1970s, 1980s, and 1990s, the volume of water in the confined system appeared to recover. While the application of policy changes in the 1960s no doubt had a significant impact on reducing the groundwater depletion, the recovery seen in the graphs was likely also a product of the growing well network in the unconfined region of the aquifer, farther up the Pecos Slope.

Another factor that should be taken into account was the potential leakage between the confined and unconfined system. Since the artesian system was first discovered, thousands of wells had punctured the confining unit to tap the pressurized system beneath. In an effort to reduce leakage between the aquifers, the PVACD had plugged more than 1500 leaky or abandoned artesian wells (Shomaker, 2003). Originally, the effort focused on preventing the upward flow from the confined aquifer into the alluvial aquifer. Recently, however, the potentiometric surface of the confined system dropped below the water table in the alluvial system. As there was currently a downward gradient in places between the two aquifers, water was potentially leaking from the alluvial system into the confined aquifer.

Previously reported estimates of annual pumping for the entire basin suggested a steady increase from the early 1900s until the early 1950s, when pumping reached approximately 400 kaf/yr. Pumping dropped off in the early 1970s and has remained steady, around 350 kaf/yr. Recharge estimates for the basin assume approximately 300 kaf/year (Barroll and Shomaker, 2003; Rawling and Newton, 2016). If we assumed pumping withdrawals had exceeded recharge by roughly 50 kaf/yr for the past 70 years, this amounted to a cumulative deficit of 3.5 Maf. This approximation was reassuringly close to our estimate of a cumulative storage loss of 4 Maf for both aquifers (Fig. 8, 10).

In the San Juan Basin, there were not enough data to provide any meaningful analysis, as there was a paucity of historical groundwater data integrating across the area of interest. Between the two decades that we have enough data to actually process water level changes, the actual change in storage was minimal. The volume change from the 1970s to 1980s, using the IDW analysis, was an estimated 0.19 Maf decline (Fig. 12).

I.5.2. Limitations of method

While the alluvial storage change method (from Rinehart et al., 2016) was successfully adapted to study the variably confined aquifers, there were several limitations that must be taken into account when applying it further. Compared with the previous method, the analysis described here required a significantly more detailed understanding of the hydrogeologic framework. When studying the alluvial aquifers, we lumped all of the sand and gravel units into one aquifer unit. For this analysis it was necessary to have a firm grasp of which units make good aquifers, as well as which aquifers were closely interconnected and could be grouped together.

Just as important as knowing which units were good aquifers was having a grasp of which units act as confining units. Having enough aquifer top picks to construct an accurate structural contour map of the contact between the aquifer and the confining unit above it was crucial in delineating where the aquifer was unconfined or confined, and as a result, determining the volume of water that had been removed from the aquifer. To construct this contact, detailed well logs were required, as well as a map of the geologic surface contact. Even in basins with an in-depth understanding of the hydrostratigraphy and robust well control, there was still a degree of error associated with contouring the elevation of the structural confinement surface map. Because of the inherently large disparity between the confined and unconfined storage coefficients, the error associated with the elevation of the confinement surface when calculating storage change was compounded.

Data coverage was another major limiting factor. One of the aims of this project was to develop a method which could be generally applied to a given well network screened in a confined aquifer. When a dataset was robust, both spatially and temporally, as was the case in the PSRAB area, a rigorous filtering protocol could be applied to the water level measurements. This included removing all records collected during the agriculture season when water levels may be impacted by pumping. Wells with fewer than three water level records were also be excluded as this typically isn't enough to determine if the measurements are consistent through time. Additionally, problematic measurements that were flagged by the measurement agency were removed to exclude possibly spurious data. These were easily automated filtering protocols that can be uniformly applied to a dataset, resulting in repeatable, non-biased datasets. In the San Juan Basin, measurements were so sparse, spatially and temporally, that a significantly more relaxed filtering protocol was applied. Even with the relaxed filtering protocol, only two decades had sufficient data to be processed. Less spatially consistent well networks were found to introduce errors at the edges of the interpolated area. IDW surface interpolation was found to reduce the effect, as opposed to the kriging interpolation.

Finally, changing well networks played a large role in complicating storage change estimates. We attempted to mitigate the impact of expanding and contracting well networks by only comparing well networks where they overlapped within the radius of correlation. Additionally, we calculated changes using both a total change from a base time and a decade-by-decade change to partially isolate well network changes from storage changes.

A chronic challenge with the method, both in Rinehart (2015 and 2016) and in this study, was the lack of uncertainty analysis. The reality was that there were very few resources available to compare our study to. This type of analysis had been uncommon, especially with the added constraint of producing results comparable across New Mexico. Rather than attempting to test the uncertainty at every step, we had chosen to use multiple interpolations and multiple differencing to characterize uncertainty within the method. We had also chosen to only use the highest quality data, both for the geologic formation top picks and for the water level measurements, and excluded lower quality information. This makes our estimates generally conservative (best-case-estimates; Rinehart, 2015).

Uncertainty analysis in spatiotemporal fields was notoriously difficult. The only general approaches were to limit study areas where a traditional regression-interpolation cycle could be used (see Part II, this study, and Rawling and Rinehart, in press, for an example), followed by incorporation of regression residuals, interpolation leave-one-out cross validations, and analysis of ordinary kriging variances. However, this would have required us to only look at small parts of basins (Rinehart, 2015). The other approach would be to construct a largely ad hoc bootstrap or Monte Carlo approach to explore the full distribution of individual well measurement effects. This would be computationally prohibited.

As mentioned above, we rather choose to use the highest quality data available, to areally restrict the area to regions with data coverage and spatial correlation, and to use multiple interpolation and differencing schemes to understand the range of uncertainty.

I.6. Summary

The method described here effectively estimated the change in water storage in variably confined aquifers, particularly in the Roswell Artesian Basin and Pecos Slope. A structural contour map was created of the top of the confined aquifer using well logs, and digital elevation and geologic maps. For each decade the potentiometric surface was contoured for the entire basin. The top of the confining unit was used to determine where the aquifer transitions from unconfined to confined. This allowed the fluctuations in the potentiometric surface of the confined portion to be processed separately from the water table fluctuations in the unconfined portion. The change in water level was determined for every decade where enough data was available. The aquifer storage change for the entire basin could then be determined by applying known hydraulic properties of the aquifers to the water level fluctuations.

In the Roswell Artesian Basin there was sufficient well coverage for the storage change analysis, both spatially and temporally. This demonstrated that, with a robust well network, the method could produce estimates of storage change. In the San Juan Basin, however, measurements were so sparse, spatially and temporally, that a significantly more relaxed filtering protocol was required. Even with the relaxed filtering protocol, only two decades had sufficient data to be processed. While we were able to apply the method to the sparse data coverage, the analysis of the San Juan should be viewed more as a proof of concept. In general, the approach for calculating storage changes in stacked or variably confined aquifers was effective, given dense enough data through time and in space.

I.7. Acknowledgments

This work was funded under NM Water Resources Research Statewide Water Assessment. Water level data was contributed by L. Sherson (U.S.G.S., Albuquerque Water Science Center) and by A. Balok from the Pecos Valley Artesian Conservancy District. We thank Talon Newton (NMBG), Michael Johnson (USGS) and Stacy Timmons (NMBG) for their thoughtful reviews.

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PART II. Historical Groundwater Storage Change in the New Mexico Southern High Plains Aquifer

Alex Rinehart, Geoffrey Rawling, Brigitte Felix, and Cathryn Pokorny.

II.1. Introduction

The aquifers underlying the Southern High Plains (SHP) of New Mexico are the primary source of water for the communities, agriculture and industry of the region, making understanding changes in groundwater storage vital for the economic health of the region. In this study, we modified the methods of Rinehart et al. (2015 and 2016) and Rawling and Rinehart (in press) to estimate changes in saturated thickness and aquifer storage from the 1950s through the 2010s at decadal time steps in the New Mexico SHP aquifer. The primary aquifer in this region was the saturated portions of the Ogallala Formation with some aquifers in the overlying alluvium and eolian sands. The Ogallala Formation was a generally unconfined, coarse-sand to gravel fluvial sandstone stretching from Nebraska and eastern Colorado to South Texas (Weeks et al., 1988).

Located along the eastern edge of New Mexico, the New Mexican SHP stretch from the Canadian River south to just north of the southern New Mexico border, and extend roughly 60 miles into New Mexico from the eastern border of the state (Hart and McAda, 1985). It includes parts of Curry, Lea, Quay and Roosevelt counties (Fig. 13). Technically, the Ogallala Aquifer north of the Canadian River is considered to be part of the Central High Plains province (Weeks et al., 1988).

Based on historically saturated regions of the SHP aquifer, our study was split into two parts: the region in Quay, Roosevelt and Curry Counties (QRC area); and the region in Lea County (Lea County area; Fig. 13). In the QRC area, the primary users of water were large-scale irrigated agriculture, with secondary users being municipalities, domestic users, livestock agriculture, and other users (Longworth et al., 2013). The Lea County area similarly had large-scale irrigated agriculture as the dominant consumptive use with secondary water users being municipalities, industry, livestock, commercial and domestic users (Longworth et al., 2013).

Critically, over 60 years of research have shown that there is little recharge entering the SHP aquifers, and what little there is had been overwhelmed by pumping (see Rawling, 2016 for review). The lack of recharge, the continued intensive pumping, and the dwindling water supply all emphasized that our estimates of storage change needed to be as precise as possible. Additionally, the 60+ years of research by the USGS and others provide a foundation for us to build off of—the SHP aquifer in New Mexico has some of the most reliable water level measurement well networks in New Mexico (Fig. 13a), and had high quality maps of historically saturated regions (Fig. 13b), specific yield, and aquifer bottom elevations (Fig. 13; Hart and McAda, 1985; and Tillery, 2008). The combination of critical supply issues and the availability of high quality data allowed us to further refine the method developed in the previous two years of work with WRRI Statewide Water Assessment (Rinehart et al., 2015 and 2016).

The goal of this study was to estimate the changes in groundwater storage in the entire New Mexican SHP split into the QRC area and the Lea County area. We began by providing an overview of the hydrology, water use and geology of the region (Section II.2), followed by an overview of the refined method (Section II.3). In Section II.4, we presented the results of our calculations in the form of maps of water

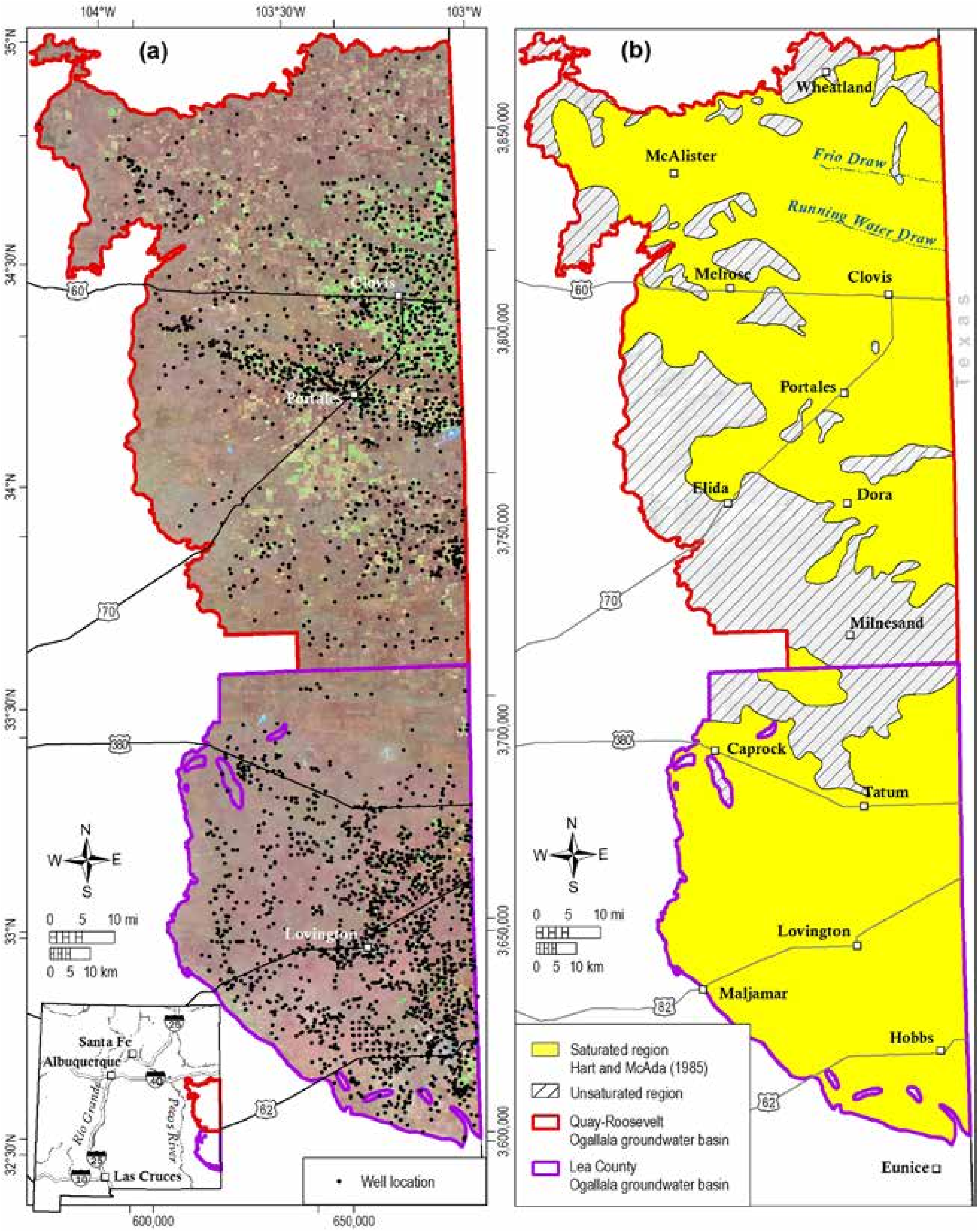

Figure 13. With study area inset, maps of (a) locations of USGS water level well measurements; and (b) historically saturated regions of the SHP aquifer in New Mexico (Hart and McAda, 1985) for the Quay-Roosevelt-Curry County (QRC) area (red outline) and the Lea County area (purple outline).

table elevation, saturated thickness and changes in saturated thickness, and total volumes of groundwater storage change. These results were compared against NM Office of the State Engineer and other groundwater storage change results, and were discussed relative to the socioeconomic drivers of groundwater use in Section II.5. We summarize the major findings of the study in Section II.6.

II.2. Hydrologic and geologic setting

The New Mexican SHP was part of the broader Southern High Plains in West Texas and New Mexico. The region was defined by the primary aquifer (Ogallala Formation), low relief topography, and semiarid climate with hot summers and occasional intense storms (Weeks et al, 1988; and Rawling, 2016). Economically, the region was driven by groundwater-fed irrigated agriculture, and oil and gas development in the various Permian-aged basins (Longworth et al, 2013). There were various military installations and, increasingly, renewable energy installations that provided an additional economic base (Longworth et al., 2013; and NMISC, 2016a,b).

In Lea County, the economy was driven largely by oil and gas development, with over 19,000 oil and gas wells existing in the county as of 2016 in many different oil pools, and by groundwater irrigated agriculture (NMISC, 2016b). The use of hydraulic fracturing in horizontal wells in Lea County had created concerns of fresh water being used for oil well development, taking water away from an already stressed aquifer (NMISC,2016b). However, the vast majority (87%) of water use had been for irrigated agriculture, with the next major use being municipal use, followed by industrial and domestic use (Longworth et al., 2013). Almost all of the fresh water was from groundwater (Longworth et al., 2013).

In the QRC region, once again, the vast majority of water use was for irrigated agriculture (97%) and greater than 99% of the freshwater was groundwater from the Ogallala aquifer (compiled in Rawling and Rinehart, in press, from Longworth et al, 2013). In the QRC region, however, the underlying rocks were not being used for oil and gas development; the only current water use were irrigated agriculture, municipal use, industrial and non-irrigated agriculture (e.g., cheese factories), and domestic water use (NMISC, 2016a).

As mentioned before, the New Mexican SHP aquifer was in the Ogallala Formation. We defined the SHP aquifer as the saturated part of the Ogallala Formation and overlying Pleistocene and younger alluvial and eolian deposits (Weeks et al., 1988). The Ogallala Formation was a thick body of sand and gravel laid down in paleovalleys from about 20 Ma until 5 Ma. These river systems were broad, coarse braided streams that downcut and backfilled through the Miocene, leaving behind coarse sands and gravels with finer grained silty sands to clays interbedded and along the margins of the paleovalleys (Hawley, 1993). The paleovalleys were generally cut into Mesozoic and older sedimentary rocks; these rocks had low hydraulic conductivities and commonly had poor water quality (Weeks et al., 1988). On top of the Ogallala Formation lay the Plio-Pleistocene eolian and alluvial Blackwater Draw Formation, and Holocene to recent deposits of wind-blown sediments and alluvium (Holliday, 1989; and Hawley, 1993). There was no barrier to flow between the Ogallala Formation and overlying younger sediments; they formed a single aquifer system when there was enough saturated thickness.

The SHP aquifer was one of the most thoroughly studied aquifer systems in the country. Over 60 years of detailed research had been undertaken by the U.S. Geological Survey (USGS) and others. The goals of these studies had been to define the geometry, sediments, hydraulic properties, recharge potential, and changing water table elevations. The High Plains aquifer stretched from Nebraska to southern Texas, with the SHP aquifer beginning in west Texas and central-east New Mexico and continuing into southwest Texas. In New Mexico, the bottom and top of the aquifer were well defined—the aquifer top was defined by the water table as measured in wells (Fig. 13a) and areas that had been historically unsaturated (Fig. 13b), and the bottom was formed by the paleovalley boundaries (Fig. 14). The New Mexican

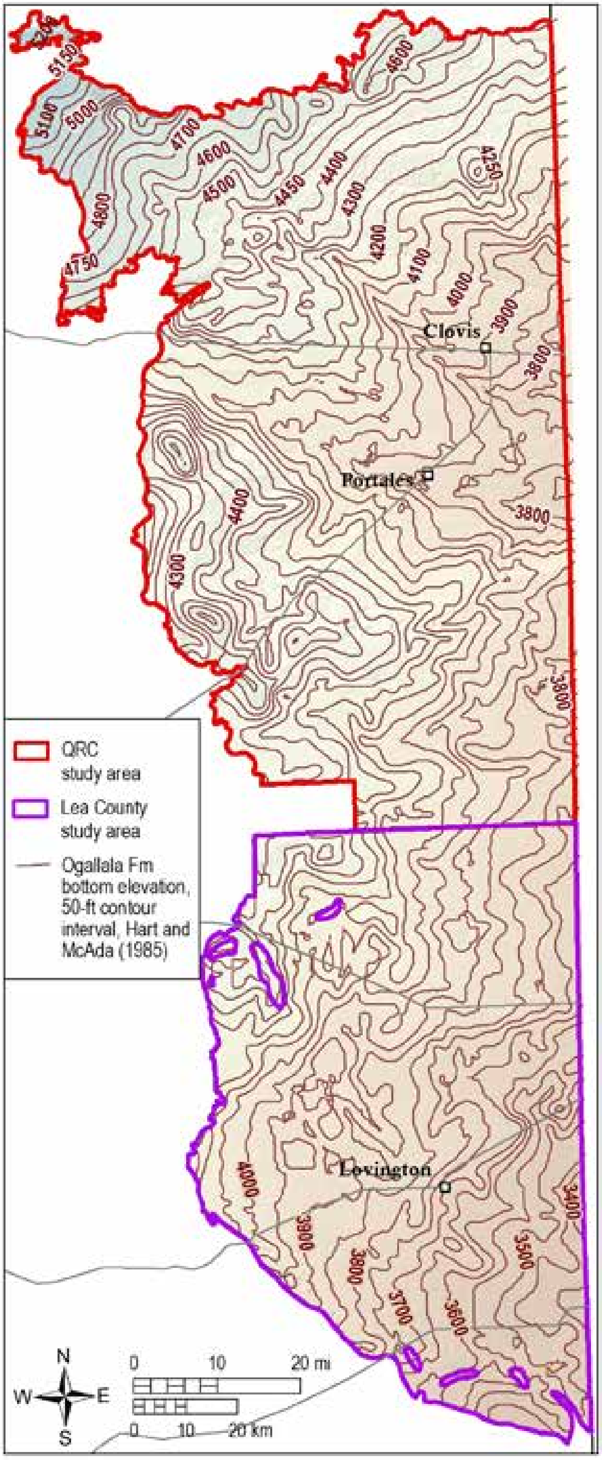

Figure 14. Map of the elevation of the bottom of the Ogallala aquifer (Hart and McAda, 1985).

Ogallala Formation and aquifer were thinner and less extensive than the further east in western Texas, Kansas and Nebraska.

Hydraulically, the Ogallala generally had very high hydraulic conductivities (10^{-2} m/s to 10^{-3}) and normal specific yields (0.10 to 0.25) for sand and gravel deposits (Hart and McAda, 1985). The highest conductivity parts of the Ogallala aquifer were the gravel deposits at the bottom and along the axis of paleovalleys (Weeks et al., 1988). Hydraulic conductivities decreased toward the edges of the paleovalleys (Hart and McAda, 1985). The high hydraulic conductivity of the Ogallala aquifer meant that large withdrawals at a single well can quickly draw water from far away (Batu, 1998). The cones of depressions for a single well likely overlapped with other wells.

As mentioned the lateral boundaries of the aquifer corresponded to historically unsaturated regions and the edges of the paleovalleys (Hart and McAda, 1985; and Weeks et al.,1988) . One of these boundaries existed between Clovis and Portales (Hart and McAda, 1985). Another boundary coincidentally fell on the northern Lea County boundary (Hart and McAda, 1985). This allowed us to justifiably separate the QRC study area from the Lea County study area.

There were no perennial streams in the New Mexican SHP. There were few relatively large dry washes that terminated in playas. Playas, or small closed basins that could be filled with water after large runoff events, were common throughout the regions. While there were one- to two-orders of magnitude greater recharge per unit area in these regions compared to the surrounding uplands, the playas formed a relatively small area of the entire SHP; recharge, on average, remained miniscule compared to the rates of pumping (see Rawling et al, 2016 for review).

II.3. Methods

The approach taken in this study built from but differed substantially from the methods of Rinehart et al. (2015 and 2016). In this study, we were able to consistently use groundwater table elevations rather than depth-to-water by using a more complicated workflow that, in regions where it was viable, produces more statistically justifiable results. First, however, the dataset was reviewed and pumping-affected data were removed.

This was done by reviewing the historical hydrograph for each well in the study area from the USGS groundwater level measurement dataset. In most cases, all irrigation season data, defined to be from March 1 to November 1, were removed. Instances of a single measurement point were also generally removed. Any measurement that was flagged as being affected by pumping, either recovering or having a well nearby being pumped or the well of interest being pumped, was always removed. We also removed wells completed more than 25 ft below the Hart and McAda (1985) 50 ft elevation contour intervals of the bottom of the Ogallala aquifer. However, regions of both the QRC area and the Lea County area had sparse data in the 2000s and 2010s. In order to get a more robust change estimates some wells with single measurements and, in the QRC region, some wells that may be partially completed out of the Ogallala Formation were allowed. The deeper-than-Ogallala-Formation completions were allowed recognizing the long screening intervals common in the region (Rawling and Rinehart, under review). Most, if not all, of the water in these wells was from the Ogallala aquifer. Also, some irrigation season measurements were included in those regions if they did not show the characteristic ‘saw-tooth’ decline-recovery curves characteristic of pumped wells.

After unreliable measurements were removed, a median depth-to-water (DTW) was estimated for each decade at each well. The decadal median DTW was converted to water table elevation by subtracting it from a 10-m DEM-derived elevation for each well. The aggregation of data to a 10-year interval was required in order to ensure enough well coverage for the region of interest, which was consistent with the work of Rinehart et al. (2015, 2016).

At this point, we departed significantly from the previous work. We performed all the work below in the R statistical software (R Core Team, 2016), using the base, *gstat* (Pebesma, 2004; and Graler et al., 2016), *raster* (Hijmans, 2016), *sp* (Pebesma and Bivand, 2005; and Bivand et al., 2013) and *rgdal* (Bivand et al., 2016) libraries; all software was open-source and available at no cost. In Rinehart et al. (2015, 2016), it was necessary to interpolate decadal median DTW, not water table elevations, because of the large topographic relief in the basins of interest. The DTW measurements were roughly normally distributed with a relatively small mean, but the large relief caused large variations in water table elevation that could not be fitted with simple surfaces or interpolated. Thus, we choose to interpolate decadal median DTWs using ordinary kriging and IDW rather than seeking a more complicated, but more theoretically grounded, approach. While the interpolation of median DTWs led to some artifacts along the edges of the basins, the use of DTW allowed us to make water level surfaces that were comparable as the well networks and DTWs changed through time.

In the New Mexican SHP, there was little topographic relief. This allowed us to use a more rigorous interpolation approach. Given the severity of water shortages anecdotally reported in the SHP, we felt urgency in estimating saturated thickness, which required water table elevations, rather than simply calculating water level changes. Also, an approach involving removing regional trends then interpolating the normally distributed residuals was well more theoretically justified (Cressie and Wikle, 2011).
The new approach to estimate water table elevations for each decade was as follows and was conceptually demonstrated in Figure 15 in one dimension (easting or northing). First, a third-order two-dimensional (easting and northing) polynomial was fit to median decadal water elevations (Fig. 15a, black dots and line). The order of polynomial was found by trial and error by examining plots of the water elevations. Then, the polynomial trend was removed from the water table elevation data at each well, with a

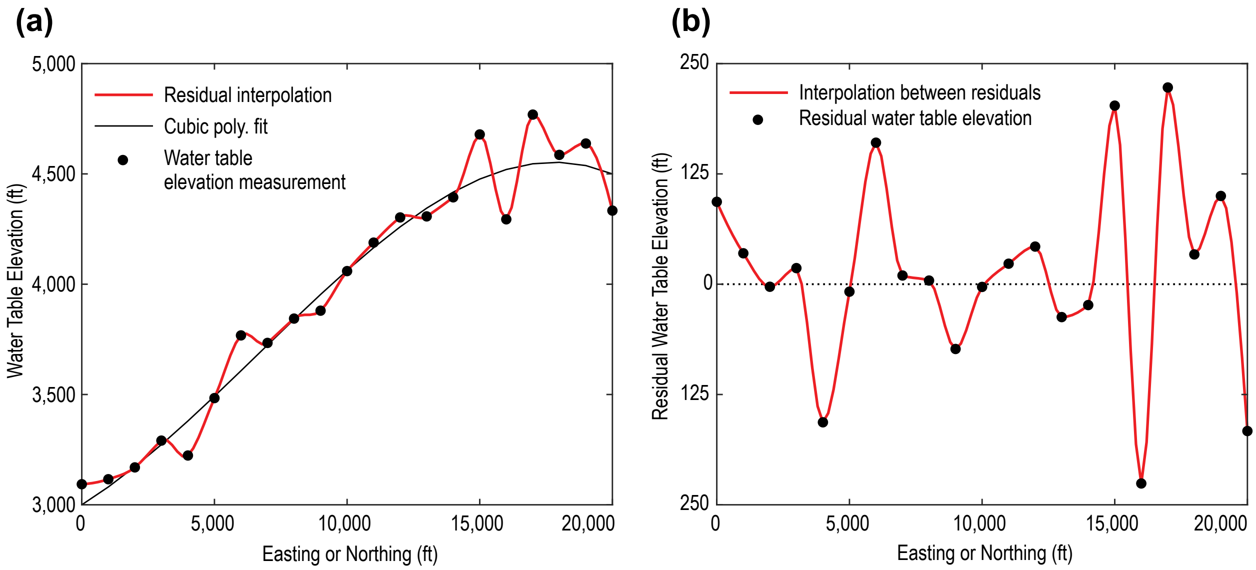

Figure 15. Conceptual image of one-dimensional regression-interpolation process to generate water elevations as a function of easting or northing, with (a) water table elevation elevations (black dots), a third order polynomial least-squares fit of the measured water table elevations (black line), and the sum of the interpolation of the residual water table elevations and the polynomial fitted to water table elevations (red line); and (b) the residual water table elevations (black dots) and interpolation of residual water table elevations (red line).

residual elevation calculated at each well (Fig. 15b, black dots). There was still spatial correlation among the residual elevations, and the residual elevations were normally distributed, which was theoretically required for ordinary kriging to be valid. The residual elevations were interpolated (Fig. 15b, red line) using ordinary kriging (Isaaks and Srivastava, 1989) after fitting the residual elevation variogram to spherical, circular or, in the QRC area, Gaussian variogram functions (Rawling and Rinehart, in press). The interpolated residual elevations were then added to the regional water table elevation trend function (Fig. 15a, black line) to find an estimated water table elevation for the study area (Fig. 15a, red line).

As part of the last step, we also used the plots of the variograms from each decade to visually estimate a general range of correlation for the residual water table elevations in each region (Bivand et al., 2013). This was done so that we could restrict the final storage and storage change calculations to a statistically reasonable area (Rinehart et al., 2015 and 2016).

The level of confidence we had in these water table elevation surfaces was captured in large part by an additional leave-one-out cross-validation (LOO-CV) analysis of the kriged surfaces (Bivand et al., 2013). Essentially, a new residual elevation surfaces were constructed leaving one well out, and then the difference between the measured residual elevation and the new surface was computed (Bivand et al., 2013). This was repeated for every well in the data set (Bivand et al., 2013). Higher values of found using LOO-CVs correspond to a larger change, and thus a greater sensitivity to the well network (Bivand et al., 2013). This, in turn, shows areas where we have less certainty in the interpolation—a single well should not dominate the interpolation.

In Figure 15, this could be visualized by imagining what would happen to the interpolated line (red) when one of the measurement points (black dots) was removed. Consider, for example, the fifth point from the left. If this point was removed, then the interpolated line would move by over 125 ft upward, introducing a large error. In areas with low sampling density (few wells), a high LOO-CV value for a well implied a strong dependence on that well for an accurate water level and water level change estimate—if you removed the well, then that entire region changes value. In areas with strong spatial correlation or closely spaced wells, then small LOO-CV values were expected. Even if there was a high LOO-CV value among closely spaced wells, the net effect would be smaller; the region was too small to be significant.

Once the total water table elevation surface was constructed for each decade, we used the Ogallala Formation bottom elevations from Hart and McAda (1985) to estimate a saturated thickness for the region and subsequently compute the changes in saturated thickness. The water table elevation maps and saturated thickness maps were restricted to areas that (1) had a water table elevation above the bottom of the Ogallala Formation; (2) were not in the historically unsaturated regions or outside of the Ogallala Formation; and (3) were within the variogram-based correlation radius from the nearest well. To compute the change in saturated thickness, the intersection of the areas meeting the three conditions for each decade above was used. We computed what Rinehart et al. (2015, 2016) called the moving and cumulative differences, where the change between one decade and its preceding decade were found.

Finally, maps of change in storage through time was found by multiplying the rasters of saturated thickness change by Gutentag et al.(1984)'s map of specific yield for the Ogallala aquifer and then the total volume change was found by summing the map of storage change.

The difference calculations were done for each decade from the 1950s through the 2010s for the QRC area. In the Lea County area, the 1950s did not have enough data to serve as a base; the analysis began in the 1960s.

II.4. Results

II.4.1. QRC area

The QRC area had been the focus of a detailed study by Rawling and Rinehart (in press), and the presented results built from that study. An average variogram range of 30 km was determined in this region, and each water table elevation surface was restricted to the historically saturated region, the region currently above the bottom of the Ogallala aquifer and the area within the correlation distance.

Figure 16 showed the full interpolation and polynomial fitted water table elevations for the 1950s through the 2010s. Overlain on the water table elevations were the absolute values of the LOO-CV analysis. Recall that regions with high residual magnitudes were areas sensitive to changes in the well network.

The water table in the QRC region sloped gradually to the southeast, from maximum elevations of a little over 5,250 ft amsl down to elevations just below 4,000 ft amsl (Fig. 16). Because of the broad topographic trend and large region, it was difficult to see changes between the surfaces. However, it was clear from the LOO-CV results that the regions with closely spaced well networks had the least sensitivity to well network effects, that the well networks in the 1970s, 1980s and 1990s were the best sampled and provide the most robust comparisons (Fig. 16b-e), and that sparseness of the measurements in the 2000s and 2010s increased the uncertainty of storage change estimates in these decades (Fig 16f,g) . Additionally, the uncertainty due to well network effects in the 2000s and 2010s was greater than the uncertainty of the regional contours of the bottom of the Ogallala Formation (±25 ft; Fig. 14). This was clear by looking at the LOO-CV residuals in Fig. 16. For these later decades, the well network uncertainty was large even in regions that have a dense well network and previously had smaller LOO-CV residuals. In earlier years (Fig. 16a-d), the vast majority of wells had uncertainties less than 25 ft, so, even if the well measurement was removed, the introduced error would have been on order with the uncertainty in the formation bottom elevation. In the 2000s and 2010s, a higher proportion of wells had greater uncertainty. These 'uncertain' measurements occurred in regions that previously had small residuals (Fig. 16). This implied that small changes in the well network in later decades could have had a disproportionate effect on the final results.

Figure 17 showed the remaining saturated thickness for each decade, calculated by taking the difference between the estimated water table elevation surface and the bottom elevation of the Ogallala Formation of Hart and McAda (1985). We highlighted areas with less than 30 feet of remaining saturated thickness, because Hecox et al. (2002) found that high production rate irrigation pumps generally required

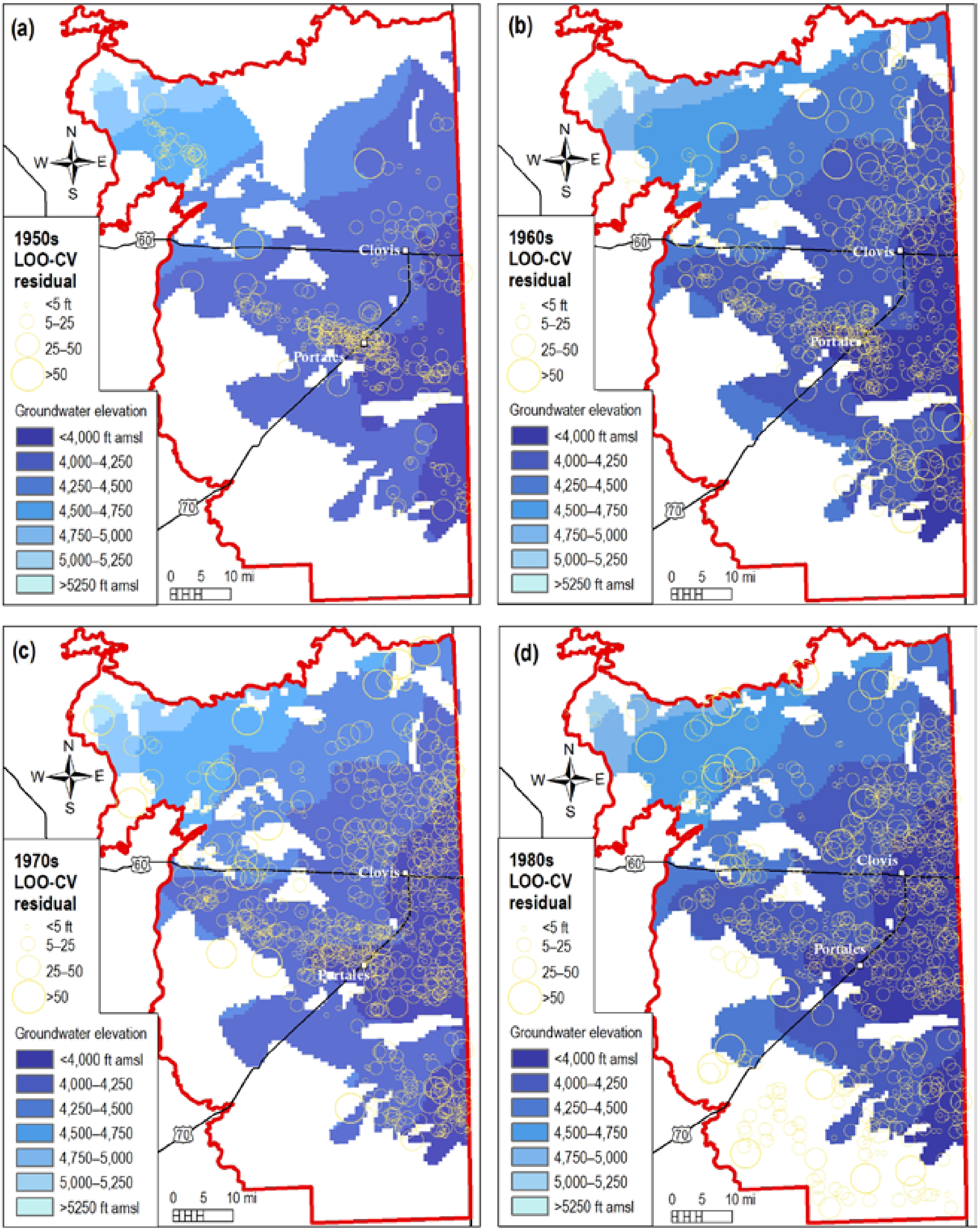

Figure 16. Estimated median water table elevation surface in the QRC area fitted by a third order polynomial with elevation residuals interpolated by ordinary kriging for the (a) 1950s, (b) 1960s, (c) 1970s, (d) 1980s, (e) 1990s, (f) 2000s, and (g) 2010s, overlain by the absolute value of the leave-one-out cross-validation (LOO-CV) residuals of the ordinary kriging-based elevation residual surface for every water level measurement location used in the analysis (yellow circles, scaled by absolute LOO-CV residual).

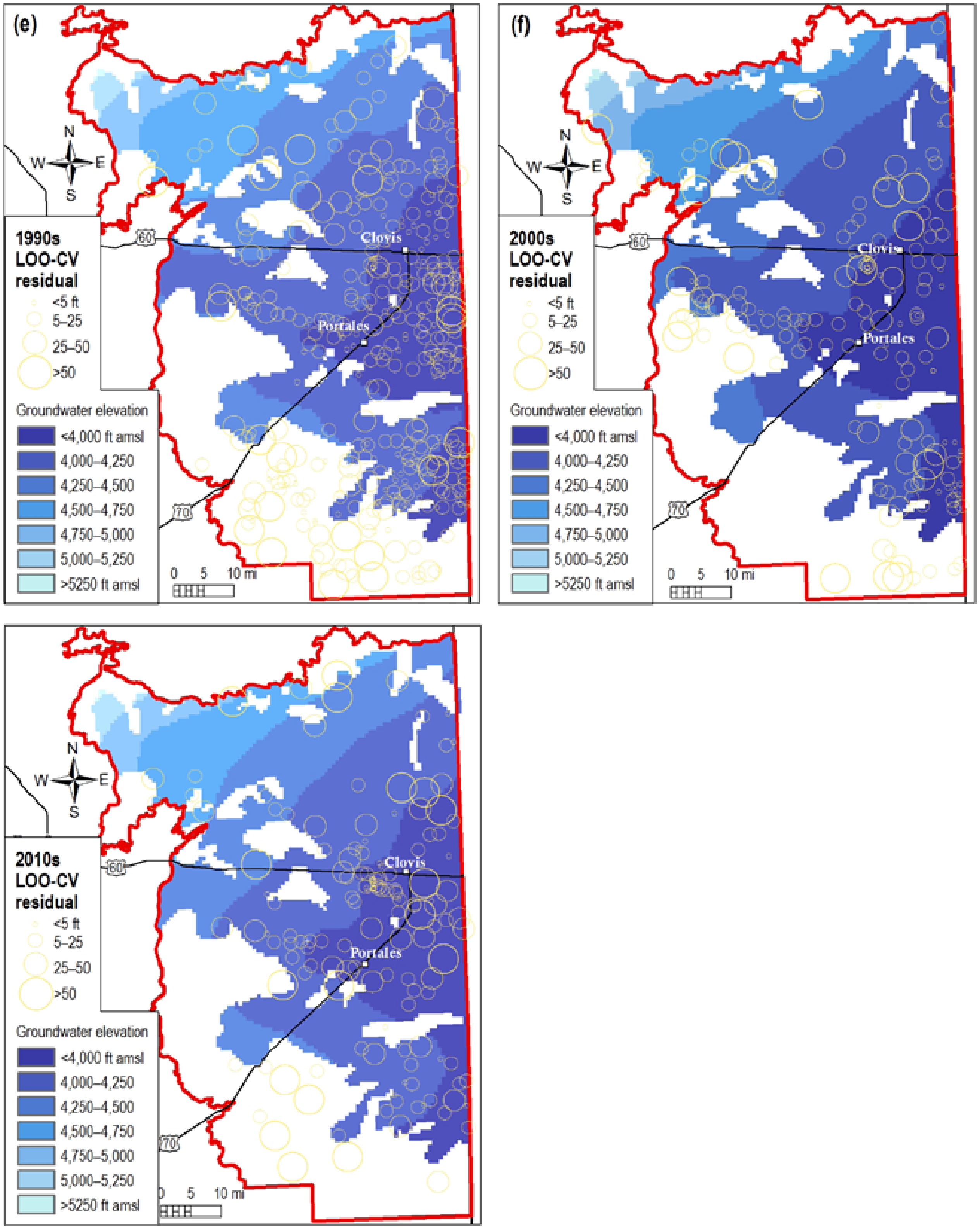

Figure 16. Continued.

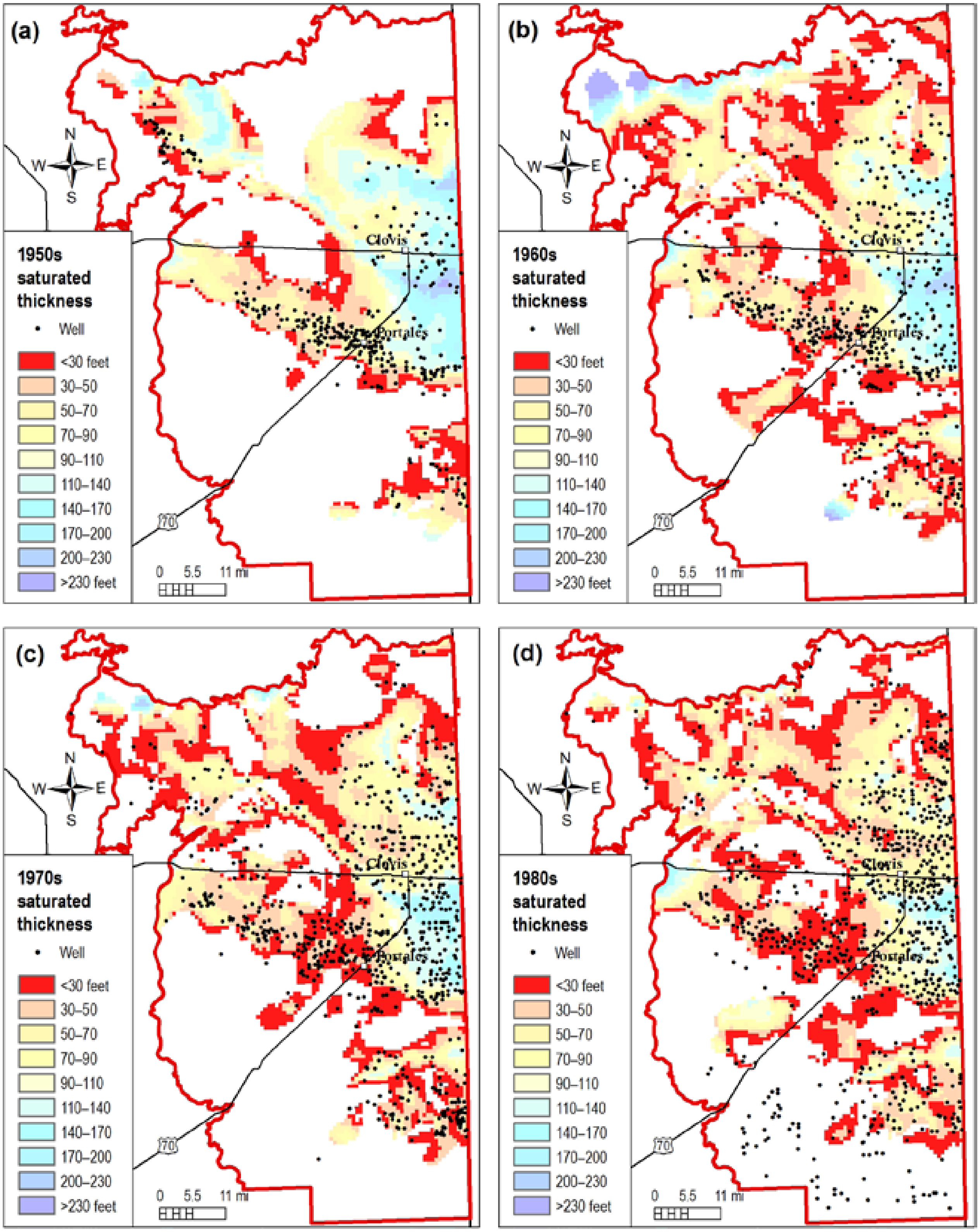

Figure 17. Maps of median saturated thickness in the QRC area for the (a) 1950s, (b) 1960s, (c) 1970s, (d) 1980s, (e) 1990s, (f) 2000s, and (g) 2010s, with thickness of less than 30 ft highlighted in red and the well network for the decade overlain.

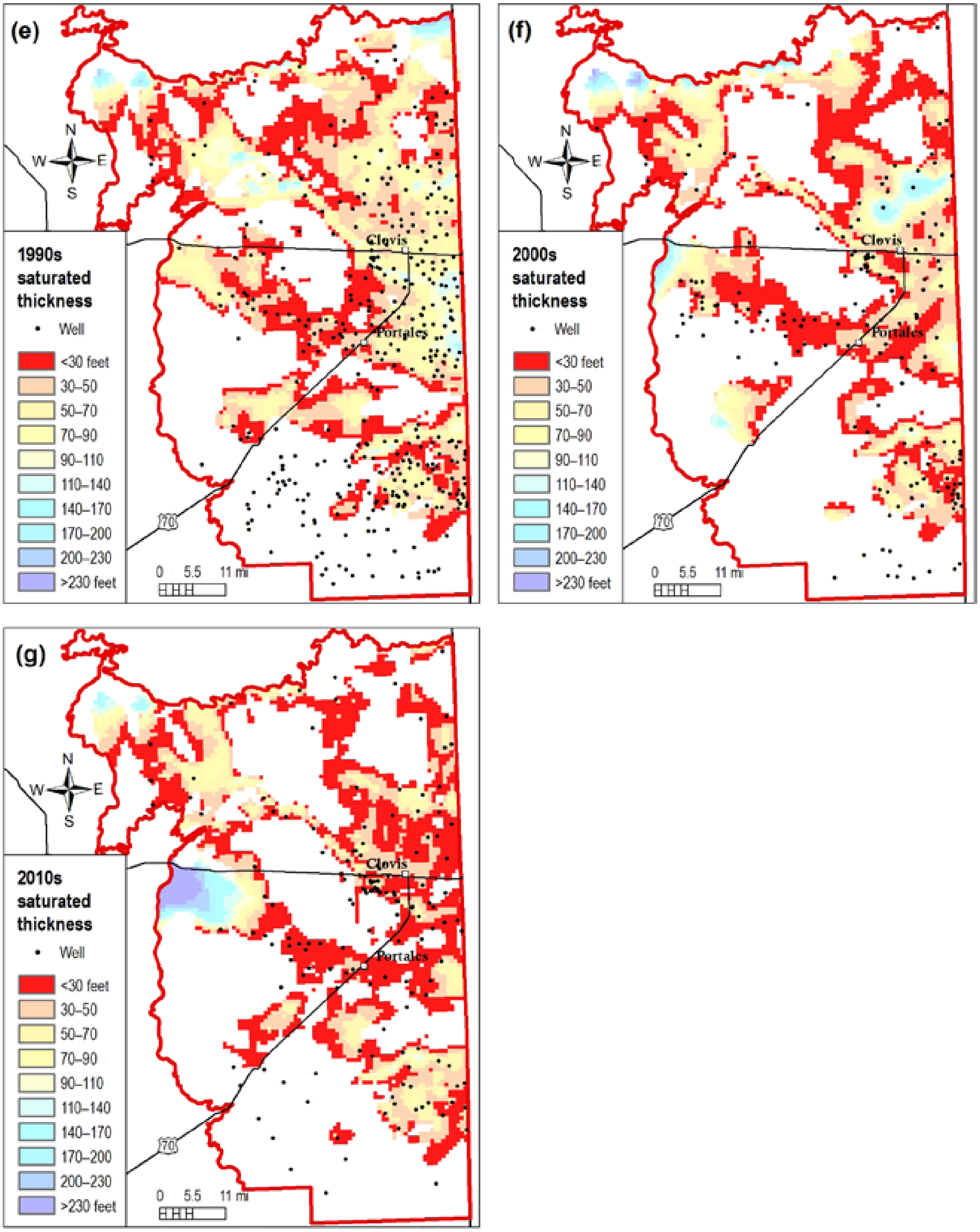

Figure 17. Continued.

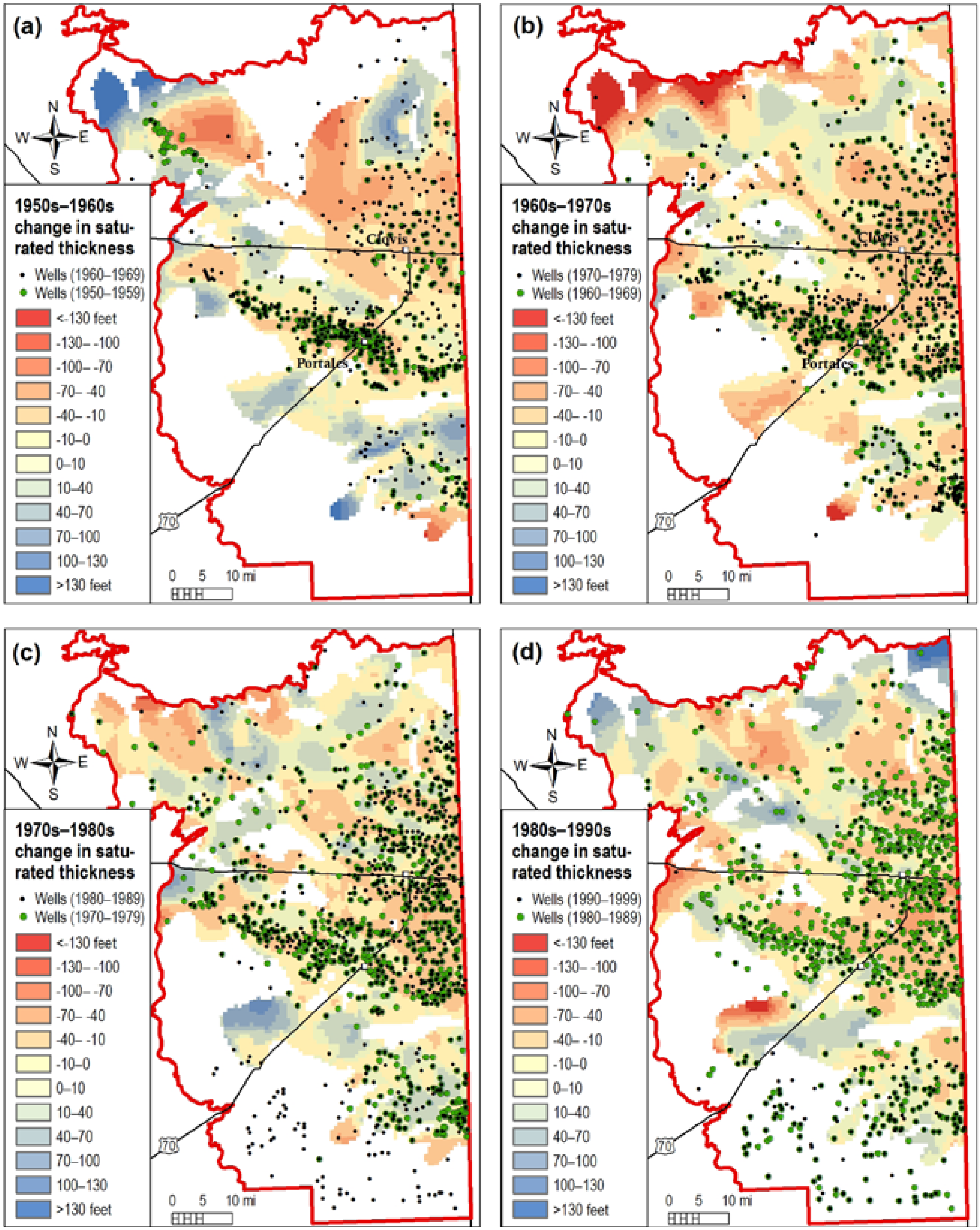

Figure 18. Maps of changes in saturated thickness in the QRC area (a) from the 1950s to the 1960s, (b) from the 1960s to the 1970s, (c) from the 1970s to the 1980s, (d) from the 1990s to the 2000s, and (e) from the 2000s to the 2010.

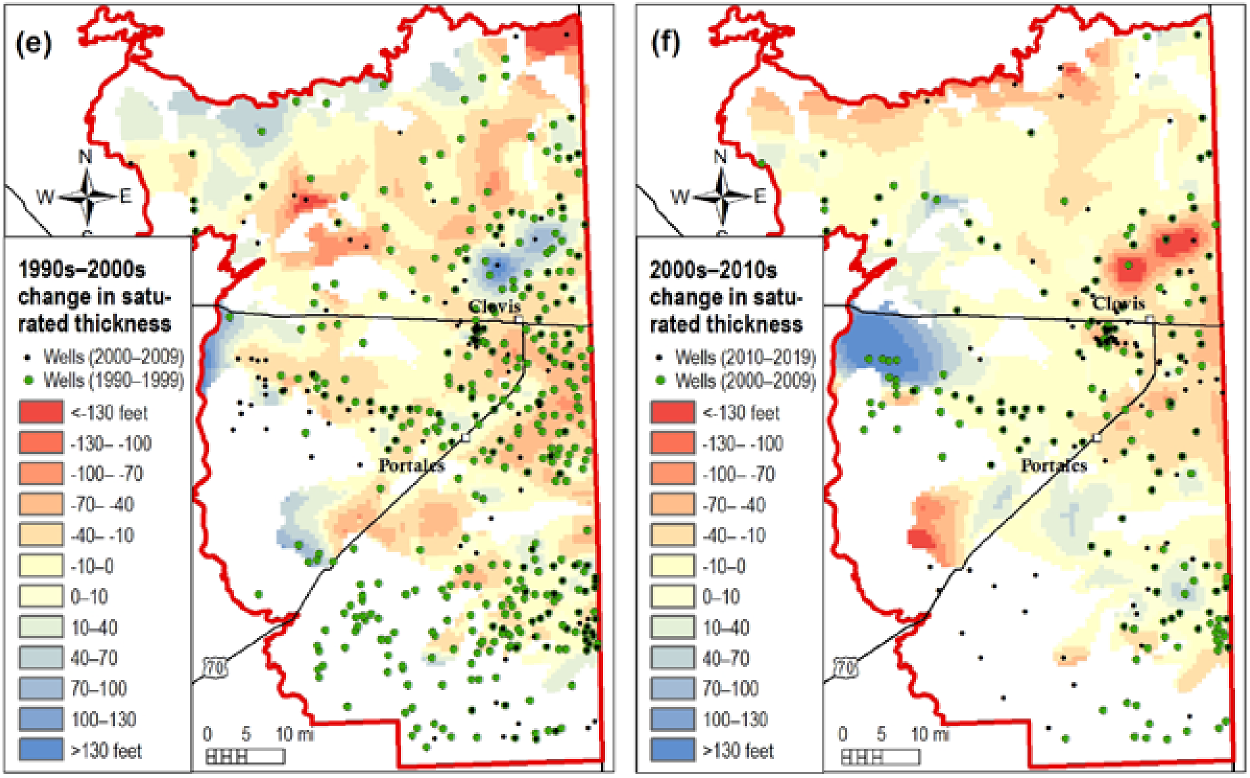

Figure 18. Continued.

a minimum of 30 ft of saturated thickness in the Ogallala Formation. Less than 30 ft of saturated thickness indicated that current irrigation and water production practices were likely problematic to implement.

From the 1950s through the 2010s, there was a systematic decrease in saturated thickness through much of the basin, and the area with water table elevations below or within 30 ft of the bottom of the Ogallala Formation grows through time. This was most clearly seen in the decades with the best well network coverage (1960s-1990s). Regions with poor well coverage or large LOO-CV residuals showed artificial increases in thickness resulting from unreliable interpolations and functional fits away from the well network.

Figure 18 showed maps of saturated thickness changes from decade to decade for the QRC region. The region had decreased saturated thickness from the 1950s until the 2010s that caused a large portion of the region to become unviable for high-intensity pumping. Areas with apparent increases were in regions either with large LOO cross-validation residuals or lacking well coverage. The most reliably sampled regions almost uniformly showed decreasing saturated thickness.

II.4.2. Lea County area

The Lea County area was analyzed with the workflow outlined above and that was used for the QRC area in Rawling and Rinehart (in press). Surprisingly, an average variogram range of 10 km was found in the Lea County area, strongly contrasting with the 30 km range in the QRC region. There was not an obvious reason for this. Possibilities included less uniform pumping practices, lower transmissivity, or simply less pumping. As in the analysis for the QRC region, maps of water table elevation and LOO-CV residuals (Figure 19), saturated thickness (Figure 20) and changes in saturated thickness (Figure 21) were made for all valid decades. Unlike the QRC region, the Lea County region did not have sufficient well coverage in the 1950s to perform an analysis consistent with the later decades. The resulting grids had been restricted to (a) the historically saturated regions, (b) where the estimated water table was above the bottom of the Ogallala aquifer, and (c) where the estimate was within the 10 km range of the nearest well.

Figure 18 showed the water table elevation raster and the LOO-CV residuals for the Lea County area. Similar to the QRC region, the Lea County area showed a broad southeast groundwater elevation gradient. Also similar to the QRC region, the well measurement density was highest for the 1960s, 1970s, 1980s and 1990s (Fig. 19a-d). The well measurements in the 2000s and 2010s (Fig. 19e,f) were sparse compared to earlier measurement networks. In general, the LOO-CV residuals are low magnitude compared to the uncertainty of the Ogallala Formation bottom elevation, and they are fairly uniformly distributed in space. The regions with the highest LOO-CV residuals were around the edge of the area (Fig. 19). However, the 2000s and especially the 2010s showed higher residuals than earlier decades, indicating greater sensitivity to the details of the well network.

Figure 20 showed the estimated saturated thickness for the Lea County area from the 1960s through the 2010s. Once again, regions with less than 30 ft remaining saturated thickness were highlighted in red. The broad pattern in saturated thickness was consistent with the aquifer geometry and groundwater use: the saturated thickness was thinner on the edges of the valley and where development was greatest around Lovington, NM (Fig. 20). In general, saturated thicknesses were thinning through time. However, in the 1990s, 2000s and 2010s, significant well network-related artifacts in the saturated thickness occurred in the western part of the aquifer (Fig. 20d-f) and along the northeastern boundary of the aquifer (Fig. 20e,f). Here, the saturated thickness increased through time solely due to a large variogram range and a loss of well coverage. The increases were artifacts of the method.

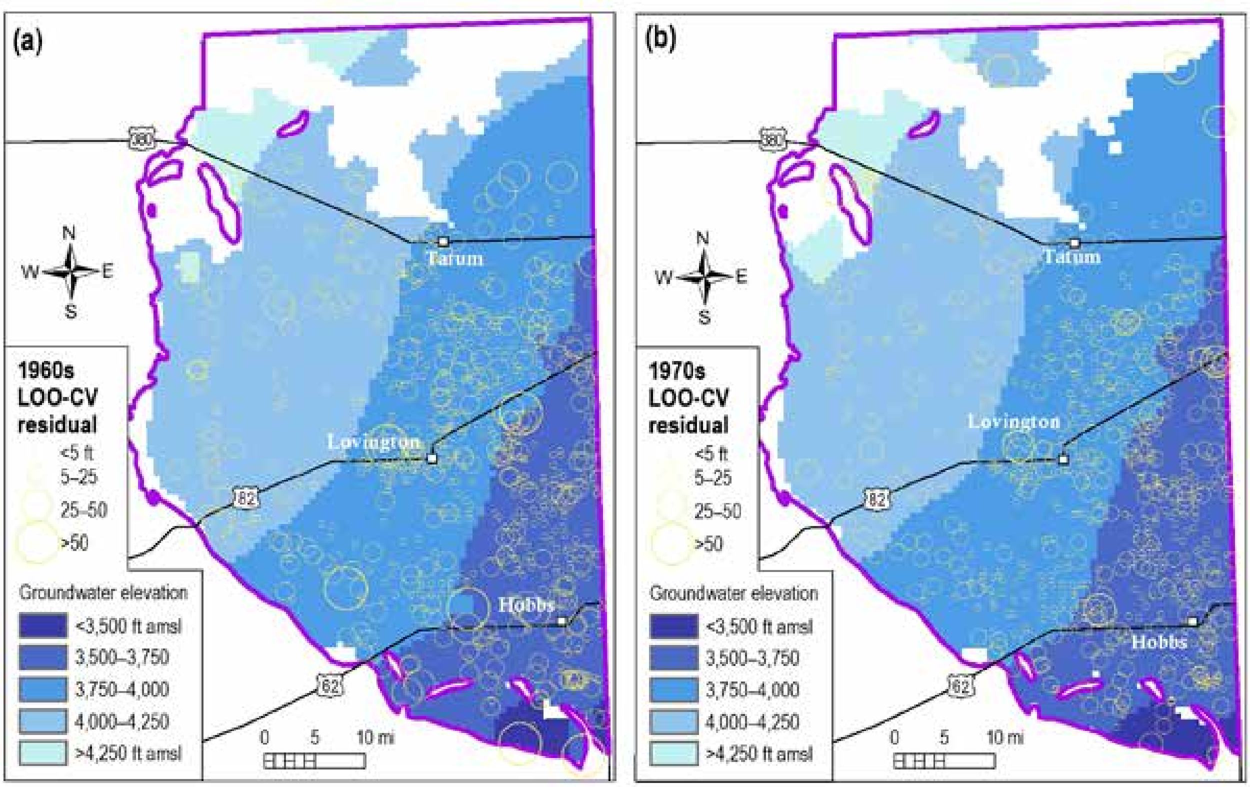

Figure 19. Water table elevations (shaded background) and leave-one-out cross-validation (LOO-CV) residuals (circles) in Lea County for (a) the 1960s, (b) the 1970s, (c) the 1980s, (d) the 1990s, (e) the 2000s, and (f) the 2010s.

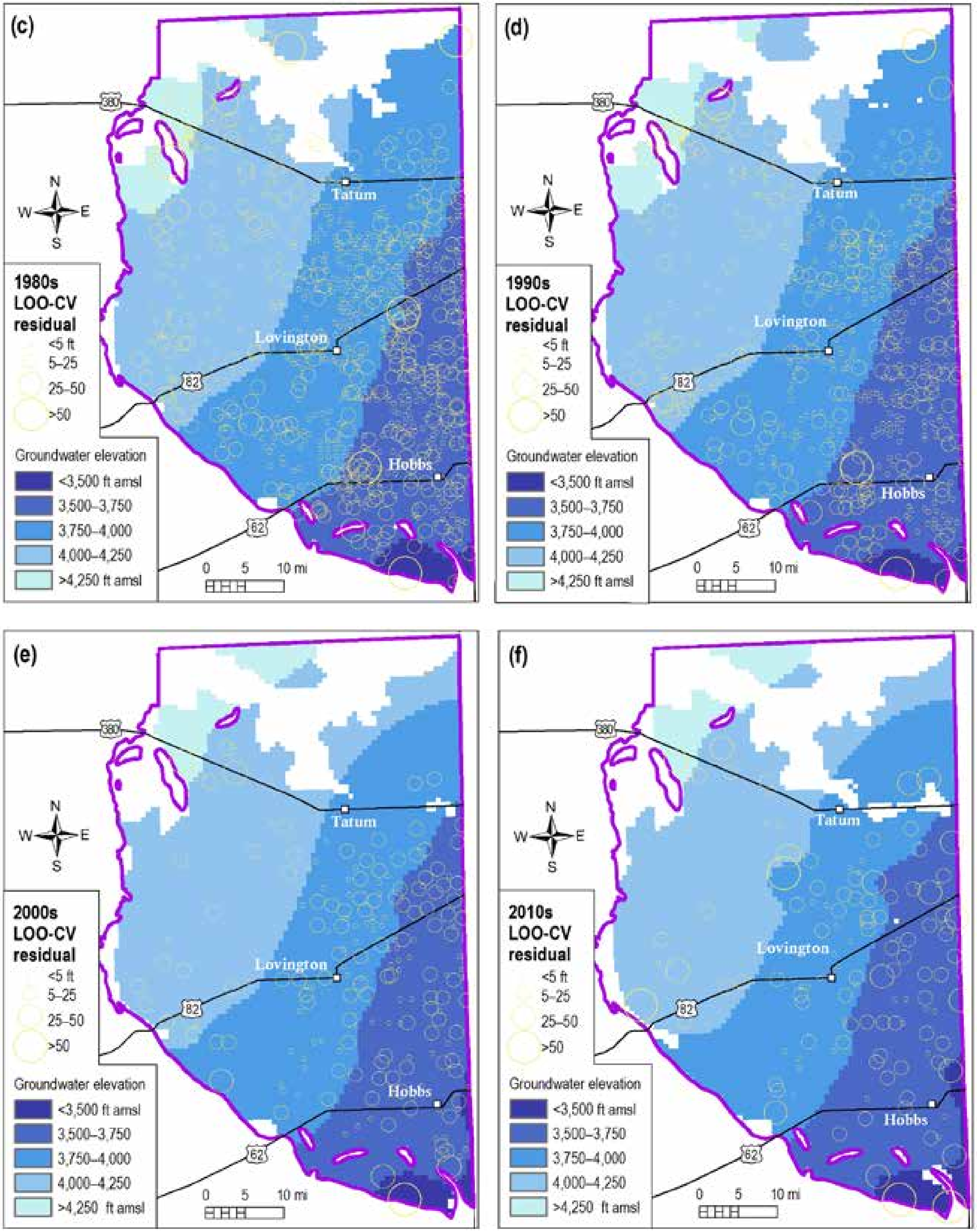

Figure 19. Continued.

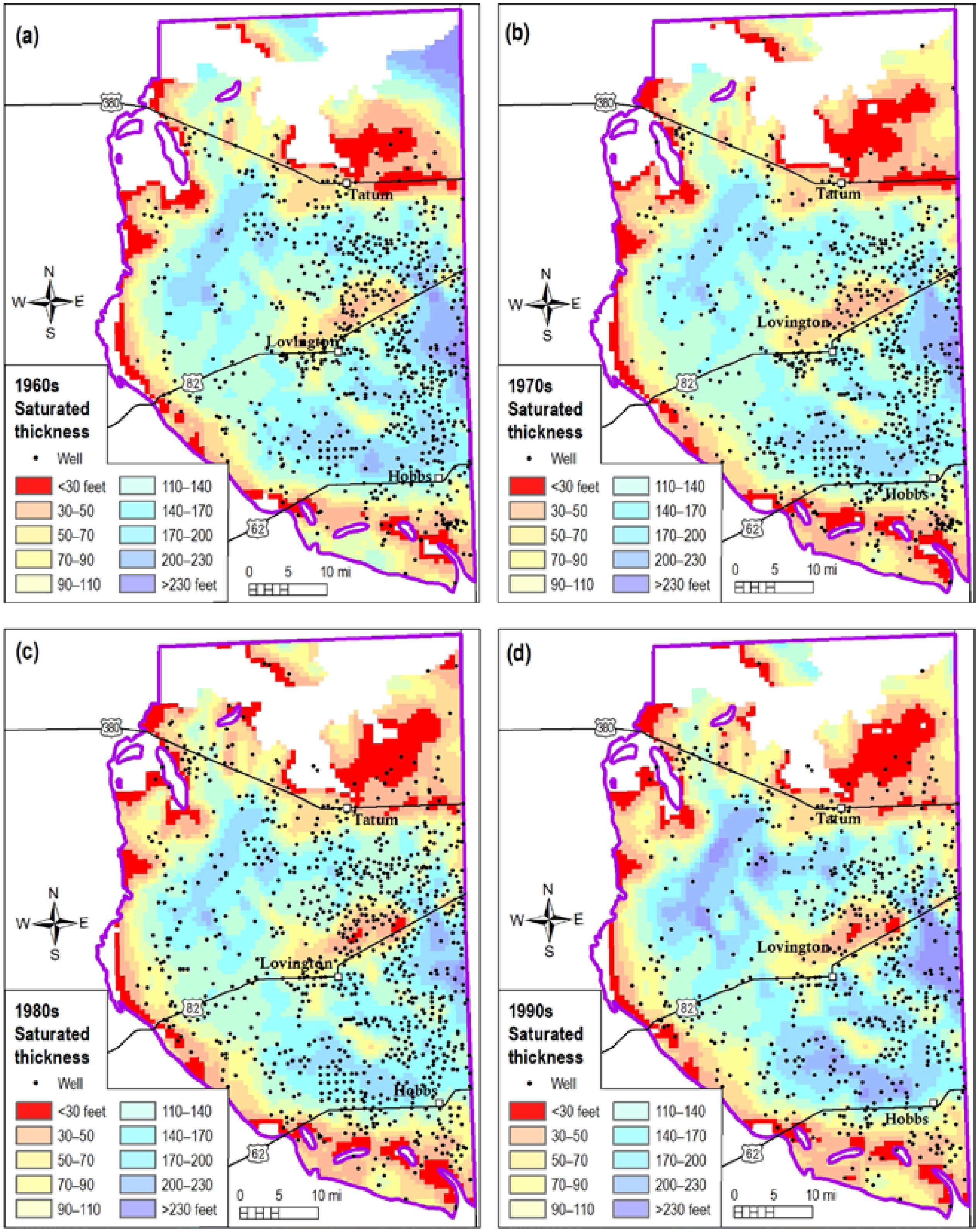

Figure 20. Decadal median saturated thicknesses and wells used to construct water table surfaces for the Lea County area for (a) the 1960s, (b) the 1970s, (c) the 1980s, (d) the 1990s, (e) the 2000s, and (f) the 2010s.

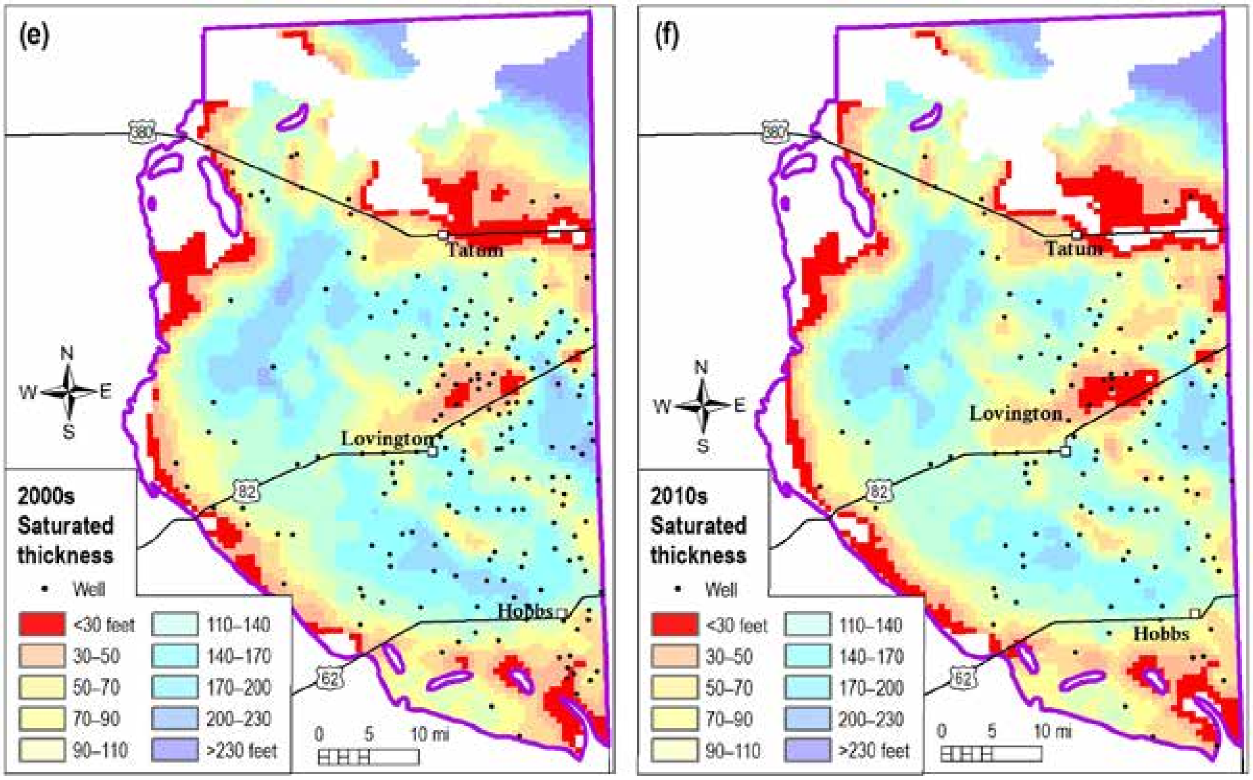

Figure 20. Continued.

Figure 21 showed the decade-by-decade changes in saturated thickness in the Lea County area, beginning with the change from the 1960s to the 1970s (Fig. 21a), and finishing with the changes from the 2000s to the 2010s (Fig. 21e). As mentioned above in the discussion of the saturated thickness results, there was a general trend of declining saturated thicknesses. These declines were especially apparent in the eastern and central parts of the area. Apparent increases in saturated thickness were seen along the northern boundary (Fig. 21d,e), in the south (Fig. 21d) and in the western part of the area (Fig. 21d,e). These increases corresponded exactly both with a decrease in well density (Fig. 21) and with a region wide increase in well network sensitivity seen in the cross-validation results (Fig. 19e,f).

II.4.3. Cumulative storage changes

Last, we presented the aggregated groundwater storage changes (Fig. 22) for both the QRC area (Fig. 22a) and for the Lea County area (Fig. 22b). These plots showed the decade-by-decade storage volume changes and the cumulative volume of the decade-by-decade changes. As mentioned in the methods, the storage change at every point was calculated by multiplying the change in saturated thickness by the specific yield at that location as defined in Gutentag et al. (1984). The regional volume change was then made by summing all of the volume changes for the region. The QRC area showed steady declines from the 1950s until the 2010s, with an approximately 8 Maf (million acre-feet) decline in storage (Fig. 22a). The interpretation of the Lea County area was complicated by the effects of the changing well networks between the 1980s and the 1990s, and the 1990s to the 2000s; the declines were compensated for by artificial areas of increase caused by losses in the well network. Nonetheless, the Lea County area showed significant declines of at least 3 Maf (Fig. 22b). The calculated storage changes were also found in Table 2.

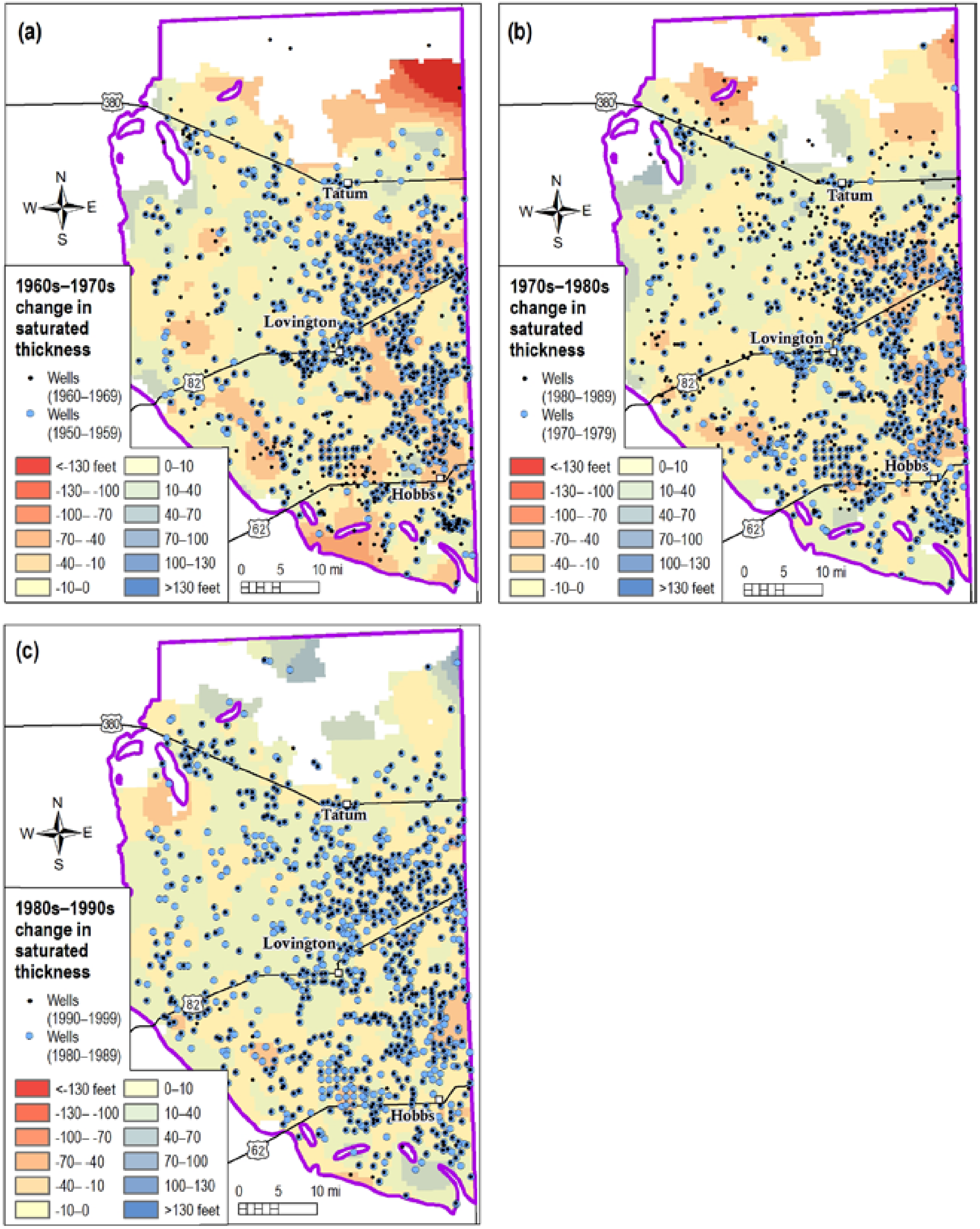

Figure 21. Decade-by-decade in median saturated thickness in the Lea County area for the intervals (a) the 1960s to the 1970s, (b) the 1970s to the 1980s, (c) the 1980s to the 1990s, (d), the 1990s to the 2000s, and (e) the 2000s to the 2010s.

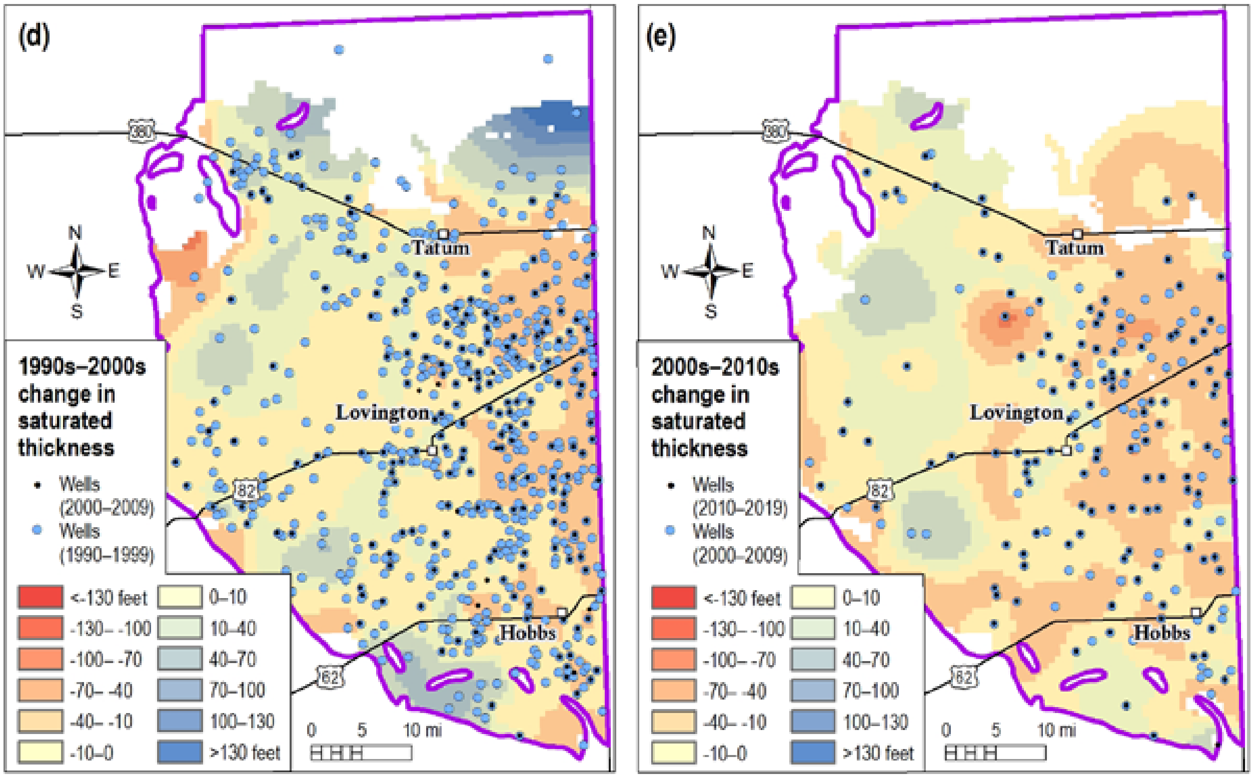

Figure 21. Continued.

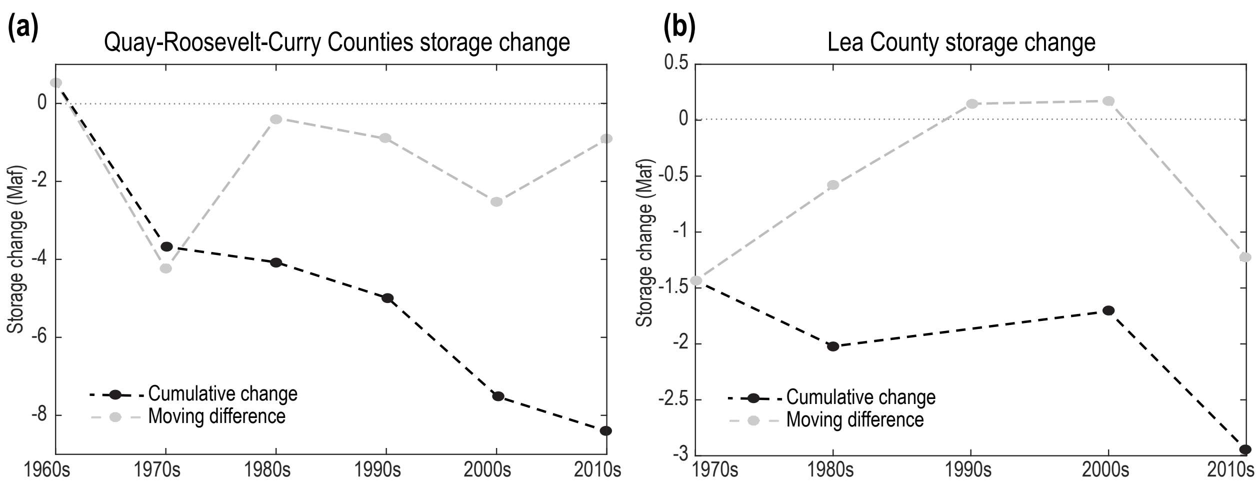

Figure 22. Total volumes of storage change for (a) the QRC area and (b) the Lea County area, with the decade-by-decade storage change (dashed grey line) and cumulative decade-by-decade storage change (dashed black line).

Table 2. Summary of calculated storage changes in SHP aquifer.

STORAGE CHANGE (Maf)				
Ending decade	**QRC area**		**Lea County area**	
	Decade-by-decade	Cumulative	Decade-by-decade	Cumulative
1960	0.550	0.550	-	-
1970	-4.250	-3.700	-1.434	-1.434
1980	-0.374	-4.074	-0.587	-2.022
1990	-0.907	-4.981	0.141	-1.88
2000	-2.532	-7.513	0.172	-1.708
2010	-0.918	-8.431	-1.235	-2.944

II.5. Discussion

The results of this study showed rapidly declining groundwater storage in the QRC area and more slowly declining groundwater storage in the Lea County area; in both regions, however, groundwater storage was declining.

In the QRC area, large swathes of the area were projected to no longer have enough saturated thickness to be viable for high volume irrigation. We estimated a little over 8 Maf of groundwater storage losses (Table 2). The storage declines had been steady through time. The regions where there were apparent increases were small and were directly correlated with decreasing well measurements and increased LOO-CV residuals. The areas where we have the largest declines were in the areas that had the greatest certainty.

In the Lea County area, the effects of the declines had been less stark, in large part due to the thicker initial saturated thickness as compared to the QRC area. The effect of changes in well networks was larger in Lea County than in the QRC area, with artificial but large positive storage changes occurring solely because of interpolation errors from well coverage losses. Throughout the Lea County area, the last two decades were substantially more sensitive to the location and measurements of individual wells (Fig. 21d,e) than in the QRC area (Fig. 18e,f).

There were a few previous studies focused on the total groundwater storage changes across the Southern High Plains in New Mexico. Specifically, McGuire (2009) and McGuire (2014) made estimates of the groundwater storage change for the SHP aquifers of New Mexico. Their estimates were roughly 15 Maf of groundwater storage declines since predevelopment (1930s; McGuire, 2009 and 2014). Our estimates using the decade by decade changes from the 1950s and the 1960s for the QRC area and Lea County came in at a total of 11.4 Maf of storage decline. This was a significant underestimate. The most likely cause is a lack of reliability in well networks from the 1980s to the 2000s in Lea County. These caused the losses to be balanced by artificial gains. Another possible cause was from large losses before the 1950s. Finally, McGuire (2009 and 2014) were based on far fewer wells, often only a handful in each region. These wells were commonly in areas with the highest depletions and correspondingly most intense land-use or highest population. It was unclear how the boundaries of the estimate were made in McGuire (2009 and 2014), but it was likely that his estimates were biased toward larger depletions, while our method provided a conservative (less change) estimate (Rinehart et al., 2015).

In Lea County, Longworth et al. (2013) and Longworth et al. (2011) estimated a groundwater withdrawal of 0.195 Maf/yr in 2010 in Lea County and a groundwater withdrawal of 0.186 Maf/yr in 2005. Correcting for return flows, which historically had been estimated at between 30 kaf and 50 ka (Wilson et al., 2003), this means that there were storage declines of approximately 1.4 Maf between 2000 and 2010, remarkably close to our decade by decade estimate (Table 1). This assumes that the estimates from individual years can be extrapolated through a decade. Our estimates for earlier decades in Lea County, however, were much less than estimated by individual years in the 1980s and 1990s by the NMOSE. We suspect that the large changes that occurred in well measurement locations and density that happened in the 1990s largely explained the disparities in Lea County between our study and the NMOSE estimates. It was important to note, however, that the individual year estimates by the NMOSE require extensive assumptions, some of which may be poorly justified, and on extrapolation from municipalities and evapotranspiration rates.

As an aside, the Lea County calculations were made with a spatially varying specific yield (~0.125) that was less than the average specific yield for the Ogallala Formation as a whole (0.2–0.25). The only area with a high specific yield, just north of Lovington, NM, had seen some of the most dramatic and consistent water level declines in the region. Preliminary estimates of storage change in Lea County were off by a factor of two, in large part because of the impact of spatially varying specific yield on the final calculations.

A similar story arises in the QRC region. Here, we found decadal declines in storage of between 0.37 Maf and 4.2 Maf, summing to just over 8 Maf over six decades. In the QRC region, Longworth et al. (2011 and 2013) and Wilson et al. (1992, 1997, and 2003) showed an extrapolated total withdrawal of 11 Maf since 1990, assuming that estimates for the years of 1990, 1995, 2000, 2005 and 2010 could be extrapolated over five years. Our analysis showed a storage decline closer to 5 Maf. This was a large discrepancy. Once again, some of this was likely related to changing well networks and incomplete coverage, some of it was related to uncertainties in the NMOSE estimates of withdrawals, and some of it was related to the difference between withdrawals and depletions. Other complications with the NMOSE estimates included not accounting for return flows from irrigation, observed in Rawling (2016). In fact, it was difficult to use the one-year estimates of NMOSE to compare to our estimate. They had been, however, one of the only sources of this type of information throughout New Mexico.

However, our calculations were internally consistent and were not based on assumptions of water use; rather, they were based on actual water level measurements and well characterized hydraulic parameters. Additionally, they were compatible and methodologically comparable with the other estimates done around the state. This compatibility was one of the primary goals for this study. Our analysis was also designed to produce conservative, or low, estimates of change by excluding poor data, and by only constructing water level surfaces using the water measurements available in a decade (not biasing areas with low data using previous decade water level estimates).

In the maps of saturated thickness changes (Fig. 18 and 21), both areas were showing a story of declining groundwater storage, though the total changes were, for some decades, more ambiguous. This was unsurprising given both areas complete reliance on groundwater for irrigation and other consumptive uses, and the lack of recharge in the SHP aquifer systems. It was mirrored across the entire SHP and Central High Plains aquifer system. Regionally, the only regions not showing steady and, in places, accelerating declines were in regions in eastern Kansas, Colorado and Nebraska where there was enough recharge to offset pumping. This was not the case in New Mexico. Here, both in the QRC area and in the Lea County area, pumping far outweighed recharge.

The declining water levels posed a serious challenge to the supporting irrigated agriculture and other industries reliant on the New Mexican SHP aquifer. The aquifer would eventually be depleted unless water use patterns change.

II.6. Conclusion

- The SHP aquifer system in both the QRC area and the Lea County area was declining rapidly in saturated thickness.
- We adapted the method from previous work to more accurately estimate decadal median water table elevations. However, this does not compensate for water level measurement network losses.
- The new method involved detailed data review, polynomial regression of the water table elevation measurements, interpolation of the residual elevations using ordinary kriging, summing the fitted trend surface and the kriged interpolation surface to derive the water table elevation surface. This was followed by detailed LOO-CV analysis.
- Rather than assuming a deep aquifer, as done in Rinehart et al. (2015,2016), we calculated saturated thickness based on the Ogallala Formation bottom elevations of Hart and McAda (1985).
- Large areas of the QRC area had saturated thickness of less than the 30 ft required for high volume pumping.

- Changes in Lea County had been less relative to the QRC, in large part due to the much thicker starting saturated thickness in Lea County than in QRC.
- Areas showing increases were almost always artifacts of decreasing or changing well network coverage.

It was vital for the users of the SHP to carefully plan for their remaining limited water resources. The purpose of this part of the 2016–2017 groundwater storage change portion of the Statewide Water Assessment was to provide estimates of the changing groundwater storage change in the SHP aquifers. The estimates were clear: the resource is limited and dwindling. Similarly stark maps as in the QRC area would eventually be seen the Lea County area unless groundwater use was changed.

II.7. Acknowledgments

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New Mexico Bureau of Geology and Mineral Resources

A division of New Mexico Institute of Mining and Technology

Socorro, NM 87801
(575) 835 5490
Fax (575) 835 6333
geoinfo.nmt.edu